The Geology of Chile
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The Geology of Chile

EDITED BY

TERESA MORENO
Earth Sciences Institute ‘Jaume Almera’,
Consejo Superior de Investigaciones Científicas, Barcelona, Spain

and

WES GIBBONS
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Email geokanda@ma.kcom.ne.jp

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Editorial acknowledgements

The editors would like to thank the authors for the huge amount of effort they have put into this book. It is never easy finding the quality time necessary to produce good written work, and book chapters can be especially onerous, especially if not written in one’s native language. We began this project late in 2003 with the intention of completing the book in 2005, and several authors quickly produced their contributions on time (special thanks!). In other cases deadlines began to slip a little: one author heroically saved a chapter from a delay that was becoming terminal, while another was especially accommodating in allowing major changes to his chapter in order to interface with another (you both know who you are, and you are much appreciated by your editors). With the finished product in hand, however, such trials of gestation seem trivial. While accepting full responsibility for any editorial oversights, such as the occasional appearance of unremediated ‘spanglish’, we are proud to have guided this project to its conclusion and congratulate the authors for producing such a comprehensive contribution to the scientific community.

One of us (W.G.) learnt his trade as a geologist pouring over fragmented, weathered exposures of ancient subduction zone rocks in southern Britain: calcalkaline plutonic complexes, blueschists, mylonitic terrane boundaries, melanges and so on. To visit Chile several decades later was something of a geological homecoming. Clambering over the low cliffs at Zapallar provided a return to the magma-mingled Cadomian plutonic rocks of the Channel Islands; photographing exposures of low grade accretionary metasediments and metabasalts on the beach at Pichilemu was a revisit to the Penmynydd Zone of NW Wales. Travelling across the country placed all these rocks within their ocean–continent plate margin setting in the most direct and exciting way possible. The whole experience was a reminder of how important it is to study both modern and ancient examples of geological processes and their products.

If the latter reminder is not enough to tempt the foreign reader to visit Chile, consider these: active, perfectly photogenic volcanoes perched on grossly thickened crust in the High Andes, the largest and strangest nitrate deposits in the world, truly vast areas of porphyry copper mineralization and an impressive history of precious metal mining, contrasting mountain belts showing modern extensional collapse and rapid compressional uplift all within a day’s drive, some of the wettest and driest climates, highest mountains and deepest oceanic trenches in the world, the chance to travel over normal versus ‘flat-slab’ ocean plate subduction zones, a seismic (and social) history that includes the largest earthquake ever recorded and a series of terrible tsunami events, an adjacent oceanography with a circulation system powerful enough to influence world climate, and a remarkable 15 000-year history of human colonization within an incomparable scenic and biogeographic setting.

During the preparation of this book the editors moved from Cardiff to Barcelona, and from hard rock geology to atmospheric pollution science. T.M. wishes to thank Xavier Querol, Andrés Alastuey and the team at CSIC Jaume Almera for helping to provide a positive working environment in which the final editing could be completed efficiently, and both editors thank Bob Pankhurst and (once again) Angharad Hills at the Geological Society for their help and patience. Finally, as Springsteen (1984) has commented ‘no puedes empezar un fuego sin una chispa’. In this context T.M. would like to thank, among many others, Rosario Lunar, Roberto Oyarzun, Hazel Prichard, Peter Fisher, Roy Richards, Tim Jones, Kelly BeruBé, and la familia Moreno Pérez. Similarly, W.G. would like to thank Greg Power, Tony Redman, Malcolm Howells, Jana Horák, Robert Shackleton, Mike Brooks, Graham Williams and his postgraduate students, Adrian Hartley, Rod Gayer, Tony Harris, Peter Burgess, Brendan Murphy, Art Goldstein, Jim McIlleland, Bruce Selleck, and the geology of Sark.

We especially wish to thank the referees who gave so much of their time to reviewing the initial manuscripts (listed below in alphabetical order).

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Contributing authors

L. Aguirre
Departamento de Geología, Universidad de Chile. Casilla 13518, Correo 21, Santiago, Chile

J. J. Armesto
Centre for Advanced Studies in Ecology and Biodiversity (CASEB), Pontificia Universidad Católica de Chile, Alameda 340, Casilla 114-D, Santiago, Chile and Institute of Ecology and Biodiversity, Universidad de Chile, Santiago, Chile

M. Atenas
Dirección General de Aguas, Ministerio de Obras Públicas, Transporte y Telecomunicaciones. Morandé 59, Santiago, Chile

S. E. Barrientos
Universidad de Chile, Facultad de Ciencias Físicas y Matemáticas, Departamento de Geofísica, Casilla: 2777, Avda. Blanco Encalada 2085, Santiago, Chile (e-mail: Sergio.Barrientos@CTBTO.ORG)

M. Calderón
Departamento de Geología, Universidad de Chile. Casilla 13518, Correo 21, Santiago, Chile

F. Camus
CODELCO-Chile Exploration, Huérfanos 1270, Santiago, Chile

A. Cecioni
Departamento Ciencias de la Tierra, Universidad de Concepción, Casilla 160-C, Concepción, Chile (e-mail: aecioni@udec.cl)

J. Cembrano
Departamento de Ciencias Geológicas Universidad Católica del Norte, Avda. Angamos 0610, Antofagasta, Chile (e-mail: jcembrano@caliche.ucn.cl)

R. Charrier
Departamento de Geología, Universidad de Chile. Casilla 13518, Correo 21, Santiago, Chile (e-mail: rcharrie@ing.uchile.cl)

G. Chong Díaz
Departamento de Ciencias Geológicas, Universidad Católica del Norte, Angamos 0610, Antofagasta, Chile (e-mail: gchong@caliche.ucn.cl)

J. E. Clavero
SERNAGEOMIN, Casilla 10465, Santiago, Chile

F. Espinoza
Departamento de Geología, Universidad de Chile. Casilla 13518, Correo 21, Santiago, Chile

V. Faundez
Departamento de Geología, Universidad de Chile. Casilla 13518, Correo 21, Santiago, Chile

G. Féraud
Géosciences Azur, UMR 6526. CNRS - Université de Nice-Sophie Antipolis, 06108 Nice Cedex 02, France

B. Fernández
Pontificia Universidad Católica de Chile, Vícuña Mackenna 4860, Casilla 306, Correo 22, Santiago, Chile

O. Figueroa
Departamento de Ciencias de la Tierra, Universidad de Concepción, Casilla 160 C, Concepción-3, Chile

S. Figueroa
Departamento de Zoología, Universidad de Concepción, Casilla 160-C, Concepción, Chile

F. Fuentes
Departamento de Geología, Universidad de Chile, Casilla 13518, Correo 21, Santiago, Chile

A. Gajardo Cubillos
Departamento de Ciencias Geológicas, Universidad Católica del Norte, Casilla 1280, Antofagasta, Chile

M. García
Servicio Nacional de Geología y Minería, Avda. Santa María 0104, Santiago, Chile

W. Gibbons
AP 23075, Barcelona 08080, Spain

S. Giglio
Programa de Magíster en Oceanografía, Escuela de Graduados, Universidad de Concepción, Chile

D. Gómez
Pontificia Universidad Católica de Chile, Vícuña Mackenna 4860, Casilla 306, Correo 22, Santiago, Chile

G. González
Departamento de Ciencias Geológicas, Universidad Católica del Norte, Avda. Angamos 0610, Antofagasta, Chile

H. González
Instituto de Biología Marina, Facultad de Ciencias, Universidad Austral de Chile, Campus Isla Teja, Casilla 567, Valdivia, Chile, and Center for Oceanographic Research in the Eastern South-Pacific (COPAS)

M. Grosjean
NCCR Climate & Institute of Geography, Bern University, Switzerland

A. Hartley
Department of Geology and Petroleum Geology, University of Aberdeen, Meston Building, Aberdeen, UK

D. Hebbeln
Geowissenschaften, Universität Bremen, Postfach 330440, D-28334 Bremen, Germany

G. Hérail
Departamento de Geología, Universidad de Chile. Casilla 13518, Correo 21, Santiago, Chile

F. Hervé
Departamento de Geología, Universidad de Chile. Casilla 13518, Correo 21, Santiago, Chile (e-mail: fherve@cec.uchile.cl)

J. C. Jofré
Comisión Nacional del Medio Ambiente. Teatinos 254, Santiago, Chile

J. Kaiser
Geowissenschaften, Universität Bremen, Postfach 330440, D-28334 Bremen, Germany

F. Lamy
GeoForschungsZentrum-Potsdam, Telegrafenberg, 14473 Potsdam, Germany

L. E. Lara
SERNAGEOMIN, Casilla 10465, Santiago, Chile

C. Latorre
Centre for Advanced Studies in Ecology and Biodiversity (CASEB), Pontificia Universidad Católica de Chile, Alameda 340, Casilla 114-D, Santiago, Chile and Institute of Ecology and Biodiversity, Universidad de Chile, Santiago, Chile (e-mail: clatorre@bio.puc.cl; clatorreh@uc.cl)

A. Lavenu
Laboratoire des Mecanismes de Transfert en Geologie, IRD-UMR 5563, 14 Avenue Edouard Belin, 31400 Toulouse, France
Chile is, geographically, an unusual and in many ways astonishing country (Fig. 1.1). It stretches north–south along the South American mainland for over 4,000 km, from 18°S, where the Altoiplano is shared with Peru, Bolivia and Argentina, to 56°S at Tierra del Fuego and the islands of Cape Horn, the next stop being Antarctica. Its western margin everywhere is the Pacific Ocean, and its eastern boundary is the summit of the Andes mountains, so that in a width of rarely more than 200 km, the topography rises from sea level to a maximum of almost seven thousand metres. Climatic variations reflect this extraordinary topography. The north is characterized by the Atacama Desert, considered to be the driest place on Earth. The south is in the temperate rainforest zone, with vegetation that struggles against the prevailing westerly gales. In this southern sector the land is moulded by recent glaciations that carved the coastal areas into fiords and archipelagos consisting of thousands of islands; the length of the Chilean coastline including these islands must exceed that of many other countries that have a larger surface area. It is the extreme variety represented by these factors that have led to Chile becoming such an attractive tourist destination, despite the isolation and comparative difficulty of access of many of its geomorphological treasures.

Figure 1.1 shows the distribution of the main tectonic and geomorphological features of Chile. The northern and central parts of the country can be reasonably divided into three north–south zones. The coastal zone (zone 1) is the Coastal Cordillera, consisting predominantly of Late Palaeozoic and Mesozoic igneous rocks, with paired belts of Palaeozoic metamorphic rocks cropping out south of Pichilemu (Fig. 1.1, zone 1a). The coastal batholith of Late Palaeozoic igneous rocks continues down to about 38°S, where it apparently passes eastwards into Argentina. In southern Chile the coastal zone (1b) is mainly represented by metasedimentary turbidites, once thought to be Palaeozoic in age but now known to be largely Mesozoic, apart from the Isla Madre de Dios area, which is interpreted as an accreted terrane with Permian limestones. The central depression (zone 2), or Central Valley, is a downwarp with a Mesozoic to Quaternary sedimentary fill; south of Santiago, this is the main agricultural zone and contains several major cities. In southern Chile the central valley is not recognized and the transition between the coastal turbidites and the Andes is dominated by the Mesozoic–Cenozoic calcalkaline igneous rocks of the Patagonian batholith — one of the longest continuous granitoid bodies in the world.

Overall, the geological evolution of Chile has resulted from the effects of east-directed subduction of Pacific (and proto-Pacific) ocean floor beneath the South American continent. This subduction is the force that generated the Andes, a chain of mountains (zone 3) whose primary uplift dates back to a Miocene event, but whose emergence continues today, as exemplified by major earthquake activity. Their elevation is accompanied by significant crustal shortening, principally accommodated by eastward thrusting. The mountain chain continues right down into Tierra del Fuego at the southern tip of Chile, albeit with continuous reduction in height. Subduction of the Nazca Plate (Fig. 1.1) is currently active, but in the southernmost sector, where the triple-junction with the Antarctic Plate has steadily migrated northwards as far as the Taitao Peninsula (c. 47°S), it has slowed to about 2 cm/year.

Present-day and recent subduction is also evidenced by an almost continuous line of active and dormant volcanoes, mostly andesitic stratovolcanoes, that are studded along the entire length of the country, accounting for some 10% of the circums-Pacific ‘ring of fire’. They comprise the Central and Southern Andean volcanic zones, separated by a short volcanic gap between latitudes c. 27° and 31°S, due to a zone of subduction with unusually shallow dip (the so-called ‘flat slab’ zone), where the Central Valley is also absent. The northern volcanoes are built on the Altoiplano, a high plateau area where the crust is up to 70 km thick, with a southern extension into the Puna of NW Argentina. Figure 1.2 shows the progress of a minor volcanic eruption in this zone. Significantly, this northern sector with thick continental crust is where the rich mineral wealth of Chile (Cu, Au, Ag) is prominently developed. The Southern Andean Volcanic Zone and the Main Cordillera reach the coast of the mainland, opposite the offshore archipelago, in the southernmost part sometimes being distinguished as the Austral Volcanic Zone due to the more basaltic composition of the volcanic products.

The pre-Andean (sensu stricto) geology, including the Coastal Cordillera and Patagonian batholiths, and the Palaeozoic igneous and metamorphic rocks, was also largely formed during east-directed subduction episodes, though in the latter case of a rather different character, with accretion of forearc and distal oceanic deposits, and even small exotic terranes. Chile is thus a most appropriate place in which to investigate models of crustal growth through magmatic and tectonic accretionary processes. The long-lived reworking of crustal material, especially where the continental crust is thickest, in a stable subduction environment is a major factor in the formation of the mineral wealth of the Chilean Andes.

The scenic diversity of this country, which lies directly above an ocean–continent subduction zone that has been continuously active since early Mesozoic times, is illustrated in Figure 1.3.

This book is intended to present a comprehensive picture of the evolution of Chilean geology, written by some of the most respected experts in the subject. However, it is appropriate to acknowledge the pioneering work of previous generations in revealing the essential background and describing features that we now have the luxury of interpreting from the standpoint of a modern state of knowledge.

One of the first to record aspects of the geology of Chile was the naturalist Charles Darwin (1809–1882), who travelled round Cape Horn and up the coast of southern Chile during the second such voyage of the Beagle in 1834/35. At the start, he made the first description of Cretaceous ammonites near Punta Arenas. Subsequently, he spent the best part of a year on land, travelling south as far as the Chonos Archipelago, where he first described the schists of the accretionary complex, and north to Copiapó, as well as crossing the Andes from Valparaíso to Mendoza. He saw Osorno in eruption and visited Concepción shortly after a major earthquake. His observations, first published in his diary (1836), show how...
thoroughly he had absorbed and applied the then revolutionary uniformitarian ideas of Charles Lyell, the first volume of whose *Principles of Geology* had been published just before the ship left Britain. In particular he remarked upon evidence for the uplift of land during earthquakes and deduced how over geological time this could explain raised terraces and, indeed, the uplift of the Andes themselves: he even noted earthquakes as the cause of tsunamis. His appreciation of geological processes extended to the volcanic activity associated with this 'upheaval' and the denudation by erosion of the uplifted crust to form the thick sedimentary deposits covering much of Patagonia.

Darwin’s contemporary, Ignaz (Ignacio) Domeyko (1802–1889), was born in Polish Lithuania and was involved in his early life in the struggle against Russian domination. He was invited to Chile in 1838 and spent the next 30 years investigating the natural history and mineralogy of his second homeland. He was the first to demonstrate the presence of Jurassic rocks in the Andean cordillera; he discovered many economic deposits of coal and copper, but was also involved in the establishment of ethical and conservationist policies for their exploitation. In 1843 he was a founder member of the Faculty of Mathematical and Physical Sciences at the University of Chile, of which he became Rector in 1867. Chilean plants, fossils, minerals, and a mountain range now bear his name.

There have been previous books on the geology of Chile as a whole, most notably those of Hans Bruggen (1950), Carlos Ruiz *et al.* (1965), Jorge Muñoz Cristi (1973) and José Frutos *et al.* (1986). Naturally, each of these illustrates and builds on the progressive increase in knowledge of Chilean earth science with the passing decades, and a growing understanding of the tectonic processes that have forged this far edge of the Americas.

Ruiz *et al.*’s *Geología y Yacimientos Metalíferos de Chile* was published in 1965 by the Instituto de Investigaciones Geológicas (IIG), of which he was director: this was the state organization which preceded the present Servicio Nacional de Geología y Minería (SERNAGEOMIN). It was the most comprehensive of these earlier books, and demonstrated the close connection between geological studies and mineral exploration, which has become a strong tradition in Chile, since the production of copper has always been the basis of the country’s economy (except for fifty years at the end of the nineteenth century).
and beginning of the twentieth centuries, when the export of naturally occurring nitrates from the evaporitic deposits of the Atacama region was paramount). Because of its economic importance, the geology of the mining areas in northern Chile has always received much more attention, and was thus far better known, than that of the vast expanses of southern Chile, where very much less mineral wealth has been found.

Geological maps of the whole country at the scale of 1:1,000,000 were published in 1960, 1968 and 1980 by IIG, and in 2002 by SERNAGEOMIN. Significantly, this last issue is also available in digital format. Larger-scale maps have existed for many areas, but still do not cover the whole extent of Chilean territory, and are concentrated in the north, where mining is widespread.

Fig. 1.2. Modern arc volcanism in Chile: ash eruption of Volcan Lascar on 20 July 2000 (see Chapters 5 and 13). View from the Quaternary evaporites of the Salar de Atacama to the west: each frame shows the approximate time during this short-lived eruption. The photographs show the flattening of the plume at the base of the stratosphere, and a second smaller pulse in the final frame. This volcano, at 23°22'S, 67°44'W, has been the most active of Chilean volcanoes in historic times, erupting over 15 times in the twentieth century; the largest eruption in 1993 when a subplinian eruption rose over 20 km rose above the crater. The most recent eruption (as at December 2005) was in May 2005 when a small explosion generated ash which dispersed towards Argentina. It is an andesite-to-dacite stratovolcano, 1400 m high, rising from the Chilean Altiplano to a summit elevation of 5592 m. The volcano erupts through the thickest continental crust in the world, in a zone of regional compression above the actively subducting Nazca Plate. The volcanic edifice has been built on Miocene dacitic ignimbrites erupted during the early evolution of the arc. Photo: Robert J. Pankhurst.
The following 12 chapters firstly deal with the basement geology (Chapter 2), examining the commonly fragmentary evidence for Pre-Andean events. This is followed by accounts of Andean tectonostratigraphy (Chapter 3) and magmatism (Chapter 4), which together provide an encyclopaedic wealth of information on Chilean igneous complexes and sedimentary successions. Chapter 4 leads seamlessly into an account of the astonishingly active recent volcanic activity (Chapter 5) and the world-class metallic ore deposits which have proven to be so critical to the welfare of the country (Chapter 6). The economic theme continues with reports on non-metallic industrial minerals, including the famous and peculiar Atacama nitrates (Chapter 7), and by an overview of Chilean water resources (Chapter 8), many of which in arid northern Chile have been severely impacted by natural and anthropogenic metal pollution.

The subject then turns to geophysics with an examination of neotectonics (Chapter 9) and earthquakes (Chapter 10), the hazardous frequency of which is a daily fact of life for the Chilean population. Chapter 11 deals with the marine geology and oceanography of the offshore Pacific, a subject that continues to attract much research, not least from those seeking to understand world climatic variations. This marine Quaternary chapter is succeeded by one written from the perspective of land-based Quaternary scientists (Chapter 12), concluding with an account examining human colonization of southernmost America. Finally, the editors offer a geological description of a drive from the mediterranean landscapes of central Chile to the hyperarid Atacama Desert (Chapter 13). This last contribution is designed to give visitors a chance to experience for themselves the geology and scenery of this extraordinary country.

Fig. 1.3. (a) ‘Piel y camanchaca’ – a wind-sculpted feature of the Atacama Desert, where the hyperarid climate can result in the surface remaining undisturbed for thousands of years. (b) Carboniferous to Early Permian limestones in the Madre de Dios archipelago, part of an exotic terrane accreted to the southwestern margin of Gondwana (photo: Fernando Sepúlveda). (c) Thrust deformation in the Andes of Central Chile. The uppermost unit on the left (west) consists of Permo-Triassic volcanic rocks of the Pastos Blancos Formation, thrust over the overturned Miocene Tilito Formation. (d) Dolerite sills (?Tertiary) cutting Cretaceous black shales of the Katterfeld Formation, near Coyhaique. (e) Evidence of recent glacier recession in the South Patagonian Batholith, Seno Iceberg, where freshly exposed granite in the foreground lacks the otherwise dominant cover of lichen, grasses or shrubs. (f) Destruction of the city of Valparaiso by a major earthquake in 1906. (g) Torres del Paine, near Puerto Natales, a Late Miocene laccolithic body of pale granite emplaced in dark-coloured Cretaceous sedimentary rocks. (h) A 20-m thick ignimbrite flow of the Divisadero Formation (Cretaceous), Aysén.
The present-day Andes have formed in response to subduction-related processes operating continuously along the western margin of South America since the Jurassic period. When these processes started, the continental margin was mainly formed of metamorphic complexes and associated magmatic rocks which evolved during Proterozoic (?), Palaeozoic and Triassic times, and which now constitute the basement to the Mesozoic and Cenozoic Andean sequences. These older units are commonly referred to in the Chilean geological literature as the ‘basement’ or the ‘crystalline basement’.

The basement rocks crop out discontinuously (Fig. 2.1) in northern Chile, both in the coastal areas and in the main cordillera. In contrast, from latitude 34°S southwards, they form an almost continuous belt within the Coastal Cordillera extending to the Strait of Magellan. In addition, sparse outcrops occur both in the main Andean cordillera as well as further east in the Aysen and Magallanes regions. In the first maps and syntheses of the geology of Chile (e.g. Ruiz 1965) these rocks were generally considered to be of Precambrian age, forming a western continuation of the Brasilian craton. Later work has demonstrated that rocks first described as metamorphic basement units show a wide range of metamorphic grades and ages extending from possible Late Proterozoic through Palaeozoic and even, in some cases, to Jurassic–Cretaceous.

With regard to previous works that have attempted to synthesize data on Chilean basement geology, the reader is referred to those by González-Bonorino (1970, 1971), González-Bonorino & Aguirre (1970), Aguirre et al. (1972), Muñoz Cristi (1973), Hervé et al. (1981a), Hervé (1988), Breitkreuz et al. (1988), Damm et al. (1990), Bahlburg & Hervé (1997), Hervé et al. (2000, 2003a) and Willner et al. (2005). These accounts reflect increasing progress in our understanding of the basement based on recent field studies, the application of radiometric dating techniques, and new ideas concerning the evolution of accretionary prisms and terrane geology.

A description of the different units that constitute the Chilean basement geology is presented below, including information about their age, metamorphic characteristics and geological settings, largely based on the above-mentioned publications as well as on more specific studies that will be cited appropriately. Particular emphasis will be on recent studies that have attempted to determine P–T–t paths of metamorphism through mineralogical observations and thermodynamic calculations which were developed in the last two decades, although studies of some of the units are still in an immature state.

As a general framework, the description of the units will be based on the terrane model as established by Bahlburg & Hervé (1997) for northern Chile and northwestern Argentina, depicted in Figure 2.1. The older metamorphic complexes of northern Chile will be treated first, the Late Palaeozoic accretionary complexes of the coastal areas of central Chile next, and finally the Late Palaeozoic to Mesozoic complexes of the Patagonian Andes.

North Chile (Norte Grande): Arequipa–Antofalla and Mejillones terrane areas

In the main Andes and in the Coastal Cordillera of northern Chile, scattered outcrops of basement rocks occur sporadically under the Mesozoic–Cenozoic cover (Fig. 2.1). The isolation of these units has hindered detailed interpretation of their geological significance and correlations between them. Interpretations can be assigned to two main types: (a) that they reflect terrane tectonics, related to the collision of Laurentia and Gondwana in Early Palaeozoic times; and (b) that these rocks represent in situ evolution of old cratonic units of the western margin of Gondwana.

Metamorphic rocks cropping out in northern Chile are grouped within the Belén, Sierra de Morena–Chojas, and Limón Verde complexes (all placed within the Arequipa–Antofalla Terrane: Fig. 2.1), and the Mejillones Metamorphic Complex, which may be a separate terrane. In addition to these metamorphic complexes, a volcanic and sedimentary sequence known as the Cordón de Lila Complex (CLC) crops out on the west side of the Andes in northern Chile.

Bahlburg & Hervé (1997) and Bahlburg et al. (2000) have observed that during the Palaeozoic era in northern Chile and northwestern Argentina, a magmatic and metamorphic lull of c. 100 million years, from the Lower Silurian to the Early Carboniferous, allows separation of the rock units into two orogenic groups: (a) Cambrian to Early Silurian rocks deformed during Lower Palaeozoic orogenic cycles in an active margin setting; (b) rocks formed after Lower Palaeozoic orogenies but affected by the Late Carboniferous to Permian Toco event, in which active margin conditions resumed after a lull during which the area evolved as a passive margin (Silurian–Early Carboniferous).

The more detailed account that follows presents lithological and geochronological data from both published and unpublished work, organizing the data within the framework of the models provided by Bahlburg & Hervé (1997) and Bahlburg et al. (2000). The detailed data available from Argentina, Bolivia and Peru are beyond the scope of this chapter (see Bahlburg & Hervé 1997; Loewy et al. 2004). The descriptions given below are organized to deal initially with the general geology, then metamorphic grade, and finally geochronology. The tectonic setting is discussed under a separate heading at the end of this chapter.

Belén Metamorphic Complex

The Belén Metamorphic Complex (BMC) (Basei et al. 1996) forms a narrow outcrop of metamorphic and igneous rocks along a high-angle west-vergent thrust system located on the western slope of the Chilean Altiplano plateau between Chapiquiña and Tignamar (Muñoz & Charrier 1996). Along
Fig. 2.1. Geological map showing the distribution of outcrops of the metamorphic complexes, the Palaeozoic and some Triassic sedimentary units, and associated plutonic belts in Chile.
these faults the BMC is thrust westward over late Cenozoic deposits. Uncomformably covering the BMC are Jurassic marine deposits, and Cenozoic volcaniclastic and continental sedimentary rocks (Pacci et al. 1980; Muñoz et al. 1988; García 1996). The BMC mainly comprises foliated amphibolites and subordinate quartz micaschists, orthogneisses and serpentinites. It is intruded by a small gabbro stock and by mafic, aplitic and felsic dykes.

Peak metamorphic conditions for the Belén Metamorphic Complex have been determined by Wörner et al. (2000) at c. 700°C and 7 kbar. The prograde metamorphic path is reflected in zoning patterns of garnets which indicate a simple, single-stage metamorphic event. Retrograde stages are marked by lower grade overprinting in amphibolite and greenschist facies. The resulting P–T–t curve is shown in Figure 2.2.

The first geochronological determination of the age of the BMC was by Pacci et al. (1980) who produced a Rb–Sr reference isochron of 1000 Ma which has since been later recalculated to 500 Ma (Damm et al. 1990). These authors determined a 1460 ± 448 Ma Nd–Sm whole-rock isochron for metabasites which was considered as the crystallization age of the protoliths. Basei et al. (1996) obtained U–Pb zircon upper intercept ages (conventional method) of 509 ± 46 Ma for the Quebrada Achacagua orthogneiss and 486 ± 32 Ma on granitic veins at Quebrada Saxamar which they interpreted as crystallization ages of the igneous precursors. These authors also reported model Sm–Nd model ages of 1746 Ma and 1543 Ma on the Quebrada Saxamar schists testifying an ancient crustal residency for the protoliths. Wörner et al. (2000a) considered that higher intercept ages of 1877 ± 139 Ma and 1745 ± 27 Ma obtained on zircons (U–Pb, conventional method) from Belén reflect crystallization ages and that lower intercepts of 366 ± 3 Ma and 456 ± 4 Ma suggest severe Palaeozoic lead loss. Loewy et al. (2004) presented a U–Pb zircon age of 473 ± 2 Ma for the Saitoco granodiorite, and revealed the presence of 1.8 to 1.9 Ga old zircons on a cross-cutting dyke in the micaschists.

In addition, Basei et al. (1996) have presented a 344 ± 22 Ma Rb–Sr whole-rock isochron, with a 87Sr/86Sr initial ratio of 0.708 for the quartz micaschists of Quebrada Saxamar. K–Ar dating on different minerals yielded ages of 536 Ma to 516 Ma (hb) in the Quebrada Saxamar schists, of 417Ma (Ms) to 365 Ma (bt) on the Saitoco orthogneiss (Basei et al. 1996), and 358 ± 10 Ma to 457 ± 7 Ma (hb) (Lucassen et al. 2000) at Belén.

All these data point towards the involvement of the BMC in Early Palaeozoic orogenies, time-equivalents to the Early Cambrian Pampian and Ordovician Ocloyic/Famatinian orogenies, both of which are more extensively represented in northwestern Argentina. The age of the protoliths seems to be Early Palaeozoic, as no undisputed evidence of Proterozoic ages has been produced (Damm et al. 1990; Wörner et al. 2000a; Loewy et al. 2004).

Sierra de Moreno – Chojas metamorphic complexes

Basement exposures of quartz micaschists, greenschists, migmatises, granites and mylonites in the Sierra de Moreno crop out as a 70 x 20 km belt orientated NNE–SSW (Skarmeta 1983). The western side of the outcrop comprises a 3-km-wide...
The Mejillones Metamorphic Complex: an independent terrane?

The Mejillones Metamorphic Complex (MMC) is an isolated 60 x 10 km outcrop of basement rocks exposed along the Mejillones Peninsula (Fig. 2.1). There is a smaller outcrop of similar lithologies at Caleta Coloso, a few kilometres to the south. Baaza (1984) has described the metamorphic rocks as consisting of orthogneiss, paragneiss, amphibolites and micaschists, intruded by mafic to silicic plutonic rocks and dykes. The MMC consists of two metamorphic areas, one with micaschists, intruded by mafic to silicic plutonic rocks and amphibolites, and the other with orthogneiss, paragneiss, amphibolites and alkali micaschists, gneisses, amphibolites, migm amphibolites, and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and alkali micaschists, gneisses, amphibolites, migmatites and algm assemblages of the pumpellyite–actinolite–chlorite zone at the base of the succession, with the metamorphic grade decreasing to higher stratigraphic levels. This burial-type metamorphism reached peak metamorphic conditions of 3 kbar and 370°C (Fig. 2.2).

A whole-rock Nd–Sm isochron from 11 basalts and andesites of the CLC yields an age of 448 ± 145 Ma (Damm et al. 1986), which is consistent with the palaeontologically derived Llanvirnian age for the CLC (Cecioni 1982). With regard to the Palaeozoic intrusives, Mpodozis et al. (1983) reported Rb–Sr whole-rock isochrons of 441 ± 8 Ma (Tučucuro pluton), 452 ± 4 Ma (Tilopozo pluton) and a 288 ± 15 Ma errorchron for the Pingo Pingo pluton, which, however, yielded K–Ar ages of 425 ± 11 (bt) and 429 ± 12 Ma (hb). Damm et al. (1990) presented U–Pb zircon ages of 502 ± 7 Ma (conventional, discordant) for the Choschas pluton, and lower intercept ages of 434 ± 2 Ma for the Cerro Lila pluton, and of 338 ± 14/−18 Ma for the Pingo Pingo pluton. All ages were interpreted as crystallization ages.

Late Palaeozoic sedimentation and volcanism

Devonian–Carboniferous passive margin sedimentation

Thick successions of shale, sandstone and rare conglomerate and limestone crop out in the Coastal Cordillera. These sequences have been given different lithostratigraphic names in different areas: El Toco Fm (TF), Sierra del Tigre Fm, Las Tortolas Fm (LTF), and the Chañaral Melange (CM: Fig. 2.1). Coeval siliciclastic shallow marine platform successions were deposited further east, in the main Andean cordillera (Zorritas Fm). The turbiditic rocks have abundantly preserved sedimentary features, such as cross- and graded bedding, obscured locally by deformation and low grade metamorphism. They have been interpreted by Bahlburg & Breitkreuz (1991) as a turbidite system representing environments ranging from a proximal depositional lobe (TF) in the north, and a distal depositional lobe to basin plain (TF and LTF) environments to the south.

Progressive synsedimentary deformation and metamorphism of these marine units took place during the Carboniferous period (El Toco event of Bahlburg & Breitkreuz 1991). The culmination of this process produced the Chañaral Melange (Bell 1982), a dismembered portion of the turbidite complex.

The fossil record in the turbidite complex is poor but indicates a Devonian to Early Carboniferous depositional age. Late Devonian plant remains occur in the El Toco Fm (Boronic 1980; Devonian brachiopods in the Sierra del Tigre Fm (Niemeyer et al. 1985, 1997b) and Lower Carboniferous conodonts (Bahlburg 1987) and brachiopods (Bell 1987b) in the Las Tortolas Fm have all contributed to the palaeontological
database. The only radiometric age determination available is a 280 Ma Rb–Sr whole-rock isochron (Brook et al. 1986) for the Las Tortolas Fm and Chañaral Melange which is in concordance with the radiometric ages of the plutonic rocks that intruded the already folded turbidite succession. Finally, the metamorphic grade of the turbidite succession is variable, low to very low grade, and little is known of the P–T conditions except in the Chañaral melange, with Marioth & Bahlburg (2003) establishing 2.2–2.8 kbar and 300–350°C for the metamorphic peak (Fig. 2.2).

Permian subaerial volcanism
Thick subaerial volcaniclastic rocks, known as the Peine Group, crop out in the Precordillera and in the high Andes in a north–south elongated discontinuous belt (Breitkreuz et al. 1986). The successions consist of felsic volcaniclastic rocks and andesitic–basaltic volcanic rocks, interbedded with limnic and shallow marine sediments. They lie with angular unconformity on the Devonian–Carboniferous turbidites, and interfinger to the east and west with the Early Permian marine carbonates of the Cerros de Cuviaetas (Breitkreuz 1986b) and Arizaro formations, in Chile and Argentina, respectively. These rocks are weakly folded and have been subjected to extensive sericitization and chloritization, particularly near plutonic intrusions.

Late Palaeozoic to Triassic batholiths
These intrusions occur in two belts, one in the high Andes and one in the Coastal Cordillera. Brown (1990) studied these plutonic belts at latitude 26°S and concluded that they were part of the same subduction-related magmatic arc, later affected by crustal extensional processes. The presence of subvolcanic domes and high level intrusive bodies led Davidson et al. (1985) to recognize the existence of former caldera systems in the Imilac area dated at 290 to 217 Ma, some of which have copper mineralization. Camus (2003) has presented a synthesis of the mineralized intrusive bodies north of 24°S in the Precordillera and High Andes, indicating that they range in age from 190 to 307 Ma (K–Ar).

The Coastal Cordillera plutons have given Rb–Sr wr isochron ages (Brook et al. 1986) in the range of 278 ± 16 Ma and 221 ± 14 Ma. In addition, U–Pb determinations on zircon have yielded 292 ± 14 Ma, 230 ± 6 Ma and 217 ± 12 Ma (Damm & Pichowiak 1981). Thus, these rocks span Early Permian and Late Triassic times, similar to those of the High Andes belt which ranges in age from 270 ± 10 Ma to 209 ± 19 Ma (Brook et al. 1986; wr Rb–Sr isochrons). A 268 ± 5/–3 Ma age (Damm et al. 1990; U–Pb zircon, lower intercept) was obtained for the Córdon Chinchuilchorro pluton, located further north, in the Córdón de Lila area. In both belts, the age groupings suggest there is a natural break in the ages during Early Triassic times.

Limón Verde Metamorphic Complex
A 12 km long and 2 km wide outcrop of metamorphic rocks is exposed on the western flank of the Sierra de Limón Verde (Baeza 1984), where it is intruded by a Late Palaeozoic batholith unconformably overlain by Triassic sedimentary and volcanic rocks. The Limón Verde Metamorphic Complex (LVMC) comprises metasandstones and metapelites with metamorphic grade varying from greenschist to amphibolite facies.

Lucassen et al. (1999), using conventional geothermobarometry and multi-equilibria calculations, established that the peak conditions of metamorphism attained by the LVMC were 14 kbar and 660 to 720°C, conditions unique in the Central Andes for metamorphic complexes of this age (Fig. 2.2). These data, combined with the age and stratigraphic data, point towards a very rapid exhumation of the LVMC during Late Permian–Early Triassic times. The tectonic environment for the attainment of these conditions is not well understood. Lucassen et al. (1999) suggested a tranpressional strike-slip environment whereas a collisional or subduction zone setting was previously suggested by Hervé et al. (1985) and Bahlburg & Hervé (1997).

Radiometric dating has not supported previous suggestions that the unit might be of Precambrian age (Baeza 1984; Rogers 1985; Hervé et al. 1985), Cordani et al. (1988) and Lucassen et al. (1999) established that the metamorphism took place in Late Palaeozoic times. Calculations of residence times using Rb–Sr isotopic data suggest that the protoliths could not have been older than 405 Ma (Silurian). Three Nd–Sm mineral isochrons indicate a range between 287 and 255 Ma (Lucassen et al. 1999). These ages are younger than the 309 ± 11 Ma and 300 ± 20 Ma ages obtained using Rb–Sr whole-rock isochrons (Hervé et al. 1985; Cordani et al. 1988). All these authors interpreted the ages as indicative of the metamorphism. Decreasing K–Ar mineral ages in hornblende, muscovite and biotite in the range 310–229 Ma indicate cooling through Permian and Early Triassic times.

North-central Chile (Norte Chico)

The basement rocks of Norte Chico comprise the Pampa Gneisses (PG), the Tránsito Metamorphic Complex (TMC) and the coastal El Teniente (ETMC) and Choapa (ChMC) metamorphic complexes. The Pampa Gneisses have been placed within the Chilenia Terrane (Ramos et al. 1986), which would have docked to South America in the Devonian, and the rest lie west of it. In addition, Mpodozis & Kay (1990, 1992) have suggested the existence of an Equis Terrane (not exposed), lying west of Chilenia, in order to explain a westerly jump in subduction-related magmatic foci from Permo-Triassic to Jurassic times (Fig. 2.1).

Pampa Gneiss
Ribba (1985) first described the presence of banded orthogneisses in a small outcrop in the upper reaches of Rio Tránsito (Fig. 2.1). They are intruded by Late Palaeozoic–Permian plutons, and thrust to the east over the TMC to produce a shear zone referred to as the El Portillo mylonites.

The metamorphic grade of the Pampa Gneiss reaches migmatitic conditions, and the mineral paragenesis suggests a high T–low P metamorphic environment. Ribba (1985) and Ribba et al. (1988) presented a Rb–Sr whole-rock isochron of 415 ± 14 Ma on the PG, as well as a Rb–Sr mineral isochron of 246 ± 18 Ma. These ages were interpreted as indicative of a Silurian metamorphic event and a Permo-Triassic resetting by a thermal event, further supported by 236 ± 6 Ma (bt) and 239 ± 10 Ma (ms) K–Ar mineral ages. The Portillo mylonites were dated at 250 ± 26 Ma (Rb–Sr, wr), a further indication of the Permo-Triassic tectonothermal event.

The Tránsito Metamorphic Complex
Along El Tránsito Valley, east of Vallenar, large exposures of quartz micaschists, metabasites, quartzites and marbles, which crop out below or are in tectonic contact with the Mesozoic cover, constitute the Tránsito Metamorphic Complex (TMC; Ribba 1985; Ribba et al. 1988). It is intruded by Late Carboniferous tonalites and is unconformably covered by sedimentary and volcanic rocks of Middle Triassic age (Reutter 1974).

The metamorphic facies of the TMC is transitional between greenschist and amphibolite facies. Hervé (1982) used amphibole mineral compositions to suggest that the TMC had been metamorphosed under an intermediate P–T regime, with a metamorphic peak at about 5 kbar and 500 ± 50°C (Fig. 2.3). Whole-rock Rb–Sr ‘errorchrons’ of 303 ± 40, 303 ± 35 and 335 ± 20 Ma suggest a Late Carboniferous metamorphic event, with this isotope system having been perturbated by later
thermal episodes related to plutonic intrusions into the TMC. K–Ar mineral ages of $238 \pm 10$, $229 \pm 6$ and $231 \pm 6$ Ma on muscovite testify to Triassic isotopic resetting or cooling.

**El Teniente Metamorphic Complex**

Discontinuous outcrops of metamorphic rocks occur in the Coastal Cordillera between Huasco (27°S) and Los Vilos (32°S). At around 31°S, Irwin et al. (1988) described the presence of a complex composed of ultrabasic and basaltic metabasites, metacherts, metasandstones and metaconglomerates which bear witness to a prolonged history of deformation and metamorphism in Late Palaeozoic through Early Mesozoic times. As used here, the El Teniente Metamorphic Complex (ETMC) encompasses the Cerro Negro and La Totora complexes of Thiele and Hervé (1984).

On the basis of the presence of staurolite, garnet, clinopyroxene, calcic plagioclase and albite, the $F_1$ episode of metamorphism is assigned to the upper greenschist-amphibolite facies transition. It is possible that an earlier higher grade metamorphism was partially overprinted by this event. Hervé (1982) used amphibole compositions to suggest that the metamorphic pressure was lower than in the ETMC and did not reach 5 kbar. Calculation of metamorphic temperature from plagioclase–hornblende pairs from Irwin et al. (1988) suggest it evolved from 700–800°C in the amphibole cores (pre-$F_1$?) to 500–550°C in the amphibole rims ($F_1$) (Fig. 2.3).

The metabasites have yielded a $311 \pm 89$ Ma Rb–Sr errorchron (Irwin et al. 1988) interpreted as representing the time of extrusion or an early metamorphic episode. The metabasites and metasediments were assembled before the 220–200 Ma metamorphic episode $F_1$, as suggested by a $201 \pm 61$ Ma Rb–Sr whole-rock isochron in the metaconglomerates and radiometric ages of the plutons emplaced into the metamorphic complex. Rb–Sr whole-rock isochrons of $220 \pm 20$ Ma (gabbro) and $200 \pm 10$ Ma (monzogranite) are comparable to six K–Ar ages (Hb) from the metabasites which range from $220 \pm 20$ Ma to $188 \pm 17$ Ma. Further deformational episodes $F_2$ (160–150 Ma) and $F_3$ (140–121 Ma) are recorded by the ETMC and in the associated intrusive bodies.

**The Choapa Metamorphic Complex**

This complex comprises intensely deformed quartz micaschists, phyllites and amphibolitic schists (Rebolledo & Charrier 1994) affected by a polyphase deformation in which up to seven deformational events have been identified. They have been interpreted by Rebolledo & Charrier (1994) as the metamorphosed equivalent of the Late Devonian–Early Carboniferous Puerto Manso Formation.

The metamorphic conditions of the Choapa Metamorphic Complex (ChMC) were considered by Rebolledo & Charrier (1994) as belonging to the greenschist facies in a low to medium P–T environment, although peak conditions of metamorphism
obtained from zoned actinolitic hornblende (Godoy & Charrier 1991) suggest a pressure not lower than 5 kbar in the greenschist-amphibolite facies transition (Fig. 2.3). Except for a K–Ar age of 359 ± 36 Ma (Am), the K–Ar ages obtained in the complex yield Jurassic ages and were probably reset by nearby Jurassic plutons.

**Batholithic intrusions**

The Norte Chico region is characterized both by the presence of the huge composite Elqui–Limari Batholith (ELB), which extends from 28°30' to 31°S in the High Andes, and by a Coastal Belt of intrusions. Following Mpdozis & Kay (1990, 1992) the ELB is viewed as comprising the subduction-related Late Carboniferous–Early Permian Elqui Superunit (ES) and the Permo-Triassic Ingaguas Superunit (IS), which resulted from crustal anatexis in a thickened continental crust generated during the San Rafael orogenic phase (Llambias & Sato 1990). Pankhurst et al. (1996) presented a zircon U–Pb age of 285.7 ± 1.5 Ma and a 256 ± 10 Ma Rb–Sr wr isochron for two of the main plutons of the ES and a map showing all the previous age determinations by different authors. Radiometric ages in the IS are generally in the range 200–230 Ma (Brook et al. 1986; Rex 1987; Parada et al. 1988) although some older K–Ar ages have been obtained as well. This allowed Parada et al. (1991) to define a Triassic-Jurassic plutonic event, the products of which are distributed in the High Andes Belt (HAB), the Ingaguas Superunit of the Elqui Limari Batholith, and in the Coastal Belt (CB). The Triassic to Lower Jurassic magmas would have formed on an extensional setting, which was replaced by subduction-related magmatism from the Middle Jurassic onwards.

**Late Palaeozoic accretionary complexes of the Coastal Cordillera of central Chile**

Between latitudes 32°S and 42°S, continuous exposures of the basement occur in the Coastal Cordillera of central Chile (Fig. 2.1). This basement comprises a metamorphic complex which is exposed continuously south of 34°S, flanked to the east by the Coastal Batholith, which reaches the coast between 32°S and 34°S, and turns east south of 38°S to reappear in the Main Andean Range at 40°S and then into Argentina (Fig. 2.1).

González-Bonorino (1970, 1971) and González-Bonorino & Aguirre (1970) distinguished three metamorphic belts or ‘series’ – the Coastal Series, the Western Series and the Eastern Series – which differ in metamorphic grade and direction of increasing metamorphism. Godoy (1970) and Aguirre et al. (1972) modified this subdivision into a Western and an Eastern series which is still widely used. The two series have been interpreted as representing a paired metamorphic belt, in the sense of Miyashiro (1961), with the Western Series being the higher P–T unit. Hervé (1977) suggested that the Western Series included accreted oceanic lithologies, and the whole was interpreted as a subduction complex by Hervé et al. (1976b, 1981a) and Forsythe (1982), and an accretionary wedge dominated by basal accretion by Glodny et al. (2005) and Willner et al. (2005). Hervé (1988) presented a synthesis of the knowledge of this subduction complex, which has increased greatly in recent years, particularly with regard to geochronology and P–T regimes of metamorphism.

**The Western Series**

The Western Series (WS) comprises highly deformed metagreywackes with intercalations of metabasite (sometimes with relict pillow structures), meta-exhalites (spessartine–quartzite, stilpnomelane quartzite, massive sulphide, tourmalinite) and serpentinites; that is, it represents a mixture of continent-derived siliciclastics and slices of dismembered upper oceanic crust. A flat-lying east-dipping transposition foliation predominates. The WS shows a transitional contact with the Eastern Series at 35°30'S, and is in fault contact in other localities. These faults are associated with later destruction of the subduction complex. The fault contact at Pichilemu (35°S) was interpreted by Ernst (1975) as a ‘coastal suture zone’, but newly interpreted by Willner et al. (2005) to represent a Cretaceous brittle fault.

The occasional occurrence of glaucophanic amphibole along the belt and of lawsonite in Chiloé (Saliot 1968) suggested to Aguirre et al. (1972) that the Western Series constituted the high P–low T belt of the paired Late Palaeozoic metamorphic belts in central Chile, an interpretation confirmed by later studies. Munizaga et al. (1973) roughly limited the metamorphism of the basement of central Chile to the interval between 273 and 342 Ma (266–334 Ma with new decay constants) on the basis of Rb–Sr whole-rock systematics over a wide area. More specific data for exposures at different latitudes follow.

34°–36°S

Willner (2005) determined the metamorphic peak at 350 ± 50°C and 7–11 kbar, followed by static recrystallization during a pressure release of 3–4 kbar and slight cooling. Hervé et al. (1974) presented a 329 ± 22 Ma K–Ar age on glaucophane from Pichilemu, thus establishing a Late Carboniferous age for the high P–T event in this area. Willner et al. (2005) dated the metamorphic peak (Ar–Ar plateau ages, phengite) in the range 319 ± 1 to 292 ± 1 Ma, whereas Ar/Ar laser ablation ages in phengite in the range 322 ± 2 to 257 ± 3 Ma indicate gradient cooling during retrograde pressure release. The further cooling/exhumation of the WS was monitored through fission track zircon (206 ± 11 to 232 ± 14 Ma) and apatite (80 ± 4 to 113 ± 8 Ma). The end of accretion is marked by a late intrusion into the WS at Constitución at 224 ± 1 Ma (U–Pb, zircon). Average exhumation velocities were c. 0.2–0.6 mm/a indicating that erosion most probably was the prime exhumation factor.

38°–43°S

In this section of the WS, which has also been called the Bahía Mansa Metamorphic Complex, Massonne et al. (1998b) recorded the second known occurrence worldwide of the high pressure mineral zussmanite in an outcrop of the WS at Ninhue. The sulphide compositions in rare massive sulphide lenses indicate metamorphic pressures of 5–7 kbar (Collao et al. 1986). Willner et al. (2001) used multivariate equilibria calculations to establish a metamorphic peak at 270–370°C and 6–8 kbar at Bahía Mansa. Glodny et al. (2005) determined 420°C, 8–9 kbar in the Valdivia area. Exotic tectonic blocks within the WS at Los Pabilos (Willner et al. 2005) exhibit a counterclockwise P–T–t path with a metamorphic culmination at 11–16.5 kbar and 600–760°C, overprinted by an epidote-blueschist re-equilibration event at 350–400°C, 10–14 kbar, not recorded in the country rocks, and a further common re-equilibration for the blocks and the country rocks at c. 300°C and 5 kbar (Fig. 2.4). These rocks are considered to represent the earliest accreted rocks beneath a still-hot mantle. Maximum P–T conditions of 7–10 kbar, 350–450°C were derived from metapelitic rocks from the Western Series in Chiloé (Massonne et al. 1999; Hufmann 2002). This range matches those conditions abundantly derived from phengites in greenschists and metabreccias in central Chile (Massonne et al. 1998a).

Hervé et al. (1990) deduced the possible presence of ages in the range 300–330 Ma, based on Rb–Sr wr dating on rocks from five localities which bear massive sulphide mineralization. Söllner et al. (2000a) presented a 293 ± 23 Ma U–Pb zircon age on a meta-ignimbrite at Caleta Parga (41°30'S), and Duhart et al. (2001) a 396 ± 1 Ma age on a trachyte body emplaced in mafic schists, a similar age range to the previously reported U–Pb conventional ages obtained on detrital zircon (between 388 and 278 Ma) in different metasedimentary rocks. These results suggest that the WS accumulated from Early Devonian to Early Permian times and are in part contemporaneous with the cooling of other parts of the complex.
Phengites of the oldest accreted rocks at Los Pabilos were dated by Kato & Godoy (1995: 304±9 Ma) and Kato et al. (1997: 323±2 Ma), whereas Willner et al. (2004b) produced Rb–Sr mineral isochrons of 305±3 and 297±5 Ma intending to date a second retrograde stage of re-equilibration. Duhart et al. (2001) produced Ar/Ar ages of white mica in the entire area in the range 260–220 Ma believed to be cooling ages, and Glodny et al. (2002) obtained a Rb–Sr mineral isochron of 285 Ma at Quidico. Younger Rb–Sr mineral isochron ages (c. 250–245 Ma) have been presented by Glodny et al. (2005) for the Valdivia area that are interpreted as dating the deformation associated with basal accretion. Rb–Sr mineral isochrons dating younger deformations during exhumation at c. 235 Ma and c. 210 Ma limit the mean exhumation rate to 0.6+/−0.2 mm/a. Zircon fission track ages are in the range 176±49 to 212±46 in the Valdivia area (Glodny et al. 2005), whereas apatite fission track ages are between 53±7 Ma and 65±8 Ma.

In summary, these data allow us to deduce that basal accretion in the WS of central Chile was active for c. 100 Ma from Late Carboniferous to Early Triassic times.

**The Eastern Series**

The Eastern Series (ES) is mainly composed of metagreywackes of turbiditic origin accompanied by minor but ubiquitous calcisilicate pods and lenses. They show stratigraphic continuity at outcrop scale as well as a non-transposed first folding of bedding planes, and represent a weakly deformed retro-wedge area (Hervé et al. 1988; Willner et al. 2001). The predominant deformation style with upright tight chevron folds folding the bedding planes points to frontal accretion of sediments of a former stable continental margin as inferred by Glodny et al. (2005). In central Chile the ES is mostly overprinted by a post-kinematic high T – low P metamorphic event. Mapped north–south trending metamorphic zones display an increasing metamorphic grade towards the batholith of the Late Palaeozoic arc which intrudes the Eastern Series in its eastern part (González Bonorino 1971; Hervé 1977). The ES has been considered the low P – high T component of the paired metamorphic belts of central Chile, and its metamorphic grade locally attains the amphibolite–granulite facies transition. Peak P–T conditions of this high T metamorphism range from

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**Fig. 2.4.** A compilation of P–T–t trajectories for the different units of the ‘metamorphic basement’ complexes of central Chile. Sources of data: Western Series 1 (WS1 = exotic blocks), Willner et al. (2004b); Western Series 2 (WS2 = greenschist) and WS3 (blueschist), Willner et al. (2005); Eastern Series 1 (ES1), Willner et al. (2005); Eastern Series 2 (ES2), Hervé (1977).
c. 400°C to 720°C at 3 ± 0.5 kbar, indicating a regional metamorphic event causing a thermal dome with the batholith in its core (Willner 2005). Hervé (1977) suggested a P–T culmination at 4–5 kbar and 600°C for the metamorphic series in the Nahuelbuta Mountains.

Hervé et al. (1984) presented Rb–Sr whole-rock isochrons of 347 ± 32 Ma (andalusite zone) and of 368 ± 42 Ma (sillimanite zone) with corresponding K–Ar ages of 299 ± 10 Ma and 278 ± 7 Ma. Muscovite Ar/Ar plateau ages between 301 ± 1 Ma and 296 ± 2 Ma were obtained by Willner et al. (2005) on a broader area in the region, indicating a short-term thermal overprint during the emplacement of the 305 Ma old coastal batholith (Fig. 2.4). The cooling/exhumation of the ES was monitored through zircon fission track ages of 221 ± 12 Ma and 215 ± 14 Ma and apatite fission track ages of 98 ± 6 Ma to 113 ± 8 Ma.

**The Coastal Batholith**

The Coastal Batholith (CB) of central Chile (Fig. 2.1) is mainly composed of calcalkaline granitoids of Late Carboniferous to Permian ages. The granitoids were intruded into the Eastern Series of the metamorphic basement, which was subject to contemporaneous deformation and metamorphism. Triassic high level plutons of limited areal extent occur as post-tectonic bodies in the high P–T Western Series north of 38°S, and similar Cretaceous plutons occur around 40°S. Jurassic plutons, in many places difficult to distinguish from the Palaeozoic ones, increase in volume towards the northern limit of the area and beyond it, and were emplaced when the tectonic and metamorphic activity had ceased in the accretionary complex to the west. Later Mesozoic and Cenozoic plutonism migrated eastward to the Main Range cross-cutting the NW–SE trend of the southern end of the Late Palaeozoic plutonic belt at 40°S.

Stratigraphic evidence for the Late Palaeozoic age of the CB is scarce, although Carnian to Rhaetian sedimentary successions lie unconformably over the CB and the ES (but not the WS) in several localities in the Coast Ranges between 34° and 37°S, and also in the Main Range exposures at 40°S. This stratigraphical relationship limits the age of the unroofing of the batholith.

Near the northern end of the Coastal Batholith, Gana & Tosdal (1996) determined a U–Pb zircon age of 299 ± 10 Ma for the Miraisol unit, and 214 ± 1 Ma for a gneissic tonalite at Cartagena. The late Carboniferous age is similar to previously reported Rb–Sr, U–Pb and K–Ar ages in the same area, and the Late Triassic pluton has equivalents north and south of it. Further dating of the CB in this area by Gana & Tosdal (1996) shows that Middle Jurassic plutons in the age range of 156–161 Ma are widespread in this northern portion of the CB. Finally, for the Nahuelbuta Central Pluton Rb–Sr whole-rock isochrons were presented by Hervé et al. (1988) (294 ± 24 Ma) and Lucassen et al. (2004) (306 ± 5 Ma), which are concordant with the 305 ± 1 Ma U–Pb age on zircon in the Pichilemu area (Willner et al. 2005). Similar U–Pb zircon ages in the range 300 ± 2 Ma to 305 ± 2 Ma were obtained by Martin et al. (1999) at 40°S.

**The metamorphic complexes of the Patagonian and Fuegan Andes**

In the Patagonian and Fuegan Andes, metamorphic rock units crop out quite extensively. They have been referred to as a ‘metamorphic basement’ to the Mesozoic and Cenozoic sedimentary and volcanic units. In the latter, it is possible to investigate the age and geological evolution by means of classic stratigraphic and palaeontological methods. The ‘metamorphic basement’, on the contrary, is for the most part composed of polydeformed rocks, where no stratigraphic controls can be established, and which contain very little or no fossil evidence for their depositional age.

The application of new methods, such as SHRIMP U–Pb determination of the detrital zircon age spectra, geochemical provenance analysis, and determination of metamorphic P–T conditions, have allowed researchers over the past decade to acquire new insights on the geological evolution of the ‘metamorphic basement’ of the Patagonian Andes. As a consequence, units differing in depositional and metamorphic ages, geodynamic setting and metamorphic characteristics have been identified. A description of their lithologies, metamorphic characteristics and geodynamic significance is given below.

**The Patagonian Andes**

The Patagonian Andes consist of a rather topographically subdued mountain belt that has had a prolonged evolution going back to Late Palaeozoic times. The backbone of these mountains is provided by the Mesozoic to Cenozoic Patagonian Batholith, whose earliest (c. 150 Ma) components intrude low grade metamorphic complexes, which at present crop out both west and east of the continuous batholithic belt. These complexes were classically considered to be time equivalents, and are represented as such on the 1:1,000,000 Geological Map of Chile (Escobar et al. 1980). Research work in the last decade, however, has modified this view and allowed a subdivision of these units, the whereabouts and extent of which are presented in Figures 2.1 and 2.5. A summary of P–T–t trajectories is given in Figure 2.6.

**Eastern Andes Metamorphic Complex**

This unit consists mainly of polydeformed turbidite successions, with minor bodies of limestones and metabasites. It includes the previously defined Cochrane and Lago General Carrera units (Legally 1975), Bahía de la Lancea and Río Lacteo formations, as well as the Staines Complex (Allen 1982). The regional metamorphic grade is in the greenschist facies or lower, with higher grade rocks appearing only in the contact aureoles of Mesozoic to Cenozoic intrusions (Calderón 2000). Hervé et al. (2003a) concluded that this unit has sedimentary components deposited during Late Devonian–Early Carboniferous times, as well as younger deposits in their western outcrop areas, ranging in age up into the Permian period. Hervé et al. (1998), Faúndez et al. (2002), Ramírez (2002), Augustsson & Bahlburg (2003) and Lacassie (2003) have suggested that the turbidites represent deposition in a passive continental margin environment. This interpretation was based mainly on provenance considerations from petrographic and geochemical data. The turbidites were derived from a cratonic source, which possibly had undergone a complex and extended sedimentary recycling history. A combination of U–Pb detrital zircon ages and fission track age data on the same zircons allowed Thomson & Hervé (2002) to conclude that these sediments were metamorphosed before Late Permian times under lower P–T metamorphic conditions than those that are typical of accretionary complexes (Ramírez 2002), as shown in Table 2.1.

**Puerto Edén Igneous and Metamorphic Complex**

This complex consists of medium to high grade metamorphic rocks, migmatites and plutonic rocks, which crop out east of the Southern Patagonian batholith (49°S). Geothermobarometric constraints indicate a nearly isobaric high T – low P metamorphic and partial melting event superimposed on earlier greenschist-facies metamorphic rocks (Calderón 2000). Metamorphic overgrowths on zircons in sillimanite paragneisses record a Late Jurassic (c. 150 Ma) age taken as evidence of local gneiss formation under in situ anatectic conditions during the emplacement of the Jurassic components of the batholith (Hervé et al. 2003a).
Coastal accretionary complexes
These comprise, from north to south, the Chonos Metamorphic Complex (CMC), the Madre de Dios Accretionary Complex (MDAC) and the Diego de Almagro Metamorphic Complex (DAMC), all of which crop out west of the Patagonian Batholith (PB) (Fig. 2.5).

The CMC consists predominantly of metaturbidites (Pimpirev et al. 1999) with restricted occurrences of mafic schists and metacherts, and broken formations are conspicuous. It has a Late Triassic depositional age, as indicated by fossil fauna (Fang et al. 1998) and by U–Pb age determinations on detrital zircon using SHRIMP (Hervé & Fanning 2001). The complex has been divided into two belts by Hervé et al. (1981b) similar to central Chile, with an Eastern Belt having well preserved primary sedimentary and volcanic structures that become progressively obliterated when passing into the more pervasively deformed and recrystallized rocks of the Western Belt. They were metamorphosed under high P–T metamorphic conditions (Willner et al. 2002) as shown in Table 2.1, before or during Early Jurassic times (Thomson & Hervé 2002).

The MDAC is composed of three tectonically interleaved lithostratigraphic units: the Tarlton limestone, the Denaro (DC) and the Duque de York (DC) complexes (Forsythe & Mpodozis 1979, 1983). The Tarlton limestone (TL), a massive pelagic limestone body, was deposited in an intra-oceanic carbonate platform during Late Carboniferous–Early Permian
times (Douglass & Nestell 1976), over a penecontemporaneous (Ling et al. 1985) oceanic substrate (the DC) composed of pillow basalts, metalliferous and radiolarian cherts, probably in an oceanic ridge environment away from the continental influence of Gondwana. This exotic terrane was later accreted to the continental margin. Large upright chevron folds of bedding planes combined with steep reverse thrusts resemble accretionary prisms formed by frontal accretion (Forsythe & Mpodozis 1979, 1983). The DYC is a turbiditic succession which was deposited uncomformably over the TL and the DC, when they reached the vicinity of the continental margin (Forsythe & Mpodozis 1979, 1983). The DYC has radiolarian cherts at Desolación Island, which indicate an Early Permian age of deposition (Yoshiaki, written communication, 2002), and detrital zircons of late Early Permian age in all main outcrops of the unit. The Duque de York Complex, and probably the underlying TL and DC, were metamorphosed before or within the earliest Jurassic (Thomson & Hervé 2002). The very low grade metamorphism (pumpellyite–actinolite facies) of the three units of the MDAC has been sparsely studied: only the metamorphic characteristics of the Denaro Complex are shown in Table 2.1, and these suggest that the MDAC was metamorphosed under a relatively high geotherm of 15–20°C/km, similar to the Eastern Belt of the CMC. Combined petrographic and geochemical analyses (Lacassie 2003) indicate that the greywackes of the DYC were derived from an intermediate (granodioritic) composition igneous source within a dissected magmatic arc tectonic setting, where erosion had enough time to expose its plutonic roots. The DYC basin was probably adjacent to the continental crust of Gondwana, in an active margin tectonic setting (Faúndez et al. 2002; Lacassie 2003).

Palaeo-magnetic data on the Tarlton limestone and the Denaro complex (Rapalini et al. 2001) indicate that these
units have undergone a very large counterclockwise rotation (117 ± 29.9°), with negligible palaeo-latitude anomaly, after Early Cretaceous remagnetization by the thermal influence of the South Patagonian Batholith. This evidence allowed the above-cited authors to conclude that the rock units involved had been accreted to the Gondwana margin from the NW rather than from the SW as had been previously suggested by Forsythe & Mpodozis (1979, 1983) on the basis of structural studies. The DAMC is composed of two subunits of differing metamorphic imprint, one composed of garnet amphibolites and blueschists, the other of quartz–mica schists and an orthogneiss. The contact between them has not been observed in the field. SHRIMP U–Pb ages in zircons in both the orthogneiss and a quartz–rich spessartine-bearing schist interleaved in the blueschists (Herve & Fanning 2003), have yielded Middle Jurassic ages interpreted as the age of crystallization of their igneous protoliths: a muscovite–garnet–bearing granite, and a rhyolitic rock respectively contemporaneous with the massif, have been studied recently by Pankhurst & Rapela (1995). The subsequent high P–T metamorphism occurred during Cretaceous times (Herve et al. 1999). This complex is in tectonic contact (Forsythe 1981, 1982) with the DYC along the mid-crustal sinistral strike-slip Seno Arcabuz shear zone (Olivares et al. 2003).

In the Diego Ramirez islands, pillow basalts and metasediments form a crush mélange (Wilson et al. 1989), with glaucophane-bearing metamorphic assemblages. A Middle Jurassic Rb–Sr whole-rock isochron (Davidson et al. 1989) suggests that these rocks can be correlated with those of the DAMC. Within this context it is relevant to note that non-basement rocks, such as recently studied metarhyolites of the Jurassic Tobifera Formation in the Magellanes fold and thrust belt (Herve et al. 2004), have also been affected by contemporaneous metamorphic events also characterized by high P–T conditions.

The Fuegan Andes

The Darwin Cordillera Metamorphic Complex

The basement rocks of the Darwin Cordillera (DCMC) consist of metasedimentary and metavolcanic units, of supposedly late Palaeozoic to early Mesozoic age, which have a Mesozoic metamorphic imprint peculiar to that area (Kohn et al. 1995). This metamorphism is characterized by the generation of biotite, stilpnomelane, kyanite and sillimanite zones, which are unique among the metamorphic basement complexes of the Patagonian and Fuegan Andes. Several authors (Dalziel & Cortes 1972; Nelson et al. 1980; Dalziel 1981, 1986) have suggested that their protoliths formed as an accretionary wedge on the pre-Middle Jurassic Pacific margin of South America. However, it is not known at present if these metamorphic rocks of the Darwin Cordillera were originally part of the Eastern Andes Metamorphic Complex, which is probably not a Late Palaeozoic accretionary complex but served as a backstop during the generation of the accretionary wedge (Augustsson & Bahlburg 2003), or if they were part of the coastal accretionary complexes of the Patagonian Andes. An orthogneiss within the DCMC showed a Middle Jurassic Rb–Sr whole-rock isochron (157 ± 7 Ma; Herve et al. 1979b, 1981c) and U–Pb zircon (164 ± 1 Ma; Mukasa & Dalziel 1996) ages affected by the main metamorphic event. Nelson et al. (1980) have suggested that part of the protolith of the complex might be the Middle to Late Jurassic Tobifera Formation silicic volcanic rocks.

Geodynamic considerations

The study of basement metamorphic rocks in Chile, and related complexes in neighbouring Argentina, provides indications about the changing tectonic environments that have existed in the area. It is clear that these metamorphic complexes have different ages and that their metamorphic evolution, though poorly known in some areas, varies widely in space and time. Probably the best known and most readily interpreted event recorded by these rocks is the Late Palaeozoic development of an accretionary prism over an east-dipping subduction zone below the southwestern Gondwana margin in central Chile. This process generated an elongated high P–low T metamorphic belt which includes accreted oceanic lithologies and a contemporaneous parallel magmatic belt in the upper plate, both of which are well recorded and exposed.

The existence of tectonostratigraphic terranes has been identified mainly by the presence of oceanic rocks flanked on both sides by blocks of older continental crust. This interpretation is

<table>
<thead>
<tr>
<th>Metamorphic complex</th>
<th>Maximum pressure (kbar)</th>
<th>Maximum temperature (°C)</th>
<th>Ages of emplacement and metamorphism (Ma)</th>
<th>Reference for peak P–T conditions</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chonos</td>
<td>(EB) 4.5–5.5</td>
<td>250–280</td>
<td>213–198*</td>
<td>Willner et al. 2002</td>
</tr>
<tr>
<td></td>
<td>(WB) 8.0–10.0</td>
<td>380–500</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Denaro</td>
<td>4.1–5.5</td>
<td>260–300</td>
<td>234–195*</td>
<td>Sepulveda (2004), Willner (unpubl. data)</td>
</tr>
<tr>
<td>Diego de Almagro</td>
<td>(BS) 9.5–13.5</td>
<td>380–450</td>
<td>157–110</td>
<td>Willner et al. (2004a)</td>
</tr>
<tr>
<td></td>
<td>(GA) 11.2–13.2</td>
<td>460–565</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>(GM) 4.9–6.5</td>
<td>580–690</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

For the Diego de Almagro Complex: BS, blueschists; GA, garnet amphibolites; GM, garnet–mica schists. The age column indicates first the depositional/emplacement ages and then the maximum metamorphic age as obtained through fission tracts zircon or Ar/Ar data. Ages marked with an asterisk are from Thomson & Hervé (2002). The age indicated for the Denaro Complex is at obtained for the Duque de York Complex. The authors cited in the last column have determined the peak P–T conditions of metamorphism.

Geodynamic considerations

The study of basement metamorphic rocks in Chile, and related complexes in neighbouring Argentina, provides indications about the changing tectonic environments that have existed in the area. It is clear that these metamorphic complexes have different ages and that their metamorphic evolution, though poorly known in some areas, varies widely in space and time. Probably the best known and most readily interpreted event recorded by these rocks is the Late Palaeozoic development of an accretionary prism over an east-dipping subduction zone below the southwestern Gondwana margin in central Chile. This process generated an elongated high P–low T metamorphic belt which includes accreted oceanic lithologies and a contemporaneous parallel magmatic belt in the upper plate, both of which are well recorded and exposed.

The existence of tectonostratigraphic terranes has been identified mainly by the presence of oceanic rocks flanked on both sides by blocks of older continental crust. This interpretation is
less straightforward, as it depends heavily on the evaluation of structural, geochemical and geochronological data, which can be imprecise and incomplete. Also, the sudden displacement of magmatic arcs in time has been considered to imply the docking of terranes to the continental margin. In contrast, a paucity or absence of magmatic and tectonic activity, together with developing sedimentary provenance analysis, has been used as an indicator of passive margin conditions. The subduction of upstanding oceanic features such as ridges and plateaux can bring subduction to a stop, or flatten the subducting slab to prevent magmatic activity near the trench. Finally, margin-parallel strike-slip environments have also been considered in some interpretations, and the process of tectonic erosion of the leading edge of the overriding plate (Stern 1991b) may destroy the continuity, or even the entire evidence for the generation of subduction complexes in a particular place and time.

Northern Chile

This is the only portion of the Andean basement in Chile where significant bodies of Early (pre-Devonian) Palaeozoic rocks exist, including latest Proterozoic (?)–Early Cambrian igneous and metamorphic suites. Dalziel (1991) and Moores (1991) suggested that the Laurentian craton was located close to Antarctica and South America in Neoproterozoic and Early Palaeozoic times. The clockwise movement of Laurentia around South America would have resulted in repeated collisional tectonic interaction between the two continents, giving rise to the Famatinian orogen in southern South America and to the Taconic orogen in Laurentia (Dalla Salda et al. 1992). The collision included two major events, resulting in the Early Cambrian Pampean Orogeny, and later the late Ordovician Famatinian orogeny. A sliver of Laurentia detached to constitute the Arequipa–Antofalla terrane, with the Belén and Chojas metamorphic complexes near its eastern margin providing a record of the Pampean orogeny. The Ocosta orogen could have been caused by the collision of this detached sliver – the Arequipa–Antofalla terrane – and not of the whole of Laurentia, if the Cordón de Lila Complex is interpreted as the product of a magmatic arc caused by eastward subduction of oceanic crust (Bahlburg & Hervé 1997) from the west (Fig. 2.7a). Alternatively, Forsythe (1982) have suggested that the Arequipa–Antofalla terrane was a sliver of the Gondwana continent, and rotated clockwise to generate the oceanic area in the Ordovician, and then counterclockwise to collide with its continent of origin, producing the Ocoyoc orogen. Lucassen et al. (2000) questioned the entire terrane concept and show that Lower Palaeozoic metamorphic crystallization ages, peak P–T data (high temperatures, low to moderate pressures, Fig. 2.2) and whole-rock Nd–Sm systematics rather favour a continuous mobile belt in northern Chile and NW Argentina, i.e. south of the Arequipa Massif at 18°S and north of the Argentine Precordillera at 28°S. This basement would have originated at a convergent margin at mid-crustal levels under a very high geothermal gradient involving intensive magmatic underplating over wide areas. In the northernmost exposures the Arequipa craton would have been partly reworked within the mobile belt. After the Silurian–Devonian lull in subduction activity, the margin of Gondwana resumed as an active margin facing and consuming an extensive oceanic plate. As a result of this, the metamorphism of the Chañaral Melange (Fig. 2.7b) took place during Late Carboniferous times within what Mariotti & Bahlburg (2003) describe as a particular type of subduction zone in which the P–T regime was not high. Around this time, the Limón Verde high pressure metamorphism occurred. As this unit crops out far from the present-day coastline and the Chañaral melange, the interpretation of its tectonic setting has been difficult. Hervé et al. (1985) and Bahlburg & Hervé (1997) favoured a subduction zone environment whereas Lucassen et al. (1999) preferred a strike-slip environment for its development.

The way in which the Limón Verde Metamorphic Complex was exhumed remains questionable, but it was broadly contemporaneous with the development of mylonitic rocks at Sierra de Moreno and in the Tránsito Metamorphic Complex, north and south respectively of Limón Verde, which exhumed the Belén and Sierra de Moreno complexes and the Pampa Gneiss, then part of the backstop of the accretionary prism. It is possible that this rapid exhumation might be related to tectonic erosion of the oceanward portions of the backstop, or by subduction of an oceanic ridge, favouring the exhumation process which brought these units to the surface in mid-Triassic times. In Figure 2.1, these units are seen aligned in what has been referred to as the Precordillera upf thrust belt, after Bahlburg & Breitkreuz (1991).

Norte Chico

As a whole this region is characterized by the presence of the (hardly exposed) Chilenia Terrane, which was amalgamated to Gondwana in Devonian times (Ramos et al. 1986). As seen further north, the Late Carboniferous Tránsito Metamorphic Complex crops out in a much more easterly position than the Choapa and El Teniente complexes (Fig. 2.7c), which both have a lower metamorphic grade. The El Teniente rocks additionally record a late Triassic metamorphic event, not recognized in those of El Tránsito, which at that time had already been exhumed and exposed at the surface. It is possible that this exhumation was contemporaneous with that of the Limón Verde rocks mentioned above, thus representing a wide-scale exhumation of the backstop and of the deeper parts of the accretionary wedge which continued to develop further west. A westward jump of magmatic activity in this area from Carboniferous to Triassic times, could be interpreted as caused by a roll-back of the subduction zone by the collision of a small terrane (Fig. 2.7d) that could have contributed to the exhumation of the early subducted rocks of the El Teniente Metamorphic Complex (‘Terrane Equis’; Mpodozis & Kay 1990, 1992). Finally, it is interesting to note the existence of the late Triassic metamorphic event in the coastal El Teniente exposures, contemporaneous with the Chonos event in Patagonia, which has no counterpart in central Chile.

Central Chile

The geological development of central Chile in Late Palaeozoic times can be best described in terms of processes occurring along a continental margin. These processes can be related to subduction of oceanic crust leading to an accretionary prism and a magmatic arc (Fig. 2.7e). The Western Series includes lithologies of oceanic parentage, mixed with detrital sediments of continental derivation, and exhibits the imprint of a high P–low T metamorphic regime. The history of metamorphism and exhumation and the ductile deformation shown by the WS can best be explained if the outcropping rocks were initially subjected to basal accretion in the accretionary complex, followed by exhumation from 25–40 km depth during ongoing accretion. The Eastern Series, mainly composed of a metamorphosed turbidite succession, was probably deposited in a forearc setting over the continental shelf. The deposition may have taken place during passive margin conditions in Devonian and Early Carboniferous times. Thus, the rocks of the ES were probably part of the uppermost accretionary wedge formed by frontal accretion as well as part of the retrowedge. Deformation took place mainly under very low-grade conditions and intermediate pressure, but the ES was later metamorphosed under low P–high T conditions close to the site where the magmatic arc developed.

Patagonian Andes

The protoliths to the eastern Andes Metamorphic Complex were deposited in a passive margin environment from Early
Devonian to Early(?) Permian times. The detrital zircon age spectra on these rocks have Gondwanan affinities. The source areas could have been located in older rocks of the Atlantic margin of Patagonia, such as the Deseado massif, or in South Africa and Antarctica (Hervé et al. 2003a). It is not known if it is in place with respect to the older continental blocks, or if it has been displaced (c. Mpodozis, personal communication).

The western accretionary complexes, on the contrary, have evolved in subduction zone environments, where accretion of ocean floor basaltic material is recorded. However, in contrast to previous assumptions, there are no indications of Late Palaeozoic subduction in the Patagonian Andes. The Chonos Metamorphic Complex reveals a subduction event near the Triassic–Jurassic boundary (Fig. 2.7f), in which the corresponding arc might have been the largely coeval Sub-Cordilleran Batholith (Rapela et al. 2003b) in Argentina. The Madre de Dios Accretionary Complex appears to involve a composite exotic terrane, probably frontally accreted to the Gondwana margin during the same Late Triassic–Early Jurassic event as the CMC. The provenance of the Duque de York Complex, characterized by a major Early Permian zircon component, is not easily attached to a contemporaneous

Fig. 2.7. Schematic cross-sections at different latitudes and time slots during the main subduction-related metamorphic and plutonic events as described in the text. Scales are not uniform.
magmatic arc in Patagonia, where Permian igneous rocks do not crop out extensively. Lacassie (2003) suggests a far-sited origin for the MDAC, which would have collided with the Gondwana margin in the southeastern Pacific area (present coordinates) and then transported along the margin to its present position, following a model proposed by Cawood et al. (2002). Alternatively, as the accretion of the MDAC was from the NW, a possibility remains that the limestones originated in an oceanic environment at the latitude of northern Chile, where thick limestone deposits of Early Permian age were deposited over continental crust, mainly in Bolivia (Copacabana Formation) but also in Chile.

Hervé & Fanning (2003) have suggested that the Late Jurassic–Cretaceous evolution of the Diego de Almagro Metamorphic Complex in a subduction zone along the western margin of Gondwana only occurred after the Antarctic Peninsula, which could have been located outboard of the present continental margin (Lawver et al. 1998), started to drift south allowing subduction to occur near the present-day continental margin.

The P–T evolution of the metamorphic complexes (Figs 2.2, 2.3, 2.4 & 2.6) varied widely in time and space. The differences between the coastal accretionary complexes, which evolved in P–T regimes characterized by geothermal gradients between 10 and 20°C/km, and the Puerto Edén and Cordillera Darwin metamorphic complexes are evident. Only the former are considered to be typical of subduction zone environments due to the derived metamorphic P–T conditions. These conditions suggest that the subduction was slow, or that the subducting oceanic lithosphere was rather young and, thus, relatively hot.

Concluding remarks

The metamorphic and plutonic basement complexes of Chile thus reveal the following history (sketched in Fig. 2.7) for a part of the southwestern Gondwana margin.

1. Early Palaeozoic metamorphic and plutonic events occurred, which can be assigned to the Pampean and Famatinian orogenic phases. The products of these events seem to be restricted to northernmost and southernmost Chile as there are no outcrops in between.
2. Conditions of a passive margin prevailed in the region during Silurian and Devonian times, when the collision of Chilenia is recorded in western Argentina.
3. A Late Carboniferous metamorphic and plutonic event, related to subduction of the Palaeopacific or Panthalassian ocean, is recorded in areas between latitudes 42°S and 23°S. This event has been called the Toco Orogeny in northern Chile. A chain of accretionary prisms is preserved with ages decreasing from north to south.
4. Late Triassic–Early Jurassic metamorphic and plutonic activities are recorded in the Norte Chico (30–33°S) and in the northern Patagonian archipelagos (43–46°S). Plutonic rocks of the same age were emplaced in the Western Series. It is not known if these two outcrop areas represent two ends of a previously continuous accretionary complex located west of the present-day continental margin that was subducted by tectonic erosion or displaced south along the Jurassic to Cretaceous left-lateral strike-slip faults.
5. Evidence for deep seated Jurassic–Cretaceous tectono-metamorphism related to an Andean setting has been found only south of 48°S. This includes the Diego de Almagro (subduction) and the Darwin (metamorphic core) complexes.

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3 Tectonostratigraphic evolution of the Andean Orogen in Chile

REYNALDO CHARRIER (coordinator), LUISA PINTO & MARÍA PÍA RODRÍGUEZ

Since the comprehensive synthesis on the Argentine–Chilean Andes by Mpodozis & Ramos (1989), important progress has been made on the stratigraphy, palaeogeographic evolution and tectonic development of the Andean Orogen in Chile. We present here an overview of this evolution considering the new information and interpretations, including some unpublished ideas of the authors. To enable the reader to delve further into the subjects treated here, we accompany the text with abundant references. In the interpretation of the stratigraphic and radioisotopic data we used the timescale of Harland et al. (1989).

During most of its history the continental margin of South America was an active plate margin. The Late Proterozoic to Late Palaeozoic evolution was punctuated by terrane accretion and westward arc migration, and can be described as a ‘collisional history’. Although accretion of some terranes has been documented for the post-Triassic history, the evolution during post-Triassic times is characterized more by the eastward retreat of the continental margin and westward arc migration, attributed to subduction erosion, and therefore can be described as an ‘erosional history’. The intermediate period, comprising the Late Permian and the Triassic, corresponds to an episode of no, or very slow, subduction activity along the continental margin, during which a totally different palaeogeographic organization was developed and a widely distributed magmatism with essentially different affinities occurred. It is therefore possible to differentiate major stages in the tectonostratigraphic evolution of the Chilean Andes, which can be related to the following episodes of supercontinent evolution: (1) post-Pangaea II break-up; (2) Gondwanaland assembly; and (3) break-up of Gondwana. These stages can in turn be subdivided into shorter tectonic cycles separated from each other by regional unconformities or by significant palaeogeographic changes that indicate the occurrence of drastic tectonic events in the continental margin. These tectonic events have been related to modifications in the arrangement and dynamics of the lithospheric plates (see James 1971; Rutland 1971; Charrier 1973a; Aguirre et al. 1974; Frutos 1981; Jordan et al. 1983a, 1997; Malumián & Ramos 1984; Ramos et al. 1986; Isacks 1988; Ramos 1988b; Mpodozis & Ramos 1989).

Morphotectonic features and subdivision of the Andean Orogen in Chile

Within the Andean Orogen, which is the first-order morphologic element in this region, it is possible to differentiate two other types of features: morphostructural units orientated parallel to the strike of the range, and oroclines bending around which are major changes in the orientation of the morphology and structure of the range (Fig. 3.1). Two oroclines are present, comprising the Bolivian and the Patagonian oroclines, in northernmost and southernmost Chile respectively. The continuity of the strike-parallel morphostructural units is interrupted in the regions where the Juan Fernández and the Chile ridges intersect the continental margin, causing segmentation of the orogen (Fig. 3.2). The region where the passive Juan Fernández Ridge is subducting the continental margin (between c. 27°S to c. 33°S) corresponds to a flat-slab subduction zone, whereas in the regions north and south of this flat-slab segment the Wadati–Benioff zone is steeper (Cahill & Isacks 1992). Further south (46–47°S), the intersection of the active Chile Ridge and the continental margin determines the existence of the Taitao triple-junction.

In the Chilean Andes north of the Taitao triple-junction, the main morphological change caused by the flat-slab subduction zone is the absence of the Central or Longitudinal Depression, a morphological unit that separates the Coastal Cordillera from the Principal or Main Cordillera (Fig. 3.2a). In this region it is therefore not possible to differentiate between these two cordilleras. This situation determines the existence of two segments, one between 18° and 27°S and the other between 33° and 46°S, in which the Central Depression is well developed, and an intermediate segment which lacks a Central Depression (27–33°S), called the zone of transverse river valleys or Norte Chico. On the Chilean side of the Andes, absence of the Central Depression in the flat-slab segment is associated with absence of recent volcanic activity, indicating that the subduction of the Juan Fernández Ridge controls morphology, magmatism and tectonics. In fact, the existence of the flat-slab subduction also causes important along-strike morphologic and tectonic variations on the Argentinian side of the Andes (see Ramos et al. 2002). South of the Taitao triple-junction a drastic change occurs in the morphostructural units and the general orographic pattern of the cordillera (Fig. 3.2b). The morphostructural units in this southern region comprise, from west to east: the Archipelago, the Patagonian Cordillera and the Precordillera.

Based on the morphologic and tectonic differences occurring at the intersections of the continental margin with oceanic ridges, Aubouin et al. (1973b) and Gansser (1973) subdivided the Andean Range into three main regions: the Northern, the Central and the Southern Andes (Fig. 3.1a). The Chilean Andes form part of the southern Central Andes, north of the Taitao triple-junction, and the Southern Andes, south of the triple-junction. North of the Taitao triple-junction, convergence between the Nazca and South American plates is essentially orthogonal,
Fig. 3.1. Location of the Central and Southern Andes relative to South America, and major morphologic elements of the Andean Cordillera and oceanic plates facing the western margin of southern South America. (a) Location of the Andean Cordillera relative to South America, major subdivisions along the mountain range, and major tectonic and morphologic elements in the oceanic plates facing western South America: the Nazca and Pacific–Antarctic plates, separated by the Chile Ridge, and from north to south the Carnegie, Nazca and Juan Fernández passive ridges (in light grey). (b) Major morphological features of the Central and Southern Central and Southern Andes in Chile (17°45′ to 56°S): Bolivian and Patagonian Oroclines, Andean segmentation, and morphostructural units: 1, Coastal Cordillera (Coastal Ranges, between 27° and 33°S); 2, Central Depression; 3, Forearc Precordillera and Western Cordillera, between 18° and 27°S, High Andean Range, between 27°S and 33°S (flat-slab subduction segment), Principal Cordillera, between 33°S and c. 42°S; 4, Patagonian Cordillera; 5, Andean foreland in the southern Patagonian Cordillera. (c) Schematic section across the northern high-angle subduction segment (Central Volcanic Zone) showing distribution of the morphostructural units. (d) Schematic section across the flat-slab segment. (e) Schematic map and section across the high-angle subduction segment (Southern Volcanic Zone) showing distribution of the morphostructural units. (f) Schematic section across the Southern Andes showing distribution of the morphostructural units. Abbreviations: AB, Arica Basin; CC, Coastal Cordillera; CD, Central Depression; EC, Eastern Cordillera; FC, Frontal Cordillera; FP, Forearc Precordillera (northern Chile); P, Precordillera (in Argentina between 27° and 33°S); PA, Patagonian Archipelago; PAC, Patagonian Cordillera; PC, Principal Cordillera; PR, Pampean Ranges; SS, Subandean Sierras; T, Trench; WC, Western Cordillera.
whereas south of the triple-junction convergence gradually changes from nearly orthogonal to oblique, and finally to parallel to the continental margin, in southernmost Chile.

Pampean tectonic cycle (Late Proterozoic–Early Cambrian)

Tectonic framework

The palaeo-tectonic reconstructions for the Late Proterozoic and Early Palaeozoic have provoked considerable debate (see Chapter 2). Some authors favour the view that at this time Laurentia was drifting northward along the western margin of Gondwana (the western margin of present-day South America) with a clockwise rotation (i.e. Dalziel et al. 1994). It has been proposed that during its displacement Laurentia interacted repeatedly with South America and that this interaction resulted in the deformation of the margin of South America and in the rifting off of fragments of eastern Laurentia, which were later accreted to the continent (Dalziel et al. 1994), i.e. the Arequipa–Antofalla Terrane (AAT) or Arequipa Massif (AM), and the Precordillera Terrane (Ramos 1994; Loewy et al. 2004). Although the exact age of these events is unclear, the main collisional event most probably occurred during Ordovician times, that is, during the next tectonic cycle (Bahlburg & Hervé 1997). Other authors envisage that the region presently located in northern Chile and Argentina was an active continental margin associated with a ‘mobile belt’ (Puncoviscana Basin) located to the east, with accretionary events occurring further south (Lucassen et al. 2000).

Geological units

Stratified rocks belonging to this cycle have not been found in Chile. Proterozoic U–Pb ages (see Chapter 2) have been obtained for scattered localities by Damm et al. (1990) on a granodiorite in the Sierra de Limón Verde (777±36−35 Ma), and on an orthogneiss and a migmatite in the Choja Metamorphic Complex in the Sierra de Moreno (1254±97−94 Ma and 1213±28−25 Ma). However, ⁸⁷Rb–⁸⁶Sr whole-rock isochrons on migmatites and schists from the Sierra de Moreno gave Early Palaeozoic ages (511±9 Ma, and 485±12 Ma) (Skarmeta 1983). A somewhat older age has been obtained in the Mejillones Peninsula, at 23–23º30’S (c. 530 Ma; ⁸⁷Rb–⁸⁶Sr, whole rock) (Díaz et al. 1985), and from a granodiorite recovered from a drill hole in extra-Andean Tierra del Fuego (intrusion U–Pb age of 529±7.5 Ma; concordia intercept on unabraded zircons) (Söllner et al. 2000b) (Fig. 3.3a).

The ⁸⁷Rb–⁸⁶Sr ages mentioned do not differ much from the ones more recently obtained by the same method in the Complejo Metamórfico de Belén (CMB) (Fig. 3.3b). This complex, exposed in the Precordillera in the forearc in northernmost Chile, has also been assigned a Proterozoic age (1000 Ma) (Pacci et al. 1980). However, new radiisotopic dating has yielded ages of 544±22 Ma (Rb–Sr whole-rock isochron) in quartz–mica schists, and of 536 to 516 Ma (K–Ar in minerals) in schists (Basei et al. 1996). Slightly younger ages (U–Pb in zircon) than the ones indicated above, although concordant
within errors, of 507 ± 48 Ma and 475 ± 31 Ma were obtained in an orthogneiss body and granitic veins respectively (Basei et al. 1996). For the orthogneiss body, Loewy et al. (2004) obtained a U–Pb in zircon age of 507 ± 48 Ma. Basei et al. (1996) consider that the older ages correspond to the regional metamorphic event of the CMB in Cambrian times and that the younger ones indicate an uplift and cooling event of the metamorphic rocks in Early Ordovician times.

The CMB is mainly composed of foliated amphibole and subordinately quartz–mica schists, gneissic schists, orthogneisses and serpentinites. It is intruded by a small gabbro stock and mafic, aplitic and felsic dykes, and forms a narrow north–south trending strip along a high-angle, west-vergent thrust system located in the western slope of the Chilean Altiplano or forearc Precordillera between Chapiquiña and Tignamar (Salas et al. 1966; Pacci et al. 1980; Basei et al. 1996; Muñoz & Charrier 1996; García 1996, 2002; García et al. 1996; Heber 1997; Lezaun 1997; Lucassen et al. 2000; Wörner et al. 2000c; García et al. 2004). The CMB has been uplifted by these faults and thrust over Cenozoic deposits.

Stratified rocks of the same age as those of the CMB are found in the Proterozoic to Lower Cambrian metaturbiditic Puncoviscana Formation, largely exposed in the Eastern Cordillera in northwestern Argentina (Turner 1960). Trace fossils in this formation indicate a westward deepening of the depositional environment (Aceñolaza et al. 1988, 1990) suggesting the existence of a west-facing marine platform east of the granite into the schist. According to the age of the granite, the schists can be assigned a pre-Silurian age, which would permit comparison with the Ordovician deposits described next; although, based on the metamorphism of these rocks, a correlation with the CMB seems more probable.

**Discussion**

Based on the Proterozoic age previously reported for these rocks, the CMB was considered to be part of the Arequipa–Antofalla Terrane (see Ramos et al. 1986; Ramos 1988b), rocks of which are exposed in southwestern Peru (Dalmayrac et al. 1980), northern Chile, the Sierra de Limón Verde and the Sierra de Moreno (Damm et al. 1990), as well as in the Argentine Altiplano and Puná (Allmendinger et al. 1982). The Early Cambrian age more recently obtained for the CMB is considerably younger than the age generally accepted for the rocks forming the AAT. However, it has been proposed that this terrane collided against Gondwana in Ordovician times, during the tectonic cycle described next. This means that the Early Cambrian CMB could be part of a collided massif, either as part of the AAT or as part of another terrane that collided later.

**Famatinian tectonic cycle (latest Cambrian–Early Devonian)**

**Tectonic framework**

Global tectonic conditions, and in particular Laurentia–Gondwana interactions, during this cycle did not differ significantly from the ones existing during the Pampean cycle (see above). The Arequipa–Antofalla Terrane (AAT) and the Precordillera in central western Argentina (which, according to Thomas & Astini (2003) had been rifted apart from Laurentia in Early Cambrian times) collided against Gondwana in Ordovician times, during the Famatinian tectonic cycle. According to Bahlburg & Hervé (1997), the collision of the AAT occurred in the middle Ordovician (Guandacol Event?), and according to Thomas & Astini (2003), the collision of the Precordillera Terrane occurred in the Late Ordovician (Ocloyic tectonic phase).

**Geological units**

Although stratified rocks formed during this cycle are well known in southwestern Bolivia and in northwestern Argentina, no stratified Late Cambrian, Silurian and Early Devonian rocks are known in northern Chile, and only very few deposits of Ordovician age have been reported from the western side of the Andes, all of them in northern Chile. The only known deposits of probable Silurian age have been reported for south-central Chile, at Lumaco (39°S) by Tavera (1979a, 1983) (Fig. 3.4).

The only stratified rocks of undisputed Ordovician age in Chile are exposed at Agua de la Perdiz, in the Altiplano next to the international boundary (García et al. 1962; Ramirez & Gardeweg 1982). The Argomedo Beds in the Sierra de Argomedo, SE of Antofagasta, originally reported as containing a trace fossil of Ordovician age (Cruziana cf. furciera: Breitkreutz 1985, 1986b), have been recently assigned a Late Devonian–Early Carboniferous age based on the existence of a Pholadomiidae indet., which has a maximum Early Devonian age (Marinovic et al. 1995). The latter authors renamed the deposits the Argomedo Formation, and assigned to them an Early Devonian–Early Carboniferous age. In the light of this revised age assignation for the Argomedo Beds Formation, the tentative Ordovician age assigned by Breitkreutz (1986b) to the Cerro Palestina Beds, based on their similarity with the Argomedo deposits, is now unsustainable. Another deposit of alleged Ordovician age is the Complejo Igneo y Sedimentario

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**Fig. 3.3.** Location map for units of the Pampean tectonic cycle in Chile. Localities in northern Chile (see inset top left) are figured on the right, based on Salas et al. (1966). García (1996, 2002), Vergara (1979a), Maksaev (1978), Damm et al. (1990), Skarmeta (1983), Díaz et al. (1985). Location of drill-hole in extra-Andean Tierra del Fuego, in southernmost Chile, from which a granodiorite core yielded a Late Cambrian age (Söllner et al. 2000b), is indicated bottom left.
de Lila (CISL) in the Cerros de Lila, south of the Salar de Atacama (Niemeyer 1989). Because of the paucity and scattered nature of exposures of Ordovician deposits in Chile, any attempt to reconstruct the palaeo-geography and the tectonic evolution at that time needs to consider evidence from the well documented Ordovician exposures of the Puna and Eastern Cordillera in adjacent Argentina (Fig. 3.4b).

**Ordovician units**
The Aguada de la Perdiz Formation consists of a rather thick (2000 to 2700 m) succession of fine to coarse detrital (with a strong volcanic component), shallow marine deposits with volcanic and volcaniclastic intercalations (Breitkreuz 1986b; Bahlburg 1990; Breitkreuz et al. 1988; Bahlburg et al. 1994). A middle Arenigian graptolite fauna was found in the Aguada de la Perdiz Formation (García et al. 1962; Bahlburg et al. 1987, 1988, 1990; Breitkreuz et al. 1988). The outcrops of the Aguada de la Perdiz Formation extend into Argentina where they have been included in the Calalaste Group (Aceñolaza & Toselli 1981; Bahlburg et al. 1988). The CISL consists of a 3000-m-thick series of basic lavas interstratified with sandstone and pelitic layers intruded by small dioritic to gabbrie and rhyolitic
subvolcanic plutons which are generally associated with lavas of similar composition (Niemyer 1989). The CISL has been subdivided into four members. The lowest two members (1000 m thick) consist of basaltic and andesitic lavas (some pillowed in the lower member) with thin turbiditic intervals. The 1800 m-thick third member consists of a thick turbiditic succession, followed by andesitic and basaltic lavas. The 1250-m-thick uppermost member is essentially composed of fine yellowish-white tuffs and dacitic and rhyolitic lavas. The CISL is considered to be of Ordovician age based on the following evidence. (1) It is intruded by granitoid stocks (Pingo-Pingo, Tilopozo and Tucúcaro) dated as close to the Ordovician-Silurian boundary (Mpodozis et al. 1983). (2) A clast containing a Llanvinián trilobite has been found in the overlying Devonian–Early Carboniferous Lila Formation (Ceccioni 1982). According to Niemyer (1989), the bimodal magmatic rocks of the CISL have tholeiitic affinities, suggesting development in an island arc. Sediment supply was mainly from the NW.

Bahlburg et al. (1994) and Bahlburg & Hervé (1997) interpret the Chilean or western Ordovician deposits as part of a volcaniclastic apron developed on the east side of the arc that extends to western Argentina. These deposits grade upward to a volcaniclastic turbidite series (Puna Turbiditic Complex), indicating deepening of the basin. Towards the east the distal volcaniclastic deposits interfinger with the Cobres Group (Bahlburg et al. 1994). Sediment supply to this group was from the east, indicating the presence of a west-dipping slope in the Ordovician basin, which Bahlburg and co-workers interpret to be the west flank of a peripheral bulge.

Based on the distribution and facies of the deposits, the sediment supply and the geochemical affinity of the magmatism, it is possible to deduce the existence of a volcanic island arc with a backarc basin lying on the east side of the arc (Niemyer 1989; Bahlburg et al. 1994; Bahlburg & Hervé 1997). Bahlburg & Hervé (1997) have recorded the existence of ultramafic assemblages apparently associated with the Ordovician deposits in the Argentinian Puna (Allmendinger et al. 1983; Forsythe et al. 1993; Blasco et al. 1996), suggesting a major extensional episode in the backarc basin. According to Bahlburg et al. (1994) and Bahlburg & Hervé (1997), deformation in the backarc occurred in Middle–Late Ordovician times (Breitkreuz 1986b; Bahlburg et al. 1994; Bahlburg & Hervé 1997) has been attributed to the Ocolyic Phase caused by the collision of the AAT with the western margin of Gondwana. As explained before, it is presently impossible to determine whether the Belén Metamorphic Complex, if it really is an allochthonous terrane (see Chapter 2), participated in this, in a previous or a later collisional event. The presence of Ordovician, mafic/ultramafic rocks (ophiolitic suite) in the Argentinian Puna has also led to considerable debate about their tectonic setting. Their presence might indicate, as is the case for Ordovician ophiolites further south in the Precordillera of San Juan in central western Argentina (Furque 1972; Mpodozis & Ramos 1989), that extension in the Ordovician backarc basin was considerable and that this basin was ocean-floored.

Silurian units

Silurian deposits have been reported by Tavera (1979a) from Lumaco in the Coastal Cordillera at 39°S (Fig. 3.4c). These deposits constitute slightly metamorphosed portions of the protolith of the eastern series of the Late Palaeozoic metamorphic complex exposed along the Coastal Cordillera in south-central Chile (Hervé 1988; Fang et al. 1998).

Apart from the Late Ordovician–Early Silurian plutons that intrude the CISL (Pingo-Pingo, Tilopozo and Tucúcaro), there are other bodies of similar age located southward and close to the others (at 24°S) that do not intrude the CISL (Fig. 3.4), and are known as the Tambillo, Alto del Inca and Choschas plutons (Mpodozis et al. 1983). To the north of this locality, in the Precordillera or western flank of the Altiplano at 21°S, a muscovite granite has been dated at 431 ± 9.97 Ma (K–Ar in muscovite) (Vergara 1978b). This north–south alignment of Early Silurian intrusives along the high cordillera in northern Chile probably represents part of a partially exposed plutonic arc of late Early Palaeozoic age. Finally, another Silurian unit in northern Chile is the La Pampa Gneisses, which form a reduced outcrop in the upper Tránsito river valley, at 29°S (Fig. 3.4b), dated at 415±4 Ma (87Rb–86Sr isochron, whole rock) (Ribba et al. 1988).

Discussion

It is important to bear in mind that the palaeogeographic and tectonic model presented here for Chilean Ordovician deposits is mainly sustained by the alleged Ordovician age of the CISL (arguments for this have been given above). In this model, the Chilean Ordovician rocks correspond to the magmatic arc. Other essential aspects for understanding the tectonic evolution at this time are the stratigraphic position and origin of the Faja Eruptiva de la Puna oriental, a series of igneous rocks located in the Argentinian Puna. According to Bahlburg et al. (1994) and Bahlburg & Hervé (1997), these rocks are intrusive and were emplaced into Upper Ordovician or Lower Silurian rocks during Early Silurian times. Although this conclusion eliminates the difficulty of having to postulate the existence of two magmatic belts during Ordovician times, it instead causes the difficulty of needing to explain the existence of two intrusive belts in Early Silurian time in the region: the western intrusives emplaced in the CISL, in Chile (Faja Eruptiva de la Puna occidental), and the Faja Eruptiva de la Puna oriental, in the Argentinian Puna.

The model deduced by Bahlburg et al. (1994) and Bahlburg & Hervé (1997) for the Ordovician deposits in northern Chile and western Argentina envisages a backarc basin evolving to a forearc basin. Deformation of the Ordovician backarc deposits in Middle–Late Ordovician times (Breitkreuz 1986b; Bahlburg et al. 1994; Bahlburg & Hervé 1997) has been attributed to the Ocolyic Phase caused by the collision of the AAT with the western margin of Gondwana. As explained before, it is presently impossible to determine whether the Belén Metamorphic Complex, if it really is an allochthonous terrane (see Chapter 2), participated in this, in a previous or a later collisional event. The presence of Ordovician, mafic/ultramafic rocks (ophiolitic suite) in the Argentinian Puna has also led to considerable debate about their tectonic setting. Their presence might indicate, as is the case for Ordovician ophiolites further south in the Precordillera of San Juan in central western Argentina (Furque 1972; Mpodozis & Ramos 1989), that extension in the Ordovician backarc basin was considerable and that this basin was ocean-floored.

Gondwanan tectonic cycle (Late Devonian–earliest Permain)

Tectonic framework

Rocks assigned to this cycle are considerably more abundant than those belonging to the previously described cycles. For this reason it is possible to make a much better reconstruction of the processes that occurred at that time along the continental margin of western Gondwana. In northern Chile (north of 33°S), rock units belonging to this cycle are exposed in western as well as eastern parts of the country. Between 33°S and 46°S, rocks formed during this cycle are almost exclusively exposed along the westernmost part of the Coastal Cordillera, whereas south of 46°S, exposures occupy regions in the Archipelago and the Precordillera on both sides of the Patagonian Batholith (Fig. 3.5). We will describe separately the evolution in these three regions.

Palaeomagnetic reconstructions indicate that the South Pole migration was very rapid during Carboniferous and Early Permian, whereas in Late Permian times the pole migration came almost to a halt (Vilas & Valencio 1978). This indicates that continental wandering was initially also occurring at considerably high rates, and this, in turn, indicates that rapid convergence and subduction were most probably occurring along active continental margins like western Gondwana. Consequently, during this cycle a magmatic arc developed essentially along the present-day high cordillera. This arc,
located partly in western Argentina, was flanked to the west by a forearc basin located in Chile, and this basin, in turn, was flanked to the west by an accretionary complex.

Tectonostratigraphic evolution north of 33°S

The exact age range of some of the units belonging to this cycle is difficult to establish because of a lack of age-diagnostic fossils, partially exposed successions, and paucity of radiometric age determinations. Despite these difficulties and based essentially on the evolution deduced from the sedimentary deposits, Upper Palaeozoic rocks within this part of Chile can reasonably be separated into the following three stages:

1. Middle/Late Devonian to Early Carboniferous

2. Late Carboniferous to earliest Permian

3. Early Permian to Middle–Late? Permian

Fig. 3.5. Location map for units of the Gondwanan tectonic cycle. North of 33°S. First stage (Middle/Late Devonian to Early Carboniferous). Key: 1, Metamorphic units: Complejo Metamórfico del Choapa (CMC) and Complejo Metamórfico El Tránsito (CMT). 2, Metasedimentary units and western sedimentary deposits: El Toco Formation (ETF), Sierra El Tigre Formation (SET), Las Tórtolas Formation or Complejo Epimetamórfico Chañaral (LTF) and Chañaral Mélange (CM), Arrayán Formation and Unidad Metasedimentaria de Agua Dulce (ARF). 3, Eastern sedimentary deposits: Quebrada Aroma Formation (QA), Lila or Icnitas Formation (LF), Zorritas Formation (ZF), Argomedo Formation (AF). Cerro del Medio Beds (CM), Chinches Formation (CF), Las Placetas Beds (LP), Hurtado Formation (HF). 4, El Cepo Metamorphic Complex (ECMC).

Second stage (Late Carboniferous to earliest Permian). Key: 5, Volcanic deposits: Quipisca Formation (QF), Collahuasi Formation (CHF), Tuina Formation (TF), El Bordo Beds (EBB), La Tabla Formation (LT) and Pantanos Formación (PF). 6, Plutonic units: Sierra del Medio and Sierra de Moreno (SM) and further south along the Domeyko Range between 24° and 27°S, Elqui Superunit (ESU). Third stage (Early Permian to Middle–Late? Permian). Key: 7, Juan de Morales Formation (JM), Cerro El Arbol Formation (CA), Las Represas Formation (LR), Cerro 2484 Beds, Huantelauquén Formation and Quebrada Mal Paso Beds (HLF). Between 33° and 40°S. First stage. Key: 8, Western Series of the Metamorphic Complex. 9, Eastern Series of the Metamorphic Complex. 10, Late Carboniferous–Early Permian granitoids, mainly forming the Coastal Batholith. South of 40°S. Eastern Metamorphic Complexes (east of the Patagonian Batholith). Key: 11, Eastern Andes Metamorphic Complex (EAMC) and Puerto Edén Igneous and Metamorphic Complex (PEIMC). 12, Cordillera Darwin Metamorphic Complex (CDMC). Western Metamorphic Complexes (west of the Patagonian Batholith). 13, Chonos Metamorphic Complex (ChMC). 14, Madre de Dios Complex (MDC).
Late Palaeozoic structures and palaeogeographic features is NNW–SSE, and, therefore, slightly oblique relative to the mainly north–south orientated younger Andean structures (Late Cretaceous to Present). Magmatism at this time apparently had a more continuous evolution, which in some regions crosses through the boundary of these stages, and according to its geochemical characteristics can be related to subduction activity along the continental margin.

First stage: Middle/Late Devonian to Early Carboniferous

Middle/Late Devonian units exposed north of 33ºS include metamorphic, metasedimentary and sedimentary rocks that form series of outcrops with NNW–SSE orientations, with a westward increase in metamorphic degree (Fig. 3.5).

Sedimentary units. The sedimentary units form two groups of outcrops: a western, essentially turbiditic marine series, and an eastern platformal marine series. The western turbiditic deposits extend from 22ºS to 32ºS and correspond to the following units, from north to south (Fig. 3.7): El Toco Formation (Harrington 1961); Sierra El Tigre Formation (Niemeyer et al. 1997b; J. Cortés 2000; González & Niemeyer 2005); Las Tórtolas Formation (Uklricksen 1979; Bell 1982, 1984, 1987a, b; Naranjo & Puig 1984) or Complejo Epimetamórfico Chañaral (Godoy & Lara 1998, 1999); the Unidad Metasedimentaria de Agua Dulce or Agua Dulce Metasedimentary Unit (Rebolledo & Charrier 1994); and Arrayán Formation (Rivano & Sepúlveda 1991). The last mentioned formation includes the Puerto Manso, Arrayán and Los Vilos formations defined in the same region by Muñoz Cristi (1942). The western turbiditic deposits consist, in general, of thick, intensively and repeatedly folded successions of monotonous, medium- to coarse-grained turbidites in the El Toco Formation (Breitkreuz & Bahlburg 1985; Breitkreuz 1986b), the Sierra El Tigre Formation (Niemeyer et al. 1997b), the Agua Dulce Metasedimentary Unit (Rebolledo & Charrier 1994) and the Arrayán Formation (Muñoz Cristi 1942, 1973). The Las Tórtolas Formation, which is also strongly folded, includes siltstone intervals, limestone and chert intercalations (Bell 1984), and some metabasites (Godoy & Lara 1998). The Las Tórtolas Formation and the Agua Dulce Metasedimentary Unit are slightly metamorphosed, although the protolith of these rocks is perfectly recognizable. Sediment supply from NNW and NW has been deduced for the Sierra El Tigre (Niemeyer et al. 1997b), Las Tórtolas (Bell 1984) and Arrayán (Rebolledo & Charrier 1994) formations.

The age of these deposits has been generally difficult to assess. The El Toco Formation has been assigned a Devonian and/or Early Carboniferous age based on fossil plant remains (fragments of Dadoxylon and Haplostigma furquei Frenguelli) (Maksaev & Marinovic 1980; Bahlburg et al. 1987), and the late Early Carboniferous age of post-tectonic bodies that intrude this formation (Skarmeta & Marinovic 1981). The Las Tórtolas Formation contains a Devonian brachiopod (Mucrospirifer sp.) (Ferraris & Di Biase 1978; Niemeyer et al. 1997b; J. Cortés 2000; González & Niemeyer 2005). Based on fossil plant remains, Cecioni (1962, 1974) suggested a Middle Devonian age, whereas Bernarde de Oliveira & Roesler (1980) favoured a Late Devonian to Early Carboniferous age. Although the fossiliferous content of these deposits indicates a Late Devonian–Early

Fig. 3.6. Three-stage subdivision of the Gondwanan tectonic cycle and location of the San Rafael tectonic phase, based on the stratigraphic information available from the region north of 33ºS and applicable to the region between 33º and 46ºS. Subdivision of geological time is based on Harland et al. (1989).

Fig. 3.7. Stratigraphic units of the three stages of the Gondwanan tectonic cycle in Chile mentioned in text. Within each of the stages eastern (right) and western (left) units are listed from north (above) to south (below) and face each other according to the latitudes of their occurrence.
Carboniferous age, it is possible that the upper part of the successions reached into the Late Carboniferous.

Apart from considerable folding affecting these units, development of local deformation, described by the different authors as a mélangé or broken formation, has been observed in the Sierra El Tigre (Niemeyer et al. 1997b) and Las Tórtilas (Mélange de Chañaral; Bell 1984, 1987a, b) formations, and in the Unidad Metasedimentaria de Aguas Dulce (Rebolledo & Charrier 1994). The latter authors considered this feature as indicative of intense shearing affecting these rocks, presumably within a subduction complex.

The eastern platformal deposits are exposed in northern Chile along the Domeyko Range and the Principal Cordillera, between 23°30' and 31°S (Fig. 3.5). These deposits correspond, from north to south, to the following units: the Lila Formation (Moraga et al. 1974) or Icnitas Formation (Niemeyer et al. 1985), the Zorritas Formation (Cecioni & Frutos 1975) and the Argomedo Formation (Marinovic et al. 1995) (previously Argomedo Beds; Breitkreutz 1985, 1986b) (Fig. 3.7). The Lila Formation is a 1000-m-thick marine (transgressive-regressive) succession composed of quartz-rich sandstones, siltstones and conglomerates (Moraga et al. 1974; Niemeyer 1989). The 3040-m-thick Zorritas Formation has been subdivided into a lower member of coarse- to fine-grained sandstones (up to 1325-m thick), a 1600-m-thick middle member consisting of mudstones, and a 40-m-thick upward-coarsening upper member comprising fine-grained sandstones (Niemeyer et al. 1997a). The 1150-m-thick exposed marine deposits of the Argomedo Formation form an alternation of pelitic and fine- to coarse-grained sandstones rich in volcanic components (Marinovic et al. 1995). The upper 200 m consist of six upward-coarsening sandstones cycles. The Lila and Zorritas formations overlie Late Ordovician–Early Silurian plutons, whereas the base of the Argomedo Formation is not exposed. All these formations are intruded by Late Carboniferous–Early Permian granitoids (Davidson et al. 1981; Isaacson et al. 1985; Breitkreuz 1986b; Niemeyer et al. 1997a; Marinovic et al. 1995).

The upper part of the Lila Formation is not exposed because of erosion, whereas the Zorritas and the Argomedo formations are unconformably overlain by rhyolitic to dacitic volcanic rocks of Carboniferous to Permian age (La Tabla Formation according to Marinovic et al. 1995). The fossiliferous content of the Lila Formation indicates an Early Devonian age for its exposed part (Cecioni 1982; Breitkreuz 1986b; Niemeyer 1989). Breitkreuz (1986b) suggested on the basis of its correlation with the Zorritas Formation that its upper portion should reach the Early Carboniferous. The Zorritas Formation has been assigned a Middle Devonian to Early Carboniferous age on the basis of its palynological (Rubinstein et al. 1996) and brachiopod content (Isaacson et al. 1985; Dutro & Isaacson 1990; Boucot et al. 1995; Niemeyer et al. 1997a). The fossil content of the Argomedo Formation (several trace fossils, and a specimen of Pholladomiidae indet., which indicates a maximum Early Devonian age for these deposits), its stratigraphic position below the Carboniferous–Permain La Tabla Formation, and its intrusion by Late Carboniferous–Early Permian granitoids, allowed its assignment to Late Devonian–Early Carboniferous (Marinovic et al. 1995).

The deposits described above correspond to a transgressive-regressive event that occurred during Middle Palaeozoic times on the western margin of Gondwana, probably associated with a global sea-level rise (see Bahlburg & Breitkreuz 1991). The depositional environment deduced for these formations corresponds to a north-trending stable and shallow westward-deepening marine platform (Bahlburg & Breitkreutz 1991, 1993; Breitkreuz 1986b; Niemeyer et al. 1997a), flanked to the east by a volcanic arc (Arco Puneño; Coira et al. 1982; Niemeyer et al. 1997a), which represented the source of the sediments (Fig. 3.8).
Other deposits of probably similar age have been reported from the Precordillera, at 18°30’S (Quicucho Beds; García et al. 2004), 19°30’S (Quebrada Aroma Formation; Harambour 1990), as well as from the Domeyko Range between 25°15’S and 25°30’S (Cerro del Medio Beds; Naranjo & Puig 1984), and between 26°25’S and 27°30’S (Chinchesa Formation; Mercado 1982; Bell 1985; Cornejo et al. 1998; Tomlinson et al. 1999) (Figs. 3.5 & 3.7). The Quebrada Aroma Formation consists of a 1000- to 1500-m-thick upward-coarsening succession of metasedimentary marine deposits separated in a lower mudstone member and an upper turbiditic member with sedimentary structures indicating westward deepening of the basin. The Cerro del Medio Beds consist of metaquartzites and phyllites. These beds are intruded by granitoids that are overlain by Triassic marine and continental deposits (Quebrada del Salitre Formation; Naranjo & Puig 1984). No fossils have been found in these beds. The Chinchesa Formation (Davidson et al. 1978; Mercado1982; Bell 1985; Cornejo et al. 1998; Tomlinson et al. 1999) is a > 1000-m-thick (>2500 m; Mercado 1982), probably lacustrine succession consisting of foliated shales, siltstones and very fine- to coarse-grained sandstones (Bell 1985) and a few tuff beds (Rivano & Sepúlveda 1991; Bell 1985), which have been subjected to contact metamorphism. The nature of the base of the formation is unknown, being covered by sedimentary and volcanic rocks of possible Carboniferous to Permian age, and intruded by Permian granitoids (Mercado 1982). The age relationships with adjacent units and the geographic position of these deposits suggest a correlation with the Lisa and Zorritas formations.

Further south, two strongly folded marine units are exposed in the High Andes next to the international boundary, between c. 29ºS and 31ºS: the Las Placetas (Reutter 1974) and the Hurtado (Mpodozis & Cornejo 1988) formations (Fig. 3.5). The Late Carboniferous Las Placetas Formation (Reutter 1974; Nasi et al. 1990) comprises a c. 500-m-thick succession of greywackes (possibly turbiditic) and shales, green sandstones, and limestone intercalations containing fossil plant remains (Lepidodendron peruvianum, Frengueli, Sigillaria sauri Brant?, Plagioclamites?). Badly preserved marine invertebrates indicate a marine environment. The so-far unfolisissurable Hurtado Formation (Mpodozis & Cornejo 1988) is slightly contact metamorphosed, and forms a sometimes rhythmic alternation of slates and sandstones at least 1500 m thick, with the composition of the sandstones suggesting a granitoid sediment source. The base of these formations is not exposed. The Las Placetas Formation is intruded by Late Permian–Triassic granitoids and the Hurtado Formation by Late Carboniferous granitoids (Elqui Superunit).

Other probable deposits of the eastern platformal deposits. Several foliated units crop out along the Precordillera and high Andean cordillera between c. 26º and 31ºS with protoliths that tentatively can be assigned to the eastern platformal deposits, e.g. the Esquitos El Jardín (Muñoz 1986, fide Tomlinson et al. 1999) and the El Cepon Metamorphic Complex (Mpodozis & Cornejo 1988). The Esquitos El Jardín form a small outcrop of foliated metapelitic rocks exposed in the El Jardín region north of Potrerillos intruded by the Sierra del Castillo Batholith of Permian age (Tomlinson et al. 1999) (see below). The El Cepon Metamorphic Complex (Complejo Metamórfico El Cepo: Mpodozis & Cornejo 1988), exposed in the High Andes between 29º and 31ºS (CMEC in Fig. 3.5), forms small outcrops that correspond to roof-pendants in plutonic rocks of the Elqui-Limari batholith. These rocks correspond to sandstones and pelitic deposits apparently affected by two episodes of metamorphism: an older dynameothermal metamorphic episode with development of andalusite + cordierite + 3 albite + biotite + quartz, and a younger episode of contact metamorphism with development of quartz + albite + biotite + muscovite + amphibole. Abundant xenoliths from this metamorphic complex are found in the Late Carboniferous to Early Permian El Volcán intrusive unit in this batholith. Because it is intruded by Late Carboniferous plutonic units its age must be older than that. The older metamorphic episode that affects these rocks is not present in other sedimentary rocks of Late Devonian to Early Carboniferous age known in this region and further north along the High Andes (see below), suggesting that the Complejo Metamórfico El Cepo might be older than these rocks. However, considering the nature of the protolith, which is similar to the one exposed by the eastern platformal deposits described above, and that the metamorphic episodes that affected these rocks could represent a local feature associated with the Late Carboniferous to Early Permian magmatic events, we prefer to assign to the original sediments or protolith a Middle/Late Devonian to Early Carboniferous, rather than an Early Palaeozoic age. In fact, Martin et al. (1995) consider that the protolith of the El Cepo Metamorphic Complex is the Hurtado Formation. Absence of metabasites and marble in this metamorphic complex permits its clear differentiation from the El Tránsito and Choapa metamorphic complexes (Complejo Metamórfico El Tránsito and Complejo Metamórfico Choapa), which correspond to an accretionary complex (see description below).

Metamorphic units. Metamorphic rocks formed during the Gondwanan cycle crop out extensively in northern Chile (see Chapter 2), mostly along the coastal ranges between Huasco (27ºS) and Los Vilos (32ºS) (Fig. 3.5). In the high Andean cordillera, metamorphic outcrops between 28º30’S and 30º30’S have been included in the El Tránsito Metamorphic Complex (Complejo Metamórfico El Tránsito: CMT) (Reutter 1974; Ribba et al. 1988), and the coastal exposures south of 30ºS have been included in the Choapa Metamorphic Complex (Complejo Metamórfico del Choapa: CMC). These metamorphic complexes comprise polyphase-deformed quartz–mica (generally grey in colour) and amphibole (green in colour) schists, marbles and some chert (Miller 1970, 1973; Maas & Roeschmann 1971; Reutter 1974; Godoy 1976; Thiele & Hervé 1984; Ribba et al. 1988; Irwin et al. 1988; García 1991; Rivano & Sepúlveda 1991; Rebollole & Charrier 1994; Bahlburg & Hervé 1997). Metamorphic facies in the CMT are transitional between greenschist and amphibolite facies (Ribba et al. 1988); in the CMC the metamorphic conditions at 31ºS correspond to the upper greenschist–amphibolite facies transition (Irwin et al. 1988), whereas at 31°30’S they correspond to the greenschist facies in a low to medium P–T environment (Rebolledo & Charrier 1994). The protolith of the quartz–mica schists corresponds to greywackes and arkoses and that of the amphibole schists to oceanic basalts (Irwin et al. 1988; García 1991; Rebollole & Charrier 1994). The very different nature of the protolith of these rocks, the metamorphism, the intense deformation of the schists and of the metasedimentary units (development of broken formation, and polyphase deformation affecting in the same process the two types of schists) has suggested an environment associated with a subduction complex (Hervé et al. 1981a). This view has been confirmed by subsequent authors (Thiele & Hervé 1984; Irwin et al. 1988; García 1991; Rivano & Sepúlveda 1991; Rebollole & Charrier 1994; Bahlburg & Hervé 1997). 1⁸⁷Rb–1⁸⁶Sr ‘errorchrons’ for the metamorphic age of the CMT are 355 ± 20 Ma and 304 ± 40 Ma (Ribba et al. 1988). Metamorphism ages (K–Ar) for CMC are 311 ± 89 Ma (Irwin et al. 1988; García 1991) and 359 ± 36 Ma (Rebolledo & Charrier 1994) on amphibole schists (metabasites). The 359 ± 36 Ma age reported by Rebolledo & Charrier (1994) for the region at 31°30’S was obtained from amphibole crystals growing parallel to the main foliation of the metamorphic complex. This foliation corresponds to the axial plane of major isoclinal folds involving the two lithologies, indicating that widespread deformation was occurring by then. All these ages fall essentially in the Carboniferous, but the errors of most of them extend to the Devonian and to the Permian periods. It is therefore only possible to
assign a Late Palaeozoic age to the metamorphism, and an older age to the sedimentary protolith, probably Early Devonian or Silurian and even Ordovician.

It is important to underline the fact that the protolith of the quartz–mica schists in the CMC corresponds to greywackes (or arkoses), as do the unmetamorphosed units of the western turbiditic deposits described above (such as the El Toco, Sierra El Tigre and Arrayán formations). Therefore, the CMC can be interpreted as consisting of metamorphosed and strongly deformed portions of an accretionary complex, including portions of ocean crust and sediments shed to the trench from the continent, and metaturbidites and turbidites deposited on the eastern side of a well developed accretionary ridge that emerged above sea level and bounded the forearc basin to the west. If metamorphism has a Carboniferous age and the turbiditic formations have a Devonian–Carboniferous age, both processes (metamorphism and turbiditic sedimentation) occurred at least partially synchronously. The protolith of the quartz–mica schists probably corresponds to deeply buried turbiditic successions accumulated in the forearc basin, exposed in the present-day coastal region. These processes were also coeval with platformal marine sedimentation as well as with arc magmatism occurring further east in the present-day High Andes (Lila, Zorritas and Chinchones formations, and Cerro del Medio Beds, all of which received their sedimentary supply from a volcanic source referred to as the Punoño Arc and located to the east and SE: see Niemeyer et al. 1997b).

Middle/Late Devonian–Early Carboniferous magmatism. There is little record in Chile for the magmatic activity that occurred in this period. Evidence derives from the pyroclastic intercalations in the Chinchones Formation (Sepúlveda & Naranjo 1982; Bell 1985), volcanic quartz crystals present in the intercalations in the Chinches Formation (Sepúlveda & et al. 1983; Niemeyer et al. 1997b). The evidence for magmatic activity at this time would indicate that the Silurian–Early Carboniferous magmatic lull mentioned by Ballburg & Hervé (1997) was probably not as important as suggested by these authors, or that the volcanic centres were located farther east in Argentinian territory.

The north–south alignment of Late Ordovician–Early Silurian plutons, probably representing a magmatic arc of that age, is apparently located west of the Punoño Arc which developed later in Late Devonian–Early Carboniferous times (Gondwanan tectonic cycle). This would indicate an eastward shift of the latter and probably reflects some as yet undefined tectonic event at the continental margin.

Second stage: Late Carboniferous to earliest Permian

This stage represents the magmatic arc activity associated with a period of rapid continental drift and high convergence rate along the western margin of Gondwana in Late Carboniferous–Early Permian times. Evidence for this activity derives from volcanic deposits and major plutonic bodies. The almost complete absence of evidence for magmatic activity during the preceding stage contrasts with the extensive and abundant evidence for Late Carboniferous to Early Permian magmatism, suggesting that magmatic activity now attained its maximum, increasing the width of the arc and allowing its products to be abundantly exposed on the western side of the Andes, although these did not reach the present-day Coastal Cordillera. The proximity to the South Pole, which was located in southern Africa during Late Carboniferous–Early Permian times, combined with the great development of the orogen and magmatic arc during this period of high convergence rates between the western margin of Gondwana and the adjacent ocean floor, resulted in alpine-type glaciation along the mountain range and deposition of glacial deposits on either side (Charrier 1986) and literature therein for evidence of glaciation in Argentina.

A thick, mainly silicic volcanic succession (e.g. the La Tabla Formation, see below) formed during this period in the Domeyko Range and the High Andes in Chile, unconformably covering the eastern platformal Argomedo Formation (Late Devonian–Early Carboniferous) (Fig. 3.8). However, the base of most of the volcanic successions is not exposed, and it is not possible to determine the exact stratigraphic relation between these and the older eastern platformal deposits. Based on the unconformable relation between the Argomedo Formation and the La Tabla Formation, we include these two types of deposits in separate stages. In fact, these two formations may have both been linked in space and time to the evolution of the same arc. Uplift of Argomedo Formation marine sediments is recorded by a regressive facies in the upper part of this platformal succession, and is thought to have been a response to more intensive activity in the magmatic arc, itself linked to enhanced plate convergence rates. Thus arc volcanism moved into the area, erupting volcanic material over the former marine platformal region, an event recorded by the Argomedo–La Tabla unconformity (Fig. 3.8).

Volcanic deposits. Volcanic deposits of this age are represented by a series of thick volcanic formations, some of which contain thick and generally coarse detrital sedimentary intercalations, and subvolcanic intrusive bodies in the Precordillera and the High Andes in northern Chile (Fig. 3.5). In some regions this volcanism was apparently continuous until, at least, Late Permian times. The volcanic rocks are all essentially silicic in composition and comprise the following units: the Quipisca Formation, at 20°S (Galli 1968); the Collahuasi Formation, at 21°15′S (Vergara 1978b; Vergara & Thomas 1984; Ladino 1998); the La Tabla Formation, between 24°S and 27°S (Garcia 1677; Marinovic et al. 1995; Cornejo et al. 1998, 2006; Tomlinson et al. 1999); the Tuina Formation (Raczynski 1963; Marinovic & Lahren 1984; Mundaca 2002) or its equivalents the Agua Dulce Formation (Garcia 1677; Marinovic & Lahren 1984), the El Bordo Beds, at 22–24°S (Ramirez & Gardeweg 1982), the Pantanoso Formation in the Copiapó region, at 27–28°S (Mercado 1982; Iriarte et al. 1996) and the lower part of the Pastos Blancos Group (Guanaco Sonso sequence; 281–260 Ma) (Martin et al. 1999b) (Fig. 3.7); and other outcrops of Late Carboniferous–Early Permian age that have often been included in the Peine Formation (i.e. Marinovic & Garcia 1999). Because of the existence of deposits of similar compositions, but of different ages in northern Chile, and the reduced radioisotopic age determinations on them, their stratigraphic assignment remains a problem. The names that we consider most adequate to describe these deposits are given above. However, some authors prefer to reserve the name Agua Dulce Formation (see Marinovic et al. 1995) for Late Permian–Triassic volcanic deposits that we prefer to include in the Peine and Cas formations/groups because U–Pb determinations at their type localities gave Late Permian ages (Breitkreuz & van Schmus 1996), thus indicating their greater relevance to the Pre-Andean rather than to the Gondwanan tectonic cycle (see section on the Pre-Andean tectonic cycle below). In northern Chile, huge relict calderas (Mariposas, Ilimac and Guanaqueros) and hydrothermal alterations (Davidson et al. 1985) have been attributed to this volcanic activity.

These volcanic deposits correspond to thick (1000 to 3000 m) dacitic, rhyolitic and subordinately andesitic volcanic successions alternating with thick fluvial and locally lacustrine intercalations. Radioisotopic determinations from the La Tabla Formation yielded ages that mainly correspond to the Early Permian: 291 ± 9 Ma, 268 ± 11 Ma, 239 ± 9 Ma, 265 ± 10 Ma with the K–Ar method (Marinovic et al. 1995) and 262.9 ± 2.0 Ma with the U–Pb method (Cornejo et al. 2006). Apart from the Permian fossiliferous (Ostracodes, plant remains and
one saurian fossil) El Bordo Beds, the other units contain no fossil remains and have not been radioisotopically dated. The geochemical pattern of the La Tabla volcanic rocks follows that of the hypabyssal intrusives with which it is associated, described next (Marinovic et al. 1995).

**Intrusive units.** Late Carboniferous to Early Permian intrusives are known southward from the Sierra del Medio and Sierra de Moreno, between 21ºS and 22ºS (Fig. 3.5). At these localities, age determinations on intrusive bodies yielded Late Carboniferous–Early Permian K–Ar ages of 296 ± 10 Ma and 271 ± 22 Ma (Hute et al. 1977). The volcanic components of the Late Carboniferous–Early Permian magmatism are not exposed south of c. 25ºS, but huge batholithic bodies of the same age witness the existence of the magmatic arc along the present-day High Andes between this latitude and 33ºS (Fig. 3.5).

Between c. 24ºS and 27ºS (Domeyko Range), intrusive bodies of late Carboniferous to early Permian age have geochemical signatures indicating production of magmas associated with subduction and of magmas of crustal origin (Marinovic et al. 1995), whereas between 27ºS and 31ºS magmatism corresponding to the different origins is clearly separated in time (Nasi et al. 1985; Mpodozis & Kay 1990). This has been considered by Mpodozis & Cornejo (1994) as evidence for the segmented nature of the Late Palaeozoic–Early Mesozoic magmatic belt.

In the Domeyko Range, between 24 and 25ºS, Marinovic et al. (1995) reported the existence of three intrusive associations: foliated granitoids, unfoliated granitoids, and hypabyssal intrusive rocks. The foliated granitoids are calcalkaline and correspond to I-type plutons, and therefore can be associated with subduction, whereas the unfoliated granitoids and the hypabyssal intrusives have geochemical signatures indicating a crustal origin. The ages of the two groups are Late Carboniferous–Early Permian, which indicates that the two types of magmas formed approximately at the same time. A similar situation has been reported for the High Andes at approximately 26–27ºS (Copiapó region). Here, the synchronous emplacement of Late Palaeozoic magmas potentially associated with subduction, and magmas with crustal affinities has been documented (Mpodozis & Cornejo 1994). The Permian (280–255 Ma) Sierra Castillo batholith consists of intrusive bodies with a partially synmagmatic and partially cataclastic foliation (Cornejo et al. 1993; Tomlinson et al. 1999), which are compositionally similar to the foliated intrusives from the Domeyko Range, between 24º and 25ºS, and the Late Carboniferous–middle Permian Pedernales batholith. The latter includes epizonal plutons of probable crustal origin, similar to the unfoliated intrusives and hypabyssal plutons further north. On the other hand, in the High Andes, between 27º30´S and 31ºS, two well exposed and successive (not synchronous) intrusive superunits correspond to magmatic activity associated with a Late Palaeozoic episode of subduction linked to rapid convergence (Elqui Superunit), and to a subsequent stationary episode (Ingaguá Superunit). The Elqui Superunit represents the intrusive component of the magmatic activity developed during Late Carboniferous–Early Permian times (Nasi et al. 1985; Mpodozis & Kay 1990). This superunit includes a series of mesozonal plutons that have been grouped into four major units that record the evolution of tectonic conditions during slowdown of continental drift and during tectonism associated with westward deformation and definitive accretion to the continent of the subduction complex. These units are, from oldest to youngest: Guanta, Montosa, Coguchiás and El Volcán. The Guanta Unit is composed of calcalkaline, metaluminous, I-type hornblende-biotite tonalites and granodiorites with subordinated gabbros and quartz diorites, formed along the active continental margin. The Montosa Unit consists of biotite (± hornblende) granodiorites that differ from the Guanta Unit in their higher SiO₂ content and lower K₂O content. The Coguchiás Unit consists of peraluminous, leucocratic granites and granodiorites, and have rare earth element (REE) signatures consistent with a high pressure residual mineralogy suggesting formation by melting of a thicker crust. The El Volcán Unit consists of cataclastic, course-grained biotite granites, and relative to the Coguchiás Unit it is richer in K₂O; its geochemistry allows comparison with S-type granitoids (Mpodozis & Kay 1990). Based on the contact relationships with neighbouring units and on radioisotopic age determinations, the Elqui Superunit has been assigned a Late Carboniferous–Early Permian age.

Because there is no direct radioisotopic age from the El Volcán Unit, it is difficult to precisely determine the age of its magmatic foliation and cataclastic fabric, which has been attributed to the effects of the San Rafael tectonic phase (Polanski 1970) during magmatic emplacement, which in Argentina has been assigned an Early Permian age (Llambias & Sato 1990). The age of the El Volcán Unit is younger than the Late Carboniferous (301 ± 4 Ma) age obtained from the Coguchiás Unit and older than the maximum Early Permian age (276 ± 4 Ma) obtained from a plutonic body intruding in the Elqui Superunit (Nasi et al. 1985), coinciding with the age of the San Rafael phase. However, in the Domeyko Range between 24ºS and 25ºS, the presence of a magmatic foliation in one granitoid association and not in the other synchronous association (see above) suggests that the magmatic foliation might be explained, at least in these latter granitoids, by local shearing along some major fault(s) located in this range. A more precise chronology on these intrusive associations is needed to determine the possible effect of the San Rafael phase during emplacement of the foliated intrusives.

**Sedimentary deposits.** These deposits correspond, from north to south, to the following units: Juan de Morales Formation (Galli 1956, 1968); Cerro El Árbol Formation (Marinovic et al. 1995; J. Cortés 2000); Las Represas Beds (Sepúlveda & Naranjo 1982); Cerro 2484 Beds (von Hillebrandt & Davidson 1979); La Corvina Beds (Irwin et al. 1988; Garcia 1991); and Huentelauquén Formation (Muñoz Cristi 1973), including the Mal Paso Beds (Mundaca et al. 1979) (Fig. 3.7).

The **Juan de Morales Formation** (Galli 1956, 1968), located in the western Precordillera, east of the Central Depression, at 20ºS, was referred to also by Zeil (1964), Harambour (1990) and Diaz-Martinez et al. (2000). This c. 150-m-thick marine formation comprises three upward-fining and thinning cycles of conglomerates (fluvial at the base), sandstones and shales. The third cycle begins with a c. 10-m-thick bioclastic limestone with a well preserved Middle Permian brachiopod fauna (Díaz-Martinez et al. 2000). The Juan de Morales Formation unconformably overlies the silicic volcanic deposits of the Quipisca Formation (Late Carboniferous–Early Permian?), and is overlain by the Diablo Formation of possible Triassic age (Galli 1956, 1968). These deposits accumulated in a platformal environment (Galli 1956, 1968; Hambour 1990; Diaz-Martinez et al. 2000), and have been correlated with the Huentelauquén Formation.

The **Cerro El Árbol Formation** (Marinovic et al. 1995; J. Cortés 2000) was deposited in a shallow, coast-near platform environment during Early Permian times, this age being based on a diverse brachiopod, bivalve and gastropod fauna (Marinovic et al. 1995; Niemeyer et al. 1997b). The formation includes the Cerros de Cuevas Formation, in the Salar de Navidad region at 23º45´S (Niemeyer et al. 1997b) (the Cerros de Cuevas Formation was previously included in the Estratos
del Salar de Navidad (Ferraris & Di Biase 1978) together with the Sierra El Tigre Formation, the Baquedano Beds, next to Baquedano, and the Cerro 1584 Beds, in Augusta Victoria, next to the Km. 99 station of the Antofagasta to Salta railway (Marinovic et al. 1995). The deposits in the Cerros de Cuevitas region consist of a 460-m-thick alternation of conglomeratic sandstones and shale intercalations; sandstones are often calcareous, and limestones predominate in the upper portion (Niemeyer et al. 1997b). Next to the Cerro El Árbol type locality, the deposits consist of a c. 370-m-thick succession of conglomeratic breccias that pass upward to tuffaceous coarse-to-medium-grained sandstones, pelitic and sandstone deposits, followed by a 150-m-thick conglomeratic level, 50-m-thick rhyolitic intercalation, and a 130-m-thick succession of shales with intercalations of fossiliferous limestones, followed by upward-coarsening sandstones. At Cerro 1584, a conglomeratic intercalation separates the deposits into a lower turbiditic member and an upper calcareous member. These deposits unform unconformably overlie the Devonian Sierra El Tigre Formation and the Carboniferous–Permian La Tabla Formation, and are overlain by Triassic–Jurassic continental and marine deposits. This Cerro El Árbol Formation has been correlated with the Las Represas Beds, in the Sierra de Fraga, Copiapó region, at 27°S, and the Huentelauquén Formation in Chile, and the Arizana Formation in Argentina (Niemeyer et al. 1997b). Equivalents of the Cerros de Cuevitas Formation in the Antofagasta region are preserved as a limestone outcrop containing productid fossil remains (Niemeyer et al. 1997b), and the Cerro 1584 Beds.

Northeast of Copiapó, at 27°S, two outcrops have been described that correspond to deposits of this age: the Las Represas Beds, next to the locality of Carrera Pinto (Sepúlveda & Naranjo 1982), and c. 30 km to the east of this locality the Cerro 2484 Beds, in the Sierra de Fraga (von Hillebrandt & Davidson 1979). The Las Represas Beds comprise a 300-m-thick folded succession that consists in its lower part of conglomerates with grey limestone intercalations, and in its upper part of sandstones with fine marl intercalations. It unconformably overlies silicic volcanic deposits of the Pantanoso Formation and is unconformably overlain by the Late Triassic La Ternera Formation. Based on its fossiliferous content (essentially brachiopods and gastropods) it has been assigned an Early Permian age. In the Sierra de Fraga, the Cerro 2484 beds consist of a 300-m-thick succession of grey limestones, marls and sandstones. They overlie granitoid bodies of possible Late Carboniferous age and are overlain by plant-bearing Triassic deposits assigned to the La Ternera Formation. This and the presence of Bellerophon sp. and other fragments of fossils allow its assignment to the Permian. The deposits of both localities have been correlated with the Huentelauquén Formation (Sepúlveda & Naranjo 1982; von Hillebrandt & Davidson 1979).

The gently folded Huentelauquén Formation (Muñoz Cristi 1973) and other equivalent deposits exposed in the coastal area between 31°S and 31°45’S (La Corvina Beds, lower portion of the Totoral Beds, and Quebrada Mal Paso Beds) consist of coarse- to fine-grained clastic marine deposits, locally turbiditic, with calcareous intervals. The Huentelauquén Formation has been subdivided into two members separated from each other by an erosional unconformity (Rivano & Sepúlveda 1983, 1985, 1991) argued for a Late Carboniferous–Early Permian age based mainly on the discovery of Foraminifera in the upper member. Díaz-Martínez et al. (2000) analysed the faunal content of the Huentelauquén Formation described by the mentioned authors and concluded that its age is late Early Permian. Rivano & Sepúlveda (1983, 1985, 1991) consider that the faunal remains contained in the Huentelauquén Formation represent a temperate to cold water association (foramol association). The coarse clastic deposits in both members of the Huentelauquén Formation suggest the existence nearby of emergent areas and a high clastic influx from them. The clastic lower member probably corresponds to a deltaic to submarine fan environment. The mainly calcareous upper member indicates development of a calcareous platform located adjacent to the emergent areas which continued to supply abundant detritus. The thin turbiditic interval and the turbiditic character of the coquina horizons observed in the lower member suggest deposition in regions distant from the coast. South–southeastward sediment supply deduced from the turbiditic intervals coincides with the sediment transport directions observed in the Late Carboniferous–Early Permian turbiditic formations in this same region.

Palaeogeographic considerations. The Early–Middle/Late? Permian deposits are located approximately in the area previously covered by the Middle/Late Devonian to Early Carboniferous forearc basin (first stage of this tectonic cycle) (see Fig. 3.8), with magmatism apparently remaining concentrated in the eastern regions, closer to the magmatic arc. Although the deposits of this final stage of the Gondwanan tectonic cycle are separated from those of the first period by an angular unconformity, indicating the existence of tectonic activity, and the depositional environment changed considerably, the marine basin remained about the same size. Strong similarities can be observed between the deposits accumulated during this stage, with all formations being typified by abundant black and grey shales. Coarse- to fine-grained clastic deposits bracket the shale intervals, whereas calcareous intercalations are concentrated in the upper portion of the formation. Depending on the locality, clastic deposits correspond either to upward-fining or upward-coarsening sequences probably indicating tectonically controlled fluctuations in the basin. Deposits in this basin indicate shallower conditions than in the previous one and the existence of emergent reliefs that facilitated the development of the calcareous deposits found at different locations across the basin. The southeastward supply of sedimentary material probably

intercalations of variable thickness. This formation unconformably overlies the strongly folded Arrayán Formation and is unconformably overlain by Triassic conglomerates.

No mention exists in the literature about the direction of sediment supply for the Huentelauquén Formation. Recent measurements by one of us of flute-cast orientations in the turbiditic succession in the lower member indicate sediment transport from NW to SE, an orientation that coincides with the sediment transport direction deduced for the underlying Arrayán Formation (Rebolledo & Charrier 1994) in the same region. An abundant fauna has been recovered from this formation and especially from its upper member (Sundt 1897; Groeber 1922; Fuenzalida 1940; Muñoz Cristi 1942, 1968; Minato & Tazawa 1977; Mundaca et al. 1979; Thiele & Hervé 1984). The La Corvina Beds have been assigned a Late Palaeozoic age, most probably Permian based on brachiopod shell fragments (Irwin et al. 1988). The lower part of the Totoral Beds has been assigned an Early Permian age by Tavera (see Thiele & Hervé 1984). The Quebrada Mal Paso Beds have been assigned a Permian age based on the existence of abundant brachiopod remains (Productus) (Mundaca et al. 1979). The age of the Huentelauquén Formation has been discussed at length (see Charrier 1977; Mundaca et al. 1979), with most authors agreeing on a Permian age. However, more recently, Rivano & Sepúlveda (1983, 1985, 1991) argued for a Late Carboniferous–Early Permian age based mainly on the discovery of Foraminifera in the upper member. Díaz-Martínez et al. (2000) analysed the faunal content of the Huentelauquén Formation described by the mentioned authors and concluded that its age is late Early Permian. Rivano & Sepúlveda (1983, 1985, 1991) considered that the faunal remains contained in the Huentelauquén Formation represent a temperate to cold water association (foramol association). The coarse clastic deposits in both members of the Huentelauquén Formation suggest the existence nearby of emergent areas and a high clastic influx from them. The clastic lower member probably corresponds to a deltaic to submarine fan environment. The mainly calcareous upper member indicates development of a calcareous platform located adjacent to the emergent areas which continued to supply abundant detritus. The thin turbiditic interval and the turbiditic character of the coquina horizons observed in the lower member suggest deposition in regions distant from the coast. South–southeastward sediment supply deduced from the turbiditic intervals coincides with the sediment transport directions observed in the Late Carboniferous–Early Permian turbiditic formations in this same region.

Palaeogeographic considerations. The Early–Middle/Late? Permian deposits are located approximately in the area previously covered by the Middle/Late Devonian to Early Carboniferous forearc basin (first stage of this tectonic cycle) (see Fig. 3.8), with magmatism apparently remaining concentrated in the eastern regions, closer to the magmatic arc. Although the deposits of this final stage of the Gondwanan tectonic cycle are separated from those of the first period by an angular unconformity, indicating the existence of tectonic activity, and the depositional environment changed considerably, the marine basin remained about the same size. Strong similarities can be observed between the deposits accumulated during this stage, with all formations being typified by abundant black and grey shales. Coarse- to fine-grained clastic deposits bracket the shale intervals, whereas calcareous intercalations are concentrated in the upper portion of the formation. Depending on the locality, clastic deposits correspond either to upward-fining or upward-coarsening sequences probably indicating tectonically controlled fluctuations in the basin. Deposits in this basin indicate shallower conditions than in the previous one and the existence of emergent reliefs that facilitated the development of the calcareous deposits found at different locations across the basin. The southeastward supply of sedimentary material probably
indicates the existence of a major relief bounding this basin to the WNW, as occurred during the first stage of this cycle (Fig. 3.9).

Rhyolitic and tuff intercalations at the type locality of the Cerro El Árbol Formation indicate siliceous explosive activity during this period although these represent probably only the westernmost occurrences during this third stage of the cycle. No volcanic intercalations have been recognized further west in the accretionary prism and in the western turbiditic sediments. However, more detailed observations are certainly needed to definitely discard their presence in the western region considering the strong recrystallization of the metamorphic rocks present in this region.

**Magmatic and tectonic evolution**

Based on the unconformities separating the deposits of the three stages, two deformation events can be identified during the Gondwanan tectonic cycle in rocks exposed north of 33ºS (Figs 3.6 & 3.7). The first event can be located approximately in mid-Carboniferous times, and is apparently responsible for the emergence of the complete forearc basin, or at least the eastern basin region or platform. This event occurred before deposition of the Late Carboniferous–Earliest Permian westward-wedge volcanic deposits, which developed westward covering the deposits of the first stage of this tectonic cycle, i.e. the Middle Devonian–Early Carboniferous Zorritas Formation. The unconformity separating, in the eastern region, the volcanic deposits, and in the western region, the accretionary prism and the turbiditic forearc sediments from the marine deposits of the next younger stage, indicates that deformation and uplift affected the two margins of the forearc basin, if not the entire forearc. This second deformation event can be correlated with the late Early Permian San Rafael tectonic phase described in western Argentina (Fig. 3.6). This phase, according to Rapalini (1989) and Rapalini et al. (1989), is responsible for the rotation of the San Rafael block in Argentina, and for the synmagmatic foliation and cataclasis affecting the El Volcán Unit of the Elqui Superunit (Mpodozis & Kay 1990). The evidence presented here indicates that this same deformation event was also responsible for the first closure episode of the Late Palaeozoic forearc basin, though apparently did not interrupted magmatic activity. The cause of the deformation would be the collision of ‘Terrane X’ invoked to explain the possible westward growth of the continental margin and the consequent westward shift of the Jurassic magmatic arc relative to the location of the previous Late Palaeozoic magmatic arc (Mpodozis & Kay 1990). In our opinion, ‘Terrane X’ corresponds to the Late Palaeozoic subduction complex or accretionary prism formed along the continental margin, remnants of which are still exposed along the coastal region in Chile. We also consider that the two described deformation events are part of a rather continuous process during which the subduction complex was gradually backthrust towards the continental margin of Gondwana and amalgamated to it. A third and local unconformity separates the two members of the Huentalauquén Formation indicating that deformation of the forearc basin deposits was still active at that time and that it finally led to the inversion of the basin.

Backthrusting of the subduction complex and amalgamation to the continent, definitive closure (inversion) of the Late Palaeozoic forearc basin, and arc magmatism all occurred during the third period of the Gondwanan tectonic cycle in the present-day Chilean side of the Andes. At the same time, further east in Argentina the Paganzo marine and continental backarc basin was developing (Salfity & Gorustovic 1983; Fernández-Seveso et al. 1993) (Fig. 3.9).

**Overview**

The three stages differentiated for the Gondwanan tectonic cycle, which are separated from each other by unconformities, indicate intense tectonic activity, a complex palaeogeographic organization, and its rapid modification. The unconformity separating the Gondwanan tectonic cycle from the preceding Famatinian cycle is well developed along the High Andes in the eastern part of the Chilean territory. To the west this unconformity has not been detected and probably never developed considering that this region corresponded to a marine domain in which later the development of the accretionary prism occurred. Similarly, the unconformity separating the first from the second stages of the Gondwanan tectonic cycle is not exposed in the western side because the Late Carboniferous to earliest Permian magmatic arc was located far to the east (in the present-day High Andes in Chile and Argentina), and the forearc basin formed a barrier to the volcanic deposits (Fig. 3.8). Finally, the unconformity separating the second from the third stages, which is ubiquitous, indicates major palaeogeographic modifications occurring in the forearc basin. These changes continued during deposition of the Huentalauquén Formation at the end of the cycle. The erosional unconformity described by Rivano & Sepúlveda (1991) separating the lower from the upper deposits of the Huentalauquén Formation is an indication of this, although it is probably a local effect during the continuous process of forearc basin inversion during accretion of the prism.

According to Ramos et al. (1984, 1986), collision of Chilenia against Gondwana (Chanic orogenic phase) occurred during Middle to Late Devonian times. However, the tectonic evolution for the accretionary prism outlined above suggests that its development began earlier than Middle or Late Devonian times and therefore prior to Chilenia’s collision, and there is no evidence for a collisional event at that time. Therefore, collision would have occurred earlier, probably at the end of the preceding cycle (Rebolledo & Charrier 1994). This view coincides with Astini’s (1996) proposition for the collision of Chilenia in Early Devonian times.

**Tectonostratigraphic evolution between 33ºS and 43º30'S**

**Middle/Late Devonian to Early Carboniferous evolution**

A major change in the distribution of tectonostratigraphic units occurs at 33ºS. Further south from this latitude Late Palaeozoic intrusive rocks form the Coastal Batholith and are exposed along the Coastal Cordillera in a continuous swath of outcrops located immediately east of Gondwanan metamorphic rocks (Fig. 3.5). At 38ºS the batholith bends SE away from the coast and becomes one of the units constituting the Principal Cordillera. We briefly describe the metamorphic units and the Coastal Batholith in this region.
Metamorphic units. South of 33ºS, the Gondwanan metamorphic complex is almost continuously exposed along the Coastal Cordillera. Between 33ºS and 38ºS, these exposures are flanked on their eastern side by the Late Carboniferous–Permian Coastal Batholith. The metamorphic complex consists of two paired metamorphic belts: the western and the eastern series (Godoy 1970; Aguirre et al. 1972; Hervé 1974, 1988). The eastern series consists mainly of polyphase deformed metaturbidites, with recognizable primary structures, and lenses of calcislicate rocks, probably deposited in a forearc basin. Eastward increasing metamorphism (towards the Coastal Batholith) developed under relative low P/T ratios, locally attaining the amphibolite–granulite facies transition (Gana & Hervé 1983; Hervé et al. 1984; see Chapter 2). Metamorphic exposures in the westernmost side of the Principal Cordillera at c. 39º30’ to 40ºS, assigned to the eastern series, form the Trañú Metamorphic Complex, Devonian to Carboniferous in age (Rodríguez et al. 1999). According to these authors, these rocks are intruded by a Late Carboniferous–Early Permian batholith, and are unconformably overlain by Late Triassic deposits.

The western series consists of polyphase deformed and metamorphosed clastic sediments (sandstones and pelites), metacherts, metabasites and serpentinites (Godoy 1970; Aguirre et al. 1972; Hervé 1974, 1988; Gana & Hervé 1983; Duhart et al. 2001). Pillow structures are sporadically recognizable in the metaturbidites. Ultramafic serpentinitized bodies were apparently tectonically emplaced (Godoy & Kato 1990; Kato & Godoy 1995). Metamorphism developed under high P/T ratios and increased westward. Crossite, glaucephane, zussmanite and lawsonite are locally developed (Saliot 1968; Hervé et al. 1984; Duhart et al. 2001). The western series has been interpreted as an east to west transition from deformed forearc basin deposits to an accretionary complex (Kato 1985; Hervé 1988; Martin et al. 1999; Willner et al. 2000). The rather low Sr isotope ratios obtained in the western series seems to confirm its oceanic nature (Hervé et al. 1984).

Age determinations by Hervé et al. (1984) in the eastern series yielded ages (87Rb–86Sr on whole rock) of 368±42 Ma for the sillimanite zone, close to the Coastal Batholith, and 347±32 Ma in the staurolite-andalusite zone farther west (away from the batholith). These Late Devonian–Early Carboniferous ages allow a chronologic comparison with the metamorphic age of the complexes in the region north of 33ºS. At the same latitude, the western series yielded younger ages (87Rb–86Sr on whole rock) of 311±10 Ma (Late Carboniferous) for glaucephane schists. The age of metamorphism of the central Chile complex is constrained between late Early Carboniferous and Late Permian times (Munizaga et al. 1973; see Chapter 2). According to Duhart et al. (2001), between 39º30’S and 42ºS, metamorphism and deformation occurred during two phases: one in Carboniferous and the other in Permian–Triassic times.

The age of the protolith of the Late Devonian to Early Carboniferous metamorphic complex is difficult to determine. However, some fossiliferous localities have been found in the eastern series. At Lumaco, a locality located at 38ºS, Silurian fossils (Gordia sp., Nerites nahueltitanus and Monograptus (Spirigraptus) aff. spiralis spiralis (Geinitz)) were found and determined by Tavera (1979a, 1983), indicating the depositional age of the protolith (Fig. 3.4c). Further south, at Buill in the Huequi Peninsula (42º15’S), a marine fauna consisting of well preserved Devonian trilobites and a coral has been reported from loose blocks by Biese (1953), Levi et al. (1986) and Fortey et al. (1992); according to these authors, these blocks most certainly come from nearby outcrops of black slates in which no fossils have yet been found. This finding indicates a Devonian age for the deposition of these low metamorphosed sediments. The age of the protolith of the metamorphic complex south of 33ºS indicated by the fossil content at Lumaco and Buill coincides with the age deduced for the protolith of the metamorphic complexes north of 33ºS (i.e. Choapa Metamorphic Complex; see above).

Between 39º30’S and 42ºS, the western series of the metamorphic complex (Bahía Mansa Metamorphic Complex, described west of Osorno at 40º45’S) includes, apart from the already mentioned lithologies, trachytic intrusive bodies and metaglimnibrites (Sollner et al. 2000a; Duhart et al. 2001) and intercalations indicating exhalative origin, like iron formations, massive sulphides, tourmalinite and spessartite quartzites (coticules) (Willner et al. 2001). U–Pb age determinations on (1) detrital zircon crystals suggest a Devonian depositional age for the protolith and on (2) a trachytic intrusive emplaced in mafic schists yielded a 396.7±1.3 Ma, which indicates a minimum Early Devonian age for the mafic schists of oceanic origin hosting the trachyte body (Duhart et al. 2000). In the northwestern part of Chiloé Island, samples from metamorphic units yielded ages of 220±6 Ma (K–Ar) and 232.5±2.7 Ma (40Ar/39Ar) on muscovite of the same sample indicating a Triassic age for the metamorphic event (greenschist facies) (Antinano et al. 2000). These dates indicate a Devonian to Triassic age for the metamorphic rocks and allow their correlation with the Bahía Mansa Metamorphic Complex (Antinano et al. 2000).

Magmatic units. In the northern part of the Coastal Batholith, Late Carboniferous plutonic activity is represented by the Mirasol Unit (299±10 Ma, U–Pb age on zircons) (Gana & Tosdal 1996). This composite calcalkaline plutonic unit, emplaced into the eastern series, comprises several coarse-grained granitoid bodies with magmatic foliation, and forms almost continuous outcrops along the Coastal Cordillera. Further south, beyond the point where the batholithic outcrops have curved SE away from the coast (Fig. 3.5), K–Ar and 87Rb–86Sr age determinations on intrusive bodies emplaced into the eastern series at 40ºS yielded ages between 242±42 Ma and 309±8 Ma (Munizaga et al. 1988). More recently U–Pb age determinations yielded ages ranging between 280 and 300 Ma (Martin et al. 1997b), confirming a Late Carboniferous to Early Permian age for this plutonic episode, which is comparable with the ages obtained along the Principal Cordillera north of 33ºS and along the Coastal Cordillera between 33ºS and 38ºS.

Discussion. In this region it is not possible to identify different depositional stages in the Late Palaeozoic evolution of the area, as has been done for the region north of 33ºS. Growth of the accretionary complex and sedimentation in the rear part of the prism were probably closely related in space and time, whereas emplacement of the Late Palaeozoic batholith occurred in later stages of this evolution. The accretionary prism is well developed in this region and there is abundant evidence for its oceanic origin. Although there is little evidence for primary sedimentary features in the protolith of the eastern series, it is probable that deposition occurred in a forearc basin, as seen more clearly further north. The sedimentation ages (Silurian and Devonian) obtained for the protolith of the eastern series coincide fairly well with the ages deduced for the sedimentary protolith of the metamorphic complexes in northern areas. According to this evidence it is probable that the palaeogeographic organization was similar or even identical to that of the region north of 33ºS.

The major differences between this region and the one located north of 33ºS are (1) the absence of Late Early to Middle/Late? Permian deposits and (2) the more western location of the Coastal Batholith, between 33ºS and 38ºS. However, the absence of Permian deposits in the southern region is probably due to deeper erosion in this area than further north, and the more westerly position of the batholith was probably caused by tectonic effects during Late Palaeozoic and/or younger evolutionary stages.
Tectonostratigraphic evolution south of 46°S

Several isolated exposures of metamorphic complexes occur south of c. 46°S. Because of their isolation and differences between them it is convenient to analyse each separately, as well as to divide them broadly into eastern and western complexes according to their distribution on one or the other side of the Patagonian Batholith (Fig. 3.5). The eastern complexes comprise the Eastern Andes Metamorphic Complex, the Puerto Edén Igneous and Metamorphic Complex, and the Cordillera Darwin Metamorphic Complex; the western complexes comprise the Madre de Dios, the Diego de Almagro and the Diego Ramírez complexes (see Chapter 2). According to recently performed radiometric dating on the Diego de Almagro Complex, its age is probably younger than Palaeozoic and we therefore analyse this complex in the Pre-Andean tectonic cycle.

According to Hervé et al. (see Chapter 2), the Eastern Andes Metamorphic Complex corresponds to the outcrops of poly-deformed and metamorphosed turbidites, limestones and pyroclastic rocks exposed east of the Patagonian Batholith between Lago General Carrera (46°S) and the southern part of the Brunswick Peninsula (54°S). In the Eastern Andes Metamorphic complex, the Puerto Edén Igneous and Metamorphic and the Staines Peninsula complexes, and the Río Lácteo Formation (Bell & Suárez 2000), metamorphism occurred under medium-grade greenschist and epidote–amphibolite facies in pre-Late Carboniferous times in an intraplate setting (Bell & Suárez 2000; see Chapter 2). Recent studies by Hervé and collaborators suggest that deposition occurred in a passive continental margin environment (see Chapter 2).

The Cordillera Darwin Metamorphic Complex (Dalziel & Cortés 1972; Nelson et al. 1980), located mainly in Tierra del Fuego, south of 54°S, consists of the Eastern metamorphic Complex of polydeformed metasedimentary and volcanic rocks. The lithologic similarity and the proximity of the Eastern Andes and the Cordillera Darwin complexes suggests that they might belong to one single unit; although, based on the distinctive metamorphic mineralogy of the latter it is preferable to maintain these two complexes as two separate units (see Chapter 2).

The Madre de Dios and the Diego Ramírez complexes are located on the westernmost margin of South America in the Patagonian Archipelago, and differ considerably from each other (Fig. 3.5). The Madre de Dios Complex is composed of three strongly deformed lithostratigraphic units: the Tarlton Limestone, the Denaro Complex and the Duque de York Complex (Forsythe & Mpodozis 1979, 1983; Mpodozis & Forsythe 1983). The Tarlton Limestone corresponds to a fusulinid-rich calcareous platform deposited in an intra-oceanic environment in Late Carboniferous to Early Permian times (Cecioni 1955b; Douglass & Nestell 1972, 1976), the Denaro Complex has been considered to be the ocean floor on which the Tarlton platform was developed, and the Duque de York Complex is a turbiditic succession accumulated next to the continental margin before accretion of the Tarlton–Denaro terrane. Recent studies revealed the existence in the Duque de York Complex of detrital zircons of late Early Permian age (Hervé et al. 2003a), and that uplift of the complete complex occurred in Early Jurassic times, suggesting that deformation and definite accretion occurred in Mesozoic times (Thompson & Hervé 2002).

The Diego Ramírez Complex is located in the southernmost tip of South America (c. 56°20′S), south of Cape Horn (Davidson et al. 1989), and is lithologically similar to the Diego de Almagro Complex. It consists mainly of polydeformed pelagic sediments, with pillow structures and pillow breccias, and minor mélangé with inclusions of chert, greywacke, tuff, basalt and limestone. A ⁴⁰Rb–⁴⁰Sr errorchron yielded a 169 ± 16 Ma age, which is interpreted as the age of metamorphism. This single age coincides with the above-reported evidence for Mesozoic metamorphic and accretionary events along the southwestern Chilean archipelago.

General discussion on the Gondwanan cycle

Considering the proposition made above in the Overview for the region north of 33°S, that collision of the Chilenia terrane occurred in Early Devonian times, the continental crust in central Chile on which the Gondwanan cycle evolved consisted of this terrane. According to Ramos & Basei (1997), this basement can be observed in erosional windows and preserved as roof pendants in the Frontal Cordillera in Argentina and Chile. The Gondwanan tectonic cycle occurred over a timespan of 40–50 Ma, as continental assembly of Gondwana took place. This assembly occurred during rapid convergence rates and the resulting interaction between oceanic lithosphere and western Gondwana determined the development of the following morphostructural units, from west to east: an accretionary prism, a forearc basin, a magmatic arc, and a backarc basin. At the final stage of assembly, complete emergence of the continental margin and accretion of the prism occurred. In Early Permian times this process is expressed by changes in the depositional environment in the forearc basin (i.e. Hueneleauquén Formation) and by the development of a synmagmatic foliation in some intrusive units, which records a modification in the stress regime along the continental margin. The effects of the new tectonic regime correspond to the San Rafael tectonic phase. However, the time of deformation associated with the final accretion of the prism does not seem to coincide everywhere along the continental margin. In the region lying between 43°30′S and 47°S, deformation occurred later (in Early Jurassic times), allowing sedimentation in the forearc basin to continue until late Triassic times. The addition of terranes (accretionary prism and forearc basin deposits) to the continent determined that, once subduction was reinitiated in Early Jurassic times after a rather long period of subduction cessation, the new magmatic arc was shifted to the west of the previous Late Palaeozoic arc, along the present-day Coastal Cordillera.

Pre-Andean tectonic cycle (Late Permian–earliest Jurassic)

Tectonic framework

We use the term ‘Pre-Andean’ for the cycle developed after the final phase of assembly of the Gondwana megacraton and before development of the Early Jurassic magmatic arc. During this cycle subduction along the continental margin was interrupted, or at least considerably diminished. This cycle does not reflect only the tectonic conditions determined by the assembly of the megacraton, but also the initial processes that later resulted in its break-up. This very distinctive period involved the development of completely different geotectonic conditions along the continental margin compared to those that prevailed before and later, i.e. Gondwanan and Andean tectonic cycles. The pre-Andean cycle may be attributed broadly to the Permian and the Triassic periods, although more accurately the tectonic conditions of subduction cessation that prevailed during this cycle began in late Early Permian and ended in earliest Jurassic times, with resumption of subduction and the beginning of associated magmatic activity.

The polar migration curves for South America and Africa (Vilas & Valencio 1978) show that, after a period of rapid continental drift in Late Palaeozoic times, a complete or almost complete pause of the continental drift of Gondwana occurred in Late Permian to earliest Jurassic times. This stationary period has been attributed to the final consolidation of the megacraton, which produced new tectonic conditions along the western continental margin of Gondwana. These conditions
favoured heat accumulation in the upper mantle, melting of the lower crust, and production of enormous volumes of magmas along the northern Chilean coast (Berg & Breitkreuz 1983; Berg et al. 1983), in the high Chilean Andes (Kay et al. 1989; Mpodozis & Kay 1990) and on the Argentinian side of the Andes (Llambias & Sato 1990; Llambias et al. 1993; Llambias 1999, 2001). As a consequence of this, crustal warping and extension of the upper, brittle part of the crust led to the development of extensional basins. The distinctive features of the pre-Andean cycle are, therefore, the development of abundant and widely distributed, essentially silicic magmatic activity, and a palaeo-geography dominated by NNW–SSE orientated extensional basins (Charrier 1979; Uliana & Biddle 1988; Mpodozis & Ramos 1989; Mpodozis & Kay 1990; Suárez & Bell 1992; Stipanicic 2001). This palaeo-geographic organization developed only at the continental margin of Gondwana, that is, in present-day Chile and adjacent Argentina (Uliana & Biddle 1988; Ramos 1994). According to Ramos (1994), the reason for such distribution of the extensional basins is the existence in this region of NW-trending weakness zones represented by the sutures that bound the allochthonous terranes accreted in Proterozoic and Palaeozoic times. Knowledge about the extensional basins and their infill in Chile derives mostly from surface studies. In Argentina important information derives from subsurface surveys.

**Distribution and general features of the deposits**

The essentially Triassic deposits of the pre-Andean tectonic cycle in Chile and adjacent Argentina generally form more or less continuous NNW–SSE orientated outcrops. As mentioned above, this rather peculiar distribution compared to that generally adopted by older and younger deposits in Chile was controlled by major faults with this orientation, some of which were activated in subsequent deformation episodes. The resulting grabens or semi-grabens and horsts were apparently oblique to the continental margin of Gondwana. This distribution of the main palaeo-geographic elements produced a coastline with embayments and peninsulas. The NNW end of the grabens that reached the pre-Andean continental margin was occupied by the sea and the deposits in these areas are marine, whereas, in these same grabens, the deposits located further SSE correspond to continental successions. Basins formed inland contain only continental deposits. Although most exposures in Chile and Argentina follow more or less continuous linear outcrops, some outcrops depart from this rather simple model indicating that the organization of the basins was in places more complex. Although rifting was mainly controlled by normal faults, strike-slip components cannot be discarded, and some of the basins might have formed by pull-apart mechanisms.

Rocks belonging to this cycle are exposed in separated regions within northern and central Chile, between 22°S and 42°S (Fig. 3.10), and in southern Chile, in the Chonos and Diego de Almagro archipelagos, between 45°S and 52°S (see Fig. 3.15). Accordingly, the two regions will be treated separately.

**Evolution in northern and central Chile, between 22°S and 42°S**

The extensional basins tentatively defined north of 42°S are, from NE to SW (Charrier 1979; Suárez & Bell 1992; Alvarez 1996; Alvarez et al. 1995) (Fig. 3.11): (A) El Profeta–La Ternera, which possibly continues south–southwestward in Argentina as the Bermejo (Ischischuca–Villa Unión) Basin, in the La Rioja and northern San Juan region (Charrier 1979; Stipanicic 2001); (B) San Félix–Cuyana, possibly extending south–southeastward in Argentina as the Cuyo (Barreal-Norte de Mendoza) Basin, in the southern San Juan and Mendoza region (Charrier 1979; Stipanicic 2001); (C) La Ramada, exposed on both sides of the international boundary; (D) El Queré–Los Molles; (E) Bio-Bio–Temuco.

Late Palaeozoic and Triassic volcanism is widely distributed (500 000 km²) in Chile and Argentina between 21°S and 44°S and is generally referred to as the ‘Permo-Triassic volcanism’. Formal names for the resulting deposits are: Choiyoí Group (Rolleri & Criado Roqué 1968) and Choiyoí Magmatic Province (Kay et al. 1989); however, it has received local names in each of the regions where it occurs. The Choiyoí Group has been correlated with the Mitu Group in Peru. This group can be subdivided into two volcanic portions with minor sedimentary intercalations (Llambias 1999). The older portion consists of Late Palaeozoic volcanic rocks of intermediate composition and calcalkaline signature developed in an arc setting in association with subduction of oceanic lithosphere. In the Frontal Cordillera, on the eastern side of the Andean range, this magmatic event formed between 272 and 260 Ma, after the San Rafael tectonic phase (post-orogenic). For this reason, it has been assigned to the third stage of the Gondwanan cycle (Fig. 3.6). The younger portion has ages between 259 and 247 Ma (latest Permian to Early Triassic) (Llambias 1999); two age determinations from the Polvaredas locality in the Mendoza river valley (33°S) yielded 240 ± 15 Ma and 238 ± 10 Ma (Ladinian) (Caminos 1970). It consists of silicic volcanic deposits, frequently ignimbritic, mainly of rhyolitic composition, which are associated with subvolcanic intrusives. The volcanic rocks are associated with major A- and S-type intrusive bodies with a similar geochemical composition displayed by the volcanic deposits. The late post-orogenic younger portion of the Choiyoí Group was developed by intense crustal melting under extensional tectonic conditions. For this reason, Llambias & Sato (1995) consider that this younger portion represents a volcanic activity transitional between arc and intraplate-type magmatism. According to its age and petrologic features, the younger portion is assigned to the early pre-Andean cycle.

In the High Andes, between c. 28 and 31°S, extensive and thick volcanic and volcanoclastic deposits have been included in the Pastos Blancos Formation (Thiele 1964), elevated to group status by Martin et al. (1999a). This group consists of at least two units: the Guanaco Sonso (rhyolitic and dacitic in composition) and the bimodal Los Tilos (basaltic to rhyolitic in composition); other outcrops of the Pastos Blancos Group are difficult to assign to either one or the other unit (Martin et al. 1999a). Radioisotopic age determinations in the Guanaco Sonso unit yielded K–Ar ages ranging between 281 ± 6 Ma and 260 ± 6 Ma and one U–Pb age of 265.8 ± 5.6 Ma that falls in the Permian. In the Los Tilos unit, one K–Ar date yielded a 235 ± 5 Ma age (Ladinian–Carnian) and U–Pb age determinations on zircon crystals gave scattered ages ranging between 225 and 210 Ma, consistent with the stratigraphic relations of this unit above the Guanaco Sonso unit and below the Early to Late Jurassic Lautaro Formation, and with the Late Triassic fossil plant remains found in this unit (Dicrodium Flora). Therefore, the Los Tilos unit can be assigned a Late Triassic age, possibly reaching to the earliest Jurassic. According to its age, lithology and geographic location in the northwestern prolongation of the Frontal Cordillera in Chile, the Pastos Blancos Group is assigned to the Choiyoí Group.

In Chile, particularly noteworthy are the thick, mainly silicic volcanic and volcanoclastic deposits of late Middle Triassic–early Late Triassic (Ladinian–Carnian) age that also form part of the Choiyoí Magmatic Province. These deposits record a widely extended volcanic pulse (the La Totora–Pichidangui volcanic pulse), which separates two major stages (Fig. 3.12) in the tectonic evolution of the central basins formed in this tectonic cycle, the San Félix and the El Queré–Los Molles basins. In the Vallenar region (28°30’ to 29°S), they form the 700–1000-m-thick La Totora Formation (Reutter 1974), which conformably covers the early to middle Anisian marine San Félix Formation. In the coastal region of central Chile (32°S), these deposits correspond to the Pichidanguí Formation.
Fig. 3.10. Distribution of pre-Andean sedimentary, volcanic and intrusive units between 20°S and 41°S. Key: 1, Older siliceous volcanic and volcaniclastic deposits; 2, marine deposits of the First Stage; 3, siliceous volcanic and volcaniclastic deposits preceding the Second Stage (La Totora–Pichidangui volcanic pulse); 4, marine deposits of the second stage; 5, continental deposits of the second stage; 6, intrusive rocks. Abbreviations for formation names: C, Cifuncho; CA, Canto del Agua; EQ, El Quereo; LL, La Ligua; LM, Los Molles; LT, La Ternera; LTT, La Totora; P, Pichidangui; PA, Pan de Azúcar; PB, Pastos Blancos; PE, Peine and Cas; PI, Pintados; PR, Profeta (lower part); QES, Quebrada El Salitre; SF, San Félix; T, Tuina. Abbreviation for granitoid unit: ISU, Ingaguás Superunit.
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(Cecioni & Westermann 1968; Vicente 1976; Vergara et al. 1991), which conformably separates the El Quereo (Late Anisian) from the Los Molles (Norian–Pliensbachian) formations. Further east in central Chile equivalent deposits are exposed as the La Ligua Formation (Thomas 1958), which is unconformably overlain by the Sinemurian marine deposits of the Quebrada del Pobre Formation. Similarly, in the Coastal Cordillera at the latitude of Curicó and Talca (35°S to 35°30’S), in the northern part of the Bio-Bio–Temuco Basin, silicic volcanic deposits are known at the base of the Late Triassic–Early Jurassic deposits in the Cruce de los Sauces Formation (Corvalán 1976), and La Patagua Formation (Hervé et al. 1976b). Further south, in the southern part of the Bio-Bio–Temuco Basin, no volcanic deposits are known at the base of the Late Triassic deposits which instead directly overlie the Palaeozoic metamorphic basement.

The basin infill in Chile also contains basaltic and andesitic intercalations. Andesitic intercalations are included in the Agua Chica Formation (Mercado 1980; Godoy & Lara 1998), basalts form the lower part of the Quebrada del Salitre Formation (‘Basaltos de Sierra Doña Inés Chica’) (Naranjo & Puig 1984; Tomlinson et al. 1999; Cornejo & Mpodozis 1996, 1997), basaltic andesites locally form major parts of the La Ternera Formation (Jensen & Vicente 1976; Mercado 1982; Sepúlveda & Naranjo 1982; Suárez & Bell 1992; Cornejo & Mpodozis 1996; Iriarte et al. 1996, 1999; Arévalo 2005b) and the El Verraco Beds (Nasi et al. 1990), basaltic exposures are included in the Crucero de los Sauces Formation (Corvalán 1976; Thiele & Morel 1981), and there are other similar deposits elsewhere. In Argentina, basaltic intercalations are known from the Bermejo Basin (Page et al. 1997; Monetta et al. 2000) and the Cuyo Basin (Ramos & Kay 1991).

With regard to the sedimentary evolution of the pre-Andean cycle it is possible to differentiate two rift stages – (1) Late Permian?–Scythian?–Late Anisian; and (2) Norian–Sinemurian – separated from each other by the above-mentioned Ladinian–Carnian silicic volcanic and volcanoclastic intercalation (Fig. 3.12). This intercalation (the La Totora–Pichidangui volcanic pulse) is apparently associated with the beginning of the second or younger stage, whereas the first or older stage seems to have been preceded by thick volcanic and volcanoclastic deposits of latest Permian and earliest Triassic ages (see description below for the Cas and Peine formations) (PE and PI in Fig. 3.10). These older volcanic deposits are coeval with the younger or upper portion of the Choiyoi Group described by Llambías (1999) for the Frontal Cordillera. The younger La Totora–Pichidangui volcanic pulse represents a still younger pulse of the Choiyoi Magmatic Province.

The sedimentary episodes separated by the Ladinian–Carnian volcanic intercalation generally, have thicknesses of several hundreds to thousands of metres, suggesting strongly subsiding conditions for these basins. The base of these series is generally formed by thick breccia deposits, which represent the beginning of a transgression–regression cycle developed over different Palaeozoic units. These deposits are well exposed in the El Quereo–Los Molles and San Félix basins. The deposits of the younger stage are marine and continental and overlie the thick Ladinian–Carnian volcanic intercalation, except for the Bio-Bio–Temuco Basin, where the silicic intercalation is present only in its northern part. The marine deposits of this stage correspond also to a transgression–regression cycle, as described for the older stage, although in most sections the upper part of the cycle is not exposed. The exposed lower portions of the marine successions also indicate rapid subsidence. The continental deposits of the younger stage correspond to alluvial, fluvial and lacustrine facies. The development of large lakes was a characteristic of the SSE prolongation of these basins into Argentina. The distribution of deposits pertaining to the each stage is given in Figure 3.10.

Fig. 3.11. Distribution of pre-Andean deposits in Chile (see text for relevant literature) and neighbouring regions of Argentina, based on Riccardi & Igleza Llanos (1999), Spalletti (2001), and Stipanicic (2001), between 22ºS and 42ºS, and tentative assignment of these deposits to major pre-Andean extensional basins developed in Chile and western Argentina: A, Profeta-La Ternera Basin, possibly extending SSE-ward in Argentina as the Bermejo (Ischischuca–Villa Unión) Basin; B, San Félix–Cuyana Basin, possibly extending in Argentina as the Cuyo (Barreal–Norte de Mendoza) Basin; C, La Ramada Basin; D, El Quereo–Los Molles Basin; E, Bio-Bio–Temuco Basin, possibly including the Chacaico and the Los Menucos deposits in Argentina. Key: 1, Marine deposits; 2, continental deposits; 3, Basin; 4, basin-bounding normal faults.
Deposits of the first or older stage
The oldest deposits of this tectonic cycle correspond to the Cas and Peine formations (Moraga et al. 1974) exposed on the eastern side of the Salar de Atacama, between 23°S and 24°S (Fig. 3.10). These formations consist of c. 2000-m-thick successions of siliceous lavas and pyroclastic deposits, subordinate mafic lavas, and red clastic deposits. The Peine Formation has a characteristic middle fluvial and lacustrine member. U–Pb age determinations of Late Permian age close to the Permian–Triassic boundary have been obtained in the Cas and Peine formations (249 ± 3 Ma is the best estimate of true age) (Breitkreuz & van Schmus 1996), indicating that these formations are rather related to the Late Permian–Triassic magmatism (younger portion of the Chojoyi Group, sensu Llambias & Sato 1990, 1995) than to the Late Carboniferous–earliest Permian magmatism as previously suggested by Breitkreuz & Zeil (1994) and Breitkreuz (1995). Based on plants remains of Noeggerathiopsis hislopi (Bumb.) and Cordaites hislopi (Bumb.), Letelier (1977) assigned a Permian age to the rhylitic Matahuaco Formation (Dedios 1967) exposed in Elqui river valley at 30°S, which unconformably underlies the Late Triassic continental Las Brees Formation (Letelier 1977; Moscoso et al. 1982a). This formation can, therefore, be correlated with the Guanaco Sonso sequence of the Pastos Blancos Group exposed further east in the High Andes (Martin et al. 1999a).

The sedimentary infill of the San Félix Basin (Fig. 3.11) is represented by the two following marine formations: San Félix (Reutter 1974; Ribba 1985; Schoener 1985; Suárez & Bell 1992) and Canto del Agua (Moscoso et al. 1982a) (Fig. 3.10). The San Félix Formation is exposed in the High Andean range in the El Tránsito drainage basin (28°30’ to 29°S). This transgressive–regressive sequence consists of > 4000 m of coarse to fine clastic sediments. It unconformably overlies the Late Palaeozoic metasedimentary deposits of the Las Tórtolas Formation and is overlain by the silicic volcanic and volcaniclastic La Totora Formation. The age of the San Félix Formation, based on its marine fossiliferous content, is early to middle Anisian. This succession begins with conglomerates and pebbly sandstones, with a crinoidal limestone intercalation, continues with a c. 3000-m-thick turbiditic sequence, and ends with cross-bedded sandstones, pebbly sandstones, and conglomerates with pelitic intercalations (Reutter 1974; Ribba 1985; Schoener 1985; Suárez & Bell 1992). The existence of a thick turbiditic central part indicates considerable subsidence of the basin; more proximal turbiditic deposits in the middle part of the turbiditic interval have been interpreted as the result of tectonic uplift of the source area (Suárez & Bell 1992). Rapid eastward to southeast thinning and absence of these deposits further east and SE suggest the existence of steep, fault-controlled margins in this basin (Reutter 1974; Suárez & Bell 1992). The Canto del Agua Formation crops out in the coastal region at the same latitude as the San Félix Formation. It covers metasedimentary Palaeozoic deposits of the Las Tórtolas Formation and is overlain by the Neocomian Bandurrias Group. It has an estimated thickness of 2100 m and is composed of sandstones, conglomerates, shales and limestone intercalations. It contains fossils of Middle Triassic age (Moscoso et al. 1982b), and Suárez & Bell (1992) have interpreted its sedimentary environment as a coarse-grained submarine fan-delta.

In the El Quereo Basin (Fig. 3.11), two closely exposed formations have been described for this stage: the Cerro Talinai Formation (Muñoz Cristi 1942; Mundaca et al. 1979) and El Quereo Formation (Muñoz Cristi 1942, 1973; Cecioni & Westermann 1968; Rivano & Sepúlveda 1991) (Fig. 3.10). The > 3000-m-thick, marine Cerro Talinai Formation consists of a succession of conglomerates, conglomeratic sandstones and sandstones, and in the uppermost exposed levels of a fossiliferous, rhythmic alternation of sandstones and shales. Sediment supply was from the SE, probably along the axis of the basin, and the great thickness of this succession indicates a strongly subsiding environment. These deposits unconformably overlie different Palaeozoic units, their top is not exposed, and fossiliferous levels indicate an Anisian age. The c. 700-m-thick marine El Quereo Formation unconformably overlies the Late Palaeozoic turbiditic Arrayán Formation and underlies the thick silicic Pichidanguí Formation of Ladinian–Carnian age (Cecioni & Westermann 1968; Rivano & Sepúlveda 1991). A 10–20-m-thick breccia is developed at the base of the formation; the very angular clasts indicate almost no transport and are derived from the underlying Arrayán Formation. Over the breccia a transgressive–regressive succession is developed, that begins and ends with conglomerates; intermediate deposits from bottom to top are sandstones, turbidites and black shales.

Deposits of the second or younger stage
Deposits of this stage of the pre-Andean tectonic cycle accumulated in marine and continental environments during Late Triassic (post-Carnian) and earliest Jurassic (Hettangian to Pliensbachian; locally, possibly Toarcian) times. Earliest Jurassic marine sedimentation followed continental Late Triassic deposition in several localities, and probably corresponds to a late phase of basin development. In localities within the present...
coastal region these marine deposits are abruptly covered by late Early to Middle Jurassic deposits of the volcanic arc developed once subduction resumed and recording the end of the pre-Andean cycle. In localities located further east (backarc domain), not reached by the volcanic deposits, marine sedimentation continued without interruption.

In the Profeta–La Ternera Basin (Fig. 3.11) different types of marine and continental deposits have been described and given different formational names. These deposits correspond to distinct environments in the basin. The marine deposits known for this basin are included in the Profeta (Chong 1977; Chong & von Hillebrandt 1985) and Pan de Azúcar (García 1967) formations, and the continental ones in the Cerro Quimal region (Chong & Gasparini 1975; Chong 1977), in the Cifuncho (García 1967) and La Ternera (Brüggen 1918, 1950) formations, and in the La Coipa Beds (Suárez & Bell 1992). Although the lacustrine La Coipa Beds have been assigned a Late Triassic age based on their stratigraphic position, palynological material of probable early Triassic age (Suárez & Bell 1992) and Triassic conchostracan rests have been recovered from them (Gallego & Covacevich 1998).

The Cifuncho and Pan de Azúcar formations (García 1967) form a succession over 1000 m thick exposed in the Coastal Cordillera at 26°S (Fig. 3.10). The Late Triassic to earliest Jurassic Cifuncho Formation consists predominantly of a coarse-grained and poorly sorted fluviatile succession, continuing to the earliest Jurassic with the shallow marine Hettangian–Sinemurian deposits of the Pan de Azúcar Formation (García 1967; Naranjo 1978; Naranjo & Puig 1984; Suárez & Bell 1992; Marinovic et al. 1995). The dominant facies of the Cifuncho Formation corresponds to braided and ephemeral river deposits, but there are also intercalations of volcaniclastic alluvial fans, small lakes and sabkha deposits. The source area was located to the NW and west, and farther west than the present coastline. The great thickness of coarse and poorly sorted clastic debris suggests proximal deposition in a subsiding present coastline. The great thickness of coarse and poorly sorted clastic debris suggests proximal deposition in a subsiding coastal plain.

The lower member, which has yielded Triassic plant remains, consists of a bimodal association of massive basalts with pillow lavas (Basaltos de Sierra Doña Inés Chica) and associated rhyolitic and dacitic sills and domes (232.9 ± 0.2 Ma; U–Pb on zircon), with red fluviatile conglomeratic and sandstone intercalations, as well as massive breccias containing blocks of Palaeozoic foliated tonalites. The upper member consists of quartz–feldspathic and fossiliferous calcareous sandstones with conglomeratic layers, and intercalations of basaltic lavas.

Immediately to the south of the Salar de Pedernales region, at 26°30’S, the late Triassic deposits correspond to the continental El Mono Beds (Mercado 1982; Cornejo et al. 1998), consisting of a thick succession of breccias and conglomerates with rhyolitic and andesitic clasts, and intercalations with huge boulders (1 m in diameter) of Palaeozoic granitoids and rhyolites (1500 m), laminated black carbonaceous lacustrine shales and sandstones containing Triassic fauna (> 1300 m), and matrix-supported conglomerates and sandstones (800–1200 m). This succession is continuous with Early Jurassic marine deposits of the Montandón Formation.

The continental La Ternera Formation, at its type locality in the Copiapó region (28°S) (Fig. 3.10), exposes a >1800-m-thick succession of clastic sediments and an upper 300-m-thick succession of andesitic and basaltic lavas (Segrestrom 1968; Davidson et al. 1978; Sepúlveda & Naranjo 1982; Suárez & Bell 1992). Based on its fossil flora and conchostracans content, and its stratigraphic position underlying Sinemurian marine limestone of the Lautaro Formation, a Late Triassic–Hettangian? age has been assigned to this formation (Solms Laubach & Steinnmann 1899; Brüggen 1918, 1950; Mercado 1982; Sepúlveda & Naranjo 1982; Herbst et al. 1998; Gallego & Covacevich 1998; Troncoso & Herbst 1999). Sedimentary facies correspond to distal alluvial plains, braided alluvial plains, distal braided rivers, and braided floodplains (Suárez & Bell 1992; Blanco 1997) and lakes (Herbst et al. 1998). The uppermost sediments consist of an alternation of sandstones, conglomerates and carbonaceous horizons; these levels contain abundant plant fossils including tree trunks in growth position. At the type locality, the La Ternera Formation culminates with an andesitic member at least 300 m thick, thickening...
southward to >1000 m (Iriarte et al. 1999). Thickness variations (2100 to zero) across only tens of kilometres suggest accu-
mulation in a strongly subsiding and fault-controlled basin (Suárez & Bell 1992). Similar features have been described for
this formation further south in the Laguna del Negro Francisco region (Mercado 1982) and in the Copiapó river valley (Jensen & Vicente 1976). The lowermost horizons of the Lautaro Formation were probably deposited during the last phase of
basin development of the pre-Andean cycle; in this sense they are equivalent to the Pan de Azucar Formation that overlies the
Late Triassic Cifuncho Formation in the present-day Coastal Cordillera. The La Ternera Formation is a southern equivalent of
the El Mono Beds described above for the Salar de Pedernales region.

The lateral relationships between the Late Triassic to earliest Jurassic continental and marine deposits of the Cifuncho, Profeta, Quebrada del Salitre, La Ternera and Pan de Azucar stratigraphic units, although not in all cases exposed in the field, are suggested in Figure 3.13.

In the High Andes of Vallenar at 29ºS, the Lautaro Formation
locally overlies with angular unconformity the La Totoro Formation, indicating that some deformation and erosion occurred in Late Triassic times (e.g. Cero Tatul; see Reutter 1974).

In the Elqui region, between c. 30ºS and 31ºS, the continental Las Breas and the conformably overlying marine Tres Cruces formations (Dedios 1967), represent the deposits of the younger stage. The Las Breas and the Tres Cruces formations are located in the prolongation of the San Félix-Cuyana Basin (Fig. 3.11) to which they are tentatively assigned. At its type locality, the Las Breas Formation consists of 300–550-m-thick continental conglomerates, sandstones and mudstones, with carbonaceous intercalations, deposited in alluvial and lacus-
trine environments. Basaltic, andesitic and rhyolitic lavas are also included in this unit (Mpodozis & Cornejo 1988; Pineda & Emparan 2006), and plant remains (Dicrodium Flora) indicate a Late Triassic age. These deposits overlie the silicic volcanic Matahuaco Formation, a local representative of the Choyoi Group, which is a probable equivalent of the Guancaco Sonso unit of the Pastos Blancos Formation (Martin et al. 1999a) exposed to the NW in the High Andes at 29º30’S. The Tres Cruces For-
mation consists of a strongly variable succession of conglom-
erates, sandstones, marls and micritic limestones that grades upwards to sandstones and conglomerates. The marine levels contain abundant fauna indicating essentially a Sinemurian to Pliensbachian (Dedios 1967; von Hillebrandt 2002), possibly Toarcian (Letelier 1977; Moscoso et al. 1982a) or even Callovian age (Mpodozis & Cornejo 1988). The overlying continental volcanic deposits that form the Punta Blanca member of the Tres Cruces Formation sensu Letelier 1977, represent in this region distal arc deposits, which are associated with the resumption of subduction at the beginning of the Andean tectonic cycle. Therefore, at least part of the lower marine portion of the Tres Cruces Formation sensu Letelier 1977 is considered to have been deposited during the last phase of pre-Andean basin development, similarly to the Pan de Azucar Formation and the lowermost levels of the Lautaro Formation.

Deposits in the La Ramada Basin (Fig. 3.11) consist of a 400-m-thick fluvial and lacustrine succession placed within the synrift Rancho de Lata Formation (Alvarez et al. 1995) and composed of conglomerates and sandstones rich in volcanic components, with marl intercalations and pyroclastic deposits (rhyolitic ignimbrites and tuffs). This essentially Late Triassic to earliest Jurassic (age based on Dicrodium Flora and micro-
flora) unit overlies rhyolitic deposits of the Choyoi Group with slight unconformity, and underlies Pliensbachian marine deposits.

In the coastal region at 32ºS, the deposits of this stage in the El Quereo–Los Molles Basin (Fig. 3.11) correspond to the 748-m-thick, late Norian to early Pliensbachian marine Los Molles Formation (Cecioni & Westermann 1968) (Fig. 3.10). A lower conglomeratic sandstone member is followed by a pelitic member; with two higher members consisting of alternating shales and greywackes, some of which are turbiditic, and a thick-bedded turbiditic succession. The facies evolution indicates an initial phase of deposition in gradually deeper environments, suggesting rapid subsidence of the basin, and a second phase of gradually shallower sedimentation. Palaeo-current indicators suggest sediment supply from the south and SSW, and a slope dipping in the opposite direction (Cecioni & Westermann 1968; Bell & Suárez 1995). An abundant fossil flora has been collected from this unit (Fuenzalida 1938; Azcárate & Fasola 1970). A probable eastward extension of this basin in Argentina is represented by the deposits containing a Triassic ammonite (Choristoceras cf. marshi) recently found by Riccardi & Igleias Llanos (1999) in the Argentinian side of the Principal Cordillera at 36ºS. About 20 km east, in La Ligua region, early Sinemurian to at least early Pliensbachian (von Hillebrandt 2002) deposits of the Quebrada del Pobre Formation (Thomas 1958) (Raíz del Cobre Beds; Piracés 1977), equivalent in age to the Early Jurassic portion of the Los Molles Formation, were deposited during the late phase of basin formation of the pre-Andean cycle that reached earliest Jurassic Jurassic times before development of the volcanic arc, which in this region is represented by the Aji Formation (Thomas 1958).

Further south, at 36ºS, close to the international boundary, the 110-m-thick Cajon de Troncoso Beds correspond to sand-
stone and shale intercalations with Late Triassic fossil plant remains covered by tuffs and rhyolitic breccias. These deposits are unconformably covered by marine Jurassic deposits of the Nacientes del Teno Formation (Muñoz & Niemeyer 1984).

The marine deposits assigned to the Bio-Bio–Temuco Basin (Fig. 3.11) are located in the Coastal Cordillera, whereas the continental ones are located in the Central Depression and east-
ern Principal Cordillera. The marine successions correspond to a transgressive series generally overlying the Palaeozoic metamorphic complex and Palaeozoic granitoids. Some of the northernmost deposits in this region (Hualañé, Curepto–Gualleco, Cerro Gupo, Cerro Pillay and Pocilla; see Fig. 3.10) are underlain by silicic volcanic deposits (Muñoz Cristi 1960; Thiele 1965; Corvalán 1965a, b, 1976; Escobar 1976; Hervé et al. 1976b; Thiele & Morel 1981; Spichiger 1993) that we assign to the La Totoro–Pichidanguí volcanic pulse. South of 36ºS, the silicic volcanic deposits are absent. As seen in the Los Molles region, the Late Triassic transgressive succession exposed in Vichuquén–Tilicura (Corvalán 1976), Hualañé (Corvalán 1976; Gutiérrez 1980; Spichiger 1993), and Curepto–Gualleco (Thiele 1965) reaches the Sinemurian stage without interruption (von Hillebrandt 2002), and is conformably covered by the Middle Jurassic volcanic deposits of the magmatic arc.

Classic localities in the Bio-Bío river valley, next to Concepción at 37ºS, in the Coastal Cordillera (Fig. 3.10), consist of a thick fossiliferous, mainly detrital succession with a marine intercalation in the middle part included in the Santa Juana Formation (Ferraris 1981). According to Nielsen (2005), these deposits accumulated in an actively subsiding basin and correspond to alluvial braided plain deposits that grade later-
ally from alluvial fan to fluvial and, more distally, to lacustrine deposits. Coal seams have been reported from the lower levels (Steinmann 1921; Jaworsky 1922; Tavera 1960; Cucurella 1978). Fossiliferous remains (fresh water and marine inverte-
brates, and plants) indicate a Late Triassic age (Steinmann 1921; Jaworsky 1922; Tavera 1960; Nielsen 2005), with the latter two authors assigning to these deposits a more precise Norian and probable Carnian age, respectively.

Exclusively continental deposits of the Bio-Bio–Temuco Basin correspond to the following localities: Cero Parra, immediately east of the described outcrops of the Bio-Bio river valley, in the western margin of the Central Depression; Llufquentué and Nielol-Huimpil, next to Temuco; and Tralcán,
next to Panguipulli, at 40°S (Fig. 3.10). At Cerro Parra the deposits overlie plutonic units of Palaeozoic age and are covered by Mesozoic lavas whereas in the Temuco region both base and top are unexposed. In the Panguipulli lake region the deposits correspond with the Panguipulli and Tralcán formations of middle to Late Triassic age (Brüggen 1951; Hauser 1970; Hervé et al. 1976a; Parada & Moreno 1980; Rodriguez et al. 1999). The Tralcán deposits (Aguirre & Levi 1964) unconformably overlies the Trafún Metamorphic Complex and are unconformably covered by Quaternary sediments and lavas. The Panguipulli and Tralcán formations are parts of the same fluvial and lacustrine depositional system. The Late Triassic age of these two formations is confirmed by their abundant flora (Brüggen 1951; Hauser 1970; Hervé et al. 1976b; Arrondo et al. 1988; Troncoso et al. 2004; Herbst et al. 2005) and palynomorph content (Askin et al. 1981; Zavatieri et al. 2003).

Considering the Late Triassic–Early Jurassic low sea-level stand, it is deduced that subsidence in the basins formed during the last or younger rifting stage must have been considerable to permit the ingress of the sea. This might explain, for instance, the presence of marine deposits in the Malargüe region in western central Argentina (Riccardi & Iglesia Llanos 1999) (Fig. 3.11). Some of these marine deposits reached regions far from the present-day coastline, which later (in the next Andean tectonic cycle) were to become part of a backarc basin, thus permitting continuous Late Triassic to Late Jurassic marine sedimentation.

**Intrusive units**

Apart from the already mentioned occurrences of volcanic deposits recorded during this cycle, extensive plutonic activity was developed during Late Permian to Early Jurassic times (Fig. 3.10). These bodies are exposed in the High Andes between 24S and 31S (Mpodozis et al. 1983; Nasi et al. 1985, 1990; Mpodozis & Kay 1990), and along the coastal region between Chañaral (26°30’S) and San Antonio (34°S), and probably continue further south along the Coastal Batholith (Berg & Breitkreuz 1983; Breitkreuz 1986a; Berg & Baumann 1985; Gana & Tosdal 1996; Godoy & Lara 1998).

According to Mpodozis & Kay (1990), the plutonic units in the High Andes (c. 27–31°S) form a continuous plutonic belt that corresponds to a post-collisional, epizonal association including granitoids derived from deep, garnet-bearing levels in a thickened crust, and hyperstilic, calcalkaline to transitional

A-type granites, indicating extensive crustal melting of a garnet-poor crust (Ingaguás Superunit; Mpodozis & Kay 1990). This plutonic belt, and associated silicic volcanics, is coeval with the younger portion of the Choiyoi Group described by Llambias (1999) for the Frontal Cordillera. The Ingaguás Superunit, which is considered an age equivalent of the Pastos Blancos Group, consists of five units: Los Carricitos, Ingaguás Superunit, which is considered an age equivalent of the Early Permian A-type granites, indicating extensive crustal melting of a garnet-poor crust (Ingaguás Superunit; Mpodozis & Kay 1990). More recently, Martin et al. (1999a) obtained for the Chollay–El León units U–Pb in zircon ages of 249.7 ± 3.2 Ma, 242.5 ± 1.5 Ma and 242 ± 1.5 Ma, which are close to the Permian–Triassic boundary. These same authors obtained K–Ar ages of 221 ± 5 Ma (moscovite) and 219 ± 5 Ma (biotite) for the El Colorado unit. These new results confirm the previously Permian to Late Triassic age assigned to the Ingaguás Superunit (Chollay–El León and El Colorado units), do not invalidate the possibility that this plutonic activity continued in the Frontal Cordillera until Early Jurassic times, and confirm its correlation with the younger Choiyoi Group.

The plutonic units reported for the coastal region form scattered exposures, which have been more intensively studied in some places than others, and dated by means of different methods. In the Chañaral–Caldera region (26–27°S), a series of plutons have been differentiated, and yielded ages that cover the complete age range of the pre-Andean tectonic cycle (Berg & Baumann 1985; Grocott et al. 1994; Godoy & Lara 1998). Late Permian intrusions are represented by the syenogranitic and granitic Quebrada del Castillo pluton and probably the Quebrada Quiscuda pluton. Triassic intrusions are represented by the monzonitic and syenogranitic Pan de Azúcar. Cerros del Yetaedo and Capitana plutons that correspond to leucocratic S-type granitoids with high 87Sr/86Sr ratios (0.7103–0.7172; Berg & Baumann 1985). Early Jurassic intrusive rocks in this region crop out as the relatively small Bufadero, Peralillo, Cerro Castillo and Barquitos plutons, with ages ranging between 204 Ma and 193 Ma (Farrar et al. 1970; Berg & Baumann 1985; Berg & Breitkreuz 1983; Godoy & Lara 1998), and the larger Flamenco pluton (26°20’S) with ages between 202 Ma and 186 Ma and a calcalkaline character (Berg & Breitkreuz 1983; Breitkreuz 1986a; Berg & Baumann 1985; Grocott et al. 1994; Dallmeyer et al. 1996). The comparatively lower Sr isotope ratios obtained in some of these plutonic rocks (0.7042–0.7053; McNutt et al. 1975) and their calcalkaline character possibly indicate that some of them already correspond to magmas associated with the beginning of subduction. No age determinations are available to confirm plutonic activity for this tectonic cycle between 298°S and 30°S (Moscoso et al. 1992a), but this absence is possibly due only to the lack of more detailed studies. However, we do not discard the possibility that in some parts of the coastal region the pre-Andean plutons might be hidden below pre-Andean deposits filling the NW–SE orientated extensional basins. A similar view has been advanced by Grocott et al. (1994). South of La Serena along the coastal ranges, plutonic activity of Triassic plutonic rocks has been reported by Irwin et al. (1988) and Gana (1991). In the Coastal Batholith, at the latitude of Santiago (33–34°S), these plutons correspond to the calcalkaline, pre-tectonic Cartagena Dioritic Gneisses and the Tejas Verdes Unit. U–Pb age determinations yielded a 214 ± 1 Ma age for the Cartagena Dioritic Gneisses, and a 212 ± 5 Ma age for the Tejas Verdes Unit (Gana & Tosdal 1996); these ages are similar to those obtained further north at c. 31°30’S by Irwin et al. (1988) and Gana (1991).

**Tectonic evolution**

From the foregoing description we emphasize the following aspects which relate to tectonic evolution during this cycle. The breccia deposits at the base of both cycles are interpreted as having formed at the bottom of cliffs formed by extensional faults. The deposits of the older or first stage are concentrated in the two central basins (San Félix and El Quereo–Los Molles), whereas the deposits of the younger or second stage are distributed in one of the central basins (El Quereo–Los Molles) as well as in the two more external ones (Profta–La Ternera and Bio-Bio–Temuco) (Fig. 3.14). The small La Ramada Basin is located between the two central basins. Deposits in each basin and in both stages accumulated in strongly subsident environments, and display a cyclic sedimentary evolution, i.e. the marine deposits correspond to transgression–regression cycles. In the Panguipulli basin only the El Ramada Unit was apparently reactivated and the external basins were formed. In the San Félix Basin no reactivation took place and some local deformation occurred causing the local angular unconformities
observed by Reutter (1974) between the La Totora and the Lautaro formations (e.g. Cerro Tatul). Based on this we deduce that the development of the basins progressed towards the sides of the previously formed central basins and that each stage corresponds to a complete cycle of rifting (tectonic and thermal subsidence); the cycles are separated from each other by an episode of intense volcanic activity mainly localized in and next to the central basins, in which the second stage continued the extension that had occurred during the first stage (Fig. 3.14). This model coincides with the evolution proposed in Argentina by Milana & Alcobe (1994) for the Bermejo Basin, and by Spalletti (2001) for the Cuyo Basin, in which they detected cycles of synrift (tectonic) subsidence and post-rift thermal subsidence. A similar evolution has been deduced by Cornejo et al. (1998) for the younger stage, based on the depositional evolution of the El Mono Beds (26°30′S to 27°S), and by Suárez & Bell (1992) for the Profeta–La Ternera Basin, and Bell & Suárez (1995) for the Los Molles Formation. The La Totora–Pichigangu volcanic pulse apparently developed early during or shortly before the second extensional episode. The reactivation of the extension in the central grabens probably facilitated the ascent of crustal melts accumulated at deeper levels in the continental crust, whereas in the external, newly formed basins, especially the Bio-Bío–Temuco Basin, crustal fracturing probably reached shallower crustal regions and did not favour ascent of magmas. Similarly, the extrusion of the older volcanic deposits of this cycle (Cas and Peine formations and other deposits) was probably associated with initial extension at the beginning of the pre-Andean tectonic cycle.

Evolution in southern Chile (Patagonia), between 43°30′S and 47°S

The geological evolution in this region during the pre-Andean cycle departs from the model described above. Until relatively recently no depositional ages pertaining to this tectonic cycle had been reported in this region, then Fang et al. (1998) reported the existence of Late Triassic marine fossils in the Chonos Archipelago, at 45°25′S, and Hervé & Fanning (2003) discovered Triassic zircons in the Diego de Almagro Metamorphic Complex, at 51°30′S (Fig. 3.15) (see also Chapter 2). These rocks had previously been considered of Palaeozoic age.

The metamorphic complex in the Chonos Archipelago consists mainly of submarine fan metaturbidites with occasional occurrences of metabasites and, subordinately, pelagic cherts on the eastern side (Hervé et al. 1981b; Thompson & Hervé 2002). These rocks are strongly deformed and present well-developed zones of broken formation. Transposition of a first, close-to-vertical axial plane foliation by a flat-lying second foliation increases rapidly towards the west. On the western side, strongly foliated mica and amphibole schists predominate, with local metacherts, metamorphosed under high P–T conditions (Willner et al. 2002). These features allowed interpretation of this unit as a subduction complex. The fossiliferous Potranca Formation is located on the eastern side of the complex, where primary structures are recognizable. More recent U–Pb SHRIMP age determinations (Hervé & Fanning 2001) confirm the biostratigraphic age for the Chonos complex (Thompson & Hervé 2002; see Chapter 2).

The Potranca Formation is clearly younger than the age deduced for the sedimentary protolith of the Late Palaeozoic metamorphic complexes described so far in this chapter. Deformation and metamorphism of the complex are, therefore, younger than Late Triassic (the age of deposition), but older than Early Cretaceous (140 ± 6 Ma), which is the oldest age obtained for the intrusion of the North Patagonian Batholith into the Chonos subduction complex (Pankhurst et al. 1999). This situation apparently suggests that subduction in this region remained active during the pre-Andean cycle and until the Early Mesozoic, whilst in northern regions subduction was quiet during this time. However, it is possible to explain this apparently anomalous situation if amalgamation of the subduction complex with the continent and the inversion or closure of the forearc basin were not completely achieved in the Late Palaeozoic. The remaining accommodation space might have permitted further sedimentation during Triassic times. In this case, amalgamation of the subduction complex with the continent and deformation of the basin fill would have occurred once subduction resumed in earliest Jurassic times as is the case in northern regions of Chile. In fact, based on U–Pb SHRIMP and fission track ages, Thompson & Hervé (2002) constrained the stratigraphic and metamorphic age of the Chonos complex to Late Triassic–Early Jurassic. The 198 Ma age (Sinemurian) obtained by Thompson & Hervé (2002) for the exhumation of the Chonos complex coincides with the age of resumption of volcanism and of subduction in northern and central Chile (e.g. La Negra and Ajiaval formations).

The Diego de Almagro is an accretory complex consisting of blueschists, quartz-mica schists, amphibole schists and orthogneisses, located to the west of the less metamorphosed rocks of the Madre de Dios Complex (see Chapter 2) (Fig. 3.15). These two units are separated from each other by a major shear zone containing a mylonitized granite (Seno Arcabuz Shear Zone) (Olivares et al. 2003). SHRIMP U–Pb dating of zircons from the Diego de Almagro Complex and the foliated granite gave Middle Jurassic ages around 170 Ma, indicating that (1) the zircons probably originated with the widely extended Middle to Late Jurassic silicic, anatectic magmatism associated with the opening of the South Atlantic Ocean, and (2) the magmatic rocks were emplaced in the recently accreted Madre de Dios Complex (Hervé & Fanning 2003). These authors concluded that the Diego de Almagro accretionary complex resulted from the subduction of slices of continental margin (tectonic erosion) together with mafic oceanic rocks, which were later dragged upwards to the surface or uplifted along the Seno Arcabuz Shear Zone.

The pre-Andean tectonic cycle: summary and discussion

The pre-Andean cycle affected all units previously formed along the western margin of Gondwana. The major faults controlling rifting during this cycle coincide with zones of weakness that, according to Ramos (1994), correspond to the suture zones of accreted Palaeozoic terranes. Considering that
the orientation of these faults is NNW–SSE, it is possible to deduce that continental growth by terrane accretion as well as by amalgamation of the accretionary prism to the continent occurred along a NNW–SSE orientated continental margin. This situation suggests that rifting in the southern part of the described continental margin affected terranes, which were amalgamated with the continent later than further north. The Early Jurassic age determined for the deformation and accretion of the Madre de Dios and the Diego de Almagro complex seems to confirm this idea.

The description of events that occurred during the pre-Andean cycle represents the evolution of the continental margin after assembly of the Gondwana megacontinent and prior to its break-up. The processes caused by the growth and existence of a megacontinent are at the same time the cause of its subsequent dismemberment. If we consider that the last appearance of magmas with an orogenic signature occurred at c. 240 Ma (Mpodozis & Cornejo 1988; Nasi et al. 1999), which is the oldest age obtained in the Ingaguás Superunit (except for an apparently anomalous older age mentioned by Mpodozis & Cornejo 1988; Mpodozis & Kay 1990), and that the first magmas associated with the renewal of subduction in Early Jurassic occurred in the Sinemurian–Pliensbachian (boundary at 194.5 Ma), it is possible to estimate a 55-million-year time span for the non-orogenic magmatic episode associated with the formation of the megacontinent.

Andean tectonic cycle (late Early Jurassic–Present)

The Andean tectonic cycle began in Early Jurassic times once the quiescent period of plate movement described above came to an end and subduction renewed its activity. This tectonic cycle reflects the evolution of the active continental margin of western Gondwana and South America during continental break-up and continental drift. The renewed subduction activity has created Andean arc magmatism that has continued almost uninterrupted right through to the present day.

In the evolution of this cycle in Chile it is possible to differentiate two regions with different palaeogeographic developments: (1) north, central and south-central Chile, north of 39°S; and (2) southern Chile (Patagonia), south of 42°S. The area between these two regions represented some kind of NNW–SSE orientated swell separating the basins developed north of 40°S and south of 42°S. The evolution of southern Chile is treated separately at the end of this chapter.

The early evolution of this cycle in northern, central and south-central Chile (north of 39°S) is characterized by the development of a magmatic arc parallel to the western margin of Gondwana with a backarc basin on its eastern side. In contrast, the later evolution (Late Cretaceous and Cenozoic) is characterized by the gradual shift to more easterly positions of the magmatic arc, and by the development of foreland basins on the eastern side of the arc. These two major periods correspond to the Early Period and Late Period, respectively.
ANDEAN TECTONIC CYCLE

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* Subdivision according to Coira et al. (1982)
** Subdivision according to this work

Fig. 3.16. Subdivision in stages of the 190-million-year-long Andean cycle, from late Early Jurassic to Present: first stage from late Early Jurassic to late Early Cretaceous (c. 90 million years); second stage from Late Cretaceous to Early Palaeogene (middle Eocene) (c. 55 million years); and third stage from Early Palaeogene to Present (c. 45 million years). The first stage can be subdivided into two substages: first substage, from late Early Jurassic to Kimmeridgian (c. 40 million years); and second substage, from Kimmeridgian to late Early Cretaceous (c. 50 million years). These subdivisions correspond to 55 to 40-million-year periods during which extensional tectonic conditions prevailed in the arc, backarc and forearc domains. These extensional periods were separated from each other by much shorter contractional episodes during which inversion of the basins took place and contractional deformation also occurred elsewhere, controlled by pre-existing deep-reaching faults. Therefore Andean evolution appears to be characterized by rather long-lasting periods of tectonic extension followed by shorter episodes of contractional deformation. For this reason, complete transgression–regression sedimentary cycles were developed in the Jurassic and Early Cretaceous marine backarc basins. The first stage corresponds to the Early Period of Andean evolution according to Coira et al. (1982) and the second and third stages correspond to Coira’s et al. (1982) Late Period.

There is general stratigraphic continuity between the pre-Andean and the Andean cycle deposits. In the region where the Late Triassic–earliest Jurassic extensional basins of the younger stage of the pre-Andean cycle were developed, the Andean volcanic arc deposits in the coastal region cover the earliest Jurassic; marine sediments of the pre-Andean cycle without major unconformity. Similarly, further east the backarc marine deposits of the Andean cycle are conformable and continuous with the marine or continental deposits of the late pre-Andean cycle. The deposits of the first stage of the Andean cycle generally lie unconformably below early Late Cretaceous deposits, which are mostly of volcanic origin developed in a continental environment assigned to the second stage of Andean evolution.

To facilitate comprehension of the abundant available information, the evolution of this tectonic cycle is described in three stages: (1) Early Jurassic–late Early Cretaceous; (2) Late Cretaceous–Early Palaeogene; and (3) Late Palaeogene–Present. Some of these stages have been subdivided into substages (Fig. 3.16).

**First stage: late Early Jurassic–late Early Cretaceous**

The first stage of the Andean cycle in the region north of 39ºS corresponds to the development of a new palaeogeographic organization characterized by the development of an essentially north–south orientated magmatic arc along the present-day Coastal Cordillera and a backarc basin to the east of the arc. These palaeogeographic features can be followed further north (NW) into southern Peru. The new magmatic arc, emplaced to the west of the previous Late Carboniferous–Early Permian magmatic arc, and the elongated backarc basin developed parallel to the continental margin. Considering that the subducting plate remained inactive for some 40–50 million years (see above), it is probable that during this first stage of the Andean tectonic cycle the coupling between the old, and therefore cold, oceanic plate and western Gondwana was rather loose. This condition, which is probably the main cause for the development of the extensional conditions during growth of the arc and the backarc basin, was maintained for most of Jurassic and Early Cretaceous time, permitting the dominance of extensional tectonic conditions on the continental margin, intense magmatic activity along the arc, and abundant sedimentation in the backarc basin. Contemporaneously with volcanism, huge batholiths were emplaced into the volcanic succession.
regions: Arica to Chañaral (18°S to 26°S), Chañaral to La Serena (26°S to 30°S) and La Serena to Valdivia (30°S to 39°S).

Northernmost Chile between Arica and Chañaral (18°S to 26°S) Archean evolution. In northernmost Chile, the Jurassic arc has generically been named La Negra Arc, based on the classic type locality of the widely outcropping La Negra Formation (García 1967), close to Antofagasta (Quebrada La Negra) (see arc deposits in Fig. 3.17). However, several names have been given to the arc deposits along the Coastal Cordillera. The most common of these are: Arica Group, including the Middle to Late Jurassic Camaraca (Cecioni & García 1960) and Los Tarros formations (Salas et al. 1966; Tobor et al. 1968; Vogel & Vila 1980; García et al. 2004), in the Arica region (18°–19°S); Oficina Viz Formation in the Coastal Cordillera between Pisagua and Iquique (Thomas 1970; Silva 1977; Kossler 1998); and La Negra Formation (García 1967) in the region lying between Antofagasta and Chañaral (c. 26°S). These deposits have been included in the Early Andean Magmatic Province (Oliveros 2005). The La Negra Formation conformably overlies: (1) the Late Triassic? to Sinemurian shallow marine calcareous sandstones and limestones assigned to the Cerros de Cuevitas Beds in the Antofagasta region (Munoz 1989) (not the Permian Cerros de Cuevitas Formation defined by Niemeyer et al. (1997b), in the same area); and (2) the Hettangian–Sinemurian Pan de Azúcar Formation in the Chañaral region (García 1967; Naranjo & Puig 1984; Téllez 1986; Marinovic et al. 1995), which corresponds to the late deposits of the pre-Andean cycle (thermal subsidence phase of the second rift stage). The La Negra Formation is unconformably overlain by the Tithonian–Neocomian continental Caleta Coloso Formation.

Volcanic activity related to the La Negra Formation began in Late Sinemurian–Early Pliensbachian times and lasted until the Late Jurassic epoch (Naranjo 1978; Munoz 1989; Munoz et al. 1988b; Godoy & Lara 1998; González & Niemeyer 2005). Intensive $^{40}$Ar/$^{39}$Ar dating of the La Negra Formation between 18°30’S and 24°S indicates that the main magmatic activity took place between 159 and 137 Ma in Late Jurassic and Early
Cretaceous times (Oliveros 2005). However, there is evidence for older activity at 175–170 Ma in the Iquique region (Oliveros 2005). U–Pb determinations from the La Negra Formation in the Antofagasta region yielded 169 ± 5.3 Ma (Basso 2004), which is in agreement with the older activity detected by Oliveros (2005).

The associated deposits form most (c. 80%) of the present-day Coastal Cordillera, and the thickness of the arc deposits reaches several thousand metres (7–10 km); (Boric et al. 1990; Buchelt & Tellez 1988; González & Niemeyer 2005). This succession is mostly composed of andesitic and basalt-andesitic lavas locally forming pillows, with subordinate intercalations of continental and marine volcanoclastic and calcareous deposits (e.g. the Rencoret Beds; Muñoz 1989). Locally (e.g. east of Antofagasta), the lowermost levels of the La Negra Formation correspond to silicic ignimbritic deposits (Muñoz et al. 1988b; Muñoz 1989). Volcanism evolved with time from an initial stage of tholeiitic to a calcalkaline affinity in more evolved stages, and more alkaline signatures in its latest stages (Losert 1974; Palacios 1978, 1984; Rogers & Hawkesworth 1989; Pichowiak 1994). Low 87Sr/86Sr ratios indicate little crustal contamination, and the existence of an attenuated crust (Rogers & Hawkesworth 1989). According to Oliveros (2005), fractional crystallization was the dominating process for the differentiation of magmas.

Evidence has been advanced favouring volcanic activity located in central vents (Muñoz 1989), although some authors, based on the geochemical signature, considered that the La Negra Formation, at least at an early stage, originated by fissural volcanism (Scheuber et al. 1994). The extrusion of such a thick volcanic succession had to be accompanied by considerable crustal subsidence probably related to an extensional setting of the whole Jurassic magmatic belt (Scheuber et al. 1994; Dallmeyer et al. 1996). Considering that deposition mostly occurred at sea level, no significant positive relief was apparently formed during volcanic activity, confirming considerable crustal thinning, as is also indicated by the Sr isotope ratios (Scheuber et al. 1994; Maksaev & Zentilli 2002). The tectonic setting of the La Negra Formation has been debated: Davidson et al. (1976) and Buchelt & Zeil (1986) suggested the development of an active continental margin volcanic arc, whereas Palacios (1978, 1984) favoured the existence of an island arc, Rogers (1985) suggested an aborted marginal basin, Rogers & Hawkesworth (1989), Grocott et al. (1994) and Grocott & Taylor (2002) suggested development of the arc system in an extensional setting, and Scheuber & Reutter (1992), Scheuber & González (1999) and Reutter (2001) considered that the arc developed in a sinistral transensional environment.

Rocks of the La Negra Formation show evidence of several separated alteration events between 160 and 100 Ma (Oliveros 2005). These rocks have been intensively affected by heating, which can be attributed to both major intrusions and burial (very low and low grade metamorphism) (Losert 1974; Palacios 1984; Sato 1984; Oliveros 2005), and enriched in sodium close to intrusive bodies and mineralized zones (Losert 1974; Marinovic et al. 1995; Dallmeyer et al. 1996; Taylor et al. 1998; González 1999; Scheuber & González 1999). The La Negra Formation hosts several strata-bound Cu–(Ag) deposits like the Mantos Blancos and Mantos de la Luna ore deposits, and the Buena Esperanza and Michilla Districts (Losert 1974; Sato 1984; Boric et al. 1990; Maksaev & Zentilli 2002).

The homoclinal dip of the La Negra volcanic succession and the total absence of tight folds in it have been considered arguments indicating extensional conditions (Scheuber et al. 1994). According to this, extension occurred after deposition of the La Negra Formation and preceded the magmatic activity of the second substage.

Activity in northernmost Chile is also known from the second substage when the geotectonic setting of the arc was similar to that during the previous substage. However, few certain Late Jurassic–Early Cretaceous volcanic deposits have been reported from this region, probably either because arc activity was more reduced relative to that developed during the Jurassic period, or because these deposits have not yet been fully differentiated from the older ones. In the Araucanian region, the Late Jurassic–Early Cretaceous Ataúna Formation (Cecioni & García 1960), comprising over 1300 m of coarse conglomerates and sandstones with volcanic intercalations, and the probably equivalent Saucine Formation, correspond to the volcanic activity of the arc during this substage. The Ataúna Formation is conformably covered by the marine Neocomian Blanco Formation (Cecioni & García 1960). Furthermore, pyroclastic components in the upper Neocomian part of the Livícar Formation (Muñoz et al. 1988a) confirm the existence of volcanic activity at this time. A similar situation exists in the Antofagasta region: here, the stratigraphic equivalent of the Ataúna Formation is the Caleta Colooso Formation, which unconformably overlies the La Negra Formation, and the equivalent of the Blanco Formation is the Neocomian calcareous El Way Formation that conformably overlies the former.

Further south in the Talalt–Chañaral region, volcanic rocks of this second substage correspond to the >1000-m-thick Aeropuerto Formation (Ulrichsen 1979; Naranjo & Puig 1984; Marínovic et al. 1995), which represents the arc–backarc transition.

**Plutonic activity**

A number of calcalkaline plutonic units of different sizes, including dyke swarms, have been emplaced into the Coastal Cordillera (see intrusive rocks in Fig. 3.17). Several of these have been studied in detail in the Antofagasta region, e.g. the Early Jurassic Cerro Coloso Gabbricóx Complex, Middle Jurassic Cerro Bolfín Complex, and Late Jurassic Cerros Plutones (Scheuber & Andriessen 1990; Andriessen & Reutter 1994; González 1996, 1999; González & Niemeyer 2005). Further south along the Coastal Cordillera, plutonic bodies have been grouped in the Vicuña Mackenná Batholith that forms the bulk of the Coastal Cordillera in this region, and consists of several units of Jurassic and Early Cretaceous ages (Marínovic et al. 1995). Although plutonic activity apparently lasted without interruption until Early Cretaceous times (Scheuber 1987, 1994; Scheuber et al. 1994; Reutter 2001), it is possible to separate plutonic units emplaced during each one of the substages of the first Andean stage. The Barazarte (biotite granodiorites and leucocratic tonalites), Paranal (gabbro-norites, gabbros and diorites and their monzonitic varieties), and Blanco Encalada (biotite and hornblende granodiorites and quartz diorites) units have late Early to Late Jurassic ages (192 Ma to 157 Ma; K–Ar on minerals and concordant 87Rb–86Sr isochrons), that coincide within the time span of the first substage. Younger plutonic activity is represented, for example, by the Late Jurassic quartz dioritic to granodioritic Cerro Cristales Pluton (Scheuber & Andriessen 1990; González 1996, 1999; González & Niemeyer 2005), which, among several other plutons, is part of the Ventarrones Unit of the Vicuña Mackenná Batholith (Marínovic et al. 1995), and the two younger units of this batholith, Remiendos and Herradura. All these units have ages (K–Ar and 87Rb–86Sr isochrons) within the time span of the second substage. The Ventarrones Unit has an age close to the Jurassic–Cretaceous boundary, the Remiendos Unit has a late Neocomian age, and the Herradura Unit has a late Early Cretaceous age.

**Backarc evolution**

The basin infill in northernmost Chile has been described under different names depending on the area. In the Arica region (18–19ºS), it corresponds to the c. 1.700-m-thick, Sinemurian to Early Cretaceous (Neocomian) Livícar Formation (Muñoz et al. 1988a). This stratigraphic unit overlies unconformably rhyolitic deposits of probable Triassic age (Riolitas del Santuario), and is intruded by Late Cretaceous bodies and overlain by Cenozoic deposits (Muñoz et al. 1988a;
Muñoz & Charrier 1996; García 1996, 2001, 2002; García et al. 2004). The lower part of this formation corresponds to a transgression–regression cycle that represents the first substage of the first Andean stage (see Fig. 3.16). After deposition of a pelitic interval indicating deposition in a rather deep environment, regression during the Kimmeridgian stage is characterized by two gypsum intercalations, and is followed by a thick shallow marine succession that reaches the Neocomian and represents the second substage (Muñoz et al. 1988a).

Between approximately Pisagua and Iquique, the marine Jurassic succession in the Coastal Cordillera that represents the transgression–regression cycle of the first substage overlies arc volcanic rocks of the Oficina Viz Formation, and is composed, from base to top, by the following diachronous units (Fig. 3.18): the late early Bajocian to Bathonian? Caleta Lígate Formation, the early late Bajocian–middle Oxfordian El Godo Formation, and the middle to Late Oxfordian Huantajalla Formation (Kossler 1998). The volcaniclastic Caleta Ligate Formation corresponds to deposition during the period of basin enlargement, and reflects the intense influence of the arc in the backarc basin. The black, thinly stratified, pelitic, bituminous El Godo Formation with its occasional intercalations of lavas and volcaniclastic deposits, corresponds to a period of reduced circulation and oxygen content in the basin, whereas the calcareous Huantajalla Formation corresponds to a period of higher salinity and indicates the tendency towards closure of the basin. No early Cretaceous deposits are known from this region. To the east, in the Precordillera at 19º30'S to 20ºS, the marine succession formed by the Sinemurian Longacho and the middle Jurassic to Early Cretaceous Chacarilla Formations (Galli 1957, 1968; Galli & Dingman 1962) represents the transgression–regression cycle of the first substage. Footprints of different dinosaur groups (teropods, sauropods, ornithopods) are exposed in the greyish red sandstones of the upper Chacarilla Formation. According to Blanco et al. (2000), the presence of tracks of very large ornithopod dinosaurs in the upper continental layers of this formation indicates an Early Cretaceous age for these levels. These deposits are unconformably covered by the latest Cretaceous andesitic volcanic and continental sedimentary Cerro Empexa Formation (Galli 1956, 1957, 1968; Galli & Dingman 1962).

In the backarc basin of the region between Arica and Iquique (Tarapacá basin), although the influence of the Jurassic volcanic arc on the backarc sediments diminishes rapidly towards the east, as almost everywhere else along the basin, and the deposits rapidly pass to rather deep marine facies, the deepest regions of the backarc basin are to be found still further east in the present-day eastern Precordillera and Altiplano (Muñoz & Charrier 1993).

The backarc deposits NE of Antofagasta in the western flank of the Sierra de Moreno have been mapped as the Quinchamale Formation (Skarmeta & Marinovic 1981; Ladino 1998) (‘Quinchamale Formation’ in Figs 3.19 & 3.20). These deposits are Sinemurian to Kimmeridgian in age. Conformably overlying these deposits is a volcanic succession named the Cuesta de Montecristo Volcanites, which overlie the Early Palaeozoic Challo Formation and Late Palaeozoic granitoids, and have been subdivided into two members. A fossiliferous, marine, c. 1000-m-thick lower member, consisting mainly of limestones and shales, has a Sinemurian to Oxfordian age, and corresponds to the transgression–regression cycle of the first substage. The upper part of this member contains a 900-m-thick succession of conglomerates and sandstone intercalations, and presents short-distance lateral variations. The upper member corresponds to the deposits of the second substage of the first Andean stage. The upper part of this unit can be correlated with the Quehuita Formation (Vergara 1978a; Vergara & Thomas 1984) and the Capella Beds (Vergara 1978b), exposed towards the NE in the eastern Precordillera.

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### Figure 3.18

Chronostratigraphic relations between arc and backarc units during the first substage of the first stage of the Andean tectonic cycle for the Coastal Cordillera (CC) and Precordillera (PC) in the Arica and Pisagua–Iquique regions (18° to 20°30'S).
At the southern end of the Sierra de Moreno and slightly to the east of the region previously described, the stratigraphic succession representing the backarc deposits is partly similar to the one described for the western flank of the Sierra de Moreno (Fig. 3.19). It is represented by: (1) the marine Cerritos Bayos Formation (Biese 1961; Baeza 1979; Lira 1989), which corresponds to the first substage deposits; (2) the San Salvador Formation, which represents the second substage deposits; (3) the Sierra de San Lorenzo Beds, which are coarse-grained intercalations; and (4) the Guacate Andesites, which are an equivalent of the Cuesta de Montecristo Volcanites (Lira 1989). The lower Moctezuma Member of the Cerritos Bayos Formation corresponds to the transgressive deposits and to the development of an Early to Middle Jurassic calcareous platform, and the upper Campamento Member represents the Late Jurassic regression with gradually shallower facies, a gypsum intercalation, and finely laminated algal limestones. The San Salvador Formation corresponds to paralic and fine-grained continental deposits.

Southeast of Antofagasta, the backarc deposits are exposed in the Precordillera or western flank of the Domeyko Range and are known as the Sierra El Cobre Formation (Tobar 1966; Muñoz 1989) (Fig. 3.19). Further east the predominantly calcareous, Bajocian to Kimmeridgian Caracoles Group (García 1967), which can be correlated with the Cerritos Bayos Formation, represents deposits accumulated on the eastern flank of the basin (Muñoz 1989) (Fig. 3.20). The lower Moctezuma Member of the Cerritos Bayos Formation corresponds to the transgressive deposits and to the development of an Early to Middle Jurassic calcareous platform, and the upper Campamento Member represents the Late Jurassic regression with gradually shallower facies, a gypsum intercalation, and finely laminated algal limestones. The San Salvador Formation corresponds to paralic and fine-grained continental deposits.

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This group comprises a >1300-m-thick fossiliferous succession of tufts, conglomerates, limestones, calcareous sandstones and pelites (Torcazas and Caracoles formations), an alternation of limestones and pelites with gypsum intercalations (Doralisa Formation), massive gypsum (Millonaria Formation), and grey limestones (Honda Formation). The overlying and continuous continental sedimentation corresponding in this region to the second substage (Tithonian to Aptian–Albian) (Fig. 3.20) is represented by: (1) the c. 1600-m-thick fluvi al and alluvial deposits of the San Manuel Beds, an equivalent of the San Lorenzo Beds in the Cerritos Bayos region; (2) the >2000-m-thick monotonous succession of marls, mudstones, limestones, red sandstones and evaporitic levels of the Llanura Colorada Formation (Muñoz 1989; Marinovic & García 1999), which is correlated with the San Salvador Formation in the Cerritos Bayos region; (3) the volcanic succession of the Río Seco Beds (breccias, sandstones and andesitic lavas) (Muñoz 1989), that can be correlated with the Peña Morada Formation (Maksaev 1978). Cuesta de Monterrico Volcanites on the western flank of the Domeyko Range (Ladino 1998) and the Guacurco–Andesites are described in the Cerritos Bayos region (Lira 1989). These deposits are unconformably covered by the Late Cretaceous Quebrada Mala Formation (Montaño 1976; Muñoz 1989; Muñoz et al. 1989) (Fig. 3.19).

Further south, the Late Triassic–Late Jurassic (Tithonian) Profeta Formation (Chong 1973) includes the backarc deposits. This richly fossiliferous formation is exposed on both sides of the Domeyko Range and consists of a >2200-m-thick succession of marine, predominantly calcareous deposits. The succession begins with conglomerates, sandstones and breccias with coral-bearing limestone intercalations, and is followed by a succession of limestones and shales that reaches up to the Kimmeridgian, with two rather thick evaporitic intercalations, one of early Bajocian and the other of late Oxfordian age (Marinovic et al. 1995). The lower portion relates to late Triassic–earliest Jurassic shallow platform deposits that are assigned to the younger stage of the pre-Andean tectonic cycle. This lower portion is thus a stratigraphic equivalent of the Cerros de Cuevitas Beds of Muñoz (1989) and of the Pan de Azúcar Formation, exposed in the Coastal Cordillera in this region. The rest of the Profeta Formation corresponds to backarc deposits, coeval with deposition of the volcanic La Negra Formation (volcanic arc), and represents the transgression–regression cycle of the first late Early Jurassic to Kimmeridgian substage. This part of the formation is therefore an equivalent of the Sierra El Cobre Formation further north, and a partial equivalent of the Bajocian to Kimmeridgian Caracoles Group. Deposits representing the easternmost extent of arc volcanic activity are included in the Candeleros Formation (Naranjo & Puig 1984), which is a western and partial equivalent of the Profeta Formation. The latter is conformably overlain by the marine and continental clastic succession of the Late Jurassic–Early Cretaceous Santa Ana Formation (Naranjo & Puig 1984). This unit consists of a 450–500-m-thick succession of marls and algal limestones followed by sandstones and siltstones, locally with andesitic intercalations (Naranjo & Puig 1984; Marinovic et al. 1995). It corresponds to deposits accumulated during the second substage, and can be correlated with the Llanura Colorada Formation and its equivalents further north (Fig. 3.19). The lava intercalations probably are comparable with the Río Seco Beds (Muñoz 1989), exposed further north.

**Region between Chañaral and La Serena (26°S to 30°S)**

**Arc evolution.** In the northern part of this region, immediately north and south of Chañaral, the arc volcanics assigned to the first substage correspond to the already described La Negra Formation (Naranjo 1978; Godoy & Lara 1998). Here, as in the region south of Antofagasta, this formation conformably overlies the late pre-Andean Hettangian–Sinemurian deposits, which are named there the Pan de Azúcar Formation. Locally it overlies the Late Devonian–Early Carboniferous Las Tórtolas Formation (Complejo Epimetamórfico Chañaral). An eastern or distal extension of the La Negra Formation is the Middle–Late Jurassic to Early Cretaceous Sierra Fraga Formation, known east of Chañaral (Tomlinson et al. 1999). According to these authors, this formation consists of a >2000-m-thick succession of andesitic and basalt-andesitic lavas and volcaniclastic deposits with marine calcareous intercalations containing Bajocian and Oxfordian fossils and corresponding to eastern or distal arc deposits (La Negra arc) that interfinger with backarc deposits (Fig. 3.21).

Further south, SE of Copiapó, during the first substage, arc deposits assigned to the Sierra Fraga Formation (Iriarte et al. 1996) interfingered with the backarc Lautaro Formation (Jensen 1976; Sofía 1989) probably correspond to the volcaniclastic apron developed on the eastern side of the arc (La Negra arc) (Figs 3.21 & 3.22). Eastward progression of the arc deposits caused regression in some regions of the backarc basin as early as late Toarcian times (Sofía 1989), comparatively earlier than in other regions. During this progression, the arc deposits overlie the transitional, marine to continental, regressive facies sediments that overlie the Lautaro Formation (Rewert 1974).

In the second substage (Late Jurassic and Early Cretaceous), the arc deposits in this region correspond to the Punta del Cobre Formation (with a position relative to the arc like that of the Sierra Fraga Formation in the preceding substage) and the Bandurrias Group (Moscoso et al. 1982a) or Formation (Segerstrom 1960; Arévalo 2005a) and their southern equivalents (Figs 3.21 & 3.22). The Late Jurassic–pre-Late Valanginian, volcanic–volcaniclastic Punta del Cobre Formation has a transitional boundary with the underlying La Negra Formation (Godoy & Lara 1998), indicating continuous Jurassic to Early Cretaceous (Valanginian) volcanic activity in this region (Arévalo 2005a), and underlies the marine backarc deposits of the Chañarillo Group (Segerstrom & Ruiz 1962; Marschik & Fontboté 2001) of the second substage. It consists of a lower, mainly volcanic member (Geraldo–Negro), and an upper, mainly volcaniclastic member (Algarrobo) (Marschik & Fontboté 2001b) and most probably corresponds to the deposits of the intermediate region between the arc and the backarc. This formation hosts the iron-oxide-rich Cu–Au–Zn–Ag deposits of the Punta del Cobre belt (Marschik & Fontboté 2001b; Arévalo et al. 2006).

The Bandurrias Group forms a 2500-m-thick predominantly volcanic and volcaniclastic succession with sedimentary detrital and marine calcareous intercalations that correspond to the Hauterivian to Early Aptian transition zone between the arc to the west and the backarc basin to the east, represented by the Chañarillo Group (Segerstrom 1960; Segerstrom & Ruiz 1962; Jurgan 1977a, b; Arévalo 1995, 2005a). The Bandurrias Group is therefore characterized by rapid, predominantly east–west facies variations. In the Copiapó region, the westernmost outcrops are exposed immediately east of the city. Further west, rock exposures correspond to intrusive bodies that represent the roots of the magmatic arc. This group unconformably overlies older rocks like the Triassic Canto del Agua Formation (Moscoso et al. 1982a). In this region, the Cerrillos Formation unconformably overlies the transitional arc–backarc Bandurrias Group and the backarc Chañarillo Group (Arévalo 2005a, b). Therefore the Cerrillos Formation belongs to the first volcanic deposits of the second stage of Andean evolution, similar to the Panjuacha, Cerro Empexa, Icanche, Quebrada Mala and Augusta Victoria formations in the region between Arica and Chañaral, further north.

In the La Serena region, Jurassic arc volcanics are represented by the Agua Salada Volcanic Complex (Emparan & Pineda 2000, 2005), which had been previously included in the Chañarillo Group by H. Thomas (1967). The deposits of outcrops located along the western coastal ranges to the west of the Romeral Fault Zone, separated from each other by Jurassic plutons. The Agua Salada Complex consists of a c. 6400-m-thick succession of andesitic lavas and tufts with limestone.
Fig. 3.21. Chronostratigraphic relations between arc and backarc during the first stage of the Andean tectonic cycle in the Chañaral, Copiapó, Vallenar and La Serena regions.
The presence of calcareous deposits indicates rapid subsidence rates during arc activity in this region, as has also been deduced further north for the La Negra Formation. The top and base of the Agua Salada Volcanic Complex are unexposed. The Early to Middle Jurassic age assigned to this complex is based on the age of intrusions emplaced in these rocks (162.9 ± 6.7 Ma, U–Pb on zircon crystals) and the existence in this region of Palaeozoic metamorphic and Triassic plutonic units considered to be older than the complex (Emparan & Pineda 2000, 2005).

East of La Serena in the Rivadavia region, arc volcanics are represented by volcanic breccias and lavas conformably overlying the marine lower part of the Tres Cruces Formation (Dedios 1967) (Figs 3.21 & 3.23). These lavas correspond to the Punta Blanca Member of the Tres Cruces Formation sensu Letelier (1977). The marine, Sinemurian to Pliensbachian (and possibly Toarcian; Letelier 1977) Tres Cruces Member of the Tres Cruces Formation, that underlies the volcanic deposits, can be assigned, like the Pan de Azúcar Formation in the Chañaral region, to the deposits accumulated in the last phase of basin development of the pre-Andean tectonic cycle, before the beginning of the volcanic activity associated with subduction. Therefore, the volcanic Punta Blanca Member is considered to correspond to arc deposits in this region. However, because of its presence relatively far to the east of the main arc deposits (Agua Salada Volcanic Complex) and plutons, the Punta Blanca Member can be considered an eastern or distal extension of the arc volcanics, similar to the Sierra de Fraga Formation further north in the Chañaral and Copiapó regions.

Early Cretaceous arc volcanism is mainly represented in this region by the Arqueros and Quebrada Marquesa formations (Aguirre & Egert 1965) in the Elqui river drainage basin (Fig. 3.23). These formations are exposed west and east of a

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**Fig. 3.22.** Schematic palaeogeographic cross-sections for the first and second substages of the first Andean stage in the Copiapó region. Abbreviations of stratigraphic units: BG, Bandurrias Group; CG, Chañarcillo Formation; LF, Lautaro Formation; LNF, La Negra Formation; PCF, Punta del Cobre Formation; SFF, Sierra Fraga Formation.

**Fig. 3.23.** Chronostratigraphic relations between arc and backarc deposits of the first and second substages of the first stage of the Andean tectonic cycle in the regions between Vallenar and La Serena (28°30’S to 30°S), and south of the Elqui River valley (30° to 31°30’S).
major structural feature, the Romeral Fault Zone (Emparan & Pineda 1999, 2000, 2005; Pineda & Emparan 2006), and comprise thick volcaniclastic deposits, and aresectic, and subordinately rhythmic, lavas with marine calcareous intercalations (Aguirre & Egert 1965; Thomas 1967; Emparan & Pineda 2000, 2005). Frequent stratiform manganese deposits are developed in the upper Arqueros Formation, which also includes highly porphyritic andesites (ocotic) containing bitumen in amydules and veinlets (Rieger 2003). The close relationship between the marine limestones and the lavas suggests that the arc at this time was rather depressed topographically, as proposed for the La Negra arc in Antofagasta. The age of the Arqueros Formation is Neocomian, based on its marine fossiliferous content and one 40Ar/39Ar plateau age on amphibole of 129±3.3 Ma. The eastern equivalent of the Arqueros Formation is the marine Rio Tascadero Formation (Rivano 1980) of Berriasian to possibly Barremian age (Rivano 1980), which represents the deposits accumulated in the backarc basin (Nasi et al. 1990; Mpodozis & Cornejo 1988; Pineda & Emparan 2006).

The Quebrada Marquesa Formation (Aguirre & Egert 1965), which to has been considered to represent the arc volcanics of the second substage (Fig. 3.21), conformably overlies the Arqueros Formation (Aguirre & Egert 1965) immediately east of the Romeral Fault Zone, and conformably overlies the backarc deposits of the Rio Tascadero Formation further east. The Quebrada Marquesa Formation is overlain by the Late Cretaceous Viiñita Formation (Aguirre & Egert 1965), although it is uncertain whether the contact is conformable or unconformable. The Quebrada Marquesa Formation consists, at the type locality, of a 1900-m-thick succession of continental coarse and fine sedimentary deposits, volcaniclastic deposits and lavas, with a marine calcareous and fossiliferous intercalation near to the base. Its general stratigraphic position, overlying deposits of the Arqueros and Rio Tascadero formations, and containing Hauverian fossils in a marine intercalation from its lower part, indicates a maximum (and possibly only locally) Hauterivian age for its lower levels (Aguirre & Egert 1962; Thomas 1967; Pineda & Emparan 2006). A radiometric age determination from a higher level in this unit yielded an age of 107±0.6 Ma (U–Pb on zircons), indicating that the age range of the Quebrada Marquesa Formation is Hauverian to middle Ablainian. According to Morata & Aguirre (2003), the high AlO3 and low MgO high-K lavas of the Quebrada Marquesa Formation correspond to shoshonitic basaltic andesites and andesites containing abundant Ca-rich plagioclase phenocrysts, with low initial Sr ratios. These geochemical characteristics are similar to the coreal volcanic deposits exposed at the latitude of Santiago (see the Lo Prado Formation below).

The Quebrada Marquesa Formation interfingers to the east with volcaniclastic deposits of the Pucalume Formation, which represent the volcaniclastic apron that interfingers eastward with the uppermost marine deposits of the backarc basin. This suggests that during deposition of the Quebrada Marquesa Formation, since middle Neocomian times, the relief formed by the arc gradually increased and this probably considerably reduced the connection between the open sea (west of the arc deposits) and the backarc. Based on this new palaeogeographic setting, we propose that the marine intercalations within the lower part of the Quebrada Marquesa Formation belong rather to the western side of the arc, facing the ocean, than to the backarc side of the arc.

Plutonic activity. Plutonic activity developed under transtensional tectonic conditions and contemporaneously with sinistral strike-slip movements along the Atacama Fault Zone (in the northern and central part of the region) and the Romeral Fault (in the southern part) (see below). Abundant intrusive bodies form most of the Coastal Cordillera both north and south of La Serena (Mercado 1978; Breitkreuz 1986b; Berg & Breitkreuz 1983; Godoy & Lara 1998), as seen in the regions located further north.

In the northern part of the region, Berg & Breitkreuz (1983) and Godoy & Lara (1998, 1999) mapped several Middle Jurassic to late Early Cretaceous plutons. Early Jurassic plutons in this region are the Bufadero, Peralillo, Cerro Castillo, Barquitos and Relincho, with ages between 204 Ma and 193 Ma (Farrar et al. 1970; Berg & Baumann 1985; Berg & Breitkreuz 1983; Godoy & Lara 1998, 1999), and the Flamenco, with ages between 202 Ma and 186 Ma (Berg & Breitkreuz 1983; Berg & Baumann 1985; Grocott et al. 1994; Dallmeyer et al. 1996; Godoy & Lara 1999). An example of a Middle to Late Jurassic pluton is provided by Las Animas, with ages between 160 Ma and 150 Ma (Lara & Godoy 1998). These subduction-related plutons have 87Sr/86Sr isotope ratios between 0.7053 and 0.7033 that indicate a mantle or deep-seated crustal source, and a geochemistry corresponding to calcalkaline, I-type granites (Berg & Breitkreuz 1983). Early Cretaceous intrusions crop out as the Las Tazas, Sierra Pastenes, Sierra Aspera and Sierra Dieciocho plutons (Godoy & Lara 1998, 1999).

Further south in the Copiapó region, several plutons are exposed along the axis of the Jurassic–Early Cretaceous arc and cover, almost without interruption, the time span of the first Andean stage. Some Jurassic plutons forming major outcrops are the Caldera Gabbro (192 Ma), Morro Copiapó Granodiorite (189–180 Ma) and Sierra El Roble Pluton (165–155 Ma) (Godoy et al. 2003). Early Cretaceous intrusions, in general located east of the Jurassic intrusives, include the Cerro Moradito (155–140 Ma), Cerro Morado (140–137 Ma) and La Breá (123–117 Ma) plutons (Lara & Godoy 1998; Godoy et al. 2003; Arévalo 2005a, b).

In the southern part of this region, next to La Serena, several plutonic units have been described by Emparan & Pineda (2000). Late Jurassic–Early Cretaceous plutons as the San Juan (148±6 Ma, K–Ar whole rock) and various dioritic intrusions, and Early Cretaceous intrusive units with ages between 131 and 110 Ma form extensive outcrops along the coast south of La Serena (Emparan & Pineda 2005).

Backarc evolution. In this region (26 to 30ºS), the backarc deposits are exposed along the Domeyko Range and the High Andes, and form broad north–south trending outcrops controlled by east-vergent thrust faults (Reutter 1974; Jensen & Vicente 1976; Pérez 1982; Moscoso & Mpodozis 1988; Soffía 1989; Nasi et al. 1990; Tomlinson et al. 1999). These deposits overlie late Palaeozoic plutons and Permo-Triassic volcanic and sedimentary rocks, assigned to the Quebrada del Salitre Formation, and form a thick succession of marine and continental deposits with volcanic intercalations.

The deposits of the first substage of Andean evolution are represented by the marine Early to Middle Jurassic Montandón and Asientos formations in the northern part of the region (Chañaral), between 26ºS and 26º30′S (Harrington 1961; García 1967; Pérez 1982; Tomlinson et al. 1999; Cornejo et al. 1998), and the Lautaro Formation (Segersrom 1959, 1968) further south, up to 30°S (Nusi et al. 1990) (Fig. 3.21). These successions are equivalent to the southern portion of the Profeta Formation. The Montandón Formation (Plenischachian to middle Bajocian) consists of a 1250-m-thick (eastward thinning) succession of fossiliferous, grey to dark grey marine limestones and marls, which are locally bituminous and are conformably overlain by the Asientos Formation (Bajocian–Callovian). The latter comprises grey marine fossiliferous limestones, grey volcaniclastic sandstones, and basaltic intercalations (most abundant in the upper portion of the succession).

Further west in the Precordillera, a succession of andesitic and basaltic andesites, which gradually passes downwards into Callovian marine deposits, has been included in the 650-m-thick Quebrada Vicuña Beds (von Hillebrandt 1973) by Tomlinson et al. (1999). These high-iron tholeiitic lavas (Tomlinson et al. 1999) possibly represent a volcanic episode developed in the backarc domain, and record a similar tectonic
setting to that of the Lagunillas or ‘Malm Continental Series’ of Jensen (1976) and Jensen & Vicente (1976); Picudo, Algarrobal, Mostazal and Río Damas formations (see discussion below). For this reason, the Quebrada Vicuña Beds are assigned to the beginning of the second substage of the first stage of Andean evolution (Fig. 3.21). The succession of lavas and the Asientos Formation are overlain by the 50–450-m-thick Pedernales Formation of Tithonian to Valanginian? age consisting of marine fossiliferous limestones and volcaniclastic sandstones (Harrington 1961; García 1967; Pérez 1982; Tomlinson et al. 1999). The Montandón and the Asientos formations therefore represent the first transgression–regression cycle (Jurassic in age), and the Pedernales Formation the second transgression–regression cycle (Late Jurassic–Early Cretaceous) of the first Andean stage, which is completed by the Quebrada Monardes Formation (the Agua Helada Formation of García 1967) (Muzzio 1980; Mercado 1982; Iriarte et al. 1996, 1999; Cornejo et al. 1998; Tomlinson et al. 1999). The Pedernales Formation is a southern equivalent of the Llanura Colorada and Santa Ana formations (see above). The marine deposits of the Pedernales Formation grade upward to the Quebrada Monardes Formation, comprising up to 1000 m of continental red coarse- to fine-grained sandstones and siltstones of late Neocomian to late Early Cretaceous age.

In the Copiapó region, the marine deposits of the second substage correspond to the late Valanginian to middle Aptian Chañarcillo Group (Segerstrom & Parker 1959), which interfingers to the west with the red volcanic and volcaniclastic deposits of the Bandurrias Group (Segerstrom 1968; Jurgan 1977a; b; Arévalo 1995, 2005a, b). Marine deposits of the Chañarcillo Group transgressively overlie the Late Jurassic to early Valanginian volcanic–volcaniclastic Punta del Cobre Formation (Segerstrom & Ruiz 1962) corresponding to the arc volcanic rocks, which at that time extended eastwards (Figs 3.21 & 3.22). The Chañarcillo Group is unconformably overlain by the continental, late Early to Late Cretaceous Cerrillos Formation (Segerstrom 1960; Segerstrom & Parker 1959; Corvalán 1974; Jurgan 1977a, b; Pérez et al. 1990; Mourès 2004; Arévalo 2005a, b). The Abundancia Formation represents the late Valanginian transgression, associated with the extensional tectonic episode that formed the backarc basin. It consists of well laminated grey mudstones and arkoses that laterally pass into the calcareous Nantoco Formation. This latter formation represents the calcareous platform developed at the Hauterivian–Barremian boundary and consists of grey mudstones and wackestones, and in its upper part, of evaporites and calcareous breccias indicating a regressive episode represented by the upper Nantoco Formation. The Totorallco Formation consists of laminated marls with chert nodules and volcaniclastic intercalations. Finally, the Pabellón Formation represents the emergence of the basin and comprises massive limestones and marine sandstones followed by continental sandstones. The discovery of a fragment of Parahoplites at this level indicates a middle Aptian age for the regression episode in the basin (Pérez et al. 1990). The Chañarcillo Group is correlated towards the north with the Santa Ana (Naranjo & Puig 1984) and the Pedernales (Harrington 1961) formations. Inversion of the Late Jurassic–Early Cretaceous basin in this region resulted in rather strong deformation of the basin infill. The contact zone between the Bandurrias Group, to the west, and the Chañarcillo Group, to the east, occurs along a major east-west-vertgent thrust (Cerrillos Thrust in the Copiapó region; Arévalo 1995, 2005a, b) that possibly corresponds to the reactivation of an extensional fault (basin-bounding fault?) that participated in the development of the basin. Further east, the Chañarcillo Group is deformed by east- and west-vergent structures that correspond to several contractional and extensional deformation episodes (Arévalo 1995, 2005a, b; Iriarte et al. 1996).

In the region between Vallenar and La Serena, the backarc deposits corresponding to the first transgression–regression cycle (first substage) are represented by marine calcareous platformal deposits assigned to the Lautaro Formation (von Hillebrandt 1970, 1971, 1973; 1981a, b, 2002; Reutter 1974; Jensen 1976; Jensen & Vicente 1976; von Hillebrandt & Schmidt-Effing 1981; Mercado 1982; Sepulveda & Vanjo 1982; Soñ Naña & Nasi et al. 1990; Iriarte et al. 1999). The Lautaro Formation is overlain by the Quebrada Vicuña (von Hillebrandt 1973) and Lagunillas (Iriarte et al. 1999) formations of Late Jurassic age (Fig. 3.21). This latter formation corresponds to the ‘Malm Continental Series’, defined by Jensen (1976) and Jensen & Vicente (1976). The Quebrada Vicuña Formation consists of 50–100-m-thick massive andesites and basaltic andesites with rare calcareous sandstone intercalations, and is exposed to the east of the Lagunillas Formation. The Lagunillas Formation consists of two members: (1) the Cocambico Member, comprising a 600-m-thick succession of red cobble and pebbly conglomerates with subrounded clasts of Palaeozoic granites and rhyolites in a sandy matrix rich in quartz, plagioclase and K-feldspar and red quartz-rich arkose intercalations; and (2) the overlying Peñasco Largo Member comprising a 500-m-thick succession of andesites and basaltic andesites (Iriarte et al. 1999). This volcanic activity was coeval with the lavas intercalated in the upper part of the Asientos Formation further north. The composition of the Cocambico Member suggests direct erosion from Late Palaeozoic plutons (e.g. Montosa pluton) exposed in the same region. The thickness of the lavas in these formations, together with their occurrence quite far to the east of the exposures of the Jurassic plutons in this region (which mark the position of the magmatic arc), suggest that the volcanic activity that produced these deposits was located in the backarc domain and probably does not correspond to arc volcanism.
In the La Serena region, the backarc deposits are represented by the Tres Cruces Formation (Dedios 1967), which is exposed about 70 km inland between c. 39°S and 31°S (Dedios 1967; Letelier 1977; Mpodozis & Cornejo 1988; Emparan & Pineda 1999; Pineda & Emparan 2006), and possibly the Quebrada El Tapado Beds, which are exposed in the high Andes at this latitude (Nasi et al. 1990; Mpodozis & Cornejo 1988) (Figs 3.21 & 3.23). At the type locality, the Early Jurassic (Pliensbachian up to possibly early Toarcian; Letelier 1977) deposits of the Tres Cruces Formation are overlain by a thick succession of volcanic breccias and lavas (Punta Blanca Member of the Tres Cruces Formation, sensu Letelier 1977) from the arc, probably its distal initial products. This eruption of magmatic materials caused an early (post-Toarcian) regression of the sea in this region of the backarc basin. The conformable relationship between the Late Triassic–earliest Jurassic Las Breas and Tres Cruces formations, and the position of this latter formation underlying early volcanic deposits of the arc, suggest that the lower marine levels of the Tres Cruces Formation and possibly the uppermost part of the Las Breas Formation (Fig. 3.23) can be assigned to the younger stage of the pre-Andean tectonic cycle. In this sense, the lower Tres Cruces Formation deposits are viewed as an equivalent of the Pan de Azúcar Formation and the earliest Jurassic levels of the Lautaro Formation. Therefore only the upper portions (Middle to Late Jurassic) of the Tres Cruces Formation exposed in this region represent the backarc deposits of the first substage. These upper portions, which reach up into Bajocian and even early Callovian times, and contain evaporitic intercalations, represent the regressive facies of the first transgressive-regressive Jurassic cycle in the backarc (Rivano 1975, 1980; Mpodozis & Cornejo 1988). The Quebrada El Tapado Beds, exposed in a more distal position relative to the volcanic arc, appear to be equivalent to the upper part of the Tres Cruces Formation, as well as to the Lautaro Formation exposed immediately to the north (Nasi et al. 1990).

The Algarrobal Formation (Dedios 1967) corresponds to the deposits overlying the marine sediments of the first substage (Fig. 3.21). This formation unconformably overlies the Lautaro and Tres Cruces formations, including the volcanic breccias and lavas that overlie the calcareous deposits of the last formation (Punta Blanca member sensu Letelier 1977), and is unconformably overlain by the Early Cretaceous continental volcaniclastic Pucalume Formation (Nasi et al. 1990) and the Berriasian–Barremian? marine Rio Tascadero Formation (Mpodozis & Cornejo 1988) (Fig. 3.23). At the type locality, the Algarrobal Formation consists of a 2000-m-thick continental succession of andesitic and dacitic lavas, volcanic breccias, conglomerates and sandstones (Nasi et al. 1990; Mpodozis & Cornejo 1988). Locally, the andesitic lavas are as much as 1000 m thick. South of the Elqui river valley (30°S), the Algarrobal Formation interfingers with the Mostazal Formation (Mpodozis & Cornejo 1988) which consists of a thick succession of volcaniclastic conglomerates and sandstones with andesitic intercalations. The enormous thickness of these deposits, in particular that of the andesitic intercalations, and the great size of the conglomeratic boulders, some of which exceed 1 m in diameter (Nasi et al. 1990), suggests that these deposits (Algarrobal and Mostazal) cannot have been derived from far away. Therefore, we follow a previous suggestion that these deposits accumulated in one or more extensional (sub basins developed within the backarc domain (see Rivano & Sepúlveda 1991, fig. 18a). Thus, the Algarrobal and Mostazal formations, as well as the Quebrada Vicuña Formation (Tomlinson et al. 1999) at the latitude of Chañaral, and the Lagunillas Formation (Iriarte et al. 1999) or ‘Malm Continental Series’ described by Jensen (1976) and Jensen & Vicente (1976) in the Copiapó river valley further north, can be visualized as rift-phase sediments and volcanism related to the initial backarc extension at the beginning of the Jurassic–Early Cretaceous regression cycle (see Rivano & Sepúlveda 1991, fig. 18) (Fig. 3.24). This episode probably reactivated the faults associated with the development of the backarc basin in late Early Jurassic times. A similar situation will later be proposed for the coeval Rio Damas Formation further south, at 34°S.

In the region between the latitudes of Vallenar and La Serena, the Algarrobal Formation is unconformably overlain by the Early Cretaceous Pucalume Formation (Dedios 1967) (Figs 3.21 & 3.23). In this region, the Barriquitas Beds and the conformably overlying Baños del Toro Formation are considered to be finer-grained sedimentary and volcanic equivalents of the Algarrobal Formation (Nasi et al. 1990; Mpodozis & Cornejo 1988).

As mentioned above, the Early Cretaceous deposits that unconformably overlie the Late Jurassic Algarrobal and Mostazal formations are the Pucalume (Dedios 1967) and the Rio Tascadero formations (Mpodozis & Cornejo 1988), formerly the Tascadero Formation of Rivano (1975). According to Nasi et al. (1990) and Mpodozis & Cornejo (1988), there is an interdigitation zone between a volcaniclastic succession to the west (Pucalume Formation) and a marine sedimentary succession to the east (Rio Tascadero Formation). Therefore, the Pucalume Formation represents an eastern, distal facies of the late Early Cretaceous volcanic arc, probably somewhat shifted towards the east relative to the Jurassic magmatic arc (Fig. 3.25). The Pucalume Formation is unconformably overlain by Late Cretaceous arc deposits of the Quebrada La Totora Beds (Emparan & Pineda 1999). Finally, given the rift setting envisaged for the Algarrobal and Mostazal formations, it is possible that the infra-Neocomian unconformity separating these formations from the overlying Early Cretaceous Pucalume and Rio Tascadero formations (Rivano & Mpodozis 1976; Mpodozis & Cornejo 1988) partly resulted from the tilting and strong erosion caused by the extensional process, with the deposition of Neocomian marine sediments on top of this palaeo-geography, rather than from a compressional event.

Central Chile (30°S to 39°S): between La Serena and Valdivia (Mendoza–Neuquén Basin)

In this region, the backarc basin gradually bends southeastward, is considerably wider than further north, and is known as the Mendoza–Neuquén Basin. This depocentre, which extends eastwards into Argentina, represents the southernmost part of the Jurassic–Early Cretaceous backarc basin, and is traceable without interruption along the eastern side of the magmatic arc from at least southern Perú to southern Chile (c. 42°S). Whereas the arc and backarc deposits of the first substage maintain in this region approximately the same location relative to the present-day coastline as their coeval equivalents further north, during the second substage (Late Jurassic–Early Cretaceous) both arc and backarc lay further east of the coastline than in the regions to the north. Thus, to the west of the arc (i.e. in the Coastal Cordillera), it becomes possible to identify another
depocentre, the Lo Prado Basin, which is located in a forearc position. As before, the following text initially examines the arc and backarc evolution of the first substage, and then the forearc, arc and backarc evolution of the second substage.

**First substage.** Arc. South of La Serena, Jurassic arc deposits correspond to a southern prolongation of the very thick Agua Salada Volcanic Complex (Emparan & Pineda 2005) (see above). In the Illapel region, between c. 31ºS and 32ºS, they probably correspond to the Pupio Beds, from which no precise age is available (Rivano & Sepúlveda 1991). These deposits, consisting of sandstones, shales and limestones with silicic lavas, are a probable equivalent of the Ajial Formation (see below). In the Coastal Cordillera south of 32ºS, the Jurassic arc activity is recorded in the Ajial Formation (Thomas 1958) and the Horqueta Formation (Piracés 1977) (Fig. 3.26). A Bajocian–Bathonian? fissiliferous marine succession of pyroclastic, detrital and calcareous rocks, known as the Cerro Calera Formation (Piracés 1976; Nasi & Thiele 1982), lies intercalated between the volcanic Ajial and Horqueta formations and represents a westward advance of the sea into the arc domain. The arc deposits conformably overlie the Sinemurian shallow marine deposits of the Quebrada del Pobre Formation, which is an equivalent of the upper part of the Los Molles Formation (Cecioni & Westermann 1968), and which we consider to be related to the late phase of the younger stage of the pre-Ancean cycle, like the Tres Cruces Formation (at the latitude of La Serena) and the Pan de Azúcar Formation (in the Chañaral region). The 750-m-thick Ajial Formation (Thomas 1958; Piracés 1976; Rivano 1996) is predominantly volcanic, consisting mainly of calcalkaline silicic lavas and pyroclastics, and subordinate basic lavas and clastic volcaniclastic deposits, but also contains deltaic, volcaniclastic turbidite and marine calcareous intercalations (Piracés 1976; Vergara et al. 1995; Rivano 1996). The presence of marine intercalations in the Ajial lavas suggests that the arc did not form a high relief, and probably also indicates high rates of subsidence in the arc domain. In Late Jurassic times, uplift or increased magmatic activity permitted deposition under continuously subaerial conditions, as indicated by the Horqueta Formation. This formation consists of a 1600–1700-m-thick continental andesitic and volcaniclastic succession (Thomas 1958; Piracés 1977; Vergara et al. 1995) that overlies the Middle Jurassic Cerro Calera Formation and unconformably underlies the Early Cretaceous Lo Prado Formation (Thomas 1958; Piracés 1977; Vergara et al. 1995). The Horqueta volcanic activity is coeval with a rapid plutonic pulse detected by Gana & Todsal (1996) in the Coastal Cordillera west of Santiago (see Plutonic activity below).

In the Curió region (35ºS), the Jurassic arc activity is represented by the Middle Jurassic Altos de Hualmau Formation (Morel 1981). Here also, the arc volcanic deposits overlie a marine succession of Late Triassic to Early Jurassic age (Estero La Higuera and Rincón de Núñez formations) (Morel 1981; Spichiger 1993; Bravo 2001), which we assign to the late phase of the pre-Ancean cycle. South of 35º30’S, these deposits have fewer exposures along the Coastal Cordillera, and disappear altogether south of 36ºS (Seranegomin 2003).

**Plutonic activity.** Between La Serena and Los Vilos (30ºS and 32ºS), Jurassic and Early Cretaceous plutonism appears to have been rather continuous, and forms a long, narrow belt of epizonal bodies along the coastal range. The plutons that can be assigned to this first-substage magmatic activity are the Millahue, Tranquilla, Puerto Osasco and Cavilóen units of Middle Jurassic to Early Cretaceous age, all of them grouped within the Mincha Superunit (Parada et al. 1988). The host rocks correspond to Late Palaeozoic to Early Jurassic units of the accretionary prism (the Choapa Metamorphic Complex), comprising the turbiditic Arrayán Formation, the Huventalquén Formation and the pre-Ancean cycle deposits. The plutons include monzogranites, syenogranites, diorites, monzodiorites, tonalites and granodiorites, and most of them are bounded by faults suggesting a passive intrusion mechanism (post-kinemetic). The Sr initial ratios obtained in the oldest Millahue and Tranquilla units of the Mincha Superunit (0.7063 and 0.7050, respectively) indicate some degree of crustal involvement at the beginning of plutonic activity, whereas the ratios obtained for the younger units, Puerto Osasco and Cavilóen (0.7034 and 0.7035, respectively), indicate direct derivation of the magmas from the upper mantle, with virtually no continental crust involvement (Parada et al. 1988). Gradual geochemical transitions between rocks of the same superunit suggest that fractionation played an important role in the intrasuperunit compositional variations.

This Jurassic plutonism extended southward along the Coastal Cordillera to the latitude of Concepción (c. 37ºS). At the latitude of Santiago, Middle to Late Jurassic intrusive rocks form a wide and extensive batholith that includes several units (Laguna Verde, Sauce, Peñuelas, Limache and Lliu-Lliu) comprising I-type, calcalkaline diorites, tonalites, granodiorites and granites (Gana & Todsal 1996; see also Godoy & Loske 1988). These authors indicate that intrusion of these units occurred in only 6 million years, between 162 Ma and 156 Ma, implying a very rapid pulse of ascent of large amounts of magma. It is interesting to note that this time interval approximately coincides with the age assigned to the Horqueta Formation (Kimmeridgian).

**Backarc.** No exposures of Jurassic deposits belonging to the first substage have been reported for the northern part of this region. The Tordillo Formation, exposed in the High Cordillera (Rivano & Sepúlveda 1991; Rivano 1996), is a lateral equivalent of the Algarrobal and Mostazal formations,
described further north (Nasi et al. 1990; Mpodozis & Cornejo 1988), as well as of red detrital continental sediments exposed to the east on the Argentine side of the Andes. These deposits were named Tordillense by Groeber (1951), and formally named as the Tordillo Formation by Yrigoyen (1976). The formation consists in this region of a 1520-m-thick succession of red conglomerates and sandstones, with an upward-fining upper part containing gypsum horizons (Rivano & Sepúlveda 1991). The base is not exposed in Chilean territory and it is unconformably overlain by the Late Tertiary volcanic Farellones Formation. The Kimmeridgian age assigned to this formation is based on the age of the eastward continuation of these outcrops in Argentina where the stratigraphic position of the Tordillo Formation is better constrained. In this region it contains no volcanic intercalations, and represents a distal facies probably deposited away from the main extensional fault(s) of the extensional basin (semigraben?)

The first transgression–regression cycle in the backarc basin between 33°S and 39°S is represented by marine deposits that crop out in the Principal Cordillera. These correspond, from north to south, to the lower member of the Lagunilla Formation (Aguirre 1960), the Nieves Negras Formation (Alvarez et al. 1997; Charrier et al. 2002a) (formerly Leñas–Espinoza Formation, defined by Klohn (1960) and redefined by Charrier (1982), and further south at 35°S to the Nacientes del Teno Formation (Klohn 1960 in Chilean territory and the Calera Formation of González & Vergara (1962) at 35°30′S, and the Nacientes del Biobío Formation (De la Cruz & Suárez 1997; Suárez & Emperán 1997) at 38°30′S (Fig. 3.27). All these formations are equivalents of the Lautaro and Tre Cruces formations further north, and are partially equivalent with the Cerro Calera Formation in the Coastal Cordillera (Fig. 3.28). The lower member (Icalma) of the Altos de Bio-Bio Formation, exposed in the Principal Cordillera, contains pillow basalts, breccias and massive basaltic lavas, locally with thicknesses of several hundred metres, suggesting that the arc in this region was located further east than in northern regions. The base of these formations is not exposed, except for the Nacientes del Teno Formation that unconformably overlies rhylitic rocks of possible Triassic age at 35°S (Davidson 1971; Davidson & Vicente 1973), and of confirmed Triassic age (Cajón de Troncoso Beds) between 36°S and 37°S (Muñoz & Niemeyer 1984) (Fig. 3.27). They correspond to a thick succession of sandstones (some of them turbiditic), marls and limestones, and represent a transgression–regression cycle that ends with thick Oxfordian evaporitic deposits, generally named ‘Yeso Principe’ (Schiller 1912) or more formally Auquilco Formation (Groeber 1946, in Argentina, and Santa Elena Member of the Nacientes del Teno Formation in Chile (Klohn 1960; Davidson 1971; Davidson & Vicente 1973). This gypsum unit, which is the middle member of the Lagunilla Formation at 33°S (Aguirre 1960), is overlain by the upper member of the Lagunilla Formation (Aguirre 1960) and its southern equivalent, the Río Damas Formation (Klohn 1960), which consists of breccias and alluvial fan deposits that grade towards the east into the red, finer-grained and thinner fluvial sandstones of the Tordillo Formation (Klohn 1960; Arcos 1987). At its type locality (Río de las Damas, next to Termas del Flaco, at 35°S), the Río Damas Formation consists of a c. 3000-m-thick red continental, detrital succession, with course and fine intercalations, that includes at the top a member comprising > 1000 m of andesitic lavas culminating in breccias containing enormous angular blocks, some over 4 m in diameter. Close to its contact with the Baños del Flaco Formation, dinosaur tracks are well exposed (Casamiquela & Fasola 1968; Moreno & Pino 2002; Moreno & Benton 2005). The great thickness and coarseness of these deposits, the volcanic composition of the sedimentary components, and the thick andesitic intercalations are similar features to those described above for the ‘Malm Continental Series’ and the Algarrobal and Mostazal formations. Thus, the upper part of the Lagunilla Formation, together with the Río Damas Formation, was deposited in an extensional basin with local intense volcanic activity that extended for several hundred kilometres along the Andean range (see Fig. 3.24). These Late Jurassic backarc deposits are conformably overlain by Late Jurassic to Early Cretaceous marine deposits corresponding to the second transgression–regression cycle of the second substage.

**Second substage.** From this region southward it becomes evident that the distribution of the marine deposits accumulated during the second substage corresponds to two clearly separated depositional areas, one in the eastern Coastal Cordillera, and the other one in the Principal Cordillera, and mostly on its

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**Fig. 3.26.** Late Triassic to Cretaceous stratigraphic succession for the Coastal Cordillera in central Chile, between 32°S and 34°S.
eastern side (Charrier 1984; Charrier & Muñoz 1994) (A and C in Fig. 3.29). The two basins are separated from each other by a volcanic domain that we propose to name the Lo Prado–Pelambres Volcanic Arc. Therefore, it is possible to identify three palaeogeographic domains at this moment, from west to east: the Lo Prado Forearc Basin, the Lo Prado–Pelambres Volcanic Arc, and the Mendoza–Neuquén Backarc Basin. Next we analyse the evolution of these three palaeogeographic domains, and later we will discuss the tectonic implications of this palaeogeographic organization.

**Forearc (Lo Prado Forearc Basin).** The marine and continental Late Jurassic to Late Neocomian Lo Prado and La Lajuela formations, exposed along the Coastal Cordillera NW, west and SW of Santiago, correspond to the older forearc basin deposits (A in Fig. 3.29). At 33°S, the Lo Prado Formation unconformably (Piracés 1976) or conformably (Rivano 1996) overlies the volcanic Late Jurassic Horqueta Formation and older units, and according to Nasi & Thiele (1982), south of the Maipo river valley (33°30′S) it conformably overlies the Horqueta Formation (Fig. 3.26). A similar view was reported by Bravo (2001) for the contact between the equivalent La Lajuela Formation and the underlying Middle Jurassic arc deposits of the Altos de Hualmapu Formation at c. 35°S. The Lo Prado Formation is conformably overlain by the continental volcanic Veta Negra Formation (Piracés 1976; Nasi & Thiele 1982; Vergara et al. 1995). The La Lajuela Formation is similarly conformably covered by volcanic deposits (El Culenar Beds), which have the same stratigraphic position as the Veta Negra Formation (Bravo 2001). These two extremely thick formations (3000 m for the Lo Prado, according to Rivano (1996), and 4100 m thick for the La Lajuela, according to Bravo (2001))

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**Fig. 3.27.** Stratigraphic succession for Jurassic to Cretaceous deposits in the Principal Cordillera in central Chile and Argentina, between 32°S and 39°S, based on: (A) Aguirre (1960); (B) González (1963), Thiele (1980); (C) Álvarez et al. (1997); (D) Klohn (1960), Davidson (1971), Davidson & Vicente (1973), Charrier (1973b, 1981b), Charrier et al. (2002a); (E) González & Vergara (1962), Muñoz & Niemeyer (1984); (F) De la Cruz & Suárez (1997), Suárez & Emparán (1997); (G) Legarreta & Gulisano (1989).
consist mainly of marine sandstones (some are turbidites), breccias, conglomerates and calcareous deposits in the lower portion, and alternating marine and continental sedimentary deposits with thick volcanic intercalations of bimodal geochemistry in the upper portion. The marine diagnostic ammonite fauna present in the Lo Prado and La Lajuela formations permits their assignment to the Lower Cretaceous (Neocomian) (Vergara 1969; Piracés 1977; Nasi & Thiele 1982; Bravo 2001) and more exactly to the Berriasian to Valanginian (Rivano 1996). The lavas in the Lo Prado Formation are high Al$_2$O$_3$ and low MgO high-K to shoshonitic porphyric basaltic andesites with low initial Sr ratios, indicating the existence of strong crustal extension in the forearc basin at this time (Morata & Aguirre 2003; Parada et al. 2005a).

The Las Chilcas Formation (Thomas 1958) consists of a 3500-m-thick succession of mostly coarse alluvial and volcanioclastic deposits with a thick calcareous intercalation (Rivano 1996). These deposits show short-distance lateral and vertical facies changes, and interfinger southward, within 40 km, with marine calcareous deposits of the Polpaico Formation (Corvalán & Vergara 1980; Gallego 1994; Martínez-Pardo et al. 1994). The Las Chilcas Formation conformably rests on the Veta Negra Formation (Rivano 1996) (see Fig. 3.26). The contact with the overlying Lo Valle Formation has been reported as conformable, unconformable and interdigitated. According to Gana & Wall (1997) it is unconformable, and the unconformity represents a 20 Ma hiatus. In contrast, Nasi & Thiele (1982) indicated that the lower member of the Lo Valle Formation interfingers with the Las Chilcas Formation, and they furthermore observed at 34ºS that the upper member of the Lo Valle Formation rests directly over deposits of the Veta Negra Formation. These observations are in agreement with the development of an extensional basin, in which the Las Chilcas (and the marine Polpaico) Formation was deposited, and with the continued existence of this basin (probably the sag phase) during deposition of the lower member of the Lo Valle Formation. In this same context, deposition of the upper member of this latter formation over the Veta Negra Formation would indicate that the relief formed by the basin had by then been completely filled so that deposition overstepped its margins.

Fossil remains in the Las Chilcas Formation are not age diagnostic, and neither are there radioisotopic age determinations. The Early Cretaceous (Neocomian) age of the Lo Prado Formation (Rivano 1996) and the Late Cretaceous (Campanian to Maastrichtian) $^{40}$Ar/$^{39}$Ar age determinations for the lower portion of the overlying Lo Valle Formation (Gana & Wall 1997) constrain the age of the Veta Negra and the Las Chilcas formations broadly to late Early to early Late Cretaceous times.

Arc. The arc stratigraphy described for the La Serena region, namely the marine, richly calcareous Arqueros and the
Marquesa formations and the transitional arc–backarc volcanioclastic Pucalume Formation (see Fig. 3.25), can be extended for some distance southward (Rivano & Sepúlveda 1991). South of c. 32° S one can clearly differentiate two outcrop belts of second substage arc rocks: one to the west, in the coastal region, and the other in the High Andes, close to the present-day Central Depression and the Coastal Cordillera (Fig. 3.28). No backarc deposits of the second substage are exposed in the Chilean cordillera south of 36° S.

At the latitude of Santiago, in the western part of the Mendoza–Neuquén Basin, the Late Jurassic–Early Cretaceous (Neocomian) transgression–regression cycle in central Chile resulted in the deposition of a thick succession of neritic to shallow marine (external platform), predominantly calcareous sediments of the San José (Aguirre 1960), Los Valdés (González 1963; Hallam et al. 1986) and Baños del Flaco (Klohn 1960; González & Vergara 1962; Covacevich et al. 1976; Charrier 1981b; Arcos 1987) formations (Fig. 3.27). At Termas del Flaco, in the Tinguiririca river valley, only the lowest portion of the Baños del Flaco is exposed because its upper portion, and probably also the overlying Colimapa Formation, have been eroded (Charrier et al. 1996). These deposits can be correlated with the Río Tascadero Formation and the Chañarcillo Group further north, and with the Mendoza Group in Argentina.

The final regressive episode led to the deposition of a second, generally thin band of gypsum (‘Yeso secundario’ or ‘Yeso Barremiano’) at the base of the 1500-m-thick, red detrital Colimapa Formation (Klohn 1960; González & Vergara 1962; González 1963; Charrier 1981b), which corresponds to the generally fine-grained continental deposits with thin calcareous intercalations containing ostracodes that followed the regression at the end of the Neocomian. Based on the presence of a charophyte oogonium it has been assigned an Aptian–Albian age (Martínez-Pardo & Osorio 1963). This formation is a lateral equivalent of the Huitrín–Rayoso Formation in western Argentina (Fig. 3.27).

Tectonic history of the first Andean stage
Regional extensional conditions generally prevailed during the first Andean stage, as evidenced both by the geochemical data and by the enormous thicknesses of arc volcanic deposits and backarc sediments, as well as in the forearc in the second substage. Within this tectonic setting, synsubduction crustal deformation occurred in both the magmatically active arc and backarc region further east. This first Andean stage was finally brought to a close by a pulse of Late Cretaceous compressive deformation that inverted the former backarc basin and created a major regional unconformity. Given the extensional tectonic setting, it is probable that the subduction boundary maintained a retreating tendency, possibly caused by subduction roll-back and/or slab steepening, throughout most of the first Andean stage, that is, between late Early Jurassic and late Early Cretaceous (c. 90 million years) times. Another possible explanation for the extension during the first Andean stage is the uplift of the asthenospheric wedge underlying the arc and backarc domains. This tendency was subsequently suppressed in latest Early Cretaceous times, probably due to the effect of increasing subduction rate along the continental margin (shallow-dipping slab) or rather to more rapid westward displacement of the continent (increase in overriding rate). More specific events linked to intra-arc deformation, backarc extension and basin inversion are discussed below.

Arc deformation: the Atacama Fault Zone
Deformation in the arc was mainly concentrated along the Atacama Fault Zone (AFZ) (Arabasz 1971), one of the major structural elements in the Chilenian Andes and which first developed during this stage. The fault is a continental-scale, trench-parallel strike-slip fault located along the Coastal Cordillera that can be traced for more than 1000 km between Iquique (20°S) and south of La Serena (Los Vilos and Los Molles at 32°S). The main trace of this fault system has been subdivided into three major curved segments,
concave to the west, which, from north to south, are Salar del Carmen, Paposo and El Salado–Vallenar (Naranjo 1987; Thiele & Pincheira 1987; Thiele & Hervé 1984; García 1991; Brown et al. 1993; Marinovic et al. 1995; Arévalo et al. 2003) (Fig. 3.30). Recent studies in the coastal region north and south of La Serena (29°30’ to 30°30’S) (Emparan & Pineda 2000, 2005) and between Los Vilos and Los Molles (32°S) (Arancibia 2004) allow the proposal of a new segment of this fault that we name Romeral–La Silla del Gobernador.

Detailed studies indicate that the fault has been at least intermittently active since Early Jurassic times (Hervé 1987a, b; Naranjo et al. 1984; Scheuber 1987, 1994; Scheuber & Andriessen 1990; Brown et al. 1993; Scheuber et al. 1994; Arévalo 1995, 2005a; Scheuber & González 1999; Armijo & Thiele 1990; Arévalo et al. 2003; González & Carrizo 2000, 2003). It runs through the plutonic rocks of the Jurassic and Early Cretaceous magmatic arc, suggesting a reduction of crustal strength caused by the high heat flow in the arc. Strike-slip movement was first proposed by Saint Amand & Allen (1960) and Arabasz (1971). More recently, sinistral as well as dextral displacements have been detected (Hervé 1987b; Naranjo et al. 1984; Scheuber 1987, 1994; Brown et al. 1993; Scheuber et al. 1994; Scheuber & González 1999; Grocott & Taylor 2002) suggesting contemporaneous oblique convergence. Vertical movements have also been reported for younger stages (Arabasz 1971; Hervé 1987a; Naranjo 1987; Armijo &

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Fig. 3.30. Traces of the different segments of the Atacama Fault Zone, possibly reaching as far south as 31°30’S, from north to south: Salar del Carmen, Paposo, El Salado–Vallenar and Romeral–La Silla del Gobernador segments. Based on Thiele & Hervé (1984), García (1991), Brown et al. (1993), Scheuber & González (1999), Emparan & Pineda (1999, 2005), Arancibia (2004) and others (see text).
Movement along the Atacama Fault Zone has resulted in both ductile and brittle deformation, with detailed studies in the Salar del Carmen and the El Salado–Vallenar segments reporting the existence of a mappable belt of foliated rocks (Scheuber 1987, 1994; Brown et al. 1993; Scheuber et al. 1994; Marinovic et al. 1995; Scheuber & González 1999; Arévalo 1995, 2005a; Cembrano et al. 2005). In the Salar del Carmen segment, a four-stage tectonic evolution of the late Early Jurassic to Early Cretaceous magmatic arc has been deduced (Scheuber 1987, 1994; Scheuber & Andriessen 1990; Scheuber et al. 1994; González 1996; Scheuber & González 1999). Arc activity appears to have begun in late Sinemurian times, which coincides well with the first appearance of lavas overlying the marine Early Jurassic deposits. The oldest structures exposed correspond to middle to deep crustal mylonitic rocks and brittle faults formed in Middle to early Late Jurassic times, and consistently indicate a sinistral sense of shearing. There is abundant evidence indicating that shearing and magmatic activity were processes acting at the same time. An extensional stage followed in Late Callovian times, which permitted emplacement of late Jurassic plutons (160–150 Ma) in turn affected by ductile normal faults with cooling ages of 152 ± 4 Ma. Younger stages correspond to an oblique tensional stress regime in Early Cretaceous time mainly represented in brittle crust by parallel-orientated dykes and a following sinistral transpressive regime represented by steeply dipping brittle faults. Fission track ages indicate that exhumation of deep-seated sheared arc units and possible uplift of the Coastal Cordillera occurred in this region in Aptian–Albian times (Maksaev 1990; Scheuber & Andriessen 1990).

South of 25°S, three arc-parallel and interconnected fault systems are developed. In the arcuate El Salado–Vallenar segment, ductile extensional activity in the AFZ has been detected for Late Jurassic times (159 and 156 Ma; 40Ar/39Ar on muscovite of mylonites), and in eastern branches of the AFZ extensional activity and ductile deformation occur associated with emplacement of Early Cretaceous plutonic complexes (Brown et al. 1993; Grocott & Taylor 2002). Steep-dipping, brittle, sinistral strike-slip faults affect the ductile shear zones. This situation differs from the one described for the Salar del Carmen segment in the overstepping nature of the several branches of the AFZ exposed in this region. However, in other respects the two fault segments are similar, with a Late Jurassic extensional episode followed by sinistral strike-slip displacement, and exhumation of the Early Cretaceous units occurring in late Early Cretaceous times. Sinistral strike-slip movement along the AFZ has been related to oblique convergence of the Aluk (Phoenix) plate relative to the South American continent (Boric et al. 1990; Scheuber & Andriessen 1990; Scheuber & González 1999; Reutter 2001) (Fig. 3.31).

**Backarc extension.** In the backarc it is possible to deduce the existence of two major extensional episodes that coincide with the beginning of each of the two substages, in late Early Jurassic and in Kimmeridgian times. The Kimmeridgian extensional event resulted in the deposition of thick, mainly coarse continental and volcanic deposits, and apparently was more intense than the Early Jurassic extension, allowing the development of abundant volcanic activity in the backarc basin (Algarrobal–Río Damas Extensional Basin). This second rift phase probably reactivated faults developed during the first extensional episode or even during the pre-Andean tectonic cycle.

The thick Kimmeridgian conglomeratic deposits that follow the marine regression at the end of the first substage in the backarc (‘Malm Continental Series’, the Algarrobal and Mostazal Formations, Tordillo, upper part of the Lagunillas and the Río Damas Formation) are exposed between 27°S and 36°30′S and have been interpreted as being the result of a compressive tectonic phase (Araucanian or Kimmeridgian orogenic phase; see Charrier & Vicente 1972; Aubouin et al. 1973b). In particular, Rivano & Mpodozis (1976) and Mpodozis & Cornejo (1988) deduced for the Ovalle region (30°30′S) the

![Fig. 3.31. Plate geometries for (a) Early Late Cretaceous; (b) Late Cretaceous to late Miocene, and (c) Late Miocene to Present, taken from Scheuber et al. (1994) and Reutter (2001). Different obliquity between (a) and (b) explains different movements along the Atacama Fault Zone in Jurassic to Early Cretaceous times (sinistral) and in Late Cretaceous times (dextral). (d) Cenozoic convergence rate curves between the Farallon and Nazca plates and South America according to Pardo-Casas & Molnar (1987) and Somoza (1998).](image-url)
existence of a younger compressive event that they named the Infra-Neocomian Tectonic Phase. They based their proposition on the existence of an unconformable relation between Neocomian calcareous deposits over marine Bajocian deposits, which are thrust over conglomerates of Kimmeridgian age. In contrast, we argue that the Kimeridgian deposits correspond to an extensional rather than to a compressional event, and that the stratigraphy in the Ovalle region reflects a palaeo- geography developed during basin extension and subsequent basin inversion. Therefore, in our view, use of the terms ‘Araucanian’ and ‘Infra-Neocomian’ (compressive) tectonic phases should be discontinued. We relate the Late Jurassic regressive episode and the associated continental sedimentation separating the two transgression–regression cycles (corresponding to the substages of the first Andean stage) to a generalized crustal warping episode probably linked to the initial development of the Atlantic Ocean.

**Basin inversion.** At the end of the first Andean stage, Early Jurassic to late Early Cretaceous deposits were deformed (thrust and folded), uplifted and eroded. This deformation episode corresponds to the so-called Subhercynian or Peruvian Phase (Steinmann 1929; see also Groeber 1951; Charrier & Vicente 1972; Vicente et al. 1973; Reutter 2001). This episode can be correlated with plate reorganization that caused the northeastward displacement of the Farallon Plate and dextral oblique convergence between the oceanic and the South American plates. Since this time, horizontal movements along essentially north–south orientated faults have shown dextral displacements. The backarc sediment fill was inverted by a generally east-vergent thrust and fold system. However, because of the presence of older faults, which are related to the Atacama Fault Zone (see above), westward inversion occurred along major faults like the Cerrillos Thrust in the Copiapó region (Arévalo 1995, 2005), the Romeral Fault Zone in the La Serena region (Emparan & Pineda 2000, 2005) and the Silla del Gobernador Fault Zone in the coastal region between Los Vilos and Los Molles (32°S) (Arancibia 2004). These faults include cataclastic as well as mylonitic zones. Radioisotopic determination on the fault rock of the Romeral Fault Zone (29°30′ to 30°30′S) yielded several ages (K–Ar on minerals and whole rock and one 40Ar/39Ar inverse isochron), between 115 ± 4 Ma and 100 ± 2 Ma (Emparan & Pineda 2000, 2005). The Silla del Gobernador Fault Zone corresponds to a steeply dipping, dextral shear zone interpreted as an inverted normal fault with a slight dextral component originally associated with the extension of the backarc basin (Arancibia 2004). Geochemical studies performed by the same author on this structure indicate a maximum 40Ar/39Ar age of 109 ± 11 Ma and a minimum age of 97.8 ± 1.5 Ma for the reverse dextral shearing, which coincide with the ages obtained in the Romeral Fault Zone. These late Early Cretaceous dates also coincide with the Aptian–Albian exhumation ages obtained by Maksaev (1990) and Scheuber & Andriessen (1990) in the northern segments of the Atacama Fault Zone and therefore have been considered to correspond to the age of the basin inversion episode at the end of the first stage of Andean evolution.

Other major faults in the coastal region, such as the Puerto Aldea (30–31°S) (Emparan & Pineda 2005) and the Quebrada Teniente Fault (31–31°30′S) (Irwin et al. 1988; García 1991) and the ductile El Cajón Fault at 33°S (S. Rivano, pers. comm.), are possibly part of the Romeral–Silla del Gobernador Segment of the Atacama Fault Zone (Fig. 3.30).

As a result of this deformation episode, a major and regional unconformity separates these deposits from overlying Late Cretaceous deposits. This unconformity separates the Early Period and Late Period of Andean evolution of Coira et al. (1982). The strong compressive regime that caused this deformation was probably triggered by a stronger coupling between the oceanic and the South American plates as a consequence of the high production of Pacific crust at this time (Larson 1991).

This plate reorganization was to produce a fundamental change in subsequent Andean palaeo-geography.

**The first Andean stage: summary and discussion.** The onset of the Andean tectonic cycle is recorded by the initiation of subduction-related magmatism, although renewal of subduction presumably took place some time before the ascent of magmas and volcanic activity in the arc, which was located along the present-day Coastal Cordillera. Arc activity began in the Early Jurassic (Sinemurian–Pliensbachian) with the intrusion and extrusion of magmas into and generally unconformably upon older units. However, in the regions where the extensional basins of the younger stage of the pre-Andean cycle were developed there is a general stratigraphic continuity between the pre-Andean and the first Andean volcanic and sedimentary deposits. In these regions, the Andean volcanic deposits conformably overlie earliest Jurassic marine deposits of the Pan de Azúcar, lower portions of the Lautaro and Tres Cruces, Los Molles and Quebrada del Pobre formations, which had been deposited in the late extensional phase of the previous pre-Andean tectonic cycle. Similarly, the deposits resulting from marine transgression in the backarc are continuous with those accumulated in the late pre-Andean stage.

The rapidly subsiding arc and its backarc basin further east, both with enormously thick successions, were the dominant palaeogeographic features during this stage, and were orientated essentially north–south (although perhaps slightly also to the west of north). The arc remained close to sea level, probably due to continuous magmatic activity, whereas the backarc remained well below sea level for most of the time. The two substages of the first Andean stage can be recognized both in the arc activity and backarc deposition.

Arc activity during the second substage was shifted slightly to the east of the locus of the arc during the preceding substage, although it remained concentrated in the Coastal Cordillera. In the backarc, the two substages are characterized by thick, essentially marine deposits corresponding to transgression–regression cycles separated from each other by an episode of abundant continental sedimentation and in some regions intense volcanic activity.

South of 30°S, in the region between La Serena and Santiago, probably either because of a general slight departure from a strictly north–south orientation of the arc–backarc system, or because of a slight bend of the palaeogeographic elements towards the SSE in this region, it is possible to detect the existence of a forearc basin, the Lo Prado Forearc Basin, during the second substage. This was located west of the magmatic arc, and to continuous magmatic activity, whereas the backarc remained relatively so that of the Mendoza–Neuquén backarc basin in this region. It is therefore probable that the arc in the La Serena region and further north is the northward prolongation of the Lo Prado–Pelambres Volcanic Arc, and its NNW–SSE orientation coincides with that of the Mendoza–Neuquén backarc basin in this region. It is therefore probable that the arc in the La Serena region and further north is the northward prolongation of the Lo Prado–Pelambres Volcanic Arc, and that the forearc basin north of La Serena disappears into the sea in the same direction (Fig. 3.29). There is no doubt that the Mendoza–Neuquén basin continues further north into the backarc in northern Chile. Based on these observations, one can propose that Andean palaeo-geography, at least for the second substage, comprised three main areas which, from west to east, were the forearc basin, arc and backarc basin. The similar geochemical characteristics displayed by the Quebrada Marquesa (La Serena region) and the Lo Prado Formations (Santiago region) (the latter deposited in the Lo Prado Forearc Basin) indicate a similarly extensional tectonic setting for their magmatism (Morata & Aguirre 2003).

The absence of evidence for a forearc basin north of La Serena and the evident reduction of the width of the presently remaining arc towards the north, and its almost complete absence in the Arica region, all support the interpretation that the structural trend in Jurassic and Early Cretaceous times was slightly orientated to the NW. We suggest that this orientation
was inherited from the structures that controlled the paleogeography (extensional basins) of the pre-Andean tectonic cycle, and probably from still older accretionary events along the continental margin of Gondwana (see discussion for the pre-Andean cycle). The reduction of width of the arc exposures towards the north has been interpreted as resulting from strong continental tectonic (subduction) erosion that occurred in Late Mesozoic and Cenozoic times (Rutland 1971; Coira et al. 1982; Parada et al. 1988; Stern 1991b).

As already indicated, stratigraphic, structural and geochemical evidence supports the existence of extensional conditions during this stage. The strong extensional conditions in Early Cretaceous times have led to the proposition of the existence at that time of an ‘aborted marginal basin’ along the coastal region of north-central Chile (c. 27ºS to c. 33ºS) (Levi & Aguirre 1981; Åberg et al. 1984; Mpodozis & Ramos 1989). Based on paleogeographic considerations, Charrier (1984) and Charrier & Muñoz (1994) suggested the existence of an Early Cretaceous intra-arc basin in the present-day Coastal Cordillera, far west of the Mendoza–Neuquén Backarc Basin, that coincides with the above-mentioned aborted marginal basin. However, according to the views presented here, we consider that the intra-arc depocentre corresponds instead to the Lo Prado Forearc Basin, and that this basin in fact represents the aborted marginal basin deduced by Åberg et al. (1984).

A major magmatic pulse in Late Jurassic times has been detected by Oliveros (2005), in northern Chile, and by Gana & Tosdal (1996) in central Chile. The first author argued for magmatism taking place across Late Jurassic–Early Cretaceous times, whereas the latter suggested that all plutons were intruded in just 6 million years, between 162 Ma and 156 Ma (Late Jurassic). This magmatic pulse coincides with deposition of the thick Horqueta Formation (Piracés 1977) in this same region, as well as with the extensional stage described for the Atacama Fault Zone in northern Chile (Scheuber & González 1999; Brown et al. 1993) that facilitated the ascent of large amounts of magma. This suggests that the magmatic pulse in central Chile probably also occurred under extensional conditions, and that the extensional stage in the evolution of the AFZ in northern Chile corresponds to a generalized event along the entire magmatic arc. Thus, in the light of our proposition that the Late Jurassic (Kimmeridgian) continental detrital and volcanic units known from the backarc domain were deposited in an extensional basin, it is possible to deduce the existence of a generalized extensional phase not only along the magmatic arc, but also along the whole continental margin. Within this context, the dominantly sinistral transtensional regime acting along the AFZ makes it probable that the Algarrobal–Río Dumas Extensional Basin (rift phase) corresponds to one or more pull-apart basins.

In the Arica region, the western position of the Ataúña (continental) and the Blanco (marine) formations, with their volcanic intercalations, are equivalent to the Caleta Coloso and El Way formations in the Antofagasta region. The apparent lack of connection between these deposits and those forming the backarc successions, located far to the east, suggests that these formations were deposited on the western side of the arc, as has been suggested for the Arqueros and Marquesa formations in the La Serena region, and the Lo Prado Formation in the region of Santiago. They therefore might be preserved remnants of a forearc basin, of which little evidence remains along the northern Chilean Andes. Marinovic et al. (1995), considering the peculiar position of these deposits in the Antofagasta region, suggested that they correspond to pull-apart basins associated with the Atacama Fault Zone.

Finally, in extreme contrast to the complex, tectonically active volcanosedimentary setting in the west, deposition on the eastern side of the backarc basin during this initial stage of Andean evolution was for much of the time undisturbed by subduction activity. Between approximately Chañaral and La Serena this eastern succession is characterized by essentially continuous late Early Jurassic to Late Neocomian calcareous deposition only briefly interrupted by evaporitic, coarse terrigenous deposits and andesitic flows in Late Jurassic (Kimmeridgian) time. This quiet, apparently continuously subsident slope has been named the Aconcagua Platform (Mpodozis & Ramos 1989).

Second stage: late Early Cretaceous–Early Palaeogene

In early Late Cretaceous times a major change in plate interactions occurred along the continental margin of southern South America. This episode can be related to the late Early to Late Cretaceous phase of very rapid ocean crust production in the primitive Pacific Ocean (Larson 1991), and probably was linked to a reduction of the subduction angle below South America (Chilean-type subduction). As a consequence of these modifications, the second regression episode of the previous stage culminated in the emergence of the continental margin during an episode of intense contractional deformation, with uplift and erosion of the pre-existing units, particularly of the Early Jurassic to Early Cretaceous backarc basin fill (backarc basin inversion). This tectonic phase (the so-called Subherrycnian or Peruvian; see Charrier & Vicente 1972; Aubouin et al. 1973b; Vicente et al. 1973; Reutter 2001) marks the separation between the early period and late period into which Coira et al. (1982) subdivided the evolution of the Andean tectonic cycle. After this episode the paleogeographic organization in this region of the Andes changes completely: the magmatic arc shifted considerably eastwards, a continental foreland basin was formed to the east of the arc instead of a backarc basin, and a rather wide forearc region west of the arc was produced as a result of eastward arc migration. Oblique subduction also prevailed at this time, although the movement of the Farallon oceanic plate towards the continent was now north–southward, producing dextral displacement along north–south orientated transcurrent faults (Fig. 3.31).

During this stage of Andean evolution two particular events occurred that introduced notable modifications to the new paleogeographic organization: (1) the development of the Salta Rift system located in the Andean foreland in the Salta region, Argentina; and (2) a high sea level stand in latest Cretaceous–earliest Cenozoic times. Although rift development in northwestern Argentina began in Early Cretaceous times (Gallinski & Viramonte 1988; Marquillas & Salfity 1988; Salfity & Marquillas 1994; Cristallini et al. 1997; Viramonte et al. 1999; Marquillas et al. 2005), its effects did not reach the newly developed foreland basin in northern Chile at the latitude of Antofagasta until Late Cretaceous times (Uliana & Biddle 1988; Salfity & Marquillas 1994). The high sea level affected mainly the western border of the Coastal Cordillera, and had little effect in eastern Chile, except for the Salar de Atacama Basin where a marine incursion of probable Atlantic origin has been detected. However, on the eastern flank of the Principal Cordillera next to the water divide between 33º30'S and 35ºS, in westernmost Argentina, the Late Cretaceous marine Saldeño and Malargüe formations are well developed (Bertels 1969, 1970; Tunik 2003).

This stage of Andean evolution is characterized by extensional episodes associated with intense magmatic activity. This activity is represented by major plutons and abundant volcanic deposits of andesitic and rhyolitic-dacitic nature, frequently associated with development of great calderas. The Late Cretaceous deposits in this stage accumulated in a series of fault-controlled extensional basins located along the magmatic arc: Cerro Empexa, Quebrada Mala and Llanta (Cornejo et al. 2003). Additionally, we propose to include in this list the basins of the town stage that hosted the Horruitos, Villa and Salamancas formations. Latest Cretaceous and/or earliest Palaeocene inversion of these basins was followed by the development of early Palaeogene depocentres further to the east where more thick volcanic and volcanioclastic successions were deposited. The end
of the stage is marked by a major deformation event in Eocene times.

The succession deposited during this stage generally contains two unconformities, both linked to tectonic events: one is located approximately at the Early–Late Cretaceous boundary (90–80 Ma), and the other separates Late Cretaceous from Early Palaeogene deposits, located approximately at the Cretaceous–Tertiary boundary. The lower of these two stratigraphic breaks, which has been well identified in the Copiapó and La Serena regions, apparently represents a discontinuity associated with a reactivation of the extensional conditions prevailing at the beginning of this second stage of Andean evolution. The second unconformity, which has a wider regional extent, represents an intense, short-lived compressional deformation event and is used here to differentiate two substages: (1) late Early to Late Cretaceous (with a first period of very intense extension in late Early to Early Late Cretaceous times); and (2) Early Palaeogene. However, to simplify the description of the tectonostratigraphic evolution we make no special separations for each substage within the text. We separate the following descriptions into the same three geographic areas used before, namely: northernmost Chile (between Arica and Chañaral); southern north Chile (region between Chañaral and La Serena); and central Chile (between La Serena and Valdivia).

**Northern Chile: between Arica and Chañaral (18°S to 26°S)**

*Arc evolution.* Late Early Cretaceous to Early Palaeogene arc deposits in northern Chile have not been reported for the Arica region in the far north. A little further south however, between Iquique and Chañaral, they form a belt of discontinuous outcrops along the Precordillera, the western flank of the Sierra de Moreno, the Domeyko Range, and, further south still, in the Central Depression (Fig. 3.32). These deposits accumulated in a series of extensional basins bounded by normal faults and consist of: (1) essentially volcanic and volcaniclastic successions of variable thickness, and (2) thick, mainly conglomeratic continental/marine deposits. Late Cretaceousinic deposits correspond to the following formations, from north to south: Panjuacha (Harambour 1990), Cerro Empexa (Galli 1968; Galli & Dingman 1962), Quebrada Blanca de Poquis, Quebrada Mala (Montaño 1976; Muñoz 1989; Marinovic & García 1999), Augusta Victoria (García 1967). In addition, the volcanic Lomas Negras Formation, with marine intercalations containing Miloliot foraminifera of undetermined age (Marinovic & Lahsen 1984)

![Fig. 3.32](image-url)
and dated at 66 Ma (Hammerschmidt et al. 1992), represents the Late Cretaceous arc volcanism in this region. Other volcanoes and volcaniclastic units from this stage, but of early Palaeogene age, are the Icanche (Maksaev 1978), Calama (lower member) (Blanco et al. 2003) and Cinchado (Montaño et al. 1976) formations (Fig. 3.32). The red continental detrital deposits correspond to the Tolar (Maksaev 1978; Ramírez & Huete 1981) and Tambillo formations (Skarmeta & Marinovic 1981; Ladino 1998). All these deposits unconformably overlie units of Jurassic–Early Cretaceous age or Peruvian Triassic rocks belonging to the Gondwanan or the pre-Andean cycle and are unconformably overlain by younger formations, frequently by the Late Oligocene–Miocene Altos de Pica Formation (Galli 1957).

The Panjuacha Formation, defined in the Tarapacá river valley, at c. 20°S, comprises >300 m of andesitic and dacitic lavas and welded lapilli tuffs with subordinate conglomerates and sandstones (Harambour 1990). This formation is an equivalent of the Cerro Empexa Formation defined immediately to the south by Galli (1968) and exposed further east and south in the Precordillera and western Atiplano (Vergara & Thomas 1984; Ladino 1998; Tomlinson et al. 2001). The latter authors differentiated two members in the Cerro Empexa Formation: (1) a lower member consisting of andesitic lavas and breccias, lahars, a few ignimbritic intercalations, and minor conglomerates and sandstones that present frequent and rapid lateral variations of thickness; and (2) an upper member consisting of andesitic lavas and breccias with subordinate dacitic tuffs. U–Pb and K–Ar radioisotopic dating in the Cerro Empexa Formation along the Precordillera and Sierra de Moreno yielded latest Cretaceous ages between 69 and 65 Ma (Tomlinson et al. 2001).

The Tolar Formation (Maksaev 1978), located on the west side of the Domeyko Range (Fig. 3.33), consists of a generally well stratified, upward fining, 1000-m-thick red succession of breccias, conglomerates and sandstones. Dark rhyolitic clasts contained in the lower part of the formation probably derive from the Peña Morada Formation. The Tolar Formation unconformably overlies the Early Cretaceous Arca Formation and probably the Peña Morada Formation, and is conformably overlain by the andesitic Icanche Formation of Eocene age (Maksaev 1978; Ramírez & Huete 1981). A western equivalent of the Tolar Formation is the Tambillo Formation (Skarmeta & Marinovic 1981). The late Early to Late Cretaceous age of the Tolar and Tambillo formations, apart from their similar stratigraphic positions overlapping strongly deformed Early Cretaceous and older formations, and unconformably overlain by Late Palaeogene and Neogene deposits, is constrained by the presence of an andesitic boulder in the Tolar deposits dated (K–Ar) at 109±4 Ma and by the existence of sills in the Tambillo deposits that have been dated (K–Ar) at 61±1 Ma and 73±1.5 Ma (see Ladino 1998). The Tolar deposits are overlain conformably by the Eocene andesitic Icanche Formation (Maksaev 1978; Ramírez & Huete 1981). A similar stratigraphic relationship exists between the Tambillo deposits and the Late Cretaceous Empexa Formation (Ladino 1998), the latter being unconformably overlain by the Altos de Pica Formation (Oligo-Miocene). This indicates that (1) these formations belong to the same depositional event and to the same tectonic stage of Andean evolution of early Late Cretaceous to Early Palaeogene age, and (2) the volcanic and the red detrital deposits, although frequently exposed separately, represent different depositional environments in the same arc domain.

In the Loa river valley in the Calama region, NE of Antofagasta, deposits of Eocene age are exposed, and have been included in the lower Calama Formation (Naranjo & Paskoff 1981). This formation has been deposited in a tectonically controlled depocentre developed along the magmatic arc in this stage. This depocentre, which is generally referred to as the Calama Basin, either corresponds to a separate basin or forms part of one of the following basins: the Empexa, located to the north, or the Quebrada Mala, located to the south of Calama. Here, the tectonic control is exerted by a major trench-parallel fault, the Argomeda–West Fissure (Falla Oeste) Fault developed along the Domeyko Range and its northern prolongation, the Sierra del Medio. The Calama Formation deposits

Fig. 3.33. Distribution of the foreland deposits of the second stage of Andean evolution in the Antofagasta–Salar de Atacama region with location of the Toconao-X1 hydrocarbon exploration well. Outcrops NW of Calama correspond to the Tolar Formation, an age equivalent of the deposits east of the Domeyko Range. The easternmost deposits in the Puna plateau close to the border correspond to small outcrops of the Late Cretaceous calcareous deposits assigned to the Quebrada Blanca de Poquis Formation (north) and to the Chojfias and Siglia formations.
are located to the east of the fault and consist of a 760-m-thick succession of alluvial conglomerates and sandstones (Blanco et al. 2003; see also May et al. 2005). The lower member of this formation contains two lower intercalations of andesitic lavas, one of which yielded an age of 51.9 ± 1.7 Ma (K–Ar, whole rock) and, according to Blanco et al. (2003) its deposition probably lasted until 47 Ma (Blanco et al. 2003). These ages constrain the age of the lower member to middle Eocene. This age coincides with the age of the Icanché Formation (53–43 Ma; Maksaev 1978) and suggests that this volcanism was related to the extensional tectonic activity that controlled development of the basin. A major change in clast composition and orientation of sediment supply has been detected for the middle member of the Calama Formation. These changes have been interpreted as resulting from a tectonic event at that time which caused reactivation in the main fault systems and major palaeogeographic reorganization.

Further south in the western flank of the Domeyko Range east of Antofagasta, the Quebrada Mala Formation unconformably overlies older deposits (Early Cretaceous) (Montaño 1976; Muñoz 1989; Muñoz et al. 1989; Marinovic & García 1999) and shows a close similarity to coeval units further north. This c. 3700 m thick succession presents in a single section all the features described above for the deposits further north of similar age. Thick intercalations of andesites and silicic volcanic deposits indicate the bimodal character of the volcanism, and K–Ar age determinations on minerals yielded Late Cretaceous ages between 76 ± 2 Ma and 66 ± 3 Ma (Marinovic & García 1999). In latest Cretaceous or Early Cenozoic times the Quebrada Mala deposits were folded and thrust over rocks of the Caracoles Formation along the imbricated El Buitre fault system that apparently had acted as a normal fault system during deposition of the Quebrada Mala Formation. This deformation event relates to the compressional event at approximately the Cretaceous–Tertiary boundary that separates the two substages of the second stage of Andean evolution. The next youngest deposits in this region have been assigned to the Cinchado Formation (Montaño 1976), and comprise a 500-m-thick volcanic and sedimentary succession with K–Ar ages lying between 63 and 55 Ma (Marinovic & García 1999).

Southeast of Antofagasta, the Augusta Victoria Formation (García 1967; revised by Boric et al. 1990) (including parts of the Chile–Alemania Formation as defined by Chong 1973) is mainly exposed in the Central Depression between Antofagasta and Chañaral (Marinovic et al. 1995). It represents an igneous complex consisting of volcanic deposits, domes and small stocks. Lithologies include basaltic to rhyolitic lavas, tuffs, tuffaceous breccias, volcanic agglomerates, and volcanioclastic deposits, and subvolcanic intrusive bodies have gabbroic to dactitic compositions. Still recognizable volcanic structures, such as the nested Cachinal calderas of Palaeocene age with frequent evaporitic intercalations. These marine levels are correlated with the Late Cretaceous, foraminifera-bearing Lomas Negras Formation (Marinovic & Lahren 1984), and with calcareous outcrops of Late Cretaceous age exposed further east in the Puna (Altiplano) next to the Chile–Argentina border (Ramirez & Gardeweg 1982; Donato & Vergani 1987) (Figs 3.32 & 3.33) and correlated with the Yacoraite Formation, which is well developed in the Eastern Cordillera in the Salta Rift Basin (Saltiff et al. 1985). These deposits have been interpreted as accumulated in a tectonically inverted extensional basin associated with the Salta Rift system.

The Salar de Atacama contains up to 9000 m of Cretaceous to Recent sediments and occasional volcanic intercalations. This basin is presently located in the Andean forearc; however, during the Late Cretaceous and Palaeocene it was in a retroarc position. Macellari et al. (1991) and Flint et al. (1993) suggested an extensional tectonic history for this basin, whereas Charrier & Reutter (1994), Muñoz et al. (1997, 2000, 2002), Arraigada (1999) andMpodozis et al. (1999, 2005) have argued for extensional evolution followed by tectonic inversion. Cretaceous basin development in this region is probably related to extensional effects of the Salta Rift, which in turn is probably linked to continental-scale processes leading to the opening of the Atlantic Ocean (Uliana & Biddle 1988; Saltiff & Marquillas 1994). Normal faults, originally formed during the extensional
stage of the Salar de Atacama Basin, were invented to produce several kilometres of vertical uplift during the Late Cretaceous and subsequent Cenozoic compression. As a consequence of the uplift of the basement blocks forming the Domeyko Range, and the thin-skinned eastward thrusting of the sedimentary infill, sedimentation and space generation in the Salar de Atacama Basin was continuously modified. Another fault that uplifts an eastern basement block (Cordon de Lila) is the east-vergent Tucúcaro Thrust located on the eastern side of the Salar de Atacama (Niemeyer 1984). The thick and coarse Late Cretaceous to Early Cenozoic terrigenous deposits accumulated next to the uplifting Domeyko Range correspond to the Purilactis Formation. The similar and coeval deposits on the western side of the Domeyko Range, which have been included in the Tolar and Tambillo formations, probably correspond to equivalent syntectonic deposits on the opposite flank of this range.

The existence in this region of Late Cretaceous (Senonian) marine calcareous deposits in the Salar de Atacama Basin fill, and possibly also in the Lomas Negras Formation (Marinovic & Lahsen 1984), presents the challenge of determining the origin of the marine ingression. There is evidence for a possible northward connection of the Salta Rift Basin(s) with the Pacific Ocean through present-day southern Peru and Bolivia in Cenomanian–Santonian times and a southward connection with the opening Atlantic Ocean in Campanian–Maastrichtian times (Marquillas & Sallify 1988). Considering the existence of the Late Cretaceous arc located approximately along the present-day Central Depression and western flank of the Domeyko Range, which was undergoing uplift at that time, it is difficult to imagine a connection between the Salar de Atacama Basin with the ocean to the west. On the other hand, a connection with a marine ingressin of the sea from the Atlantic Ocean seems more probable, as has been proposed by Charrier & Reutter (1994) and Muñoz et al. (2002), based on the existence of Late Cretaceous marine limestones further east in the Altiplano (El Molino Formation) as well as in the Argentine Puna and Eastern Cordillera where they are well developed (Yacoraite Formation) in the Salta Rift System (Sallify et al. 1985). These deposits have been correlated with parts of the Purilactis Formation (Ramírez & Gardeweg 1982; Donato & Vergani 1987) (Fig. 3.33).

Southernmost Chile: region between Chañaral and La Serena (26°S to 30°S)

In this region, differentiation between late Early to early Late Cretaceous, Late Cretaceous and Early Palaeogene deposits is well defined. The late Early to Late Cretaceous deposits can be rather easily traced along the region. However, the younger magmatic events form a series of different volcanic units and associated intrusive bodies, mostly subvolcanic, corresponding to stratovolcanoes as well as to caldera events. It is therefore impossible to give a thorough account of all these units. In Figure 3.34 only some of the most representative of such units have been considered.

In the Chañaral region (26°–26°30′S), the 3500-m-thick volcanic and volcanioclastic Llantac Formation represents the Late Cretaceous deposits of this stage. This unit is exposed east of the Sierra Castillo Fault and is unconformably overlain by Cretaceous deposits of this stage. This unit is exposed east of this region (Ramírez & Gardeweg 1982; Donato & Vergani 1987) (Fig. 3.33).

Plutonic rocks crop out essentially in the Central Depression and its southward prolongation in the flat-slab segment, continuing the outcrops described in the region north of 26°S. In the Copiapó region (27°–27°30′S), conspicuous Late Cretaceous to Early Cenozoic plutons of this stage are exposed (see Fig. 3.32), e.g. the Cachiyuyo and the Cabeza de Vacas (Arévalo 1995, 2005a; Iriarte et al. 1996). In the northern part of the region, between Vallenar and La Serena (29°–30°S), Late Cretaceous deposits correspond to the Cerrillos Formation (Nasi et al. 1990) and in the southern part they correspond to the Quebrada La Totora Beds (Emparan & Pineda 1999). These deposits are unconformably and paraconformably overlain by the Late...
Cretaceous deposits of the Hornitos Formation, in the northern part, and the Viñita Formation, in the La Serena region respectively. As previously indicated, the Cerrillos and Quebrada La Totora formations were deposited during the major extensional episode developed at the beginning of this stage, whereas the Hornitos and Viñita formations were formed during a second extensional episode (extensional reactivation) in Late Cretaceous times. In this region the Cerrillos succession consists of 2000 m of continental andesitic to rhyolitic lavas, pyroclastic deposits, and coarse- to fine-grained sedimentary deposits. The Quebrada La Totora Beds comprise a 1000-m-thick succession of continental coarse and fine sediments and pyroclastic breccias containing dinosaur bones, the remains of crocodiles and turtles, and abundant silicified tree stems. The age assigned to this formation is late Albian to Cenomanian (Pineda & Emparan 2006).

The Viñita Formation (Aguirre & Egert 1965), which paraconformably overlies the Quebrada La Totora Beds and the Pucalume Formation, consists of andesites and basaltic andesites and abundant coarse and fine pyroclastic intercalations. The age of the Viñita Formation is constrained by a U–Pb determination that yielded 86.77 ± 0.84 Ma, and its stratigraphic position overlying the Quebrada La Totora Beds and underlying the Quebrada Yungay Beds with a U–Pb age determination of 68.5 ± 6.4 Ma (Pineda & Emparan 2006). This formation is correlated to the north with the following formations, from north to south: Quebrada Mala in the Antofagasta region, Llanta in the Chañaral region, and Hornitos in the Copiapó region (Fig. 3.34). Contemporaneously with or shortly after deposition of the Viñita Formation, were formed the Condoriaco and the Cerro El Indio calderas. Somewhat later, but still in Late Cretaceous times, the Los Elquinos Formation and the Quebrada Yungay Beds were deposited, and the Cerro Tololo, Llano Perrada and Tierras Blancas calderas were formed. Although the Los Elquinos Formation (Dedios 1967) and the Quebrada Yungay Beds (Emparan & Pineda 1999) probably form a continuous volcanic succession, the outcrops of these units are in fault contact. The Los Elquinos Formation consists of basaltic to rhyolitic lavas, tuffs and breccias, and the Quebrada Yungay Beds, of massive ignimbrites, tuffs and andesitic lavas. Several intrusive bodies of late Early to Late Cretaceous age occur in this region.

Early Palaeogene units are represented by Palaeocene volcanic complexes (La Corina and Cerro del Inca, c. 60 Ma) and
intrusive bodies of dioritic to granitic composition. The Llano Perrada and Tierras Blancas calderas, probably nested in the major and older Condoriaco Caldera, were also formed at this time.

As described previously, at the beginning of this stage, following inversion of the Late Jurassic–Early Cretaceous basin, approximately between Aptian and Cenomanian times (late Early to early Late Cretaceous), there was an extensional tectonic event that is particularly well recorded in the Copiapó region (Mpodozis & Allmendinger 1992, 1993; Arévalo 1995, 2005a; Iriarte et al. 1996). This event produced major décollement surfaces separating allochthonous units (the Bandurrias and Chañarcillo groups) from a para-autochthonous unit (the Punta del Cobre Formation). In the La Serena region, late Early Cretaceous extension, though clearly recognized, caused much less spectacular deformation and was mainly controlled by the El Chape Fault that bounded the extensional basin to the west and the Rivadavia Fault further east (Pineda & Emparan 2006).

Contractional deformation during this stage in Late Cretaceous–Early Paleocene (Incaic I phase) and at the end of the stage in Eocene times (main Incaic phase) was mainly controlled by inversion of major faults developed along the axis of the arc domain, some of which participated in the previous extensional movements in late Early Cretaceous times. In the La Serena region, the Vicuña and Rivadavia faults show evidence of Eocene deformation (Pineda & Emparan 2006). Fault rocks developed in several units along the Rivadavia Fault have yielded K–Ar ages of 40.5 ± 2.2 Ma and 40.3 ± 1.6 Ma (whole rock) and of 40.9 ± 1.9 Ma (amphibole) (Emparan & Pineda 1999), indicating that fault activity was related to the main Incaic phase.

**Central Chile: region between La Serena and Valdivia (30°S to 39°S)**

As indicated above, in La Serena region the Late Cretaceous to early Palaeogene deposits correspond to the Quebrada La Totora Beds and to the Viñita and Los Elquinos formations. South of 30°S, in the Illapel region, between 31°S and 32°S, the deposits corresponding to this stage are represented by the Salamanca and Estero Cenicero formations. According to Rivano & Sepúlveda (1991), the Salamanca Formation is an age-equivalent of the Viñita Formation, and consists of a c. 1300-m-thick lower alluvial sedimentary member (Santa Virginia), and an upper volcanic and volcanioclastic member (Río Manque). The 2000-m-thick succession of the Estero Cenicero Formation is comparable with that of the Los Elquinos Formation, and, according to Mpodozis & Cornejo (1988), these formations probably represent the southernmost outcrops of the Late Cretaceous volcanic arc that are active from Iquique southwards.

Finally, plutonic activity in this northern part of central Chile (30°–33°S) is represented by the Cogotí Superunit which has yielded K–Ar ages between 67 ± 2 Ma and 38 ± 1 Ma, indicating intrusive activity from latest Cretaceous to Eocene times (Parada et al. 1988; Rivano & Sepúlveda 1991).

Further south in central Chile, between 33°S and 39°S, the morphostructural units developed on the Chilean side of the cordillera are, from west to east, the Coastal Cordillera, the Central Depression and the Principal Cordillera. Late Cretaceous to Early Cenozoic deposits are exposed on the western and eastern flanks of the Coastal Cordillera, and in the Principal Cordillera. Each of these is now considered in turn (Fig. 3.32).

On the **western flank of the Coastal Cordillera** south of Santiago, deposits of both Late Cretaceous to Palaeocene age and of late Palaeocene to Eocene age have been reported. The Late Cretaceous to Palaeocene outcrops are fissiliferous marine deposits related to the eustatic high stand developed at this time, and are exposed in the following localities from north to south: Algarrobo (Levi & Aguirre 1960), Topocalma (Tavares 1979b; Charrier 1973b), Faro Carranza (Chanco Formation) and in the Arauco region (Quiriquina Formation) (based on Sernageomin 2003). Outcrops in Algarrobo and Topocalma are too small for representation in Figure 3.32. Equivalent deposits on the eastern side of the Principal Cordillera (Argentina) correspond to the Saldeño Formation (Tunik 2003) and Malargüe Group (Bertels 1969, 1970).
Fig. 3.36. Late Mesozoic and Cenozoic sedimentary deposits in the Arauco region on the western side of the Coastal Cordillera. Continental (right-hand side) and marine (left-hand side) deposits build up the Concepción Group. Major hiatuses separating the Cenozoic deposits are indicated.

to Eocene times. It consists of the following succession of formations, some of which interfinger with each other: Pilpilco (Early Eocene, littoral marine sequence, partly continental), Curanilahue (Early Eocene, a mainly continental sequence, coal-bearing strata), Boca Lebu (Early Eocene, marine transgressive sequence), Trihueco (Middle Eocene, a mainly continental sequence, coal-bearing strata), and Millongue (middle to late Eocene, marine sequence) (Tavera 1942; Muñoz Cristi 1946, 1973; Pineda 1983a, b; Arévalo 1984). It is unconformably overlain by the Miocene sedimentary Ranquil Formation, which is an equivalent of the Navidad Formation exposed along the coast at the latitude of La Serena and Coquimbo. No Oligocene deposits exist along the coastal region in Chile, probably because of the eustatic low stand at this time, and a major hiatus separates Eocene from Miocene deposits (Fig. 3.36).

Along the eastern flank of the Coastal Cordillera, Late Cretaceous deposits between 32°S and 35°15′S are represented by the Lo Valle Formation (Thomas 1958; Godoy 1982; Moscoso et al. 1982; Rivano 1996; Gana & Wall 1997; Bravo 2001). At its type locality this formation comprises a 700-m-thick succession of silicic, often welded, pyroclastic deposits, with intercalations of lavas and continental sediments. Further south it reaches a thickness of 3500 m (Nasi & Thiele 1982), and still further south its exposed thickness is 750 m (Bravo 2001) (see Fig. 3.26). According to the variable thickness and the different types of contacts with the underlying units observed by the different authors, it is probable that it corresponds to the deposits of the late phase of an extensional episode (see previous Andean stage). K–Ar age determinations from samples collected at 33°S yielded 70.5 ± 2.5 Ma and 64.6 ± 3.5 Ma (Vergara & Drake 1978; Drake et al. 1976), and 40Ar/39Ar determinations yielded ages between 72.4 ± 1.4 Ma and 71.4 ± 1.4 Ma (Gana & Wall 1997). This calcalkaline, essentially pyroclastic formation represents the deposits of the Late Cretaceous volcanic arc.

In the Principal Cordillera at 35°S, a small outcrop of upward fining and th inning red-coloured fluvial deposits that unconformably rest on the Early Cretaceous Baños del Flaco Formation and unconformably underlie the Early Oligocene mammal-bearing levels of the Oligo-Miocene Abanico Formation has been informally named Brownish-red Clastic Unit (BRCU) by Charrier et al. (1996). Apparently similar deposits crop out next to the international boundary at 36°S overlying Middle to Late Jurassic rocks of the Nacientes del Teno Formation and underlying the Late Cenozoic volcanic deposits assigned to the Campanario Formation (Drake 1976; Hildreth et al. 1998). These have been included in the Estero Cristales Beds by Muñoz & Niemeyer (1984). According to their stratigraphic position, these deposits can be assigned a Late Cretaceous age, and based on this and on their sedimentological features they have been correlated with the Late Cretaceous Neuquén Group in Argentina (Charrier et al. 1996). These units can also be correlated with the volcanic Lo Valle, on the eastern side of the Coastal Cordillera, and the Viñita formations, further north.

No marine deposits of this stage have been found on the Chilean side of the cordillera, although in westernmost Argentina, between 33°S and 35°S on the eastern side of the Principal Cordillera, marine deposits of the Saldeño Formation (Tunik 2003) and the Malargüe Group (Bertels 1969, 1970) testify to the far-reaching nature of the Late Cretaceous to early Tertiary marine transgression over the more stable part of the continent. The absence of these deposits in Chile suggests the existence of a relief that stopped the advance of the sea further west.

Late Cretaceous to Early Eocene plutonic activity between 33°S and 39°S is concentrated along the eastern flank of the Coastal Cordillera where it forms discontinuous outcrops intruding Mesozoic units (Bravo 2001).
times indicates that the general environment during this stage was strongly subsiding. It suggests the imposition of extensional conditions shortly after the late Early Cretaceous regression episode in the backarc basin and the generalized deformation event at the end of the previous Andean stage. Thus a major extensional tectonic episode has been deduced for Aptian–Conomanian times in the Copiapó region by Mpodozis & Allmendinger (1992, 1993). According to these and other authors (Arévalo 1995, 2005a; Iriarte et al. 1996), at that time the deposits of the Bandurrias and Chañarcillo groups were deformed into allochthonous extensional wedges and detached from each other and from the para-autochthonous Punta del Cobre and older formations by subhorizontal normal faults (Sierra de Fraga–Puquios Extensional Complex), with local stratigraphically and structurally chaotic areas such as the so-called ‘Positas in Cauc’. As a consequence of this extensional event were deposited the Cerrillos Formation (Arévalo 1995, 2005a; Iriarte et al. 1996) in the Copiapó region, and the Quebrada la Totora Beds in the La Serena region (Emparan & Pineda 1999; Pineda & Emparan 2006). Although this event has been correlated by some authors with the development further south in the Santiago region of an ‘Aborted Marginal Basin’, proposed by Abery et al. (1984), we consider that it corresponds to a different, younger extensional episode. Further extension, imposed on the preceding one, developed the accumulation space for thick Late Cretaceous deposits (e.g. Llanta, Hormitos and Viñita formations between Chañaral and La Serena regions). The evolution of the geochemical characteristics of the arc deposits in the La Serena region indicates considerable reduction of crustal thickness since Early Cretaceous times (Martínez 2002; Morata et al. 2003) and supports the concept of extreme extensional conditions having been present during Late Cretaceous times.

The dominance of extensional/transtensional tectonic conditions during this stage can be related to a period of very oblique and low convergence rate between the Farallon Plate and South America during Late Cretaceous and Early Cenozoic times (Pardo-Casas & Molnar 1987). Extensional conditions also occurred approximately at the same time (Late Cretaceous) in the foreland in northern Chile (Salar de Atacama region). However, these conditions have been related to the development of the Salta Rift in the Andean foreland in western Argentina. Considering that the origin of the Salta Rift has been related to a continental-scale extensional event linked to the opening of the Atlantic Ocean, we suggest that such an event might have also favoured extensional conditions in the arc domain.

The generalized extension characterizing this stage probably accounts for the separation of the Late Palaeozoic Batholith north and south of Valparaíso (33°S). North of this latitude, the batholith’s exposures are located along the High Andes (ESU in Fig. 3.5), whereas south of Valparaíso they are abruptly shifted to the west and located along the Coastal Cordillera (Coastal Batholith south of 33°S in Fig. 3.5). Based on evidence indicating that the batholith, though not exposed, exists further north of Valparaíso underneath the Choapa Metamorphic Complex, Rebolloredo & Charrier (1994) suggested the possibility that the two parts of the batholith (one to the north and the other to the south of Valparaíso) were separated by an event of extensional tectonics oriented slightly oblique (NNW–SSE) to the main batholith trend and probably controlled by pre-existing weakness zones. The major Aptian–Conomanian extensional event described by Mpodozis & Allmendinger (1992, 1993) for north-central Chile may have been the one responsible for this tectonic separation.

The forearc domain apparently formed a positive relief that represented an obstacle to eastward marine Pacific ingression during oceanic high stand in Late Cretaceous and earliest Cenozoic times. Late Cretaceous fission track ages from the Coastal Cordillera at c. 33°S obtained on Late Palaeozoic and Late Triassic metamorphic and plutonic rocks (106 ± 8.6 Ma, 106 ± 7.4 Ma and 98 ± 10 Ma) and from the late Early–Late Cretaceous Caleu pluton (91 ± 3.2 Ma (Gana & Zentilli 2000), and between 95 ± 5.4 Ma and 82.0 ± 5.6 Ma (Parada et al. 2005a)) suggest that the forearc was uplifted and formed a topographic barrier by Late Cretaceous times. Similarly, the arc itself also represented an obstacle for the eastward incursion of Atlantic waters.

The third Andean compressive event occurred at the end of this stage mainly in Eocene times when activity in the magmatic arc or intra-arc ceased. The precise timing of this event has been debated: 38.5 Ma according to Hammerschmidt et al. (1992), c. 44 Ma according to Tomlinson & Blanco (1997a), and over a longer time span according to Maksae & Zentilli (1999). The tectonics of this event shows strain-partitioning to shortening and longitudinal strike-slip movements causing a generalized tectonic inversion of the arc (or intra-arc) and of the extensional basins developed during the last substage in early Palaeogene time, and deformation of the infill deposits. Inversion of the arc caused uplift of NNE–SSW orientated blocks with cores of Palaeozoic rocks that were thrust west and eastwards over Mesozoic and Early Palaeogene deposits. This thrusting caused deformation in the foreland, e.g. eastward thrusting and further growth of accommodation space in the foreland basin (Muñoz et al. 2002) and triggering of the eastern thrust-fold belt in Mesozoic backarc deposits developed on the eastern flank of the Domeyko Range in the Chañaral and Copiapó region (Tomlinson et al. 1999) and resulting in considerable crustal thickening. Exhumation of the uplifted blocks occurred according to Maksae & Zentilli (1999) between c. 50 and 30 Ma preceding and overlapping porphyry copper emplacement along the Domeyko Range. Associated with this tectonic event, there were also clockwise rotations in the north Chilean forearc (Arriagada et al. 2003) and counterclockwise rotations in the south Peruvian forearc (Roperch et al. 2006), and dextral and sinistral strike-slip activity along the axis of the Late Cretaceous–Early Cenozoic magmatic arc, namely the Domeyko Fault System in the present-day Domeyko Range (Argomedo–West Fissure [Falla Oeste] Fault System and its southward prolongation as the Agua Amarga–Sierra Castillo Fault) (Reutter et al. 1991, 1996; Tomlinson et al. 1994, 1999; Niemeyer 1999; Conrojo et al. 2006).

This deformation episode, which separates the second from the third stages of Andean evolution and corresponds to the main Incaic phase or Incaic phase II (Steinmann 1929; Charrier & Vicente 1972; Charrier & Malumian 1975; Maksae 1978; Conrojo et al. 2003; Reutter 2001), coincides fairly well with a peak of high convergence rate associated with a considerable reduction of the obliquity of convergence at approximately 45 Ma (Pilger 1984; Pardo-Casas & Molnar 1987; Somozoa 1998; see also Reutter 2001, Fig. 5).

The 20-million-year long exhumation period detected in the Domeyko Range by Maksae & Zentilli (1999), during which 4 to 5 km of the crust were eroded away, suggests that phases like the main Incaic phase represent long-lasting events affecting broad regions of the lithosphere. Therefore the rather sharp chronology determined for these phases based on structural features that reflect them next to the Earth’s surface (unconformities, fault rocks, deformation, etc.), probably corresponds to shorter events of intense strain and erosion associated with the major event. Although this chronology is essential for establishing the tectonostratigraphic evolution of the orogen, thorough understanding of the tectonic event requires consideration of the much broader effect caused by these phases.

**Third stage: Late Palaeogene–Recent**

This stage corresponds to the last phase of Andean development, during which the Argentine–Chilean Andes adopted their present configuration. During this stage, among other important events, Andean uplift took place, the morphostructural units were developed, the volcanic arc reached its present position, and the post-Incaic porphyry copper deposits
were emplaced. Also during this stage, in the early Miocene, the c. NE-orientated segment of the Juan Fernández Ridge collided against the continental margin in southern Perú and northern Chile, and, because of almost east–west orientated convergence since 26 Ma, the locus of the collision between this ridge segment and the continental margin migrated southward at the speed of 200 km/Ma to its present position (Yáñez et al. 2002).

The formation of the Isthmus of Panama, which began in late Miocene and was completed in Pliocene times, favoured the Great American Biotic Interchange (between South and North America) and caused the end of the long-lasting faunistic isolation of South America after its separation from Africa. Thus the characteristic endemism of the South American faunas came to an end.

During this stage, tectonic evolution north of 47ºS was controlled by the relative movements between South America and the Nazca Plate. Deformation of the deposits of the previous stage occurred in middle Eocene times at the end of an episode of increasing convergence rate (attaining a rate of > 10 cm/year, between 49.5 Ma and 42 Ma) (Pilger 1994; Pardo-Casas & Molnar 1987), that coincides with the main Incaic phase or Incaic phase II (Charrier & Vicente 1972; Coira et al. 1982; Cornejo et al. 2003). According to Pilger (1994), Pardo-Casas and Molnar (1987) and Somoza (1998), after this episode of high rate convergence, relative displacement of the oceanic and continental plates was almost parallel to the continental margin, and the convergence rate diminished until 26 Ma. At this moment, another plate readjustment occurred and the oceanic plate adopted an almost orthogonal motion relative to the continental margin (c. N80ºE; Yáñez et al. 2002), and convergence rate was increased until 12 Ma (Pilger 1994; Pardo-Casas & Molnar 1987) (see Fig. 3.31c). These modifications of the plate kinematics and convergence rates caused important variations of the tectonic regime on the continental margin, expressed in the forearc (i.e. the western side of the Andes) by uplift and synchronous extensional, strike-slip, and compressional deformation (Hartley et al. 2000).

Palaeo-geography at the beginning of this stage was characterized by the Incaic Range, which was formed during inversion of the magmatic arc or buckarc domain existing during early Palaeogene times (main Incaic phase). Inversion and uplift of the Incaic Range in middle to late Eocene times was controlled by several NNE–SSW trending faults with both east and west vergence. Block movement along these faults controlled development of basins located on the west and east sides of the uplifting range that received abundant sediments from the eroding uplifted areas. The units formed during late Eocene to Present times are exposed in all morphostructural units. However, distribution of the deposits is not uniform and, furthermore, their continuity in the Central Depression is interrupted in the flat-slab segment as a consequence of major uplift caused by the present-day subduction of the Juan Fernández Ridge. In this intermediate Andean segment, which lies between c. 27ºS and c. 33ºS, no Central Depression is developed, volcanic activity is absent along the axis of the cordillera, and the Frontal Cordillera and the thick-skinned Pampean Ranges are developed further east in the Andean foreland in Argentine territory, instead of the thin-skinned thrust-fold belts developed north and south of the segment (Fig. 3.37).

In the following analysis, we describe the evolution along different sections across the cordillera from north to south. In each case we indicate the morphostructural units existing in the section and the location of the deposits. However, considering that there are two morphostructural units that are developed without major interruption along the continental margin, the Continental Platform and the Chile Trench, we first describe briefly their principal features.

**The Continental Platform and the Chile Trench (Fig. 3.38)**

Although the Continental Platform has been relatively poorly studied, the available information indicates that it corresponds to a succession of basins more than 60 km wide, filled with very thick sedimentary successions. These deposits mostly overlie metamorphic rocks of the Late Palaeozoic accretionary prism, and marine sediments accumulated along the coast in Late Cretaceous to early Tertiary times, e.g. the Quiriquina Formation. The general structural features associated with these basins indicate that they developed under extensional tectonic conditions (Muñoz & Fuenzalida 1997; Mordojovic 1976, 1981;
González 1989; Hinz et al. 1998). At the latitude of Valparaiso (33ºS), in the continental slope, a poorly developed accretionary prism has been interpreted from seismic profiles (Ranero et al. 1997). The Chile Trench corresponds to a narrow elongated basin developed at the foot of the continental slope that locally reaches depths of > 7000 m in northern Chile, where the basin is almost depleted of sediments (sediment-starved) and northward-channelized sedimentation forms isolated ponded basins. South of 27ºS the trench fill forms a narrow, though continuous band of sediments. South of 30ºS (Elqui river), the major rivers form submarine canyons in the continental platform (Thornburg & Kulm 1987a, b; Thornburg et al. 1990; Pineda & Fanucci 1994). South of 41ºS deposition is not channelized and the basin is filled with sheet turbidites, with a well developed accretionary prism continuously flanking the trench up to the south of Tierra del Fuego. Here, the forearc basin is almost completely filled with sediments and, along with the continental platform, forms a wide flat surface known as the Fueguian Terrace (Herron et al. 1977; Díaz 1997; Díaz et al. 1997).

Palaeogeographic development, sedimentation and uplift of the Altiplano in this region are closely related to two thrust systems: (1) a north–south to NNW–SSE orientated, high-angle, westward propagating, west-vergent thrust-fault system (WTS) developed along the Precordillera (Mortimer & Saric 1975; Naranjo & Paskoff 1985; Muñoz & Sepúlveda 1992; Muñoz & Charrier 1996; Charrier & Muñoz 1997; García 1996, 2002; García et al. 1996, 2004; Parraguez 1997; Riquelme 1998; Victor & Oncken 1999; Pinto 1999; Victor 2000; Farias 2003; Pinto et al. 2004a, b; Victor et al. 2004; Farias et al. 2005a); and (2) an east-vergent thrust system developed to the east of the WTS in the Western Cordillera (Riquelme 1998; Hérail & Riquelme 1997; Riquelme & Hérail 1997; David et al. 2002; Charrier et al. 2005b) (Fig. 3.39). The features described for the

**Fig. 3.38.** Morphology of the Continental Platform and Chile trench between 24ºS and 42ºS (Antofagasta to Chiloé). Note northward diminution of the trench filling and close association between the main rivers and the development of marine canyons.
Forearc Precordillera can be followed to the Iquique region and even further south, although in the Western Cordillera at the latitude of Iquique (Altos de Pica) a pull-apart system has been proposed by Victor (2000), Victor & Oncken (1999) and Victor et al. (2004) in addition to the east-vergent contractional features.

Arica cross-section (18–19°S) (Fig. 3.40)

Central Depression. Deposits in the Central Depression correspond to an almost flat-lying, sedimentary and volcaniclastic continental succession, up to 1000–1500 m thick, of early to middle Oligocene to Holocene age. This succession consists of conglomerates and sandstones of alluvial and fluvial facies in the lower portions that grade upwards to alluvial conglomerates (Salas et al. 1966; Vogel & Vila 1980; Naranjo & Paskoff 1985; Schröder & Wörner 1986; Parraguez 1997; García 2001, 2002; García et al. 1996, 2004; Victor 2000; Farias et al. 2005a).
et al. 1966; Parraguez 1997; García 2001, 2002) (Fig. 3.40). This feature indicates that already in Oligocene times the Coastal Cordillera was forming a relief to the west of a wide accumulation space located in the present-day Central Depression (Parraguez 1997; García 2001, 2002). These deposits correspond in the Arica and Iquique cross-sections to syntectonic accumulations associated with the activity of a west- and an east-vergent thrust system (WTS and ETS respectively).

Forearc Precordillera. In this region, the WTS consists of two sets of thrust faults: one, formed by the Ausipar Thrust, defines the boundary between the Central Depression and the Forearc Precordillera, and the other, the Belén thrust system, in the eastern Precordillera (Figs 3.39 & 3.40), is formed, from east to west, by the Chapiquiña–Belén, the Cerro Lagunas–Belén–Tignamar and the Copaquiña–Tignamár thrusts. The Belén system essentially thrusts the Early Palaeozoic Belén Metamorphic Complex over Cenozoic deposits to the west, and its activity generated syntectonic deposits of different ages: the Joracane Formation, between 18.2 ± 0.8 Ma and c. 16 Ma, and the Huaylas Formation, between 11.7 ± 0.7 Ma and 10.7 ± 0.3 Ma (García 1996, 2002) (Fig. 3.41). In this same region, on the western side of the Altiplano or Western Cordillera, another contemporaneous though east-vergent thrust system (ETS) is developed, the Chucal Thrust System (Riquelme & Hérail 1997; Charrier et al. 2002c, 2005b) (Figs 3.40 & 3.42). As the two thrust systems have opposite vergences, tectonic activity developed an uplifted block between them, the Chapiquiña–Belén Ridge (Charrier et al. 1999, 2000) (Fig. 3.40). This ridge is probably continued southward by the Sierra de Moreno up to the latitude of Antofagasta and further south by the Domeyko Range.

In the Arica cross-section the stratigraphic succession in the Central Depression has been subdivided, from older to younger, into the following traditional stratigraphic units (Fig. 3.43): (1) the early to middle Oligocene Azapa Formation, of which probably only the uppermost part is exposed; (2) the 80–100-m-thick late Oligocene to middle Miocene Oxaya Formation; and (3) the 250-m-thick early to middle Miocene El Diablo Formation (Montecinos 1963; Salas et al. 1966; García 1967; Tobar et al. 1968; Vogel 1975; Vogel & Vila 1980; Naranjo & Paskoff 1985; García 1996, 2001, 2002; Parraguez...
were dated by cosmogenic $^{21}\text{Ne}$ and yielded Oligocene to Miocene ages that agree with the sedimentation age obtained on tuff intercalations within the Azapa Formation (Dunai et al. 2005). According to these authors, the dated surfaces have been practically unaffected by erosion since 25 Ma ago and these ages probably indicate the onset of hyperaridity in this region.

In the Precordillera, which relative to the Central Depression corresponds to a series of blocks thrust up to the west, a different stratigraphic succession is developed. Immediately east of the Ausipar fault, overlying the Early Jurassic Livícar Formation, exposed in deeply incised valleys (Muñoz et al. 1988a), rests the Oxaya Formation (Fig. 3.40). This c. 1000-m-thick ignimbritic Oxaya Formation forms a broad anticline or flexure on the hanging wall of the Ausipar Thrust. It consists of dacitic to rhyolitic welded ignimbritic flows containing pumice blocks that reach 20 cm in diameter (García, 1996), and lithic fragments are all of volcanic origin. Abundant K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ age determinations constrain the age of the Oxaya Formation to between 19.0 ± 0.6 Ma and 25.6 ± 0.9 Ma (Naranjo & Paskoff 1985; Aguirre 1990; Walford et al. 1995; Muñoz & Charrier 1996; García et al. 1996, 2000, 2004; Wörner et al. 2000b). These ages are in good agreement with magnetostratigraphic data obtained in this formation (Parraguz 1997; Tapia et al. 2000). Geochronologic and palaeo-magnetic data permit the recognition of several welded ignimbritic flows in the Oxaya Formation, emplaced over a period of 5–6 million years. These ignimbritic deposits originated from volcanic vents located along the Western Cordillera and were expelled to the west, across the Pampa Oxaya area and downward into the Central Depression, where they almost reached the Coastal Cordillera (Vogel & Vila 1980; Walford et al. 1995; García et al. 2000, 2004; Wörner et al. 2000b). Similar deposits are also known in southern Peru and south of the study region in northern Chile (Galli 1968; Galli & Dingman 1962; Lahsen 1982). The bending of the Oxaya and equivalent deposits covering the outcrop of the Precordillera formed major, essentially north–south oriented extensional cracks along the hinge of these structures that represent weakness zones controlling the development of major landslides (Naranjo 1997; Pinto 1999; García 2002; García & Hérail 2005; Pinto et al. 2004a). This is the case for the LLuta landslide or collapse cut in the Oxaya Formation, along the LLuta river valley, at c. 18°25'S, and the Moquella landslide (Pinto et al. 2004c) further south, in the Moquella region (19°15'S). Strasser & Schlunegger (2005) made a thorough description of the LLuta landslide or collapse, however, without considering the above-mentioned structural control along the crest of the Oxaya Anticline.

Further south, at 19°30'S, exposure ages of quartz clasts from the sediment surface of deposits assigned to the Azapa Formation, overlying Mesozoic units forming the Coastal Cordillera, were dated by cosmogenic $^{21}\text{Ne}$ and yielded Oligocene to Miocene ages that agree with the sedimentation age obtained on tuff intercalations within the Azapa Formation (Dunai et al. 2005). According to these authors, the dated surfaces have been practically unaffected by erosion since 25 Ma ago and these ages probably indicate the onset of hyperaridity in this region.

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East of the Belén Thrust system, in the next upthrust block, the Lupica Formation crops out, overlying the Belén Metamorphic Complex and sediments known as the Quichoco Beds, which have badly preserved marine fauna of Permian age (Montecinos 1963; Salas et al. 1966; Muñoz 1991; Pacci et al. 1980; García et al. 1996, 2004; Wörner et al. 2000b,c) (Fig. 3.43). According to recent radioisotopic age determinations, the Lupica and Oxaya formations are equivalent in age (García 1996, 2002; García et al. 1996, 2004). The 1800-m-thick continental Lupica Formation (Montecinos 1963; Salas et al. 1966), which has also been observed in the Western Cordillera, comprises thick siliceous tuff deposits and ignimbrites with volcaniclastic and lacustrine intercalations, developed in a very active arc setting consisting of stratovolcanoes and collapse calderas (García 1996, 2001; García et al. 2004). Several radioisotopic age determinations indicate a late Oligocene to middle Miocene age for the Lupica Formation (García 1996, 2002; Riquelme 1998; García et al. 2004).

The El Diablo Formation in this region is only exposed in the westernmost Precordillera because further east it has been completely eroded (Fig. 3.43). Further east in the Precordillera, on the east-dipping eastern limb of the Oxaya Anticline (Fig. 3.40), the Oxaya Formation is conformably covered by the middle Miocene (K–Ar ages between 15.1 ± 0.1 Ma and 12.3 ± 0.4 Ma) Zapahuira Formation (García 1996, 2002), which represents deposits of one of the oldest andesitic stratovolcanoes in the region. Syntectonic gravels of the Huaylas Formation (Salas et al. 1966), associated with the activity of the Copaquiña–Tinguñámar Thrust, contain mammal remains of post-Friasian/pre-Huayquerian age (Bargo & Reguero 1989; Salinas et al. 1991; Flynn et al. 2005), and onlap over the Oxaya and Zapahuira deposits. A 10.7 ± 0.3 Ma age from a tuff intercalation in the upper part of the Huaylas Formation indicates that anticline formation occurred at c. 12–11 Ma (Fig. 3.41).

**Altiplano.** At the same time in the Western Cordillera, deformation to the east of the Chapiquiña–Belén Ridge produced an east-vergent thrust system especially well developed in the Chucal region (Charrier et al. 1999, 2000) (Fig. 3.40). Thrusting was initiated during early Miocene times, continuing into the Pliocene epoch, and probably is still active today. This deformation has caused the development of several progressive unconformities, located mainly on the eastern limb of the
Chucal Anticline, and with associated syntectonic sedimentary deposits that contain an abundant fossil mammal fauna (Charrier et al. 1999, 2000, 2002c; Chávez 2001; Flynn et al. 2002a; Bond & García 2002) (Fig. 3.42). The resulting Late Cenozoic compressive structures (the Jaropilla Fault and the Chucal Anticline) control the present-day north–south orientated relief defined by various structural/topographic highs. The upper beds of the Lupica, the fluvial and lacustrine Chucal, and tuffaceous Quebrada Macusa or Macusa formations (García 1967; Muñoz 1991; Riquelme 1998; Chávez 2001; García et al. 2004) form a stratigraphic succession ranging in age between about 21.7 ± 0.8 Ma and 10.4 ± 0.7 Ma (from dated horizons within the lowermost and uppermost units), and are all deformed (Charrier et al. 2002c, 2005b) (C in Fig. 3.43). The age of 10.4 ± 0.7 Ma corresponds to an undeformed lava from the Anocarire Volcano, covering the upper brown tuff of the Quebrada Macusa Formation (Riquelme 1998). The late Miocene to Pliocene lacustrine Lauca Formation (Aguirre 1990; Muñoz & Charrier 1996; Kött et al. 1995; Riquelme 1998; Gaupp et al. 1999) which extends further east towards Bolivia, is only deformed in its westernmost outcrops, next to the Chucal Anticline. An ignimbritic intercalation located above
the middle part of the Laua Formation in the Laua Basin (Laua Igunimbrite of Kött et al. 1995, and Muñoz & Charrier 1996). Laucaper Perez—Ignimbrite of Wörner et al. (2000b) yielded three \(^{40}\)Ar/\(^{39}\)Ar dates on feldspar crystals of 2.67 ± 0.25, 2.32 ± 0.18 and 2.88 ± 0.13 Ma (Kött et al. 1995, table 2; Wörner et al. 2000b), and a \(^{40}\)K—\(^{40}\)Ar date on whole rock of 2.3 ± 0.7 Ma (Muñoz & Charrier 1996). Based on this age, of roughly 2.7 Ma, this intercalation has been correlated with the Pérez Igunimbrite by Kött et al. 1995, Riquelme (1998) and Wörner et al. (2000b). The Pérez Igunimbrite is well known in the Bolivian part of the Altiplano (Evernden et al. 1977; Lavenu et al. 1989; Marshall et al. 1992), and its occurrence in Chile suggests an even broader extent. Below this ignimbrite a

Lavenu roughly 2.7 Ma, this intercalation has been correlated with the 2.3

Wörner (1996); Lauca–Pérez Ignimbrite of Wörner (1996) in southern Peru (Flores 2004). The Chucal Formation can be correlated in Chile is correlative with the Pachía Ignimbrite in southern Peru where they receive the following names: (e.g. Choquelimpie; Aguirre 1990), which have been included in

the Altiplano–western Cordillera polymetallic metallogenetic province of the central Andes (Zappattini et al. 2001)

The Azapa, Oxaya and Huaylas formations described above for northernmost Chile (Tarapacá Region) also crop out in southern Peru where they receive the following names: Moquegua Inferior and Huayllillas formations, and Calientes Units respectively (Marcocci et al. 1985; Flores 2004). The Laucaper Pérez—Huaylas Igunimbrite in northern Chile is correlative with the Pachía Igunimbrite in southern Peru (Flores 2004). The Chucal Formation can be correlated in Bolivia with the Abaroa and Mauri formations (Lavenu et al. 1989; Hérail et al. 1997) and the Laua Formation with undeetermined ‘lacustrine deposits’ (Lavenu et al. 1989). A southern equivalent of the Azapa Formation is the Sichal Formation exposed further south in the Precordillera between 21°30’ and 22°S (Maksaev 1978; Skarmeta & Marinovic 1981). Deposition of the coarse alluvial Sichal Formation began in middle Eocene times in an intramontane basin formed by the uplift of Palaeozoic blocks almost immediately after the main Incaic deformation. Deposition of the Sichal Formation registered three stages of syntectonic sedimentation until early Oligocene times, and they are unconformably overlain by Miocene gravels of the Altos de Pica Formation (Blanco & Tomlinson 2006).

Deformation of the Oxaya Formation (Oxaya Anticline) in the Precordillera caused modification of the pre-existent parallel drainage network which became concentrated in a few deeply incised valleys that record a total incision of c. 1600 m. Post-folding incision in the major river valleys (11 to 0 Ma) attained rates of 56 to 58 m/Ma (García & Héral 2005). Shortening in the western edge of the Andes during Neogene times is only around 7.5 km because of the high dip angle of the reverse faults (García 2001; García et al. 2004).

Pisagua cross-section (19–20°S)

Central Depression. With regard to the morphostructural subdivision this section does not differ much from the Arica cross-section; however, it contains widely outcropping lacustrine deposits, and on its west side abundant nitrate accumulations (see the description below of the Central Depression for the Iquique–Chañaral region).

Forearc Precordillera. In the Precordillera, between 19°S and 20°S, the ignimbritic flows that form the Oxaya Formation further north are much thinner and are separated from each other by conglomeratic and sandstone deposits. In the Moquella area, these deposits have been partly included in the Latagualla Formation (Pinto 1999; Pinto et al. 2004a), and immediately south of this area, they have been included in the late Oligocene to late Miocene Altos de Pica Formation (Galli & Dingman 1962; Victor & Onccken 1999; Victor 2000; Farias 2003; Victor et al. 2004; Farias et al. 2005a) (Fig. 3.44). Both the Latagualla and the Altos de Pica formations include in their lowermost portions deposits that in the Arica region have been assigned to the Azapa Formation. The Altos de Pica Formation consists of a succession at least 600 m thick of ignimbrites and tuffs intercalated with conglomerates and breccias. The ignimbritic lavas interfinger to the east with rhyolitic-dacitic lavas, which are restricted to the eastern Precordillera adjacent to their associated volcanic vents. The upper part of the Latagualla Formation and the deposits covering the Altos de Pica Formation have been assigned to the El Diablo Formation (Farias 2003; Farias et al. 2005a), already described for the Arica cross-section. These deposits consist of thick alluvial westward-finning dark and brown breccias, conglomerates, coarse sandstones and thin evaporitic lenses. They contain dark volcanic bombs and the clasts correspond to andesitic and basaltic rocks, and interfinger to the east with andesitic and basaltic lavas erupted low on partly the volcanoes (Cerro de Sotoca) (Fig. 3.45). One age determination on one of these volcanic intercalations yielded a K–Ar age of 11.7 ± 0.4 Ma (Farias 2003; Farias et al. 2005a). The Camiña Igunimbrite (Muñoz & Sepúlveda 1992) or Pampa Tana lava (Pinto 1999), which overlies the El Diablo Formation, yielded K–Ar ages of 9.0 ± 1.0 Ma (Naranjo & Paskoff 1985) and 8.2 ± 0.7 Ma (Muñoz & Sepúlveda 1992), providing a minimum age for deposition of the El Diablo Formation and for the formation of the pediment surface on top of the latter.

At this latitude in the Precordillera, the west-vergent thrust system corresponds to high-angle reverse faults that cut through the more brittle Palaeozoic and Mesozoic basement but only fold the overlying Cenozoic cover (Fig. 3.45). These flexures affecting the cover rocks correspond to the southward prolongation of the WTS (Ausipar Thrust, other blind faults beneath the thick Oxaya deposits, and the Belén Thrust System). At 19°S, the Moquella flexure appears to be the prolongation of the WTS (Ausipar Thrust, Culmatización) with the Aroma Flexure being connected with the Moquella Flexure by a NW-striking thrust with left-lateral component. At 20°30’S, east of Iquique, deformation associated with this episode is represented by the Altos de Pica Flexure and other, more minor folds (Galli & Dingman 1962; Victor & Oncken 1999; Victor 2000; Victor et al. 2004). The throw along the Moquella Flexure has been determined at 700 m (Pinto et al. 2004a). Farias (2003) and Farias et al. (2005a) calculated a total relative surface uplift for the Calacala, Aroma and Soga Faults between 1200 and 1000 m, and if we add to this altitude the 2100 m of accumulated uplift in the three flexures, it leaves only 600–700 m to reach the average altitude of the Altiplano, which is c. 3800 m a.s.l. Farias (2003) and Farias et al. (2005a) proposed that the final 600–700 m uplift of the Altiplano was attained by a regional westward tilt of the Precordillera, which they attribute to east-vergent thrusting in the Subandean Sierras on the eastern flank of the Andes, where the shortening reaches > 200 km since the Oligocene epoch (Sempere et al. 1990; Sheffield 1990; Baby et al. 1997). As a result of this tilting the rivers flowing from the glaciers formed on the uplifted volcanoes began to cut the presently deeply incised river valleys. Considering the 8–9 Ma age of the Camiña Igunimbrite, which is the last deposit covering the pediplain before river incision, it is possible to conclude that final uplift of the Altiplano occurred after deposition of this lava, confirming the proposition of Gregory-Wodzicky et al. (1998) and Gregory-Wodzicky (2000) that the most important part of the uplift of the Altiplano occurred over the last 10 million years.
Fig. 3.44. Geological map for northernmost Chile between Arica and Iquique and correlation of Cenozoic successions along the Precordillera and Central Depression based on Pinto et al. (2004). Geological map: 1, Belén Metamorphic Basement; 2, Mesozoic units in the Coastal Cordillera; 3, Mesozoic units in the Precordillera; 4, Azapa, Lupica and Chucal formations (early–middle Oligocene to early Miocene); 5, Oxaya and Altos de Pica formations (late Oligocene to early Miocene); 6, Joracane, Huaylas and El Diablo formations (middle Miocene to Pliocene); 7, andesitic volcanic range (middle to late Miocene); 8, present-day volcanic arc; 9, Quaternary sedimentary deposits (alluvial, lacustrine and evaporitic); 10, anticline; 11, flexure; 12, thrust fault; 13, normal faults; 14, location of stratigraphic columns. Stratigraphic columns based on Salas et al. (1966), Galli & Dingman (1962), Parraguez (1997), Riquelme (1998), Pinto (1999), Garcia (2002), Farias et al. (2005a), Victor et al. (2004). BMC: Belén Metamorphic Complex. Stratigraphic columns: 1, lacustrine limestones; 2, marine calcareous deposits; 3, sandstones; 4, coarse sandstone; 5, siltstone, sandstone and conglomerate; 6, conglomerate; 7, tuff; 8, pyroclastic breccia and lahar; 9, ignimbrite; 10, andesite; 11, Palaeozoic and Mesozoic units; 12, radioisotopic ages (in Ma). Thick bar near E on map corresponds to cross-section in Figure 3.45.
Altiplano. In this region, the western Altiplano is mostly covered by volcanic arc deposits.

Iquique–Chañaral (20°S to 26°S)
The morphostructural subdivisions in this region are: the Mejillones Peninsula, the Coastal Cordillera, the Central Depression, the Domeyko Range (Sierra de Moreno, north of the Loa river valley) or Forearc Precordillera, the Preandean or Salars Depression, the Altiplano–Puna Plateau, the Eastern Cordillera, and the Subandean Ranges (Figs 3.37 & 3.46). The Western Cordillera located on the western border of the Altiplano–Puna corresponds to the present-day volcanic arc, and essentially delineates both the international boundary and the watershed between the Pacific Ocean and basins of internal drainage of the Altiplano–Puna to the east of the Western Cordillera. The peculiar Subandean or Salars Depression corresponds to an elongated tectonic basin located between the Domeyko Range and the Altiplano–Puna that can be traced from 21°S to 27°30'S, at the latitude of Copiapó. North of 21°S the Subandean Depression disappears and the northward prolongation of the Domeyko Range, called the Sierra de Moreno in this region, becomes part of the Altiplano. It contains, from north to south, the following salars: Carcote, Ascotan, Atacama, Imilac, Punta Negra, Pajonales, Pedernales, Marcungu and Negro Francisco, among others. South of c. 22°S, the characteristic flat surface forming the Altiplano tends to disappear and is replaced by the Puna, which is considered the southern termination of the Altiplano plateau and consists of a more rugged terrain with a higher mean altitude (4400 m a.s.l.). Cenozoic deposits in this region are exposed in most of the morphostructural units, each of which will be described in turn.

Mejillones Peninsula
The Mejillones Peninsula represents a tectonically rotated block of the continental platform that has been collapsing since Miocene times, and it exposes Neogene deposits that in parts of the platform are found below sea level (Niemeyer et al. 1996). NNW–SSE trending faults controlled the development and subsidence of a sedimentary depocentre, and later the sedimentary evolution of basins trapped in the middle and eastern side of the peninsula, as recorded by the existence of coarse sedimentary facies, successive tilting and unconformities next to the faults. Drastic facies changes and erosional unconformities in Pliocene times indicate a general tectonic uplift of the peninsula that determined the development of a late Pliocene upper terrace. Incision of the alluvial fans developed next to the faults indicates that tectonic activity is still active (Okada 1971; Niemeyer et al. 1996). Well exposed marine deposits in basins located peripherally to the Mejillones Peninsula (La Portada in Moreno Bay, Caleta Herradura and Mejillones Bay) have been assigned to the Miocene–Pliocene La Portada Formation (Ferraris & Di Biase 1978; M. Cortés 2000; Marquardt et al. 2003). These deposits record early Miocene initiation of extension of the platform, and subsequent late early Miocene–early Middle Miocene subsidence (coinciding with a marine highstand), possible late middle Miocene uplift, a late Miocene highstand, and further uplift in Pliocene times (Marquardt et al. 2003). Younger terraces record further vertical movement. The dating of these terraces, the evaluation of the relative participation of tectonic uplift and sea-level changes in their development and present-day location above sea level, and estimation of the rates involved in uplift have been strongly hindered by the difficulty in dating these surfaces and their associated deposits. However, recent significant progress has been made...
and it has been possible to determine the existence of slower uplift rates during early Pleistocene and first half of the middle Pleistocene and more rapid movements (240 mm/ka) in late Pleistocene times (Ortlieb et al. 1996b, 2003). Mejillones Bay is a small and shallow sedimentary basin dominated by hemipelagic sedimentation that represents an environment highly suited both to the study of palaeo-seismicity (Vargas et al. 2005; Le Roux & Vargas 2005), and the main physical ocean-climate factors driving sedimentation processes in the coastal region (Vargas et al. 2004).

Coastal Cordillera
The Coastal Cordillera represents a 1000–2000-m-high mountain range with a very steep (45º) coastal cliff, and a well preserved flat upper surface. The coastal escarpment is essentially inactive, and is the result of marine degradation of a continental margin actively uplifting since Pliocene times (Hartley & Jolley 1995; Niemeyer et al. 1996). Prominent Pleistocene terraces attest to the marine erosion that has affected the coastal escarpment (Martínez & Niemeyer 1982; Leonard & Wehmiller 1991; Ortlieb et al. 1996a; González et al. 2003). The flat top corresponds to an ancient erosional surface resulting from two erosional episodes, one in Oligocene–Miocene (Coastal Tarapacá Pediplain; Mortimer et al. 1974), and the other in late Miocene–early Pliocene times (González et al. 2003).

Central Depression
The Central Depression corresponds to a trench-parallel basin locally containing over 1000 m of Late Cenozoic to Quaternary alluvial, fluvial, lacustrine and evaporitic deposits with rapid variations of facies and thickness. In the Central Depression, south of Iquique and next to Quillagua (21º30’S) (Fig. 3.47), and extending upward along the Loa river valley into the Calama Basin, lacustrine, fluvial and alluvial deposits of early Palaeogene to Pleistocene age are exposed. Although these deposits extend eastwards away from the Central Depression across the Domeyko Range into the Salars Depression, north of Calama (Calama Basin), we will treat them all together in this section.

Apart from the already described deposits of Eocene age that form the 195-m-thick lower member of the Calama Formation (Blanco et al. 2003) (see second stage of Andean evolution), the Calama Basin hosts a sedimentary succession several hundred metres thick pertaining to the third Andean stage. The upper part of this succession represents an eastward extension of the deposits exposed in the Central Depression, with which, according to May et al. (1999), they share a similar sedimentary and tectonic history. The basin fill has been described by several authors including Maksav & Marinovic (1980), Naranjo & Paskoff (1981), Skarmeta & Marinovic (1981), Marinovic & Lahsen (1984) and May et al. (1999, 2005). The latter authors included the Calama Basin deposits in three major units: Calama Formation, El Loa Group and Chiu-Chiu Formation. More recently, N. Blanco (pers. comm.) partially modified the lithostratigraphic description by May et al. (1999, 2005) (see Fig. 3.48). 40Ar/39Ar age determinations from ignimbritic intercalations correlated with the Artola, Sifón and Carcote ignimbrites yielded late Miocene ages in the range 9.53±0.36 Ma, 8.27±0.13 Ma and 7.82±0.10 Ma, respectively, for the upper

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**Fig. 3.47.** Simplified geological map from the Quillagua region showing the relationship between the El Loa Formation (Quillagua lacustrine deposits), the Salar (evaporitic) deposits and the huge alluvial cones descending from the Sierra de Moreno. Salar development is controlled in this region by uplift of the Coastal Cordillera along the Atacama Fault Zone. Key: 1, El Toco Fm (Devonian); 2, Mesozoic units; 3, El Loa Formation; 4, Alluvial cones; 5, other Cenozoic deposits; 6, Salar de Llamara and other older salars (Soledad Formation of Skarmeta & Marinovic 1981); 7, trace of the Atacama Fault (highway runs along the fault trace).
part of El Loa Group (May et al. 2005). These ignimbrites are also known from the Salar de Atacama region to the east of the Calama Basin (see Fig. 3.49). In the Calama basin the El Loa Group is overlain with slight angular unconformity by the Chiu-Chiu Formation (Naranjo & Paskoff 1981; N. Blanco, pers. comm.), consisting of a 50-m-thick lacustrine succession of silt and mudstones, diatom muds and evaporites (Fig. 3.48), with fossil mastodont remains (Maksaev & Marinovic 1980). Its age has been assigned to the late Pliocene-Pleistocene (Marinovic & Lahsen 1984; May et al. 1999, 2005) and, in the Central Depression in the Quillagua region, it is conformably overlain by Pleistocene-Holocene lacustrine deposits forming several major salars (Salar de Pintados–Bellavista, and Salar de Llamara (Fig. 3.47). The deposits in these salars have been assigned to the Soledad Formation by Skarmeta & Marinovic (1981) (Lago Soledad Formation of Hoffstetter et al. 1957), which has an age range that apparently extends from the late Palaeogene to the Holocene. The Salar Grande represents the oldest salar deposits in this region and comprises c.15 m of gypsum–anhydrite covered by a halite crust (Maksaev & Marinovic 1980; Skarmeta & Marinovic 1981). The halite crust in the Salar Grande reaches 200 m, whereas in the Salar de Llamara it is very thin. The development of the lakes in which the Soledad and El Loa formations accumulated was probably controlled by the uplift of the western part of the Coastal Cordillera along the Atacama Fault Zone, as seen further south in the Chañaral region (26–27ºS) (see Riquelme 2003; Riquelme et al. 2003).
Alluvial fan deposits form extensive deposits along the western piedmont of the Domeyko Range, known as the Sierra de Moreno in this region. The Arcas Fan (Skarmeta & Marinovic 1981), developed since late Miocene times, is according to Kiefer et al. (1997) one of the largest known alluvial fans in the world. These enormous alluvial fan deposits are good evidence for the intense erosion associated with the late Neogene uplift of the Domeyko Range, which was probably associated with further thrusting of this range over the Salar de Atacama (see Muñoz et al. 2002). The interfingered relationship between these syntectonic deposits and the Soledad and El Loa formations records the complex tectonic and sedimentary evolution in the forearc in this region.

Along the west side of the Central Depression, north and south of Antofagasta and along the NW-orientated depression between Antofagasta and Calama (Antofagasta–Calama Lineament), abundant nitrate deposits are developed in the meteorized levels of the outcropping units (see Chong 1984, and Chapter 7). Intensive exploitation of these deposits and their associated boron, chlorine and iodine content has caused strong alteration of the original landscape surface in these regions.

Domeyko Range and Salars (or Preandean) Depression

South of the Loa river valley the Domeyko Range forms a well defined essentially north–south trending morphological element that locally reaches altitudes of 4000 m, separated from the Coastal Cordillera by the Central Depression and from the Puna (the southern prolongation of the Altiplano) by the Salars Depression. The west side of the Domeyko Range forms the Forearc Precordillera. The Salars Depression is here mainly represented by the Salar de Atacama Basin. The Domeyko Range consists of elongated (30 to > 100 km) and broad (c. 10 km) basement ridges that form the core of major anticlinal structures bounded by parallel steep reverse faults, whereas minor folds not involving the basement are developed in the cover rocks (Reutter et al. 1996) (Fig. 3.46). This structural pattern, and the formation of the major north–south trending Domeyko Fault System developed along the axis of the Domeyko Range, were mainly formed during the Incaic phase (Maksaev 1978; Maksaev & Zentilli 1999; Reutter et al. 1996) in Eocene times.

The Salar de Atacama is a 120-km-long, 60–90 km wide, north–south trending depression, bounded to the west by the Domeyko Range and to the east by the Western Cordillera or the present-day volcanic arc, located on the western side of the Altiplano. In this region the volcanic arc is deflected to the east along strike for c. 300 km around the salar. At present, the Salar de Atacama, at 2300 m altitude, is a desiccated, essentially flat plain in the bottom of a closed drainage basin (Muñoz et al. 2002).

The eastward thrusting of the Domeyko Range over the Salar de Atacama Basin infill continued after deposition of the Late Cretaceous–Early Cenozoic Purilactis Formation. On the eastern flank of the Domeyko Range, syntectonic deposits (progressive unconformity) associated with this tectonic activity correspond to the Oligo-Miocene 1300-m-thick, coarse detrital and evaporitic Tambores Formation (Pampa de Mulas Formation, further south; Gardeweg et al. 1994) that
interfingers to the east with the finer grained San Pedro Formation (Brüggen 1934, 1942, 1950; Ramirez & Gardegew 1982; Marinovic & Lahn 1984; Wilkes & Görler 1988; Hartley et al. 1992; Flint et al. 1993; Naranjo et al. 1994b; Mpodozis et al. 2000; Muñoz et al. 2000) (Fig. 3.38). Reactivation next to the top of the San Pedro Formation yielded two K–Ar ages on biotite of 26.6 ± 0.8 Ma and 26.8 ± 1.4 Ma (Mpodozis et al. 2000). After a non-magmatic interval from 38 Ma to 28 Ma, magmatism resumed in the Western Cordillera (see Reutter 2001). Younger deposits on the eastern flank of the Domeyko Range and in the en-echelon patterned fold system forming the Cordillera de la Sal include a series of Miocene ignimbritic intercalations that form part of the Altiplano–Puna Volcanic Complex (Tomlinson et al. 2004) and have yielded K–Ar ages on biotite of between 10 and 6 Ma, e.g. Artola (9.8 ± 0.3 Ma), Sifón (8.8 ± 0.5 Ma), Yerba Buena (8.2 ± 0.7 Ma) and Pelón (7.0 ± 1.0 Ma and 6.9 ± 0.8 Ma). A more recent determination (40Ar/39Ar reverse isochron on hornblende) yielded an age of 7.05 ± 0.10 Ma for the Sifón Ignimbrite.

In addition there are coarse to fine detrital and evaporitic deposits such as the Plocine Vilama Formation, dated at 3.7 ± 1.0 Ma (see Mpodozis et al. 2000) (Fig. 3.49). The ignimbritic deposits correspond to the similar deposits exposed further north in the Sierra de Moreno (and to intercalations in the Calama Basin deposits) assigned to the Ichuno Formation (Maksaev 1978; Skarmeta & Marinovic 1981) and the Ujina Ignimbrite (Vergara 1978a). Whereas the sedimentary deposits derive from the erosion of the Domeyko Range to the west, the ignimbritic flows originated in the Western Cordillera to the east. Some of these flows crop out at low levels along the eastern flank of the Domeyko Range and dip to the east, indicating that subsidence in the Salar de Atacama and/or uplift of the Domeyko Range was active after their deposition.

The western slope of the Altiplano–Puna plateau is draped by Miocene and Pliocene ignimbrites, which in turn are covered near the eastern margin of the Salar de Atacama by alluvial fan deposits (Muñoz et al. 2002). This situation makes difficult the structural interpretation of the eastern margin of the Salars Depression. Eastward, in the plateau, the ignimbrites are covered by the andesitic volcanic centres that form the present volcanic arc (Ramirez & Gardegew 1982).

Tectonic aspects

In addition to the above-mentioned generalized uplift of the Mejillones Peninsula and the Coastal Cordillera since Pliocene times, Pliocene–Quaternary near-surface extensional deformation along the Coastal Cordillera associated with activity along the Atacama Fault Zone (AFZ) has been recently demonstrated for the region between Iquique and Antofagasta (González et al. 2003). The inherited geometry of the AFZ consists here of a series of strike-slip duplexes formed by north–south striking faults and NW–striking splay faults (González et al. 2003; Cembrano et al. 2005) (Fig. 3.30). Cenozoic activity/reactivation along the described fault pattern is evidenced by prominent scarps (30 to 100 m high) that control the horst and graben topography of the Coastal Cordillera (Arabasz 1971; Okada 1971; Hervé 1987a; Naranjo 1987; González & Carrizo 2000, 2003). According to González et al. (2003), half-grabens are formed along the north–south striking and east-dipping structures, while associated dextral movement occurs along the splay faults that connect the north–south striking normal faults. Recent extensional activity is recorded by north–south oriented open cracks hundreds of metres long, some of which are spatially associated with the major north–south orientated faults.

The structural pattern of the Domeyko Range (including the Sierra de Moreno) is determined by a number of blocks separated from each other by essentially north–south trending faults. The most important of these is the major strike-slip Domeyko Fault System (DFS) or Argomedo–West Fissure Fault (or Falla Oeste) System that extends north and south of the Loa river valley, and cuts the Calama basin fill, influencing its sediment distribution (May et al. 1999, 2005; Blanco et al. 2003). This fault system is located along the axis of the previous Late Cretaceous to Eocene volcanic arc and shows evidence for the inversion of normal faults, and for dextral as well as sinistral strike-slip movement (Mpodozis et al. 1993; Charrier & Reutter 1994; Tomlinson et al. 1994; Reutter et al. 1991, 1996; Dilles et al. 1997; Tomlinson & Blanco 1997a, b; Reutter 2001). North of Calama, activity along this fault system has been detected up until mid-Miocene times. In the eastern flank of the Domeyko Range, west of the Salar de Atacama, confirming evidence revealed by the seismic profiles (Muñoz et al. 2002), most of the basement blocks have been thrust eastward over the Purilactis Group, which is exposed along the El Bordo Escarpment, inducing thin-skinned deformation in the Salar de Atacama Basin infill (Fig. 3.50b). Deformation of the Purilactis sediments is especially significant close to these basement blocks, with the red sediments displaying isoclinal to chevron folds with subvertical axial surfaces (Arriagada et al. 2000). A number of important middle Eocene to earliest Oligocene porphyry copper deposits are distributed along the east side of this >200-km-long fault system (e.g. Rosario, Collahuasi–Ujina, Quebrada Blanca, El Abra, Chuquicamata, Mina Sur, M&M, Zaldivar, La Escondida, and, further south, Exploradora and Potrerrillos) (Zappettini et al. 2001) (Fig. 3.51). The west-fissure next to the Chuquicamata mine juxtaposes a western non-mineralised granodiorite of c. 36 Ma against a mineralized eastern block consisting of a Palaeanoic granite intruded by porphyries (dated at 32–30 Ma) and their alteration products (Maksaev et al. 1988b; Maksaev 1990; Camus 2003). According to Reutter (2001), dextral movement along the West-fissure was associated with the oblique (north-eastward) convergence, whereas sinistral movements were probably caused by local conditions, i.e. inflection of the Bolivian Orocline.

Contractional deformation episodes since late Oligocene times in this region, as in the Arica and Iquique region, resulted in reduced shortening of the forearc (Lamb et al. 1997; Kley & Monaldi 1998; Hartley et al. 2000). Uplift is driven by accretion and subcrustal duplexing of material removed by subduction erosion beneath the forearc (Rutland 1971; Delouis et al. 1998; Hartley et al. 2000). The lack of important amounts of crustal material in the continental margin when modelling a balanced cross-section of the Central Andes (Schmitz 1994) supports this idea. In the Central Depression, sediment accumulation was controlled by uplift of the Coastal Cordillera; this process prevented sediment transfer directly from the High Andes to the Pacific. Accommodation space in the Salars Depression is due to superimposition of an Oligocene–Miocene basin bounded by uplifted margins (the Precordillera and Puna together with the Western Cordillera) on an older extensional backarc basin (Hartley et al. 2000).

The higher altitude of the Puna relative to the Altiplano (4400 m a.s.l. versus 3800 m a.s.l. in the Altiplano) together with the thinner crust of the Puna relative to the Altiplano has been interpreted as caused by the existence of a cooler and denser lithosphere below the Puna (Gerbault et al. 2005).

**Chañaral and La Serena (26°5’ to 30°5’)**

This region essentially corresponds to the flat-slab segment and the Central Depression is not developed (see Figs 3.1b & 3.37). However, in the northern part of this region it is possible to differentiate a coastal range lying to the west of the Atacama Fault System and a Forearc Precordillera (west flank of the Domeyko Range) to the east. Also in the north the Domeyko Range is separated from the southern end of the high Andean plateau (Puna in this region) by the Salars Depression that also ends in this region, at latitude 28°S. Further south, it is not possible to differentiate a Coastal Cordillera from a Precordillera.

**West side of the coastal range.** The geological evolution in the coastal region is recorded by sedimentary deposits and abrasion marine terraces (Marquardt et al. 2000, 2004; Ortlieb et al. 2005).
Fig. 3.50. Geological and structural setting of the Atacama Salar area, based on Muñoz et al. (2002). (a) Geological map with location of structural section. Key: 1, Late Miocene to Quaternary deposits; 2, Pliocene to Quaternary ignimbritic flows; 3, Miocene deposits; 4, Oligocene–Miocene deposits; 5, Cretaceous to Eocene deposits; 6, Late Cretaceous intrusive; 7, Palaeozoic basement; 8, Fold; 9, thrust fault; 10, Toconao-X1 exploration well. (b) Structural section across the El Bordo Escarpment and the Salar de Atacama. Basement blocks, which have been uplifted along previous normal faults associated with the extension of the Salar de Atacama (thick-skinned), thrust the infill deposits of the Salar and induced thin-skinned deformation in them.
The deposits, exposed in wide tectonically controlled embayments along the coast (Caldera and Coquimbo regions; Fig. 3.52), correspond to the classic middle to late Miocene Navidad and Pliocene Coquimbo formations. In the Caldera–Bahía Inglesa region, the deposits have been included in two units (Bahía Inglesa Formation of late Miocene to early Pliocene age and Agua Amarga Beds of early to late Pliocene age) (Marquardt et al. 2000, 2004). These deposits rest on Palaeozoic metamorphic rocks and Mesozoic granitoids and coarse continental gravels (Quebrada Totoral Gravels), and are overlain by continental and littoral deposits. The latter are included in the Caldera Beds and correspond to the deposits accumulated on seven or eight uplifted marine terraces. Sedimentological and structural analyses on these deposits...
Fig. 3.52. Marine Neogene coastal deposits and morphological features on the western side of the Coastal Range in northern, central and southern Chile up to the triple-junction at 47ºS. (A) Distribution of coastal Neogene basin deposits, based on Martínez-Pardo (1990) and Sernageomin (2003). (B) Coastal terrace development along the central Chilean coast, based on observations by Paskoff (1970) and Paskoff et al. (1995) in the Coquimbo region.
indicate that a marine ingress occurred in the middle Miocene coinciding with a marine highstand and that extensional deformation occurred during late Miocene times. This middle Miocene marine ingress might be correlated with the similar ingress deduced for the Mejillones Peninsula after initiation of the tectonic extension in the marine platform (Marquardt et al. 2003). Chronostratigraphic analyses on the middle to early Pleistocene terrace levels indicate uplift rates of 0.3–0.4 m/ka for the last 430 ka (Marquardt et al. 2004). However, more chronological information is needed to establish more precise correlations along the coast between the platformal marine deposits as well as the marine terraces to determine the chronology of the tectonic evolution. A correlation based on the faunistic content can be made between the 100-m-high terrace and one of the early Pleistocene terraces (400 ka; Ortlieb et al. 1997) in the Mejillones Peninsula. Recent studies in the outlet area of Quebrada Carrizalillo (29ºS), in the southern part of the region, included these deposits in the Coquimbo Formation and concluded that they accumulated under an arid climate on the continental platform at depths corresponding to the beach zone and, possibly, the continental talus (Gómez 2003). Sedimentation was controlled by both eustatic changes and tectonism of the coastal region associated with the Atacama Fault Zone. Transgression associated with the middle Miocene (15.5 Ma) eustatic high occurred in this region somewhat later (14.6 Ma) due to uplift associated with this fault. Abrasion marine terraces were developed (c. 3, 5 and 103 m a.s.l.) that can be regionally correlated with well developed terraces around Caldera, further north, and Coquimbo–Talcahuano, further south.

Coastal range. The main tectonic feature in the coastal region corresponds to the Atacama Fault System (AFS), which consists in the north of three north–south striking, subvertical and overlapping branches (Fig. 3.30). Middle to late Miocene vertical movements along the AFS accommodated the relative uplift of the western side of the coastal range (Arabasz 1971; Okada 1971; Mortimer 1980; Naranjo 1987; Hervé 1987a; Armiño & Thiele 1990; Delous et al. 1998; Riquelme et al. 2003), and so controlled Neogene deposition both in the Forearc Precordillera, Atacama Gravels; Sillitoe et al. 1968) and immediately east of the fault (Riquelme et al. 2003). Recent tectonic activity on the AFS has also been observed which is consistent with a localized uplift of the crustal block west of the AFS (Riquelme et al. 2003).

Forearc Precordillera or western flank of the Domeyko Range. The main tectonic feature in the Domeyko Range in this region is the Agua Amarga–Sierra del Castillo Fault system (Tomlinson et al. 1994, 1999; Niemeyer 1999; Cornejo et al. 2006). Activity along this fault system occurred since Palaeozoic times with several Mesozoic and Cenozoic reactivations. A particularly important sinistral transpressive reactivation occurred during the Eocene epoch and presumably controlled the north–south trending belt of late Eocene–Oligocene porphyry copper deposits, and a later (Late Cenozoic) reactivation caused displacement of the Atacama Gravel deposits (Tomlinson et al. 1999).

In the northern part of this region, between 26ºS to 28ºS, the late Oligocene to Miocene magmatic front was located on the eastern flank of the Domeyko Range, in the Salars Depression, and in the western border of the Atacama Basaltic Province (Fig. 3.37). The products of this magmatic activity, which was associated with rich Cu, Ag and Au epithermal mineralization, form the Maricunga Belt (Franja de Maricunga) (Mpdoodoiz et al. 1995). The climatic hyperaridization has allowed preservation of the original shape of the oldest late Oligocene volcanic domes, and volcaniclastic events have been recognized between 26 Ma and 6.5 Ma, occurring under alternating extensional and contractional tectonic conditions. Extensional tectonic conditions prevailed during the first event (26 Ma to 21 Ma) and at the end of the third event (c. 13 Ma to 12 Ma), whereas contractional conditions associated with crustal thickening occurred during the second (20 Ma and 17 Ma) and fourth (11 Ma to 7 Ma) events. The fifth and last volcanic event developed on a thickened crust (Mpdoodoiz et al. 1995).

In the southern part of this region, Reutter (1974) described, on both sides of the water divide at c. 29ºS, the Rio de la Sal and the Potrerillos formations. The former consists of a >1000-m-thick succession of andesitic breccias (some are possibly lahars) followed by epioritic deposits (gypsum and intercalated halite), and the latter comprises <500 m of gravels forming a 1º to 2º westward-dipping surface (‘mantling gravels’) with intercalated tuffs. The Rio de la Sal deposits were later assigned to the east of the AFS up to the summit line of the Domeyko Range, excluding involvement of the present-day Salars Depression. East of the AFZ (in what some authors name the Central Depression) a ~300-m-thick, fine-grained playa succession was deposited that grades laterally eastward into the Atacama Gravels. K–Ar age determinations from intercalated tuffs in the Precordillera yielded 17 to 13 Ma (Cornejo & Mpdoodoiz 1996; Tomlinson et al. 1999; Cornejo et al. 2006) and 12 Ma in the sediments next to the Atacama Fault System (Clark et al. 1967), indicating essentially a middle Miocene age for these deposits. The top of the Atacama Gravels and the deposits next to the AFZ, as well as bare rock surfaces, form a single surface throughout the southern Atacama Desert, known as the Atacama Pediplain (Clark et al. 1967; Sillitoe et al. 1968; Mortimer 1973; Paskoff & Naranjo 1979; Naranjo & Paskoff 1980; Riquelme 2003). The Atacama Pediplain is covered by the San Andrés Ignimbrite that yielded K–Ar ages of 11.5±0.5 Ma, 9.5±0.5 Ma, 9.0±0.3 Ma (Clark et al. 1967; Sillitoe et al. 1968; Mortimer 1973). 10.2±0.9 (Cornejo & Mpdoodoiz 1996) and a recent U–Pb age of 9.12±0.08 Ma (S. Matthews, pers. comm.). The pedimentation event can be attributed to three causes: (1) blocking of the valleys flowing from the Pediplain to the ocean by uplift of the western Coastal Cordillera; (2) consequent abundant gravel sedimentation on the west-dipping Pediplain; and (3) mid-Miocene climatic hyperaridization which prevented erosion of the gravel deposits in late middle Miocene times once the connection with the ocean was resumed (Riquelme 2003). However, shortly afterwards (c. 10 Ma) strong incision affected the Pediplain and formed the present-day canyon system. The cause for incision of the plain surface and the Atacama Gravels has been explained by a regional westward tilt (c. 1º) of the entire forearc and by the availability of melt water from the high-seated glaciers once the altitude necessary for their existence was acquired from the tilting process (Riquelme 2003). This explanation is similar to the one given by Farias (2003) and Farias et al. (2005a) for the incision of the Altos de Pica and El Diablo formations, and equivalent deposits, in the Pediplain of Arica and Iquique. Tilting has been attributed to eastward thrusting in the Andean foreland (Subandean Sierras) and is associated with an important episode of Andean uplift (Gregory-Wodzicki 2000).

The Domeyko Range and Salars Depression expose a Cenozoic, basement-involved belt of folds and mostly steep-dipping reverse faults showing both east and west vergence.

High Andes. In the northern part of this region, between 26ºS to 28ºS, the late Oligocene to Miocene magmatic front was located on the eastern flank of the Domeyko Range, in the Salars Depression, and in the western border of the Atacama Basaltic Province (Fig. 3.37). The products of this magmatic activity, which was associated with rich Cu, Ag and Au epithermal mineralization, form the Maricunga Belt (Franja de Maricunga) (Mpdoodoiz et al. 1995). The climatic hyperaridization has allowed preservation of the original shape of the oldest late Oligocene volcanic domes, and volcaniclastic events have been recognized between 26 Ma and 6.5 Ma, occurring under alternating extensional and contractional tectonic conditions. Extensional tectonic conditions prevailed during the first event (26 Ma to 21 Ma) and at the end of the third event (c. 13 Ma to 12 Ma), whereas contractional conditions associated with crustal thickening occurred during the second (20 Ma and 17 Ma) and fourth (11 Ma to 7 Ma) events. The fifth and last volcanic event developed on a thickened crust (Mpdoodoiz et al. 1995).

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to the Escabroso Member of the Doña Ana Formation by Nasi et al. (1990). The overlying Potrerillos Formation (exposed north of 29ºS) is correlated with the Miocene Atacama Gravels (see above) and represents the southernmost outcrops of these gravels (see Fig. 3.53). Between 29ºS and 30ºS, scattered outcrops of the late Eocene–Miocene volcanic, volcaniclastic and sedimentary Doña Ana, Las Tórtolas and Vallecito formations are exposed (Nasi et al. 1990), which form the prolongation of similar exposures further south, described in more detail in the next section. Finally, between 29º and 30ºS, hosted in the late Eocene–Miocene deposits, is the El Indio Belt (including the Pascua and El Indio mines), one of the large epithermal precious metal districts in the Central Andes (Maksaev et al. 1984; Martin et al. 1995, 1999a).

La Serena to Santiago Region (southern part of the flat-slab segment)

In this region, located in the flat-slab Andean segment, the Central Depression and the volcanic arc are not developed. Late Palaeogene to Recent deposits are mainly located along the coast and form narrow, discontinuous outcrops in the High Andes.

Coastal region. Along the coast of central Chile Pliocene events are recorded by marine sediments of the Coquimbo Formation into which have been cut five well developed and preserved (and virtually continuous) marine terraces (wave-cut platforms) (Paskoff 1970, 1977; Herm & Paskoff 1967; Paskoff et al. 1995; Fuenzalida et al. 1965; Benado 2000) (Fig. 3.52B). These terraces are the result of Pleo-Quaternary transgressions and regressions linked with sea-level fluctuations of glacio-eustatic origin and Andean uplift (Paskoff et al. 1995). The wave-cut platforms are covered by thin beach sediments in which sands, pebbles and shells are mixed (Herm 1969). The age of the terraces is as follows: the two highest (Serena I and Serena II) correspond to the Early Pleistocene interglaciation, Herradura I corresponds to the penultimate interglaciation, Herradura II corresponds to the ultimate interglaciation, and Vega corresponds to the Holocene. Finally, there is what is known as the Cachagua level, corresponding to an interstadial transgression during the last glaciation (Paskoff et al. 1995).

High Andes. In the High Andes, between c. 28º45’S and 30º30’S, late Eocene to late Miocene volcanic deposits have been designated as the Doña Ana Group (Eocene–Oligocene) (equivalent to the Río de la Sal Formation defined by Reutter (1974) between c. 28º45’ and 29º15’S), which consists of the Tilito and the Escabroso formations, the Las Tórtolas Formation (early to middle Miocene), the Tambo Formation (middle Miocene) and the Vallecito Formation (late Miocene) (Mpodozis & Cornejo 1988; Nasi et al. 1990; Martin et al. 1995, 1997b; Bissig et al. 2001). The Doña Ana Group is thus coeval with the Abanico Formation, and the Las Tórtolas, Tambo and Vallecito formations are coeval with the Farellones Formation, further south (see below) (Fig. 3.54). The geochemical signature of the lavas in these units indicates a gradual ‘enrichment’ of the magmas from late Oligocene to early late Miocene time, indicating a gradual increase of crustal thickness respective to the older parts of the Doña Ana Group (Kay & Abbruzzi 1996). The Doña Ana Group was probably deposited in an extensional basin, which further south (see below) is named the Abanico Extensional Basin (Fig. 3.55). Thus the Las Tórtolas, Tambo and Vallecito formations correspond to essentially volcanic deposits accumulated during tectonic inversion of the extensional basin in this region (Charrier et al. 2005a).

Cenozoic intrusive units are located exclusively in a region bounded by two fault zones that appear to correspond with the bounding faults of the Abanico Extensional Basin (except for the Palaeocene–Eocene Fredes Unit, which is also exposed west of the western boundary fault: Charrier et al. 2005a). These units have late Eocene to Miocene ages, and are known as El Maitén–Junquillar, Bocatoma (40K–40Ar, 40Ar/39Ar and one U–Pb ages range from 39.5 ± 1.3 Ma to 31.1 ± 1.2 Ma), Rio Grande and Infiernillo (16.7 ± 0.6 Ma), and young to the east (Mpodozis & Cornejo 1988; Martin et al. 1995, 1997b; Bissig et al. 2001). The age of the intrusive rocks matches the age of the Abanico and Farellones formations between 33ºS and 36ºS, this is older than those emplaced in these formations, as well as in Mesozoic deposits forming the eastern Principal Cordillera (between 33ºS and 36ºS), which have been dated to between 21.6 ± 4.9 Ma and 5.5 ± 0.2 Ma (see Kay & Kurtz 1995; Kurtz et al. 1997; Charrier et al. 2002c).

Central Chile between Santiago and Concepción

South of 33ºS both the Central Depression and basalt–andesite volcanism (Southern Andean Volcanic Zone – SVZ) reappear. Thus, the major morphostructural units are, from west to east: Coastal Cordillera, Central Depression and Principal Cordillera.

Coastal Cordillera. Marine terraces are once again found along the coast and record the same glacio-eustatic events described for the La Serena–Coquimbo region (Fuenzalida et al. 1965). More distinctively, along the west side of the Coastal Cordillera, between 33º40’S and 34º15’S, are extensive exposures of late Miocene marine sediments known as the Navidad Formation (Darwin 1846; Tavares 1979b; Encinas et al. 2003; Finger et al. 2003) (Fig. 3.52). This formation is overlain by the
PlIOCENE, transitional marine to continental La Cueva Formation (Tavera 1979b) which contains abundant pyroclastic material. These formations interfinger and are overlain to the east by continental deposits belonging to the Potrero Alto Beds of uncertain Miocene–Pliocene to Pleistocene age (Wall et al. 1996). The Navidad Formation was deposited in a rapidly subsiding basin. According to Encinas et al. (2003) the basin reached depths of 1500 m, and it was at this time that the basal member (La Boca) was deposited. Deposition of the two higher members (Lincancheo and Rapel) occurred in much shallower waters as was the case during the deposition of the La Cueva Formation. However, short pulses of rapid subsidence of the continental platform occurred between deposition of the Lincancheo and Rapel members and between the Rapel Member and the La Cueva Formation (Encinas et al. 2003). The Navidad and La Cueva formations can be correlated to the north of 33ºS with the Caleta Horcón Formation (Rivano 1996) and to the south with the Ranquil (Arauco region) and Lacui (Chiloé) formations. The Potrero Alto Beds have been correlated with the Confluencia Formation that interfingers with the Caleta Horcón Formation.

At 33º30’S, in the western side of the Coastal Cordillera and immediately north of the exposures of the Navidad Formation, Miocene andesitic lavas of the La Patagua Formation are exposed in the lower Maipo Valley. No relation between the lavas and the Miocene deposits has been observed. These lavas are probably associated with a major NW–SE orientated lineament that has been interpreted as an accommodation fault related to the development of the Abanico Extensional Basin (see below) in the Principal Cordillera (Rivera & Cembrano 2000).

Central Depression. Essentially Quaternary alluvial deposits derived from the Principal Cordillera are especially well developed at the mouth of the main river valleys in the Central Depression. Explosive volcanic activity in the volcanic arc (in the Principal Cordillera) expelled abundant lahars and volcanic avalanche deposits such as La Cueva Fm that reached the Central Depression and even the coast. Between 33º30’S and 34º30’S (Santiago and Rancagua) the Pudahuel and Machalí ash tuffs covered most of the Central Depression and extended towards Argentina, in the Yaucha and Papagayos river valleys (Stern et al. 1984a). The origin of these tuff deposits is apparently associated with the Maipo Caldera which is situated at 34ºS (Stern et al. 1984a). The wide distribution of such tuff deposits suggest that this pyroclastic flow covered most of the Central Depression around 34ºS, including the Santiago basin. The tuff deposits located next to Rancagua (Machalí Tuff) and Santiago (Pudahuel Tuff) yielded apatite fission track ages of 0.44±0.08 Ma and 0.47±0.007 Ma respectively. Further south, the Central Depression contains several examples of pyroclastic flows from volcanic centres located in the Principal
The younger Farellones Formation is a thick, gently folded, almost entirely volcanic unit forming a north–south trending zone of outcrops between approximately 32° and 35°S (Vergara et al. 1988). In this region, it reaches a thickness of 2400 m and is composed of andesitic to rhyolitic lavas, volcaniclastic deposits, and limited sedimentary deposits (Thiele 1980; Charrier 1981b; Vergara et al. 1988). The Abanico Formation appears to have been deposited in an extensional basin that underwent subsequent tectonic inversion. Geochemical composition and thermal maturity data of the Abanico Formation indicating a relatively thin crust during early basin development and high heat flow conditions during burial support a major extensional episode of the crust. The basin began to form before 36 Ma, while the crust was relatively thin, persisting throughout the Oligocene epoch and into early Miocene times. Contraction occurred both during and after late depositional stages of the Abanico Formation and was controlled by inverted extensional faults originally associated with basin development. This contractional event began before 21 Ma and ended by c. 16 Ma, occurring asynchronously throughout the region, and was associated with crustal thickening. Sedimentation and volcanism continued along the axis of this elongate basin throughout contraction, creating the Farellones Formation, while along its eastern and western margins inversion and exhumation led to erosion of the Abanico Formation.

The oldest radioisotopic ages for the base of the Farellones Formation (25 and 21 Ma) are from the northern part of the study region while the youngest ages for the base of the Abanico Formation (16.1 Ma) are from the south, implying a north to south progression of magmatism (and possibly deformation) associated with the initiation of the Farellones Formation. This progression may be related to the southward shift of the Juan Fernández Ridge along the continental margin during this time. The kinematics of major faults bounding the Abanico deposits to the east and west are consistent with the proposed inversion of an extensional basin with vertical displacements of 1000 to >2000 m. This, together with abundant shallow seismicity below the Principal Cordillera (which partly aligns with the trace of one of the aforementioned faults), suggests that these faults (and probably other minor ones associated with the basin development and inversion) contributed significantly to Andean uplift and that they are involved in ongoing crustal accommodations. The former southward prolongation of the basin (>38°S) in Chile is reflected by the presence of thick volcanic series equivalent in age to the ones exposed between 33°S and 36°S. A paucity of Cenozoic outcrops obscures whether the Abanico Extensional Basin extended north of 33°S. However, fault alignments that extend the traces of the major San Ramón and Chacayes–Yesillo and El Diablo fault systems exposed at the latitude of Santiago can be followed to at least 30°S (see Charrier et al. 2005a), well within the flat-slab Andean segment. In this region, two thrust systems separated by c. 50 km and with opposite vergences, cut an igneous Palaeozoic basement. Cenozoic plutons, which intrude the basement between these two fault systems, show ages that decrease eastwards and approximate (though are somewhat older) to the ages of the Abanico Formation (and its intrusives further south). This may indicate that the Abanico Extensional Basin once reached this latitude, but that subsequent basin inversion and uplift caused...
by subduction of the Juan Fernández Ridge, exhumed its base-
ment exposing deep-seated portions of the intrusive bodies
originally associated with basin development. The Cenozoic
Abanico Extensional Basin represents a major tectonic system
in the Southern Central Andes, particularly along the Chilean
side of the Principal Cordillera between 30° and 44°S. The find-
ing of Eocene fossil mammal remains (El Tapado Fauna)
within the Abanico Formation, in the Tinguiririca river valley
at 35°S west of Termas del Flaco (Wyss et al. 1994), indicates a
late Eocene onset for deposition (Charrier et al. 2002b). This
fauna is considerably older than the one collected at Termas del
Flaco next to the basal unconformity (Tinguiririca Fauna;
Wyss 1999).

The deposits of the Farellones Formation typically cover the
Abanico Formation. The contact has generally been described
as unconformable (Aguirre 1960; Kholo 1960; Jaros & Zelman
1967; Charrier 1973b, 1981b; Thiele 1980; Moscoso et al.
1982b), but other authors (Godoy 1988, 1991; Godoy & Lara
1994; Godoy et al. 1999) view the contact as (1) being confor-
able or pseudo-conformable, or (2) corresponding to a regional
low-angle thrust that they connect with the El Fierrero Fault at
Termas del Flaco described by Charrier et al. (2002b). The age
of the basal Farellones deposits that cover the Abanico Forma-
tion and the contact between these units varies considerably at
different localities. Considering that the youngest age found
for the Farellones Formation in Central Chile (33–36°S) corre-
spond to the late Miocene, one may deduce a Miocene age
for this unit. However, Miocene ages have also been obtained
for the upper part of the Abanico Formation (Charrier et al.
2002b).

The Abanico Formation is correlated southward with the
Cura-Mallín Formation (Charrier et al. 2002b). The outcrops
of the Farellones Formation end at 35°S; to the east it can be
related with the Contreras Formation and the infill deposits of
the Alto Tunuyán Foreland Basin on the east side of the
Principal Cordillera (Giambiagi 1999; Giambiagi et al. 2001).
South of 35°S, the Pleistocene andesitic and basaltic Cola de
Zorro Formation overlies the Abanico Formation (González &
Vergara 1962).

The aforementioned deformation of the basin fill produced
syntectonic deposits and progressive unconformities. It is
unlikely that deformation simultaneously affected the whole
region, while depositional processes (including volcanism)
probably continued in the basin during the contractional
episode, as suggested by the lack of a clear stratigraphic ‘break’
between the Abanico and the Farellones formations in some
areas. In other parts of the basin, however, localized folding
and erosion formed more than one angular unconformity, as
in the Farellones region, east of Santiago. Deformation along
the region between 31°30′S and 34°00′S occurred between
25.2 ± 0.2 Ma and 16.1 ± 0.5 Ma (Charríer et al. 2002c). In
Miocene times, during the inversion of the Abanico Exten-
sional Basin, movement along the San Ramón Fault that con-
stitutes the boundary between the Central Depression and the
Principal Cordillera, caused west-vergent thrusting of the
Abanico Formation deposits over the Central Depression. This
situation suggests that the mechanism that formed the basin
constituting the Central Depression is similar to that of a
foreland basin and that the geometry of the sedimentary infill
forms a westward-thinning wedge.

With regard to intrusive igneous activity, Miocene plutons are
scattered across the Principal Cordillera (Aguirre 1960;
González & Vergara 1962; Thiele 1980; Charrier 1981b; Kurtz
et al. 1997; Maksaev et al. 2003). They are associated with
supergiant late Miocene to Pliocene porphyry Cu-Mo ore
bodies such as Río Blanco–Los Bronces and El Teniente, devel-
oped within hydrothermal alteration zones linked to multi-
phase stocks, breccia pipes and diatreme structures in rocks
of the Abanico Formation (Camus 1975, 2003; Cuadra 1986;
Serrano et al. 1996; Vivallo et al. 1999; Skewes et al. 2002;
Maksaev et al. 2004).

Scattered Plio-Pleistocene volcanic activity on the western
Principal Cordillera has been reported for areas located next
to the El Teniente ore deposit at 34°S (Camus 1977; Charríer &
Munizaga 1979; Charríer 1981b; Cuadra 1986; Godoy et al.
1994; Gómez 2001), Sierras de Bellavista at 34°45′S (Kholo
1960; Vergara 1969; Charrier 1973b; Malbran 1986), and possi-
bly also at 35°S (Corona del Fraile Formation; González &
Vergara 1962). These areas form a north–south alignment
suggesting a tectonic control for this activity. Finally, the volca-
noes of the present-day magmatic arc lie east of the eastern
outcrops of the Abanico Formation, covering Mesozoic units
and forming the northern part of the Southern Volcanic Zone
(López-Escobar et al. 1995a). From north to south the most
important of these volcanoes are named Tupungato, San José,
Maipo and Maipo caldera, Tinguiririca, Planchón–Peteroa,
Descabezado Grande, Cerro Azul, Descabezado Chico,
San Pedro, Longavi, Chillán, Lonquimay and Llaima (see
Chapter 5).

South-central Chile (Concepción–Longquimay region, 37°S to
39°S)

West side of the Coastal Cordillera (Arauco). The best exposed
series of continental shelf deposits is located in the Arauco Basin
(Mordojovic 1981) (Fig. 3.52) where they comprise Late Cretaceous (Quirquihue Formation), Early Palaeogene (Concepción Group) (see previous stage) and Neogene marine
and continental sequences lying unconformably on Late
Palaeozoic metamorphic rocks (Fig. 3.36). No deposits of late
Palaeogene (Oligocene) age have been reported in this region,
probably because of the low eustatic stand at that time. The fol-
lowing Neogene to Pliocene formations have been recognized:
Ranquil (Miocene, marine sequence), and Tubul (Pliocene, mar-
ine sequence) (Muñoz Cristi 1946, 1973; Wenzel et al. 1975;
Pineda 1983a, b; Arévalo 1984). The Ranquil and Tubul forma-
tions can be correlated to the north with the Navidad and
Coquimbo formations respectively. The evolution of this out-
standing sedimentary sequence, which contains hydrocarbon
and important coal reserves, was characterized by an alter-
nation of transgressive and regressive episodes, controlled by
eustatic changes, local subsidence and uplift of tectonic blocks
(Wenzel et al. 1975; Pineda 1983a, b), and general uplift of the
Andean range. The interaction of these factors resulted in
an alternatively prograding and retrograding continental
sedimentary wedge. The Arauco–Concepción area today is
actively uplifting and the Arauco Peninsula is probably
growing towards the north (Melnick et al. 2003a). Palaeozoic
inherited crustal-scale structures seem to have controlled the
behaviour of these basins since Late Cretaceous times (Echtler
et al. 2003). The Trihueco Formation in particular represents a
period of marine regression and transgression of second-order
duration, during which barrier island complexes developed on a
muddy shelf (Le Roux & Elgueta 1997).

Central Depression. The mid-Tertiary Coastal Magmatic Belt
(Vergara & Munizaga 1974; López-Escobar & Vergara 1997;
Muñoz 1997; Stern et al. 2000) in south central Chile
(36°–43°S) formed when the locus of Andean magmatic activ-
ity was located in the current location of the Central Depression
and western Principal Cordillera (Muñoz et al. 2000). This
expansion of the arc domain occurred in conjunction with a
regionally widespread episode of late Oligocene (29 Ma) to
early Miocene (18.8 Ma) extension which thinned the crust
below the proto-Central Valley in south-central Chile and gen-
erated sedimentary basins west, within, and east of the Main
Cordillera (Muñoz et al. 2000). Continuing extension accompa-
inied a transient steepening of subduction angle, as indicated by
westward migration of the volcanic front during the formation of
the mid-Tertiary Coastal Magmatic Belt (Muñoz et al. 2000).
Late early Miocene inversion of the Cenozoic extensional basin (Abanico Basin; Charrier et al. 2005a), which in this region contains the Cura-Mallín and Trapa-Trapa formations, probably also caused the westward thrusting of the western uplifted blocks over the Central Depression.

Principal Cordillera. Along the Principal Cordillera, further south of 36°S, the Cura-Mallín Formation represents the prolongation of the younger part of the Abanico Formation down to 39°S (Niemeyer & Muñoz 1983; Muñoz & Niemeyer 1984; Suárez & Emparan 1995, 1997; Radic et al. 2002) (see Fig. 3.54). Between 36°S and 37°S, the Cura-Mallín Formation consists of a thick series of lavas and volcaniclastic deposits (Río Queuco Member) followed by a predominantly fluvial sedimentary series (the Mallá-Malla Member; Niemeyer & Muñoz 1983; Muñoz & Niemeyer 1984). Between 37°S and 39°S, the volcaniclastic series (termed Guapitrio Member at these latitudes) interfingers to the east with the thick fluvial and lacustrine Río Pedregoso Member of the Cura-Mallín Formation (Suárez & Emparan 1995, 1997; Radic et al. 2002). Between 36°S and 38°S, the Cura-Mallín Formation is conformably overlain by the andesitic and conglomeratic Trapa-Trapa Formation, which in turn is unconformably overlain by the late Miocene–early Pliocene Campanario and the Pliocene–early Pleistocene Cola de Zorro formations (Muñoz & Niemeyer 1984). A nearly coeval stratigraphic series has been recognized in the Andacollo region on the Argentine side of the Andes at 37°S (Jordan et al. 2001), with these authors applying the same formational names as used in Chile. These Argentine deposits unconformably overlie the Early Cenozoic Serie Andesítica, and are overlain by a Late Miocene sedimentary and volcaniclastic unit (the Pichi Neuquén Formation). The Cura-Mallín Formation accumulated in an extensional basin, which was the southern prolongation of the Abanico Extensional Basin which corresponded to an equivalent basin development further south (Fig. 3.55), that became inverted (Elgueta 1990; Vergara et al. 1997; Jordan et al. 2001; Charrier et al. 2002b; Radic et al. 2002) during late Miocene times (Burns & Jordan 1999; Radic et al. 2002). Seismic reflection data from the Cura-Mallín Formation provides clear evidence for contraction along an inverted, basin-bounding normal fault (Burns & Jordan 1999). Prior to this inversion, the depocentre corresponded to two asymmetric and diachronous half-grabens with opposite polarities. One of these half-grabens developed north of 38°S and the other one south of this latitude; both half-grabens were linked by an accommodation zone that coincides with Pliocene volcanic activity (Radic et al. 2002; Croft et al. 2003).

A surprisingly young 40K–40Ar radioisotopic date of 14.5 ± 1.4 Ma was initially obtained for the Cura-Mallín Formation in Chile, between 36°S and 38°S (Drake 1976; see Niemeyer & Muñoz 1983; Muñoz & Niemeyer 1984). In contrast Jordan et al. (2001) reported two 40Ar/39Ar ages of 24.6 ± 1.8 Ma and 22.8 ± 0.7 Ma for deposits of the upper part of a series correlated with the Cura-Mallín Formation exposed in the Andacollo region on the Argentine versant at 37ºS (Jordan et al. 2001) and 36ºS; Eocene Metalqui Pluton; 6, Photolineaments.

In this region, after inversion of the basin, Plio-Pleistocene transtensional activity occurred in the backarc associated with asthenospheric magmatic activity (Folguera et al. 2004; Ramos & Folguera 2005).

Longquimay to Chiloé (39°S to 43°S)

Coastal Cordillera. Cenozoic deposits in Chiloé Island (41–43°30′S) (see Fig. 3.52B) are exposed on both sides of a north–south orientated axis of metamorphic rocks (Cordillera del Pichón) (Fig. 3.56) which in this region contains the Coastal Cordillera west of Osorno (40°45′S) (Antinao et al. 2000). Along the west side of this metamorphic axis, sedimentary (mostly marine) and volcanic sequences of Oligocene to Pliocene ages overlie the Palaeozoic basement. In contrast, along the eastern side of the island there is an extensive cover of mostly glacial and sedimentary marine Quaternary deposits (Muñoz et al. 1999; Sernageomin 2003). The oldest sedimentary unit, which is exposed on the western side of the metamorphic...
ridge, corresponds to the continental Caleta Chonos Beds, and is probably older than late Oligocene (Antiaño et al. 2000). The stratigraphic succession continues with the Chonchi Beds (Quirroz et al. 2003), exposed along the coast next to Chonchi on the eastern side of the island, which contain fossil marine invertebrates and trunks indicating an Oligocene–Miocene age.

The next younger deposits correspond to the early to middle Miocene marine Lacuï Formation (Valenzuela 1982) and the Cucao Beds on the western side (Quirroz et al. 2003), which are correlative with the Chequemó Beds and Santo Domingo Formation exposed further north in the Valdivia–Osorno basin (Duhart et al. 2000). The youngest sedimentary unit along the central western coast is a marine sequence forming a thin and almost continuous coastal belt of subhorizontal deposits, assigned to the Pliocene epoch (Watters & Fleming 1972). Volcanic rocks of late Oligocene to Miocene age, including basalts, basaltic andesites and andesites, are exposed in the northernmost (Ancud Volcanic Complex) and the southwestern (Guapi Quilán Complex) tips of the island. Volcanism was partially synchronous with the marine sedimentation so that, in the north the volcanic rocks are interstratified with the Miocene marine sequences (Quirroz et al. 2003). The east–west segmentation of the island is related to north–south trending normal faults that controlled the uplift of the Palaeozoic basement in the central segment (Piuchén Range). Volcanic activity was controlled by the intersection of such lineaments with NNW-oriented strike-slip faults.

Central Depression. This region corresponds to the ‘lakes region’ in Chile. Here most of the Central Depression is covered by Pleistocene distal volcanic flows and ash-falls, and abundant deposits originated in three Pleistocene pulses of glaciation (Laugenie 1982).

Principal Cordillera and the Liquiñe–Ofqui Fault Zone. The magmatic arc in this region is represented by the Mesozoic–Cenozoic North Patagonian Batholith (NPB) and the present-day Southern Volcanic Zone (SVZ) (composed of two segments: the Central Southern Volcanic Zone and the South Southern Volcanic Zone). The Liquiñe–Ofqui Fault Zone (LOFZ) (Hervé 1976) is the main structural feature in this region (Hervé et al. 1979a; López-Escobar et al. 1995a; Cembrano et al. 1996, 2000; Lavenu & Cembrano 1996; Folguera et al. 2001; Melnick et al. 2002, 2003b; Roseau et al. 2003) (Fig. 3.57). The LOFZ is a trench-linked dextral strike-slip structure (Cembrano & Hervé 1993; Cembrano et al. 1996, 2000) that reaches a length of >1000 km between 40°S and 47°S, and runs near the centre of the NPB, parallel to the present-day volcanic arc (Hervé 1994). The general orientation of this fault zone is NNE–SSW, although at its southern extremity it bends to the SW and reaches the margin of the continent at the Golfo de Penas pull-apart basin immediately south of the Taitao Peninsula (Forsythe & Nelson 1985). The zone exposes both ductile and brittle fault rocks (Hervé 1976; Cembrano 1990; Pankhurst et al. 1992; Cembrano & Hervé 1993; Cembrano et al. 1996, 2000). On a regional scale it is characterized by two main NNE–SSW trending segments joined by NE–SW orientated en echelon lineaments and interpreted as a strike-slip duplex (Cembrano et al. 1996). This fault zone appears to have formed the locus of magmatic activity since Mesozoic times, controlling the emplacement of Neogene syntectonic plutons and the location of most of the Quaternary volcanoes (Hervé 1994) (see Fig. 3.57).

Between 42°S and 46°S there are many outcrops of Neogene pillow basalts in the western border of the Principal Cordillera, the emergent islands of the Central Depression, and the eastern border of the Coastal Cordillera (Silva 2003). According to Silva, the metabasalts were deposited in two stages. During the first stage, between 36 and 17 Ma, oblique subduction inhibited arc magmatism, and gave rise to the LOFZ and to extensional basins in which the basalts were generated to the west of the structure. These basins are asymmetric, having just one border defined by the fault (Hervé 1994; Silva 2003). As a result of the change in orientation of the convergence vector from oblique to orthogonal at 25 Ma during the second stage, between 22 and 13 Ma, the locus of the magmatic activity expanded east of its previous position towards the present day western Principal Cordillera (Muñoz et al. 2000). The eruption of lavas occurred in deeper basins along with deposition of turbiditic sequences and detrital flows (Hervé et al. 1994, 1995). According to the volcanosedimentary record in these basins (Traiguern and Ayacara formations), the LOFZ has been active at least since the Eocene epoch (Hervé 1994).
Quaternary volcanism has produced abundant composite stratovolcanoes and hundreds of minor eruptive centres, with mostly basalts and basaltic andesites being erupted. The minor centres, which can also erupt more intermediate to silicic products, are spatially associated with the main NNE-trending lineaments parallel and with NE-trending lineaments linked to the LOFZ, forming alignments oblique to the overall trend of the volcanic arc. In contrast, the stratovolcanoes developed along NW- or NE-trending alignments (Fig. 3.57), with a preference for the NE alignments possibly reflecting a transpressional tectonic regime resulting from a combination of dextral strike-slip and shortening across the arc (López-Escobar et al. 1995a). According to the same author, the lineaments only facilitated the ascent of magmas, whereas their geochemical features were determined by subcrustal processes. Considering that deformation of the orogen east of the LOFZ is considerably less than further north where this fault zone is not developed, it has been suggested that a considerable portion of the strain induced by subduction in this region has been dissipated by the transcurrent movements (Hervé 1994).

The third Andean stage: summary and discussion

Palaeogeography and tectonic evolution. The regional late Palaeogene landscape in northern Chile following the Incaic tectonic phase at the end of the second Andean stage was dominated by the Incaic Range (or ‘proto-Domeyko Range’), the positive topographic feature formed by the inversion and consequent uplift of the former arc and its associated intra-arc basins) and the eastwardly shifted magmatic arc/intra-arc on the east side of the Incaic Range. Neither the Incaic Range nor the new arc/intra-arc were topographically highly mountainous because the tectonic setting was primarily extensional; the onset of the third phase of Andean tectonic evolution coincided with the end of increasing convergence rate between the oceanic Farallon and South American plates (Fig. 3.31). Thus the arc/intra-arc area formed an extensional basin (known in central Chile as the Abanico Basin) containing active volcanoes erupting magmas with tholeiitic affinities through a thinned continental crust. The thick volcanic, volcaniclastic and sedimentary successions deposited in this basin include, from north to south, the Rio de la Sal Formation, the Doña Ana Group and the Las Tórtolas, Abanico and Cura-Mallín formations. Further north from central Chile, similar deposits were probably deposited as far north as the southern Altiplano, but their exact distribution is obscured by an extensive volcanic cover. In the opposite direction, south of Lonquimay, Abanico-like sediments extend into Argentina. Along the Chilean side of the cordillera, a series of smaller Late Cenozoic basins developed in association with splay of the Liquiñe–Ofqui Fault Zone. The latter lineament not only influence sedimentation but also channelled arc magmatism, thus acting as the major structural control in the south.

In northern Chile it is possible to trace the location of the NNE–SSW trending positive topographic feature resulting from the third compressive tectonic event, which we refer to here as the Incaic Range (or ‘proto-Domeyko Range’), a precursor to the modern Domeyko Range, based on: (1) the location of the Late Cretaceous and Early Palaeogene arc/intra-arc magmatic units and deposits accumulated in the associated extensional basins; (2) the detrital deposits resulting from the erosion of this relief; and (3) the location of the major faults and their fault rocks that participated in the opening and inversion of the extensional basins that characterized the second stage. This range probably formed the watershed at the end of this Andean stage. Further south of 30ºS, i.e. within and south of the flat-slab source of the modern Domeyko Range becomes harder to establish. At these latitudes the detrital deposits on both sides of the range are either no longer exposed or were never laid down, the southern prolongation of the major faults is difficult to establish, and neither the modern Domeyko Range nor the Salars Depression are topographically expressed. However, it is possible to trace its position along the Late Cretaceous and early Palaeogene arc volcanic deposits and intrusive rocks as a NNE–SSW trending strip running southward from the southern Domeyko Range and along the eastern side of the present-day Coastal Cordillera down to around 35ºS (Fig. 3.32). This indicates that the southward prolongation of the Incaic relief across the flat-slab segment was located west of the present-day Principal Cordillera (Fig. 3.33). Therefore, this relief formed in central Chile a palaeogeographic element slightly oblique to the present-day morphostructural configuration of the Andean range. The late Eocene–Oligocene Abanico Extensional Basin (located along the west flank of the present-day Principal Cordillera, at least between La Serena and Lonquimay) and the relief formed after its tectonic inversion, were developed to the east of the Incaic relief (Fig. 3.58). The NNE orientation suggests that the prolongation (north of 30ºS) of the Abanico Basin continued in that direction along the international boundary (mostly west, but probably also east of it) and reached regions presently located east of the Domeyko Range further north in the Salars (preandean) Depression and the Puma or southern Altiplano below the Plio-Quaternary volcanic cover.

In northern Chile the Incaic Range was deeply incised in late Eocene to early Oligocene times, with resulting sedimentation to both east and west. Some of the materials transported to the west probably reached the ocean to form part of the sediments presently accumulated in the continental platform. In other areas, the positive relief of the Coastal Cordillera blocked sediment transport. In the Arica region, for example, by early Oligocene times sediments of the Azapa Formation were already onlapping Mesozoic rocks of the Coastal Cordillera. Those transported to the east accumulated in basins which possibly corresponded to the northward prolongation of the Abanico Basin (possibly the Eocene–Oligocene Calama Formation in the Calama Basin), at this latitude located below the extensive and thick volcanic cover of the Altiplano and most of it probably in Bolivian territory, e.g. the fine-grained detrital sediments of the Neogene Potocito and San Vicente formations (see Rochat 2000; Horton et al. 2001).

By Oligocene to early Miocene times continued erosion of the remaining Incaic Range in the northern Chile forearc region was coeval with sedimentation in the Central Depression (e.g. Azapa and Sichal formations). Where rivers could reach the ocean, sedimentation occurred on the continental platform and in structurally controlled embayments or basins located along the western border of the Coastal Cordillera, as seen in the Arauco area in south-central Chile. Miocene activity along the Atacama Fault Zone uplifted the western part of the Coastal Cordillera in the north to create closed sedimentary basins to the east. Coarse sediments began to cover the western flank of the Domeyko Range (Precordillera of the Arica–Iquique and Antofagasta–Chañaral regions), infilling the deep drainage system previously generated to create units such as the Altos de Pica Formation and the Atacama Gravels. Mainly rhyolitic explosive activity, particularly active in the Arica area (Oxaya Fm), formed intercalations within these gravelly deposits. Deposition on the western flank of the Precordillera lasted until the beginning of incision of the gravel deposits, favoured by renewed access of the rivers to the ocean, availability of meltwater from the glaciers formed at high altitude on the uplifted surface, and tilting of the forearc at c. 10 Ma, which was caused by eastward thrusting in the Andean foreland.
A reversion to an essentially compressive tectonic setting in early Miocene times caused regional uplift of both the remnant Incaic Range and the arc/intra-arc area, east of the latter, where continuing magmatism through thickening crust produced thick, calcalkaline successions such as the Maricunga Belt, Vallecito, Farellones and Trapa-Trapa formations. Contraction within the Domeyko Range has thrust it both east and west to form a regional 'pop-up' structure. Further east, strong erosion of the uplifting high Andes (representing the inverted former basin of the still active arc/inter-arc area) has supplied abundant sediments both west towards the coast (e.g. Navidad and Ranquil formations) and east to the Alto Tunuyán foreland basin. By late middle Miocene times, renewed passage of rivers to the ocean in northern Chile promoted incision of the gravelly deposits on the western side of the Domeyko Range. This erosive episode was enhanced by late Miocene westward tilting of the whole forearc area. Finally, a pulse of particularly rapid uplift of the Andean Range has been continuing since 10 Ma, during which the arc has shifted further east, erupting through a greatly thickened continental crust. This arc includes abundant active Quaternary volcanoes and giant calderas located mainly along the crest of the mountain range and developed along three zones in the Andean region under consideration: the Central Volcanic Zone (extending from 14°S, in southern Peru to 27°S), the Southern Volcanic Zone (33°S to 47°S) and the Austral Volcanic Zone (south of 52°S) (see Stern 2004). With regard to the forearc region, although contraction continues in the east, the western side is currently under extension. The recent uplift of the Altiplano has favoured the growth of glaciers, meltwaters from which have promoted further incision of the Cenozoic gravels and renewed access of the rivers to the sea.

Andean oroclines. Oroclinal bends are characteristic features of subduction margins where convergence passes gradually from essentially normal (orthogonal) to oblique, as seen elsewhere in the world in places such as Sumatra, the Aleutian Islands, Bolivia and Patagonia. The mechanism by which the continental margin or the island-arc accommodates to the gradually varying stress tensor differs in different regions. In Sumatra, for example, much of the strain is accommodated by movement along a trench-parallel major fault located along the magmatic arc (Huchon & Le Pichon 1984), whereas in the Aleutian island arc block rotation seems to be the accommodation mechanism (Ryan & Scholl 1989).

Along the oroclines of the Central (Bolivia) and Southern (Patagonia) Andes (Fig. 3.1b), there is a gradual change from normal to oblique convergence. The Bolivian Orocline shows obliquity increasing north and southward of a line of symmetry (Gephart line; Gephart 1994), whereas in the Patagonian Orocline the obliquity of convergence increases gradually southward passing from normal to parallel to the continental margin through an angle of c. 90°. Isacks (1988) suggested that the origin of the Bolivian Orocline is directly related to the development of the Altiplano and that the north to south variations of Altiplano width were caused by differential shortening during the plateau uplift. According to Isacks, a slight initial bend of the continental margin has been increased by subsequent Neogene shortening to form the oroclinal bend. Palaeomagnetic studies confirm that rotation in the Altiplano has occurred during late Miocene–Pliocene times (Roperch et al. 2000). Palaeomagnetic studies indicate, however, that in the forearc in southern Peru (Roperch et al. 2006) and northern Chile (Arriagada 2003; Arriagada et al. 2003) counter clockwise and clockwise rotation have occurred, respectively, in Early Neogene times and that no significant rotation occurred in Neogene times (Fig. 3.59). Thus the Bolivian Orocline appears to be a rather old feature directly associated with pre-Neogene (Incaic) shortening, and rotation in the Altiplano probably occurred during uplift in Late Neogene time. Absence of post-Miocene rotation across the area suggests that other
mechanisms might have been operating since early Miocene times in that region.

Palaeo-magnetic studies of the Patagonian Orocline have established that this feature is, at least in part, a secondary bend due to counterclockwise tectonic rotation (Dalziel et al. 1973; Burns et al. 1987; Cunningham et al. 1991) induced by left-lateral shearing between the South American and the Scotia-Antarctic plates (Carey 1958; Hamilton 1964; Winslow 1982). More recent studies by Cunningham (1993) in the Beagle Channel region demonstrated the existence of Cenozoic brittle and Mesozoic ductile fabrics containing left-lateral kinematic indicators and provided abundant additional structural and geomorphological information supporting left-lateral slip (Fig. 3.60). This evidence is consistent with seismic activity along the Magallanes–Fagnano Fault Zone, located along the Strait of Magellan and the Fagnano Lake in Tierra del Fuego, with left-lateral strike-slip mechanisms (Fuenzalida 1974) and both vertical and horizontal displacements in the field (Schwartz et al. 2002). Cunningham (1993) proposed that the break that initiated formation of the Patagonian Orocline occurred in the weak zone represented by the ‘Rocas Verdes Marginal Basin’ (discussed below) and that with continued opening of the South Atlantic Ocean and westward motion of South America relative to Antarctica, transpression continued and new strike-slip faults developed through time as the transpression migrated northward. Compared to other strain accommodation mechanisms mentioned above for the Sumatra and Aleutian arcs, this represents another model for strain accommodation along a margin along which the plate convergence angle increases gradually, in this case influenced by a fossil obduction zone or terrane suture.

Plate tectonic controls. Andean segmentation is controlled by the subduction of the Juan Fernández Ridge below the continental margin between c. 27°S and 33°S (Fig. 3.2). The Juan Fernández Ridge is part of the Nazca Plate and, accordingly, its convergence vector is the same as that of the oceanic plate, that is, N76°E at 85 mm/year (Yáñez et al. 2001). This convergence vector was initiated 25 million years ago as a consequence of the major plate readjustment after break-up of the Farallon Plate. The Juan Fernández aseismic ridge originates at the stationary Juan Fernández hot-spot, and its segmented shape is the result of previous variations of the movement vector of the oceanic plate during Late Cenozoic times. According to the reconstruction of the Juan Fernández Ridge by Yáñez et al. (2001, 2002), the younger, N76°E-orientated segment, began to form 25 million years ago, and intersected the continental margin c. 12 million years ago at 33° S (Valparaiso). Ridge subduction since then has caused diminution of the subduction angle (flat-slab subduction) and considerable thickness increase of the lithosphere (e.g. Pardo et al. 2002a). This situation explains the characteristic morphological and geological features of the flat-slab segment of the Argentine–Chilean Andes: absence of the Central Depression, a gap in active arc volcanism in the main Andes, and eastward shift of the deformation front and volcanic activity in the Andean foreland (Sierras Pampeanas) (e.g. Jordan et al. 1983a; Yáñez et al. 2002; Ramos et al. 2002) (Fig. 3.61). Apart from the complication introduced by the subduction of the Juan Fernández Ridge, third Andean stage tectonic evolution and the present-day morphostructural units result from a long-lasting episode of Chilean-type subduction, i.e. an essentially shallow plate dip and a high degree of interplate coupling, accompanied by the occurrence of great interplate earthquakes up to M8.

The development of short-lived phases of tectonic shortening in the orogen has generally been attributed to episodes of increased relative plate motion. Similarly, periods of tectonic extension have been attributed to episodes of low convergence.
rate. In this region of the Andes, episodes of high convergence rates and plate coupling in Cenozoic times correlate with contractional deformation episodes in the orogen (e.g. the Eocene Incaic tectonic phase separating the second from the third stages of Andean evolution, and the Quechua tectonic phase in late Miocene times) and have generally been followed by an eastward shift of the magmatic arc. In contrast, episodes of low convergence rates (during which oblique convergence and reorganizations of the plate system occurred) have coincided with periods of tectonic extension and intense magmatic activity together with deposition in subsiding basins. A more detailed analysis shows these extensional and compressional phases to correspond more to periods of decreasing and increasing convergence rates, respectively, rather than to peaks of minimum and maximum rates (Charrier et al. 2002b). Slow convergence along the continental margin last occurred during the Oligocene epoch (Pardo-Casas & Molnar 1987) and coincided with extensional basin development (involving thick volcanic, volcaniclastic and sedimentary deposition), illustrating this link between subduction and deformation in the upper crust along the continental margin (Charrier et al. 1994c, 1996, 1997, 1999; Godoy & Lara 1994; Jordan et al. 2001).

Although contraction and uplift are apparently occurring presently in the eastern forearc and arc domain, evidence for east–west extensional conditions, at least in the external, more brittle continental crust of the Coastal Cordillera in northern Chile, demonstrates a more complex distribution of strain across the orogen. This situation has probably resulted from a combination of tectonic erosion processes, accumulation of materials in the subduction zone (underplating), and coseismic deformation associated with subduction earthquakes (Hartley et al. 2000; González et al. 2003). Local north–south tectonic shortening in the Coastal Cordillera in northern Chile has been related either to the development of the Bolivian Orocline or to subduction of oceanic crust in a curved continental margin (González et al. 2003).

Andean uplift. Several attempts have been made to estimate when the Andean range acquired its present altitude and how rapid this process was. Available palaeo-botanical information from northernmost Chile and southwestern Bolivia indicates that the Andean Range in early to early late Miocene times had a moderately low elevation (Charrier et al. 1994a; Gregory-Wodzicky et al. 1998; Gregory-Wodzicky 2000, 2002). According to the latter author, there has been a 2000–3000-m uplift of the Altiplano to its current elevation (average of 3800–4000 m) since 10 Ma, representing uplift rates of around 0.3–0.25 mm/year. During this uplift phase, the deep incision of the gravelly deposits on the western flank of the Altiplano in northern Chile at the latitude of Pisagua–Iquique (Pinto et al. 2004a; Farías et al. 2005a) and of the Domeyko Range at the latitude of Chañaral (Riquelme 2003) is the consequence of uplift of the eastern forearc region after 8–9 Ma, with westward tilting of the Andean forearc prior to deep erosional incision.

Miocene magmatism in central Chile developed during an episode of crustal thickening and tectonic uplift associated with increasing convergence rates. Attempts to determine exhumation and uplift on Miocene plutons in the Principal Cordillera between 33°S and 35°S have been made by Kurtz et al. (1997)
Fig. 3.61. Morphological features associated with the flat-slab subduction segment (zone), distribution of volcanoes (triangles) and contours of depth of Wadati–Benioff plane and (for location along the Chilean continental margin, see Fig. 3.1b). Central Depression (CD) is developed only north and south of the flat-slab zone, while the Sierras Pampeanas in the Andean foreland lie within it. Based on Cahill & Isacks (1992) and Ramos et al. (2002). Numbers indicate depth in kilometres of the Wadati–Benioff plane. MB, Maricunga Belt; EIB, El Indio Belt.

Fig. 3.62. Correlation between periods of decreasing and increasing convergence rates, according to Pardo-Casas & Molnar (1987) (discontinuous line) and Somoza (1998) (continuous line), and periods of extension and compression in the central part of the Southern Andes, based on Charrier et al. (2002b). A, Period of extensional basin development; B, period of tectonic inversion of the basin.

and Maksae et al. (2003). Kurtz et al. (1997) modelled the exhumation of plutons on the base of 40Ar/39Ar ages on different minerals, obtaining rather high exhumation rates of 0.55 mm/year between 19.6 and 16.2 Ma for the La Obra pluton, and 3 mm/year between 8.4 and 7.7 Ma for the Nacientes de Cortaderal pluton in the eastern forearc (see Fig. 3.63A). This modelling was based on several assumptions about the closure temperatures of the dated minerals, intrusion depth, and dimensions of the plutons. Recent fission track dating onapatites from Miocene plutons in the same region of the Principal Cordillera yielded cooling ages around 4 Ma (5.6 to 3.7 Ma), which are significantly younger than 40Ar/39Ar ages in the same bodies, suggesting an episode of accelerated denudation at the end of Miocene and beginning of Pliocene times (Maksae et al. 2003). According to these authors, the cooling model for plutons in the eastern forearc is compatible with exhumation rates of 1.41–1.70 mm/year between 5.6 Ma and 3.5 Ma (Santa Rosa de Rengo pluton), 2.2–2.6 mm/year between 3.7 Ma and 2.8 Ma, and 0.53–0.64 mm/year from 2.8 Ma to Present (Nacientes del Cortaderal pluton). Notwithstanding the possible bias introduced by the assumptions considered by Kurtz et al. (1997), both studies support the idea of rapid uplift of the Andean range after late Miocene times (c. 10 Ma) (Fig. 3.63B).

According to Maksae et al. (2003), the denudation episode between 5.6 and 3.7 Ma (probably reflecting tectonic uplift of the cordillera related to crustal shortening) overlaps with the ages of formation of the Cu–Mo megadeposits of Rio Blanco–Los Bronces (Serrano et al. 1996) and El Teniente (Skewes et al. 2002), suggesting that the ore bodies were formed during this uplift episode (see Fig. 3.63A for location of ore deposits). If denudation is the cause for sudden Cu-rich exsolution of magmatic fluids and brecciation at high crustal levels, as has
been proposed by previous authors, then intense mass removal of material would have been needed to induce the necessary rapid decompression (Maksaev et al. 2003). In the southern central Andes, the available evidence suggests uplift is young and continuing, with currently intense shallow seismic activity concentrated along major faults (Barrientos et al. 2004a; Charrier et al. 2004, 2005a; Farias et al. 2005a; Fock 2005; Fock et al. 2005).

The late Neogene age (after 10–9 Ma) deduced by the studies for uplift in the eastern forearc and arc regions in northern and central Chile coincides fairly well with the middle Miocene age deduced for uplift of the marine platform in the Mejillones Peninsula (see Marquardt et al. 2003). This time coincidence stresses the regional extent of the Andean uplift phase.

**Evolution in southern Chile (Patagonia), south of 42°S**

The evolution of the Patagonian Cordillera, located south of c. 42°S, occurred on the west side of the Patagonia Terrane (Ramos 1984, 1988b; Ramos et al. 1986) (that forms the eastern, non-orogenic side of Patagonia south of c. 39°S) and was initially associated with the development of the Patagonian backarc basin in latest Jurassic to late Early Cretaceous times. Later the area became a foreland basin, forming a separate entity from coeval backarc basins developed further north (i.e. Neuquén Basin, Fig. 3.29). The NNW–SSE orientation deduced for the main structural grain of the continental margin during the early stages of the Andean tectonic cycle (and for previous cycles) is oblique to the present-day coastline. This suggests that the Patagonian basin was developed over...
terran(s) located further SW than those on which the Mendoza–Neuquén Basin was developed. Although some episodes in the evolution of the Magallanes Basin can be correlated with similar episodes in the basins further north, the following important magmatic and tectonic events permit its differentiation from regions located further north: (1) a general trend that culminated with the opening of the Atlantic Ocean, affecting the southern parts of South America and Africa, as well as part of Antarctica. This event produced extensional basins in the Patagonian region. The NNW–SSE orientation of these basins suggests control by pre-existing fractures probably formed during the ‘Permo-Triassic’ or pre-Andean extensional event. In association with crustal extension, an extensional fossil volcanic episode of calcalkaline affinities and I-type rhyolitic– dacitic plutonism derived from crustal anatexis affected the entire Patagonian region (Feruglio 1949; Suárez & Pettigrew 1976; Bruhn et al. 1978; Riccardi & Rullier 1980; Baker et al. 1981; Allen 1982; Fuenzalida 1984; Niemeyer et al. 1984; Gust et al. 1985; Hanson & Wilson 1991). These deposits filled the extensional basins developed previously and interfinger with thick successions of sedimentary breccias, with clasts of mica schists and quartzites accumulated at the base of the cliffs formed in response to extensional faulting (e.g. the Poca Esperanza Formation; Prieto 1993). The volcanic deposits accumulated in marine environments contain macro- and micro-fossils that indicate an age between Oxfordian and Tithonian–Berriasian (Feruglio 1949; Sigal et al. 1970; Natland et al. 1974; Riccardi & Rullier 1980; Fuenzalida & Covachevic 1988; Covachevic et al. 1994). In the subsurface towards the foreland, the existence of similar basins forming half-grabens has also been detected, though probably here filled only by continental deposits (Uliana & Biddle 1988; Moraga 1996). South of 50ºS and in the western regions, bimodal volcanic activity in marine environments produced abundant silicic lavas and pyroclastics associated with ophiolitic pillow lavas and tuffs, sheeted dykes and gabbros, which form elongated, north–south orientated swaths strongly affected by east-vergent thrusts (Dalziel et al. 1974; Stern et al. 1976a; De Wit & Stern 1978; Calderón 2006). Basalt geochemistry indicates typical oceanic tholeiites to transitional-type basalts and associated differentiates (Stern et al. 1976a). These rocks form the Sarmiento (between 51ºS and 52ºS) and Tortuga (55ºS) ophiolite complexes (Allen 1982; Godoy 1978). The new tectonic setting leading to bimodal magmatism localized along the axis of the present-day Patagonian Cordillera has been interpreted as the development of a marginal backarc basin (Dalziel et al. 1974; Dalziel 1981) or a later aborted branch (aulacogen) of the rift system formed during the opening of the Atlantic Ocean (Godoy 1978). This basin is commonly known as the ‘Rocas Verdes Marginal Basin’. Recent magmatic and detrital zircon SHRIMP U–Pb analyses yielded ages of 152–147 Ma for the initiation of the ‘rocas verdes’ extensional basin (Calderón 2006), which coincides with the age determined on the basis of the fossiliferous content further north (see above). The components of this basin underwent ocean-floor-type metamorphism (Elthon & Stern 1978).

**Second stage: Latest Jurassic to mid-Cretaceous.** After the major regional extensional events that led to the development of the ‘Complejo Porfríctico de la Patagonia’ and the more localized extension associated with the Rocas Verdes Basin (rift phase), thermally driven subsidence began in latest Jurassic and/or earliest Cretaceous times (Harambour & Soffia 1988; Volcanic activity produced thick deposits included under the general term of ‘Complejo Porfríctico de la Patagonia’ (Quensel 1913). These rocks have been assigned to the Ibáñez Formation or Group between 44ºS and 49ºS, in the Palena (De la Cruz et al. 1996) and Aisén regions (Niemeyer et al. 1984; De la Cruz et al. 2003, 2004; Pankhurst et al. 1998, 2000; Calderón 2006) and has been assigned to the Chon Aike Terrane(s) located further SW than those on which the Mendoza–Neuquén Basin was developed. Although some episodes in the evolution of the Magallanes Basin can be correlated with similar episodes in the basins further north, the following important magmatic and tectonic events permit its differentiation from regions located further north: (1) a general trend that culminated with the opening of the Atlantic Ocean, affecting the southern parts of South America and Africa, as well as part of Antarctica. This event produced extensional basins in the Patagonian region. The NNW–SSE orientation of these basins suggests control by pre-existing fractures probably formed during the ‘Permo-Triassic’ or pre-Andean extensional event. In association with crustal extension, an extensional fossil volcanic episode of calcalkaline affinities and I-type rhyolitic–dacitic plutonism derived from crustal anatexis affected the entire Patagonian region (Feruglio 1949; Suárez & Pettigrew 1976; Bruhn et al. 1978; Riccardi & Rullier 1980; Baker et al. 1981; Allen 1982; Fuenzalida 1984; Niemeyer et al. 1984; Gust et al. 1985; Uliana et al. 1986; Mpodozis & Kay 1990; De la Cruz et al. 1996, 2003, 2004; Pankhurst et al. 1998, 2000; Calderón 2006) and has been assigned to the Chon Aike Large Magmatic Province (Mpodozis & Kay 1990) (Fig. 3.66).
Fig. 3.64. Digital elevation model of the Chilean Andes south of 42°S with indication of the main geographic, tectonic and morphostructural features. Abbreviations: AF, Andean Foreland; LOFZ, Lique–Ofqui Fault Zone; MFZ, Magallanes Fault Zone (main trace); NIF, northern ice-field; NPM, North-Patagonia Massif; PA, Patagonian Archipelago; PC, Patagonian Cordillera; SIF, southern ice-field; South-Patagonia Massif.
* Complejo Porfirítico de la Patagonia

Fig. 3.65. Stages of geological evolution in Patagonia, stratigraphic successions exposed in surface in the Palena, Aisén and Magallanes regions (Última Esperanza and Riesco Island sections and Springhill Platform) and tectonic interpretation, based on several authors (see text).
Soffia & Harambour 1989; Skarmeta & Castelli 1997; Mella 2001). At this time, magmatic calcalkaline activity on the westernmost edge of the continent was beginning, and therefore the subsident basin was now located in a backarc position (Fig. 3.67). A generalized marine ingression occurred along this basin (floored by continental crust north of c. 50°S and by oceanic crust south of that latitude), reaching from the Palena region at 43°S in northern Patagonia to the western tip of Tierra del Fuego, and occupied areas presently located in the Patagonian Cordillera and internal Andean foreland. This latest Jurassic–earliest Cretaceous marine ingression occurred essentially at the same time as the marine ingression in the backarc basin at the beginning of the second substage of the first stage of Andean (Tithonian) evolution in northern and central Chile.

Magmatic arc activity generated the Patagonian Batholith (Figs 3.67 & 3.68), as well as andesitic lavas and volcaniclastic deposits presently exposed in the westernmost successions of the backarc basin (Hervé et al. 1984; Niemeyer et al. 1984; Soffia & Harambour 1989; Mpodozis & Ramos 1989; Bell & Suárez 1997).

In the Palena region (43–44°S), the backarc basin deposits consist of several marine formations of Tithonian to Hauterivian age (Fuenzalida 1968) which are included in the

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**Fig. 3.66.** Distribution of the deposits of the first stage of Patagonian evolution. Key: 1, Complejo Porfirítico de la Patagonia (Ibañez Group, Quemado Group, Tobífera Formation) or Chon Aike magmatic province (see Mpodozis & Kay 1990). 2, Ophiolitic suite of the 'Rocas Verdes Basin', south of 50°S. 3, Western limit of exposures. The Complejo Porfirítico de la Patagonia most probably reached the westernmost regions of present-day Patagonia, i.e. Seno Arcabuz Shear Zone (Hervé & Fanning 2003).
De la Cruz et al. (1996) proposed a different stratigraphic organization for these deposits, using the same formation names known for equivalent deposits in the Aisén region further south. The stratigraphic succession proposed by these authors is Toqui Formation (Tithonian–Berriasian), Cerro Díaz–Monte Palena Volcanic Complex (Post-Berriasian–Valanginian?), and the Katterfeld and Apeleg formations (Hauterivian); this corresponds to a transgression–regression cycle. These deposits are overlain by a thick continental volcanic succession of andesites, tuffs and sediments assigned to the Divisadero Group (Hall & Lapido 1980), which forms very extensive outcrops in northern Patagonia. The Divisadero Group, also exposed in the Aisén region, includes the Arroyo Pedregoso and the overlying Cordón de las Tobas formations (De la Cruz et al. 1996) (Fig. 3.65). The Arroyo Pedregoso Formation overlies the Hauterivian Apeleg Formation, although the contact between both stratigraphic units has not been observed (De la Cruz et al. 1996). In this region, the lower Divisadero Group consists of an essentially volcanic succession of dacitic ignimbrites, and dacitic and andesitic lavas intruded by andesitic hypabyssal bodies (Arroyo Pedregoso Formation), and in its upper part, of tuffs and andesitic lavas with volcaniclastic intercalations (Cordón de las Tobas Formation). In the Palena region, the Divisadero Group is unconformably overlain by the La Casita Formation (Thiele et al. 1978), which comprises a well stratified succession of shallow marine sandstones containing invertebrates of Miocene age. The La Cascada Formation has been correlated with the Guadal Formation in the Aisén region south of General Carrera Lake (De la Cruz et al. 1996). Recent magmatic arc activity in this region is represented by the Melimoyu Volcano.

In the Aisén region (44–46°S), the backarc Aisén Basin (=Río Mayo Embayment) contains the Colcaique Group (De la Cruz et al. 2003, 2004), formerly Coyhaique Formation (see Cecioni & Charrier 1974; Skarmeta 1976) (Fig. 3.65). This group comprises the three following marine formations, from bottom to top (Suárez et al. 1996; Bell & Suárez 1997): Toqui (the Códidiano or Tres Lagunas Formation in Argentina; Ploszkiewicz & Ramos 1978), Katterfeld and Apeleg. The Toqui Formation is the product of a marine transgression over subaerial rocks of the Ibáñez Group and consists of calcareous sandstones containing invertebrates of Miocene age. The La Cascada Formation has been correlated with the Guadal Formation in the Aisén region south of General Carrera Lake (De la Cruz et al. 1996). Recent magmatic arc activity in this region is represented by the Melimoyu Volcano.

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In the Última Esperanza section, the stratigraphic succession consists of the following units that overlie the Late Palaeozoic Staines Metamorphic Complex (Forsythe & Allen 1980); the synextensional breccias of the Poca Esperanza Formation (Prieto 1993) (further south, Complejo Brechoso Basal) and the El Quemado Group. These units are overlain by transgressive deposits (e.g. the Tithonian–Berriasian Calcareaeous Sandstones of the Bellota Creek; Cortés 1965), which have been correlated with the Springhill Formation in the Springhill Platform (Harmourn & Soffia 1988; Soffia & Harmourn 1989). These transgressive deposits are overlain in turn by the Early Cretaceous Ercezano Formation (Cecioni 1955a, 1956, 1957a, 1958) or Zapata Formation (Katz 1963, 1964) which comprises 1000–2000-m-thick deep water black shale deposits with a few thin greywacke intercalations in its upper portion. These deposits have also been recognized in western Argentinian territory where they correspond to the Río Mayer Formation (Hatcher 1897; Riccardi 1971; Riccardi & Rolleri 1980).

In the Riesco Island section, the lower deposits covering the siliceous volcanic rocks in the Patagonian Cordillera (Seno Rodriguez Formation; Cecioni 1955a, 1958) comprise the Sutherland and Ercezano formations (Cecioni 1956) (Fig. 3.65). The Sutherland deposits consist of a 360-m-thick succession of conglomerates and sandstones deposited in a littoral environment with calcareous siltstone intercalations containing marine fauna of Tithonian age. This formation has been correlated in the subsurface to the east with the oil-productive Tithonian–Berriasian Springhill Formation (Cecioni 1955a; Riccardi 1976, 1977; Riccardi & Rolleri 1980) in the Springhill Platform. According to Cecioni (1955a, 1957a, 1958), the Ercezano Formation consists of a thick black shale succession with greywacke intercalations at its upper part. This formation has been more recently subdivided into two members (Prieto 1993; Mella 2001), the lower member corresponding to low-energy deposits accumulated in a deep-water environment (2000 m deep; Biddle et al. 1986) in Berriasian to Aptian times, and the upper (Ercezano) member to the greywacke-bearing part of Cecioni’s (1956) definition. The lower member is correlated to the east with the Favarella Beds known from drillings in the Springhill Platform (Fig. 3.65). The upper member will be considered in the next stage of evolution (see below). In Aptian–Albian times, the entire backarc area was affected by a tectonic pulse that initiated basin inversion. Contractual deformation, uplift and erosion began in the arc and internal backarc, causing closure (obduction) of the ocean floor south...
Fig. 3.67. Second stage of Andean evolution in Patagonia. Distribution of the Aisén and Magallanes basins and the Patagonian Batholith. Connection between the two basins is uncertain because of lack of outcrops. Última Esperanza section (A) Riesco Island section (B) correspond to the regions described in text. Figure is based on González et al. (1965), Bell & Suárez (1997) and Sernageomin (2003).
Fig. 3.68. Distribution of the Patagonian Batholith and relative position of the deposits pertaining to the first and second stages of Andean evolution in the Aisén and Magallanes backarc basins (see Figs 3.66 & 3.67) and in pull-apart basins associated with the Liquiñe–Ofqui Fault Zone (LOFZ) (see Figs 3.57 & 3.64). Key: 1, Patagonian Batholith; 2, deposits of the first stage of Andean evolution in Patagonia; 3, deposits of the second stage; 4, main traces of the Liquiñe–Ofqui Fault Zone.
of 50ºS (Dalziel & Cortés 1972; Bruhn 1979; Dalziel and Palmer 1979; Nelson et al. 1980; Harambour & Soffia 1988; Soffia & Harambour 1989; Farfán 1994). Continued foreland-directed contraction caused thick-skinned thrusting of the arc over the internal backarc regions along reactivated normal faults, and induced both development of a foreland basin (‘Magallanes foredeep’; Cecioni 1958) as well as thin-skinned thrusting on the previous external regions of the backarc and in the foreland (Patagonian thrust-fold belt; e.g. Ramos 1989; Alvarez-Marron et al. 1993; Farfán 1994). Whereas this contractional deformation phase was accompanied by marine regression from the forearc basin in the Palena and Aisén areas, south of 50ºS marine conditions still prevailed, although facies changes record the differing tectonic regime.

**Third Stage: Late Early Cretaceous to Present.** We now analyse the late Early Cretaceous to Present evolution in the Patagonian region in the three regions considered above – Palena, Aisén and Magallanes – with once again separate discussion for the sections across Ultima Esperanza and Riesco Island (Fig. 3.67).

In the Palena region deposits of this age have not been reported. In the Aisén region, Cenozoic outcrops are exposed west and east of the Patagonian Cordillera. On the west side, exposures of rocks belonging to the third stage of Andean evolution have been mentioned for the Lonquimay to Chiloé region (39–43ºS) in the third stage of Andean evolution and correspond to the Neogene volcanic (essentially basaltic) and sedimentary deposits accumulated in extensional basins associated with the Liquiñe–Ofqui Fault Zone (see Fig. 3.57). These extend as far south as the Taitao Peninsula (47ºS) in the Aisén region, and include such units as the Traiguén and Ayacara formations (Silva 2003). Further west, on the western side of the Chonos Archipelago, Pliocene external platform marine deposits exposed in Guamblín Island (see Fig. 3.57) can be correlated with the Tubul Formation known further north in the western side of the Coastal Cordillera of the Arauco region (Frassinetti & Covacevich 1995) (Fig. 3.36).

Cenozoic outcrops located east of the Patagonian Cordillera occur north and south of General Carrera Lake. Those north of the lake are calcalkaline and alkaline andesites, basalts, dacites and rhyolites, and include minor intrusive bodies of possible Late Cretaceous (Skarmeta 1978; Niemeyer et al. 1984; Suárez et al. 1996; De la Cruz et al. 2003) and/or late Oligocene age (Morata et al. 2005), as well as continental clastic deposits of the Galera Formation overlying the Divisadero Group (Niemeyer et al. 1984; Skarmeta 1978) (Fig. 3.65). These deposits, exposed NE, east and SE of Coihaique, record continuing subaerial conditions in the region. To the south of General Carrera Lake, well developed Cenozoic deposits crop out (Fig. 3.69). Here, the late Palaeocene–early Eocene Ligorio Márquez Formation unconformably rests over the Early

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**Table:**

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<th>Ma</th>
<th>Era/Period/Epoch</th>
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<th>AISÉN REGION (S of Lake General Carrera)</th>
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<td>Buenos Aires Meseta</td>
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<td>Galera Fm.</td>
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<td></td>
<td>Eocene</td>
<td>Lower basalts</td>
<td>San José Fm.</td>
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<td>Ligorio Márquez Fm.</td>
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<td>Córdon de las Tobas Fm.</td>
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<td>Albian</td>
<td>Arroyo Pedregoso Fm.</td>
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<td>Barremian</td>
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<td>Hauterivian</td>
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**Fig. 3.69.** Stratigraphic succession and correlation of Mesozoic and Cenozoic deposits in the Palena and Aisén (south of Lake General Carrera) regions. For the Aisén region, stratigraphic succession is given for the eastern region next to the borderline (Buenos Aires Meseta) and for the western region (Guadal Meseta) (compare with stratigraphy north of Lake General Carrera in Fig. 3.65).
Cretaceous Cerro Colorado Formation and the Flamencos Tufts (Suárez & De la Cruz 1996; Suárez et al. 2000; Troncoso et al. 2002). The Ligorio Márquez Formation consists of subhorizontal fluvial and floodplain coal-bearing deposits containing tropical–subtropical plants (Suárez et al. 2000; Troncoso et al. 2002). This formation is, in turn, overlain with gentle erosional unconformity by alkaline flood basalts of middle Eocene age (lower basalts) (Charrier et al. 1978, 1979; Petford et al. 1996; Espinoza 2003; Espinoza & Morata 2003a) in the Buenos Aires Meseta south and SW of the locality of Chile Chico. These basalts are overlain by the late Oligocene–early Miocene marine Guadal Formation, which in turn is overlain by the continental clastic Galera Formation (Río Zeballos Group of Busterso & Lapido 1983).

Further west and SE of General Carrera Lake, in the Guadal or Cosmelli Meseta (Fig. 3.69), fluvial deposits of the San José Formation (Flint et al. 1994) that overlie the Ligorio Márquez Formation are probable equivalents of the Eocene basalts in the Buenos Aires Meseta. These deposits underlie the shallow marine to nearshore Guadal Formation (Niemeier et al. 1984) (late Oligocene–early Miocene; Frassinetti & Covacevich 1999) deposited in a warm to subtropical environment containing marine invertebrates of Atlantic affinities. The Guadal Formation is in turn overlain by the continental Galera Formation. In Pampa Castillo, the terrestrial deposits overlying the marine succession contain an abundant fossil mammal fauna that can be assigned to the Santacrucian South American Land Mammal Age (SALMA) (Flynn et al. 2002b).

The highest stratigraphic units exposed in the Buenos Aires Meseta correspond to subhorizontal late Miocene alkali basalts probably associated with a subducted segment of the Chile Ridge (Charrier et al. 1978, 1979; Petford et al. 1996; Suárez et al. 2000; Suárez & De la Cruz 2000; Espinoza 2003; Espinoza & Morata 2003a). Recent magmatic arc activity in this region is, however, hardly developed, with just the Maca (45°S) and Hudson (46°S) volcanoes testifying to the subduction of the southernmost Nazca Plate underneath the South American continent. In the region between 46°S and 49°S, south of Aisén, there is a complete gap in recent volcanic activity.

In the Ultima Esperanza sector of the Magallanes region, the deposits that record the initiation of basin inversion correspond to the deep-water turbidite lobes of the Cenomanian Punta Barrosa Formation (Cecioni 1956, 1958; Katz 1963, 1964; Cortés 1965; Scott 1966; Riccardi & Roller 1980; Dott et al. 1982; Wilson 1991; Fildani et al. 2003; Fildani & Hessler 2005) (Fig. 3.65). These sediments comprise a thick succession of turbiditic sandstones with thin shale intercalations (Cecioni 1956, 1957a, 1958; Katz 1963, 1964; Cortés 1965; Scott 1966). According to Riccardi & Roller (1980), the Punta Barrosa Formation is an equivalent of the Piedra Clavada Formation in Argentina.

Conformably overlying the Punta Barrosa Formation is the 2000-m-thick Santonian–Campanian Cerro Toro Formation with a widely outcropping prominent deep-water conglomeratic intercalation (Lago Sofia Conglomerates or Lago Sofia Member) (Cecioni 1956, 1957a, 1958; Katz 1963, 1964; Cortés 1965; Scott 1966; Riccardi & Roller 1980; Dott et al. 1976, 1982; Winn & Dott 1979; Shultz 2002; Beaubouef 2004). This formation corresponds to the ‘Chondrites Flysch’ of Cecioni (1957a) and extends eastwards into western Argentinian territory (Riccardi & Roller 1980). The sediments originated in the uplifting cordillera to the west and were deposited by turbidity currents flowing from north to south, which indicates the trend of the basin axis, orthogonal to the east-dipping western palaeo-slope indicated by slump-folds and to the supply direction into the basin (Scott 1966). Most of the Cerro Toro Formation consists of a rhythmic alternation of black shales and turbidites, the latter being thinner than in the Punta Barrosa Formation. The ‘Lago Sofia Conglomerates’ consist of...
and coarse sandstone wedges of the Rosa and Rocallosa formations. Clasts in these latter two formations consist of metamorphic and intrusive rocks, rhyolites, tuffs and shales, indicating a westerly provenance. These formations correspond to the ‘Grey-greenish Shales’ and the ‘Sandly Shales’ in the Springhill Platform (Mella 2001).

The next overlying Chorrillo Chico Formation (Thomas 1949) conformably overlies the latest Cretaceous Rocallosa Formation and contains the Cretaceous-Tertiary boundary, probably in its lower part (Charrier & Lahsen 1968, 1969; Lahsen & Charrier 1972). This 275-m-thick formation was deposited in an outer shelf environment (Mella 2001), consists of mudstones with fine glauconitic sandstone, siltstone and thin limestone intercalations, and contains abundant calcareous concretions. The conformably overlying Late Palaeocene to Eocene Agua Fresca Formation (Hoffstetter et al. 1957; Charrier & Lahsen 1968, 1969) consists of a variably thick (2290 m at its type locality) succession of grey mudstones with glauconite pellets and big calcareous concretions with abundant sandstone lenses in its upper portion. The sedimentary features and benthonic microfossils indicate deposition in a quiet shallow environment grading upward to littoral conditions. The Chorrillo Chico and Agua Fresca formation have been correlated with the ‘Zona Glauconítica’ in the Springhill Platform (Charrier & Lahsen 1968, 1969; Espinoza 2003; Espinoza et al. 1985; D’Orazio et al. 2005). Spreading oblique to the continental margin in the Chile Ridge, NW of the triple junction, apparently produced the dextral shear zone and associated fault pattern of the LOFZ (Hervé 1976; Hervé et al. 1979a; Murdie et al. 1993; Cembrano et al. 1996). Close to the triple-junction, ridge subduction resulted in a distinctive Plio-Pleistocene geological evolution on the westernmost continental margin resulting in an obducted ophiolite complex (Taitao ophiolite) and associated volcanic (basaltic to andesitic pillow lavas), volcaniclastic and marine sedimentary deposits, silicic plutons of calcalkaline affinity with K–Ar ages from 3.2 to 5.5 Ma, and a thick Cenozoic sedimentary succession in the Golfo de Pasas, a possible pull-apart basin linked to the movement of the LOFZ (Forsythe & Nelson 1985; Forsythe et al. 1985, 1986; Mpodozis et al. 1985; Bourgeois et al. 1993, 1996; Nelson et al. 1993; Kaeding et al. 1990; Behrmann et al. 1994; Lagabrielle et al. 1994).

**Uplift of the Patagonian Andes**

Apatic fission track ages reveal (1) initiation of accelerated cooling and denudation at c. 30 Ma along the western margin of Patagonia, and (2) subsequent eastward migration of c. 200 km of the locus of maximum denudation that ceased at c. 12–8 Ma at the position of the present-day main topographic divide (Thompson et al. 2001). According to the latter authors, this migration is related to either coeval eastward migration of foreland basin deformation, the effects of subduction erosion in the overriding plate at the trench or, less likely, shallowing of the angle of subduction. Accelerated denudation on the western side of the Patagonian Andes, contrasting with a rather low rate of denudation east of the topographic divide (≤ 3 km), has been interpreted to be the result of increased tectonic uplift driven by a large increase in convergence rates at c. 28–26 Ma that might have triggered orographically enhanced precipitation on the west side of the Patagonian Andes thus allowing increased erosion by fluvial incision and mass transport processes (Thompson et al. 2001). These results indicate a close relationship between tectonic evolution and climatic conditions across the Andean range (see also Montgomery et al. 2001).

**Southern Chile: summary and discussion**

The oldest Mesozoic record in this region is represented by the Late Triassic deposits of the Potranca Formation in the eastern Chonos metamorphic complex (Patagonian Archipelago). Although there is no subsequent geological information until Middle Jurassic times, Late Jurassic ages for metamorphism and exhumation of the Chonos complex (Thompson & Hervé 2002) suggest that subduction activity was occurring by then.

Three stages of Mesozoic and Cenozoic tectonic evolution can be defined for the Patagonian Andes: regional tectonic extension or Rift Phase (Middle to Late Jurassic); thermal subsidence (latest Jurassic to Early Cretaceous); and tectonic inversion with development of an asymmetric foreland basin (Late Cretaceous to Present) (Biddle et al. 1986; Harambourg & Soffia 1988; Soffia & Harambourg 1989; Skarmeta & Castelli 1997; Mella 2001) (Fig. 3.65). These three stages are broadly correlative with events further north in Chile. The Middle to Late Jurassic regional extension during the first stage of Andean evolution in Patagonia could be interpreted as recording passive-margin conditions and an absence of subduction. However this is not necessarily the case because an extensional setting during active Jurassic subduction existed north of 42S; this suggests that the general tectonic environment here was
probably different. The latest Jurassic–Early Cretaceous thermal subsidence phase in Patagonia coincides chronologically with the second substage of the first Andean stage and shows a similar arc–backarc palaeogeographic organization. In southern Chile low sedimentation rates resulted in a sediment-starved basin, a situation that was to change dramatically with the late mid-Cretaceous onset of tectonic inversion. The initiation of this compressive phase coincides with the beginning of the second stage of Andean evolution north of 42°S (Subhercynian or Peruvian phase), although the palaeogeography south and north of 42°S was considerably different. In southern Chile an eastward migrating foreland basin received huge amounts of westerly derived turbiditic sediments which pass up into shallow marine and continental clastic deposits as the basin filled in Late Cenozoic times.

The Patagonian Cordillera can be subdivided into the Palena–Aisén and Magallanes regions, north and south of 50°S respectively. The major difference in the evolution of these two regions consists in the development of a basin characterized by extensive bimodal magmatic activity in the south, this being a marginal basin or a branch of the rift system that formed the Atlantic Ocean. Closure (obduction) of this basin occurred at the beginning of tectonic inversion (third stage), with the suture zone forming a zone of weakness that favoured later development of the Patagonian Orocline (see discussion above). The separation zone between these two regions currently coincides with a gap in recent volcanic activity.

During the second stage of thermal subsidence in Early Cretaceous times, low sedimentation rates resulted in a depleted (sediment-starved) basin, whereas with the beginning of tectonism (tectonic inversion) and consequent development of the eastward shifting foreland basin, huge amounts of sediments supplied by uplifting and eroding areas to the west determined the development of thick turbiditic (flysch) successions and later shallow marine and continental clastic deposits that gradually filled the basin in Late Cenozoic times. Filling of the basin with continental deposits also occurred southwards, e.g. in the Ultima Esperanza section regression occurred in Maastrichtian–Palaeocene times (Cerro Dorotea Formation) whereas in the Riesco section regression occurred in Oligocene times (upper part of the Loreto Formation) (Fig. 3.66).

Final overview

The evolution of the Andean Orogen in Chile is the c. 550 million years geological history of a continental margin over 4000 km long and reflects the interaction between lithospheric plates during continental assembly and break-up. In this evolution it is possible to distinguish five separate main periods, the tectonic settings of which were controlled by lithospheric dynamics. In this chapter we have used the term ‘tectonic cycle’ for these periods because in each one of them the palaeogeographic, magmatic and sedimentological evolution is cyclic. These cycles, which have been further divided into stages and substages as appropriate, are as follows. 1. Pampean (Precambrian to Early Cambrian), about which there is very little information in Chile. 2. Famatinian (Ordovician), characterized in southwestern South America by several deformation events assigned to repeated collisional events between Gondwana and Laurentia. These two cycles correspond to the post-Pangaea II break-up episode of supercontinent evolution. 3. Gondwanan (Late Palaeozoic), corresponding to the assembly phase of Gondwanaland. 4. Pre-Andean (late Permian to earliest Jurassic), corresponding to an episode of interruption of subduction after the final assembly of Gondwana and during the first stages of break-up of the Gondwanan supercontinent. 5. Andean (late Early Jurassic to Present), characterized by continental break-up and representing the archetypal example of a subduction-related mountain belt (Dewey & Bird 1970). Superimposed on what may be considered the normal evolution of this typical subduction-related orogen were more unusual events such as the subduction of passive and active ridges, changes in plate geometry and dynamics, and development of an aulacogen and a huge magmatic province in southern Chile.

Analysis of the long history of the Andean Orogen, especially its most recent phase (the Andean tectonic cycle), clearly reveals...
the existence of extensional episodes separated from each other by shorter episodes of contractional deformation associated with modifications in the dynamics and geometry of the oceanic plate. These contractional and extensional episodes acting on the continental margin appear to be controlled by increases and decreases in plate convergence rate. Continental accretion and erosion, and positioning of the magmatic arc and its associated forearc and backarc areas are strongly dependent on these fluctuations in plate dynamics.

Major strike-slip fault systems appear to have been controlled by the location of the magmatic arc, and their movement mainly controlled by the obliquity of convergence. These faults represented favourable zones for concentrating ore-rich solutions and development of ore deposits, and, once instigated, they remained as lines of weakness that acted as a control on later deformation and mineralization. Finally, emphasizing the general theme of overall plate tectonic control of the geological evolution and palaeogeographic landscape of the Chilean margin, there appears additionally to be a close relationship between tectonic evolution and climate along and across the Andean range.

Several grants from different institutions, including the current Fondecyt grant no. 1030965 and Anillo ACT no. 18, both from CONICYT, have supported R.C. over many years in stratigraphic and tectonic research within different Chilean regions. The experience gathered during these research efforts proved invaluable to the preparation of this chapter. Discussions of R.C. with students in class and in the field, and with graduate students while preparing their thesis have greatly increased and deepened comprehension of Chilean geology and its relationship to neighbouring regions, and has led to the conception of new interpretations, some of which have already been published and others are proposed here. We acknowledge the valuable assistance of V. Flores in the preparation of the figures and the helpful comments and suggestions by C. Arévalo (Servicio Nacional de Geología y Minería (Sernageomin), Santiago), M. Farias and C. Arriagada (Departamento de Geología, Universidad de Chile, Santiago), F. Hervé (Departamento de Geología, Universidad de Chile, Santiago) helped by supplying an early draft of Chapter 2. J. Muñoz (Sernageomin, Puerto Varas) provided several recent publications on the Triassic of the Lake Region in south central Chile. Extremely useful and important were the comments by P. Cornejo (Head of Regional Geology at Sernageomin, Santiago), who reviewed the manuscript and provided abundant additional recent geological information/literature close to publication and/or to public release on different aspects of Chilean geology, and transmitted the enthusiastic support of members of her work team to this geological synthesis. N. Blanco (Sernageomin) provided unpublished chrono- and lithostratigraphic information about the Calama Basin. K.-J. Reutter (Freie Universität Berlin) made an extremely careful review of the manuscript; his comments and suggestions permitted great overall improvement of the text and figures. We also wish to acknowledge the editors, W. Gibbons for excellent reviews and corrections of the English style, and T. Moreno for efficient coordination of the many authors and reviewers in the different phases during preparation of this book. Finally, the authors wish to point out the excellent quality of the geological maps and associated explanation texts, including the Geological Map of Chile, issued by (the Servicio Nacional de Geología y Minería (SERNAGEOMIN) and the previous Instituto de Investigaciones Geológicas (IIG), Chile). These documents contain a great amount of very valuable information and detailed analyses without which it would have been impossible to achieve a thorough and coherent synthesis of the tectonostratigraphic evolution of the Andean Orogen.
Andean magmatism

MIGUEL A. PARADA (coordinator), LEOPOLDO LÓPEZ-ESCOBAR, VERÓNICA OLIVEROS, FRANCISCO FUENTES, DIEGO MORATA, MAURICIO CALDERÓN, LUIS AGUIRRE, GILBERT FÉRAUD, FELIPE ESPINOZA, HUGO MORENO, OSCAR FIGUEROA, JORGE MUÑOZ BRAVO, ROSA TRONCOSO VÁSQUEZ & CHARLES R. STERN

Magmatism in the Chilean Andes has taken place since about 300 Ma as a consequence of protracted subduction, although with significant spatial and temporal variations due to changes in ocean-floor geodynamics controlling distinct large-scale magmatic events. Early subduction along the Chilean segment of the Gondwana active margin took place during Late Palaeozoic times and generated typical arc magmatism and a subduction complex in the forearc environment. This tectonomagmatic regime was interrupted by mid-Permain contractional tectonics (collisional?) giving rise to a thickening of the crust that allowed deep crustal melt generation. Following this the entire Mesozoic history of the area became dominated by subduction-related extensional tectonics with mostly bimodal magmatism reflecting the involvement, to different degrees, of both crust and mantle as magma sources. Mesozoic volcanism and plutonism appear to have been independent of each other. Subsequent Cenozoic magmatism records changing geodynamic conditions from Palaeogene–early Neogene extension to late Neogene compression. The Neogene magmatic episodes are interpreted as an indirect consequence of oceanic ridge subduction: the Juan Fernández Ridge along the north-central Chilean margin, and the Chile Ridge along the southernmost Chilean border. Modern volcanism is also influenced by these ridge subductions, either by generating gaps in the Quaternary volcanic chain, or adakitic volcanism derived from slab melting.

Despite the essentially tectonic control outlined above, this chapter is subdivided geographically into four Andean segments, each of which exhibits distinct magmatic features. These segments are: 18–28°S, 28–38°S, 40–47°S and 47–55°S. The chapter is subdivided geographically into four Andean segments: 18–28°S, 28–38°S, 40–47°S and 47–55°S. The exception to this approach is the section on Quaternary volcanism, which is treated as a whole and acts as an introduction to the understanding of the magmatic evolution.

Magmatism in northern Chile (18–28°S): the Jurassic extensional volcanism of La Negra Formation (V.O., L.A. & G.F.)

La Negra Formation (and its stratigraphic equivalents, the Oficina Viz and Camaraca formations near Iquique and Arica respectively) is a thick sequence of volcanic extrusive rocks which crops out in several locations for c.1000 km along the Coastal Cordillera of northern Chile (Fig. 4.1). The volcanic rocks from this unit, together with huge plutons and other, smaller intrusive bodies, are interpreted as representing magmatic arc activity in the first stages of the Andean cycle (Suárez et al. 1985; Dallmeyer et al. 1985, Schueber & González 1999). Near the type locality of La Negra Formation (Quebrada La Negra south of Antofagasta), the volcanic pile reaches 10 km in thickness and rocks belonging to this unit are found up to 1200 m above sea level. Although many faults cross the volcanic sequence it is unlikely that its thickness could be the result of tectonic events (Buchelt & Tellez 1988). Sequences up to 4 km thick crop out in the Coastal Cordillera (e.g. south of Antofagasta) where they consist of lava flows, pyroclastic rocks and minor volcaniclastic sediments, with thicknesses ranging from 5 to 40 m, lying conformably and dipping (20–55°) north, west or east depending on the area. In the outcrops which consist mainly of lava flows, thickness variations or lateral unconformities are rarely observed and eruptive centres have not been identified (Buchelt & Tellez 1988), suggesting fissure eruption as the most probably mechanism for their emplacement (Losert 1974; Rogers & Hakesworth 1989). In some localities, interbedded shallow-level sediments with marine fossils, and pillow structures, indicate that the volcanic materials were emplaced into a shallow sea level environment (Tobar et al. 1968; Suárez et al. 1985, Muñoz et al. 1988b).

Based on their lithological characteristics, Hilldebrandt et al. (2000) divided the Jurassic deposits of northern Chile into three segments: (1) 26°20'–25°20' S, basic lavas and pyroclastic rocks with intercalations of volcaniclastic terrestrial sediments, followed by ignimbrites and acid to intermediate lavas; (2) 25–21°40' S, intermediate lava flows with interbedded epiclastic sandstones; (3) 21–18°30' S, amphibole-bearing intermediate lava flows, followed by bimodal explosive volcanic rocks and intermediate to acid lava flows, subordinate ignimbrites, tuffs and volcaniclastic sediments, and finally intermediate composition lava flows (Kossler 1998).

With regard to stratigraphic ages, in the areas where the base of this unit is exposed (27–26° and 23°S) the volcanic rocks overlie Hettangian–Sinemurian sediments and are interbedded with red sedimentary rocks which contain marine fossils of Aalenian–Bajocian and Pliensbachian age (Suárez et al. 1985, Naranjo et al. 1982) or Sinemurian age (Buchelt & Tellez 1988). The top of the formation is mainly in contact with Lower Cretaceous units (24°S, Charrier & Muñoz 1994) except between 21°30' and 18°30' S, where the lavas from the Oficina Viz Formation are overlain by middle Bajocian sedimentary rocks (Thomas 1970). More explosive volcanism is recorded in this area until Callovian times and, south of Arica (18°30') there is evidence of andesitic lava flows of Oxfordian age (Kossler...
The magmatic foci of the volcanic rocks are thought to have migrated eastward between 22°S and 27°S from Jurassic to Early Cretaceous times (Dallmeyer et al. 1996). In northernmost Chile, between 18°S and 21°S, a Late Jurassic transgression from the east of the backarc sea, coupled with the development of explosive volcanism to the west of the older volcanic arc (Oficina Viz Formation), indicates a westward migration of the arc (Fig. 4.1, Hilldebrandt et al. 2000; Kramer et al. 2005).

The Coastal Batholith consists of numerous plutonic complexes, mainly hornblende–biotite gabbros, diorites, tonalites/granodiorites and minor granites, which intrude the Jurassic volcanic sequences. The emplacement of huge plutonic bodies was largely controlled by NS and NW–SE faults, with both sinistral strike-slip and normal dip-slip movements (Scheuber et al. 1995; González 1996; Dallmeyer et al. 1996). Long-lasting cooling of high T–low P metabasites related to plutonic emplacement during Early Jurassic times records a long...
residence period of accreted magmatic intrusive rocks at mid-crustal levels and a high thermal gradient in the arc (Lucassen & Thrillwall 1998).

At least two generations of dykes and other small intrusive bodies can be identified (Dalmau et al. 1996; Scheuber & González 1999) and are associated with Cu-Au stratatbound deposits in Jurassic volcanic rocks. Some authors have interpreted these minor intrusive rocks as feeder conduits of the volcanism (Palacios & Definis 1981b; Pichowiak et al. 1990; Grocott et al. 1994; Espinoza et al. 1996).

Widespread low-grade alteration events affected both volcanic and, to a minor extent, plutonic rocks. The common mineral products of these events are epidote, chlorite, albite, white mica, clays, quartz, calcite, prehnite, pumpellylite, potassic feldspar, actinolite and zeolites. These secondary minerals occur mainly in the porous tops of the lava flows, breccia and tuff matrix, and as cement in sedimentary rocks (Losert 1974; Oliveros 2002). On a local scale, hydrothermal alteration events are associated with the formation of the Cu–Ag stratatbound deposits.

**Petrology**

The products of the Jurassic volcanism in northern Chile are mainly porphyritic lavas with up to 20% phenocrysts. Volcanic breccias, tuffs and sedimentary rocks, and epilastic sandstone lenses also occur but are less abundant. Basaltic andesites and andesites are by far the main compositional types, but basalts, dacites and rhyolites (ignimbrites) have been reported. At 23°30'S in the Mantos Blancos mining area, a sequence thought to be Jurassic consists mostly of intermediate and acid rocks such as andesites and rhyolites as along with dacitic dykes (Boric et al. 1990). Hildebrandt et al. (2000) proposed a bimodal magmatism based on the occurrence of acid volcanic rocks within the main intermediate lava flow succession at 25–21°S.

The lava flows have a basal zone with anphititic texture and flattened amygdales, a more massive central part where phenocrysts are often large, and an upper part which is brecciated and highly amygdaloidal. Commonly, the contact surface with the overlying flow is glassy, very oxidized, and contains oriented amygdales or flux textures. Epilastic sandstone lenses, from a few centimetres to more than one metre thick, often finely laminated, can be observed between flows. Plagioclase is the most frequent phenocryst with crystals commonly normally zoned (with An30 rims), reaching up to 1 cm in size and with a labradoritic composition, although bytownite is present in basalts and andesine–oligoclase in more silicic rocks (Buchelt & Tellez 1988; Rogers 1985). Ferromagnesian minerals are rare as phenocrysts and are commonly altered, augite and minor diopside being the commonest phases among them. Enstatite and olivine occur in basaltic andesites but they are completely altered to mafic phyllosilicates. Hornblende and rare Fe-rich biotite occur in andesites and dacites.

The groundmass comprises plagioclase microlites, clinopyroxene, magnetite or titanomagnetite, glass, sanidine and quartz in more silicic rocks. Microlites are not generally orientated except in the flow tops where fluid textures are frequent and an increase in glass content is observed. The explosive rocks such as breccias and tuffs have high lithic contents and their main mineral phases are Na- and K-feldspars.

Low-grade alteration minerals occur replacing primary crystal phases and glass in the groundmass. They are also found filling amygdales and veins. The degree of alteration varies from area to area, albitionation of Ca-plagioclase and total/partial chloritization of mafic minerals being the commonest expression. As a whole, alteration was pervasive making unaltered outcrops extremely rare, with the products of explosive volcanicity, the epilastic sandstones, and the brecciated tops of lava flows normally being the most altered lithologies.

**Geochemistry**

The volcanic rocks of La Negra Formation are predominantly mantle-derived, with their major and trace element signatures showing a subduction pattern largely unaffected by crustal contamination (Rogers & Hawkesworth 1989; Kramer et al. 2005). Major element compositions show that they have mainly high-K to calcalkaline affinities (Buchelt & Tellez 1988; Lucassen & Franzen 1994), although tholeiitic affinities have been found in rocks representing initial stages in the arc evolution (Palacios 1978; Pichowiak et al. 1990). Fe, Mg, Cr, Ni, V and Ti contents, as well as Eu anomalies, clearly indicate that fractional crystallization of plagioclase was the main process in the generation of these rocks from the parental magma, whereas clinopyroxene, olivine and titanomagnetite crystallization had only a secondary influence in the magma evolution except during the initial stages (Rogers 1985; Buchelt & Tellez 1988).

It should be pointed out that, in many samples, Ca, K or Na contents are highly variable over a small silica range (Rogers 1985). Large-scale alteration processes affecting the igneous rocks are interpreted as mobilizing Ca, K, Na, Fe, Rb and Cu (Losert 1974; Oliveros 2002). Thus, many of the observed variations in chemical composition are likely due to alteration events.

Rare earth element (REE) patterns are uniform for the flows and dykes and coincide with those for the plutonic rocks suggesting a similar source region (Lucassen & Franzen 1994). Enrichment in light REE relative to chondrite composition, La/Yb normalized ratios of 3–5 and Eu negative anomalies are striking features. Their trace-element compositions are typical of subduction-related rocks with depletion in high-field-strength elements (HFSE) with respect to mid-ocean-ridge basalts (MORB) and no significant variations in large-ion lithophile elements (LILE). Contents of Sr, Rb, K, Ba and Th are much higher than in MORB but published data show large variations for samples from the same locality. High Th/Ta ratios and Ta/Nb>1 are typical for these volcanic rocks together with depletion in Nb, Ti and P (Rogers 1985; Rogers & Hawkesworth 1989; Lucassen & Franzen 1994; Pichowiak 1994; Kramer & Ehrlichmann, 1996).

Initial 87Sr/86Sr ratios for these volcanic rocks and dykes range from 0.7030 to 0.7040 and show no significant correlation with differentiation degree or stratigraphical position. εNd values range between 0.512850 and 0.512950, so that the samples plot on the mantle array near the depleted mantle field (Rogers & Hawkesworth 1989; Pichowiak 1994; Lucassen & Thrillwall 1998). Pb isotopic contents show only small variations, with most being in the following ranges: 206Pb/204Pb, 17.96–18.42; 207Pb/204Pb, 15.55–15.63; 208Pb/204Pb, 37.99–39.73. They plot closer to average Pacific MORB than to average Palaeozoic crust. Isotopic compositions for volcanic rocks, dykes and some plutons are reasonably uniform and reflect a common source. Assimilation of Palaeozoic crust is unlikely as Sr-Nd-Pb isotopic compositions clearly indicate a mantle source (Lucassen et al. 2002).

**Geochronology**

Geochronological age data are scarce for the volcanic rocks, but the ages of the plutonic rocks and dykes, as well as sinistral shear zones associated with the Atacama Fault Zone, are well documented (Table 4.1, Fig. 4.2). Rb–Sr ages from the volcanic rocks are known from three localities, with all rocks having similar chemical compositions and trace element trends. Rogers & Hawkesworth (1989) obtained a whole-rock isochron of 186.5±13.6 Ma for a thick succession of lava flows along an east–west profile at 22°S; Venegas et al. (1991) reported a Rb–Sr ‘errorchron’ age of 173±19 Ma for andesitic lava flows from a copper deposit in the Mirchilla mining area; and finally Pichowiak (1994) obtained a whole-rock isochron of
<table>
<thead>
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<th>Region</th>
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<th>Rock type</th>
<th>Material</th>
<th>Method</th>
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<td><em>Ar</em></td>
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P, 40Ar/39Ar plateau age

186 ± 3.5 Ma for a suite of lavas from the La Negra Formation and plutonic rocks of the Coloso Gabbro Complex at around 24°S. Two K–Ar ages of 164 and 157 Ma are known for whole-rock samples from lava flows between latitudes 14°30′S and 19°S (Table 4.1, Fig. 4.2). 40Ar/39Ar plateau ages (2 sigma) have been obtained from unaltered plagioclase in Jurassic volcanic rocks from several localities with the following results: 157.9 ± 0.8 and 159.4 ± 0.6 Ma at 18°30′S (Arica), 161.2 ± 1.5 and 164.9 ± 1.7 Ma at 22°S (Tocopilla), and 150.9 ± 1.8 to 152.9 ± 2.0 Ma at 23°40′S (Antofagasta). These ages are significantly younger than those obtained using Rb–Sr methods, consistent with Jurassic volcanic activity lasting for only a short time (Oliveros et al. 2004).

Plutonic rocks belonging to the Coastal Batholith have been dated by Rb–Sr, 40Ar/39Ar, K–Ar, and Sm–Nd methods. The ages obtained indicate that they were emplaced between 199 and 19 Ma (Table 4.1, Fig. 4.2). Rather than having continuous magmatic activity, pulses of plutonic intrusions appear to have occurred. Dallmeyer et al. (1996) obtained 40Ar/39Ar plateau and isochron correlation ages in hornblende of c. 199.3 ± 0.6 Ma to 188.8 ± 1.2 Ma, 153.0 ± 1.0, 140.1 ± 0.8, 129.2 ± 1.0, 127.2 ± 1.0 and 107.1 ± 0.5 Ma for different plutonic complexes at 26°–27°30′S, which suggests a magmatic gap between c. 190 and 150 Ma. Hervé & Marinovic (1989) obtained Rb–Sr and K–Ar ages ranging from 192 to 98 Ma for several plutonic complexes between 24 and 25°S. These ages indicate that successive pulses of magmatic activity occurred from Early Jurassic to Early Cretaceous times. These pulses record an eastward migration for the plutonic rocks. Several plutonic bodies have been dated between 23 and 25°S yielding ages of 172 ± 8 Ma to 164 ± 6 Ma, 147 ± 4 Ma, 140 ± 5 Ma, 136 ± 5 Ma (K–Ar; Chávez 1985, Scheuber & González 1999), 160 ± 23 Ma (Sm–Nd; Lucassen & Thirlwall 1998) and 137.0 ± 2.2 Ma and 138.0 ± 1.7 Ma (40Ar/39Ar plateau ages; Maksaev 1990; Scheuber & González 1999). Rb–Sr ages of 155 ± 13 Ma and 158 ± 6 Ma and 40Ar/39Ar plateau ages of 159.9 ± 0.7 Ma and 158.8 ± 0.7 Ma have been reported for plutonic rocks between 22 and 22°30′S (Rogers & Hawkesworth 1989, Maksaev 1990). South of Arica (19°S) two K–Ar biotite samples from plutonic rocks have yielded ages of 164 ± 5 Ma and 164 ± 4 Ma (Garcia et al. 2004). Finally, dykes intruding both plutonic and volcanic rocks have been dated by various methods.
methods (K–Ar, \(^{40}\)Ar/\(^{39}\)Ar, Rb–Sr) and have yielded ages between c. 155 and 120 Ma (Table 4.1, Fig. 4.2).

Concerning the alteration of these rocks, recent \(^{40}\)Ar/\(^{39}\)Ar analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages at analyses on secondary mineral phases such as sericite, K-feldspar and actinolite yielded plateau ages. By comparison with volcanic and plutonic events, most of these ages appear valid, because (1) they are mostly duplicated on different rock samples, and (2) they often correspond to precisely dated magmatic events (Oliveros et al. 2004). Whether these alteration events are related to burial metamorphism or to hydrothermal activity generated by the intrusion of large plutons is not always clear.

**Tectonic setting**

The magmatic activity described is believed to have depended strongly on the rate of subduction and type of convergence. High-angle oblique subduction is thought to have been responsible for an extensional–transtensional tectonic regime along the whole arc (Dallmeyer et al. 1996; Scheuber & González 1999; Grocott & Taylor 2002); this agrees with plate configurations for South America during the Mesozoic era (Zonenshayn et al. 1984; Jaillard et al. 1990).

The volcanic rocks of La Negra Formation were emplaced in an intra-arc basin setting related to an oblique convergent margin. They are thought to have extruded either during (1) a sinistral relative movement between the forearc sliver and the backarc when the forearc moved with the same sense as the convergence obliquity (Fig. 4.3a; Scheuber & González 1999); and/or (2) in an extensional fault system linked to a retreating subduction boundary (Fig. 4.3b; Grocott et al. 1994; Taylor et al. 1998). The fact that the thick volcanic sequence was entirely deposited close to sea level implies that subsidence (extension) and crustal growth were well balanced (Lucassen & Franz 1994). Tilting of the volcanic sequence is likely to have taken place during arc-normal extension (between c. 150 and 160 Ma; Fig. 4.3a) after the deposition of the whole sequence and prior to Kimeridge times as suggested by angular unconformity with the overlying sedimentary rocks of this age (Scheuber et al. 1995; Scheuber & González 1999).

A trench-linked structure in the arc also controlled the emplacement of plutons in different episodes from Jurassic to Early Cretaceous times as recorded by the brittle and ductile deformation that affected these rocks (Dallmeyer et al. 1996; González 1996; Scheuber & González 1999). Early Jurassic (or older) plutons were emplaced at mid-crustal levels and some had very slow cooling rates (50 Ma for K–Ar and Sm–Nd systems) while the Middle Jurassic to Early Cretaceous intrusions would have been emplaced and cooled at shallower levels. A late phase of cooling and later uplift of the Coastal Cordillera in the region started around 120 Ma (Maksaae 1990, 2000; Scheuber et al. 1994), coincident with the beginning of major sinistral strike-slip motion along the 1000-km-long Atacama Fault Zone at c. 125 Ma.

Crustal block rotations, partly affecting the Mesozoic igneous rocks, took place in post-Early Cretaceous time (Taylor et al. 1998) and/or during the Incaic orogenic event (Late Eocene–Early Oligocene) (Arriagada et al. 2003). Since Neogene times, the whole forearc has behaved as one tectonic block (Arriagada et al. 2003).

**The Jurassic volcanosedimentary units of southern Peru: equivalents of La Negra Formation**

In southern coastal Peru, between latitudes 14°S and 18°S, volcanic rocks of Jurassic age directly correlated with La Negra Formation are well displayed (Fig. 4.1). The main successions present there correspond to the Río Grande and the Chala formations.

In its type locality, the Río Grande Formation consists of two units separated by a slight angular unconformity. The lower unit, c. 500 m thick, consists of red agglomerates, conglomerates and fine- to medium-grained red volcanicogenic sandstones accompanied by brecciated silicic lava flows, silicic ignimbrites, fossiliferous limestones, calcareous sandstones and greenish tuffs. The upper unit, c. 2000 m thick, largely consists

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**Fig. 4.3.** Models for the Jurassic tectonic evolution of the Andean arc. (a) Sinistral movement in a high-stress regime, volcanism and deep-seated plutons at c. 195–155 Ma, followed by shallow level plutonism without volcanism, magmatic crustal growth and arc-normal extension due to low convergence rate in a low-stress subduction regime at c. 160–150 Ma. Modified after Scheuber & González (1999). (b) In a retreating subduction boundary, a breakaway zone is generated which acts as the locus of magma ascent. Changes in convergence rate would favour either volcanism or plutonism, leading to an eastward migration of plutonic complexes. Modified after Grocott et al. (1994).
of highly porphyritic, partly pillowal, basaltic andesites composed of large labradoritic plagioclase phenocrysts, augite, olivine ghosts, and iron–titanium oxides. Minor intercalations of reddish conglomerates and brick-red, cross-laminated, volcanogenic sandstones exist (Aguirre & Offler 1985; Aguirre 1988). The whole succession was affected by very low-grade metamorphism of the hydrothermal–burial type. A Middle Jurassic (Dogger) age was assigned to the Rio Grande Formation by Ruegg (1956, 1961) based on fossil flora and fauna representing the Aalenian–Bajocian interval (175–168 Ma; 2004 Geologic Time Scale, ISC). A K–Ar (whole-rock) analysis on a weakly altered sample from the middle part of a lava flow of the Rio Grande upper unit gave an age of 164 ± 4 Ma which corresponds to the Bathonian–Callovian boundary, and a 87Sr/86Sr ratio of 0.70516 (at 160 Ma) (Aguirre 1988).

The Chala Formation at its type locality (15°52'S), is characterized by a basal sedimentary level consisting of sandstones and lutites and a thick upper section, c. 1500 m, made up of reddish-grey, strongly porphyritic flows of basalts and andesites with centimetric phenocrysts of labradoritic plagioclase plus augite and altered olivine. The formation is locally intruded by sills and porphyritic dykes petrographically akin to the flows (Romeuf 1994). As in the case of the Rio Grande Formation, the rocks of the Chala Formation show a widespread metamorphic alteration. Roperch & Carlier (1992) obtained 40Ar/39Ar ages (whole rock) of 177.0 ± 2.0 Ma for a Chala basaltic flow and of 157.0 ± 1.0 Ma for two basaltic flows of the Chala Formation. Romeuf (1994) used 40Ar/39Ar on plagioclase bulk samples to date a basaltic anesite flow from the Chala Formation and a porphyritic basaltic dyke cross-cutting rocks of the Guaneros Formation, an equivalent of the Chala Formation. A high-temperature mini-plug plateau age of 165.8 ± 0.9 Ma (14 steps corresponding to 51% of 39Ar released) was obtained on the lava flow whereas the dyke gave a high-temperature mini-plug plateau age of 157.2 ± 0.9 Ma (12 steps, 48% of 39Ar released: Romeuf et al. 1995). These ages are probably valid when compared to the corresponding 39Ar/40Ar ratios, and are close to those obtained in northern Chile (Oliveros et al. 2004).

The chemical composition of the Rio Grande and Chala lavas correspond to the K-rich calcalkaline series typical of volcanic arcs emplaced at a continental margin. Their REE patterns show a strong light REE (LREE) enrichment compared with heavy REEC to (HREE) (La/Yb chondrite-normalized ratios of 5.6–8.3), low TiO2 (< 5%), and high Al2O3 (16.8–17.6) and Zr (150–180 ppm) contents. Compared with lavas from La Negra Formation they are poorer in MnO, TiO2, Y, Zr, V, Sm, Nb, Yb, Hf, Ta, Th and Sc. Richer in K2O, MgO, Al2O3, Sr and Ba, and similar in FeO content (Romeuf 1994; Romeuf et al. 1995).

The presence of calcalkaline volcanism of arc type and Middle Jurassic age in southern coastal Peru has been interpreted as the result of WNW–ESE obduction of the Phoenix oceanic plate along this segment of the South American Pacific margin at that time (Romeuf et al. 1995).

Concluding remarks

The thick, subaerial to shallow marine volcanic and sedimentary sequences in the Coastal Cordillera of northern Chile define intra-arc basins that existed during Jurassic–Early Cretaceous times. The change from marine conditions in some localities, e.g. southern Peru (Rio Grande Formation volcanic and marine sedimentary rocks of the Aalenian–Bajocian interval) and northernmost Chile (e.g. volcanic and marine sedimentary rocks of Callovian age at Arica and Iquique), contrasting with the continental conditions in some other localities (e.g. Tocopilla and Quebrada La Negra in the Antofagasta area), suggests the existence of passages connecting the backarc basin and the forearc sea, breaking through the arc barrier (Hildebrandt et al. 2000).

Volcanic and sedimentary successions, together with huge plutons and other, smaller intrusive bodies contributed to crustal growth despite the extensional–transpression regime and the thinning of the continental crust that dominated during Jurassic–Early Cretaceous arc construction (Taylor et al. 1998). This extensional–transpression tectonic setting is reflected in the geochemistry which indicates little or no crustal contamination.

The hypothesis of long-lasting, broadly continuous Jurassic to Early Cretaceous magmatism in northern Chile involving extension, large-scale crustal thinning, basin subsidence, and extrusion of huge volumes of magma covering large segments of the continental margin, seems to be strongly supported by the geological record. The existence of discrete pulses of magmatic activity, however, cannot be dismissed as it is implied by the 40Ar/39Ar ages on both volcanic and plutonic rocks.

Magmatism in central Chile (28°–38°S) (M.A.P. & F.F.)

Magmatism within the central segment of the Chilean Pacific margin involved numerous and discrete intensive episodes of intrusion and volcanic emissions from Late Palaeozoic to Holocene times that gave rise to the north–south trending belts that occupy about 90% of the Coastal Cordillera and High Andes. The onset and duration of the magmatic episodes are poorly constrained and have been estimated by a limited number of reliable age determinations. On a broad scale, pre-Cenozoic magmatism took place during three main stages that started with a Carboniferous–Early Permian subduction regime along the western margin of Gondwana. This initial phase was followed by Late Permian–Early Jurassic period of arrested subduction associated with extensional tectonics, large-scale crustal partial melting (e.g. Kay et al. 1989; Parada et al. 1991; Martin et al. 1999) and production of only minor volumes of mantle-derived rocks. From Middle Jurassic to Early Cretaceous times, bimodal volcanism and calcalkaline plutonic activity, with progressively more mantle participation, took place along a rifted continental margin, generating an igneous belt about 1200 km long. After a period of magmatic quiescence that followed mid-Cretaceous contractional deformation (Mpodozis & Ramos 1989; see summary in Ramos & Aleman 2000), a new stage of extensional arc magmatism developed during the Oligocene–middle Miocene interval (Nystroem et al. 2003). After this (i.e. post-20 Ma) arc magmatism progressively decreased in intensity, as a consequence of shallowing of the subducting slab (cf. Kay & Mpodozis 2002), and ended at about 5 Ma. The subduction of the Juan Fernández Ridge caused extreme shallowing of the oceanic slab beneath the continental margin of central Chile and the development of a modern volcanic gap between 27°S and 34°S. Today, active volcanoes are restricted to the High Andes, forming part of the northern segment of the Southern Volcanic Zone (SVZ), which extends from 33°S to 46°S.

One of the most striking features of the central Chilean Andes is the Meso-Cenozoic distribution of the plutonic belts characterized by an eastward decreasing age (Parada et al. 1988). The origin of this distribution remains poorly understood, although the lateral (westward) displacement of the continental margin due to successive plutonic–volcanic belt emplacement, analogous to oceanic spreading, has been invoked (Åberg et al. 1984). This ensialic spreading would explain the c. 100 km separating the pre-Middle Jurassic plutonic components of the Elqui-Limarí Batholith, in the High Andes, and the coeval plutonic rocks of the Coastal Batholith. Both batholiths and associated volcanic rocks are examined below.
patterns, small Eu anomalies and high La/Yb ratios, suggesting the presence of residual garnet in the source rocks (Mpodozis & Kay 1992). Both units are isotopically (Sr–Nd) radiogenic, with εNd in the range of −3.1 to −5.8 and initial 87Sr/86Sr ratios between 0.70603 and 0.70965 indicating either a provenance from an enriched lithospheric mantle and crust or an origin by contamination of a depleted mantle by radiogenic crust (Mpodozis & Kay 1992). Because no magmatic event older than Late Carboniferous–Early Permian has been identified in central-north Chile, the existence of an enriched lithospheric mantle before the onset of the Elqui Complex plutonism is tenable.

This complex resulted from subduction beneath the continental margin of Gondwana after the docking of the Chilenia terrane during Devonian times (Ramos et al. 1986). The shift from the metaluminous Guanta unit to the peraluminous Cochiguás unit is explained by a mid-Permian crustal thickening attributable to a contractional (collisional?) event (the San Rafael event of Polanski 1970; Azcuy & Caminos 1987; Llambías & Sato 1990), that allowed deep crustal melting and the formation of the Cochiguás intrusives.

Post-orogenic Early Triassic El León–Chollay Complex (ELCC)

Coarse-grained biotite granites and felsic porphyries of the ELCC intruded the Permian volcanic sequence of the Pastos Blancos Group and crop out along the axis and eastern flank of the Andes. U–Pb zircon ages of 249.7 ± 3.2 and 242.5 ± 1.5 Ma (Martin et al. 1999) indicate an Early Triassic plutonic event, partially coeval with the Cochiguás unit and the 267–247 Ma Colanguil Batholith (Llambías & Sato 1990) located further east in the Frontal Cordillera of Argentina (29–31°S), although an older biotite K–Ar age of 276±4.0 Ma has also been obtained (Nasi et al. 1985). This complex includes typical calcalkaline metaluminous I-type granites, although S-type and A-type granites are also present. The REE patterns are commonly flat with large negative Eu anomalies consistent with a low-pressure, garnet-free source region (Mpodozis & Kay 1992). Like the Colanguil Batholith, this complex developed during the tectonic relaxation that followed the mid-Permian San Rafael deformational event.

Late Triassic–Early Jurassic bimodal intrusions

These intrusions represent the youngest event in the construction of the Elqui–Limari Batholith and include isolated bodies epizonally emplaced into the Elqui Complex. They comprise metaluminous to alkaline granites (A-type granites) and slightly peraluminous cordierite-bearing two-mica granodiorites and granites (S-type granitoids), that previously were assigned to the Monte Grande and Hacienda Vieja suites, respectively (Parada 1988). Los Colorados brick-red felsic porphyries and some plutons previously assigned to the El León unit (Mpodozis & Kay 1992) also belong to this intrusive complex. Small gabbro intrusions are also recognized, providing a bimodal character to the complex similar to that recognized in its volcanic counterpart, the youngest succession of the Pastos Blancos Group. The Hacienda Vieja bimodal intrusions were generated during an extensional regime, attributed to arrested subduction, which promoted partial melting and emplacement of crustal-derived and mantle-derived magmas. K–Ar ages in biotite and muscovite of about 220 Ma and one imprecise Rb–Sr isochron age of 197±5 Ma have been obtained from these rocks (Mpodozis & Kay 1992; Parada et al. 1981). The metaluminous and alkaline granites are high-SiO2, low-CaO A-type granites, with moderately fractionated REE patterns and large negative Eu anomalies. On the contrary, the peraluminous granitoids have highly fractionated REE patterns with small or no Eu negative anomalies, suggesting a crustal source with residual garnet. Contrasting Sr–Nd isotopic features are recognized among both the A-type and S-type
components. A metaluminous sample has an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.70692 and $\varepsilon\text{Nd}$ of 0.0, whereas a peraluminous sample provided values of 0.71450 and $-4$ for the same isotopic parameters. Granitic and metapelitic source rocks have been invoked to explain the contrasting geochemical features between the A- and S-type granitoids, respectively (Parada et al. 1991).

**Mafic dyke swarm**

Mafic dykes (and minor felsic dykes) are widely distributed within the Elqui–Limari Batholith, intruding both it and the overlying Triassic to Jurassic sedimentary cover successions. Mutual cross-cutting relationships with some Triassic–Early Jurassic granites are locally observed. The mafic dykes are basaltic (45–47% SiO$_2$) to andesitic (54–64% SiO$_2$) in composition and have high K$_2$O (shoshonitic affinity). In terms of Zr/TiO$_2$ and Nb/Y (Winchester & Floyd 1977), some lamprophyric dykes have alkaline affinities (Parada et al. 1991). The dyke swarm represents a protracted bimodal event precursor of the Mesozoic ensialic expansion. A similar bimodal dyke swarm has been recognized hosted in the Colangiü Batholith (Llambías & Sato 1990).

**The Coastal Batholith and Mesozoic volcanic rocks between 28°S and 38°S**

This segment of the Coastal Batholith is made up of four north–south trending belts with eastward decreasing age: Late Palaeozoic, Late Triassic–Early Jurassic, Middle Jurassic and Early Cretaceous. Between 28°S and 33°S, the batholith includes only the Mesozoic granitoids, whereas south of 33°S Late Palaeozoic granitoids constitute its main component (Fig. 4.5). Most of the information given below (and in Table 4.2) comes from studies carried out between 31°S and 33°S and at 37–38°S.

**Late Palaeozoic plutonic component**

Rock exposures of Late Palaeozoic granitoids form a north–south trending belt more than 750 km long from 33°S to 38°S (Fig. 4.5) and are usually continuous along the coastline, but limited inland. These granitoids appear to be the southern segment of a once-continuous Late Palaeozoic belt, now displaced c. 100 km west as a consequence of Mesozoic crustal spreading. Most of the studies on the coastal Late Palaeozoic granitoid belt are restricted to its northern and southern extremes, where the Santo Domingo Complex and Nahuelbuta Complex crop out, respectively.

**Santo Domingo Complex.** Exposed along the coastline from 33°S to 34°S approximately, this unit includes coarse-grained hornblende–biotite-bearing granodiorites and tonalites and minor amounts of granites with 1–4-cm-long microcline megacrysts. Tonalites commonly have ovoidal and lens-shaped mafic enclaves (gabbroic to dioritic compositions), schlieren and synplutonic dykes, all of them attributed to a process of mingling between two contrasting magmas represented by the enclave-free granites and the more mafic enclaves and synplutonic dykes (Parada et al. 1999). The northern and southern exposures of this complex consist of very coarse-grained tonalites and granodiorites with a NW-trending foliation and deformed fabrics that were generated during or soon after emplacement. Al-in-hornblende geobarometry obtained from the Santo Domingo tonalites gave a depth of emplacement equivalent to a pressure of about 4.5 kbar (Sial et al. 1999). Like the Elqui–Limari Batholith, this complex is intruded by a Mesozoic mafic dyke swarm.

Rb–Sr isochrons obtained in three localities (Reñaca, Algarrobo and Santo Domingo) gave similar ages between 292±2 and 308±15 Ma (Hervé et al. 1988). These ages are concordant with two zircon U–Pb ages of 309 and 290 Ma (Godoy & Loske 1988) obtained c. 20 km south of Valparaíso. Lithologically and geochronologically, this complex is equivalent to the Early Permian Guanta unit of the Elqui–Limari Batholith.

The granitoids and enclaves form a calcalkaline suite covering a wide compositional spectrum from 50 to 70% SiO$_2$ and show field relationships and linear trends in variation diagrams for major elements, typical of a suite derived from the mixing of two contrasting magmas (Parada 1990). Both granitoids and enclaves are isotopically (Sr–Nd) enriched. The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios fall in a broad range between 0.7057 and 0.7098, the lowest values corresponding to the enclaves. Similar $\varepsilon\text{Nd}$ values between $-2$ and $-4$ were obtained for both granitoids and enclaves (Parada et al. 1999).
Table 4.2. Summary of the main features of the Coastal Batholith and Mesozoic volcanic rocks of central Chile based on references given in the text

<table>
<thead>
<tr>
<th>Rock type</th>
<th>Late Palaeozoic plutonism</th>
<th>Triassic–Early Jurassic plutonism</th>
<th>Mid-Jurassic plutonism</th>
<th>Early Cretaceous plutonism</th>
<th>Jurassic volcanism</th>
<th>Early Cretaceous volcanism</th>
</tr>
</thead>
<tbody>
<tr>
<td>Age of magmatic events</td>
<td>c. 290 Ma</td>
<td>c. 200 Ma</td>
<td>c. 165 Ma</td>
<td>c. 100–95 Ma</td>
<td></td>
<td>119–117 Ma</td>
</tr>
<tr>
<td>Textural and structural features</td>
<td>Coarse and tectonically oriented minerals.</td>
<td>Coarse-grained to porphyric leucogranites.</td>
<td>Medium to coarse-grained rocks.</td>
<td>Medium to high-K calcalkaline suite.</td>
<td>Equigranular to porphyric granodiorites.</td>
<td>Equigranular to porphyric rhyolites</td>
</tr>
<tr>
<td>Isotopic signatures</td>
<td>Enriched Sr-Nd isotope values.</td>
<td>Enriched granites: Sr: 0.715–0.712 and εNd: −0.6, −0.7.</td>
<td>Depleted Sr-Nd isotope values. Sr: 0.7034–0.7044, εNd: +0.9 to +3.7.</td>
<td>Depleted Sr-Nd isotope values. Sr: 0.7031–0.7050, εNd: c. +2.0 to +4.5.</td>
<td>Depleted signature: Sr: 0.703–0.705, εNd: 0 to +6.0.</td>
<td>Depleted to strongly depleted isotopic signature.</td>
</tr>
<tr>
<td>Tectonic regime</td>
<td>Contractual</td>
<td>Extensional</td>
<td>Extensional</td>
<td>Extensional</td>
<td>Extensional/contractional</td>
<td></td>
</tr>
</tbody>
</table>

Hb, hornblende; Px, pyroxene; Bt, biotite

Nahuelbuta Complex. This is hosted in the accreted low P–T metasedimentary belt of the Late Palaeozoic subduction complex and is overlain by Late Triassic sedimentary successions. Granite is the most abundant lithology (followed by diorite) and is usually slightly peraluminous, particularly when near metasedimentary host rocks, and has relatively flat REE patterns (Lucassen et al. 2001b). Like the Santo Domingo Complex, the Nahuelbuta Complex has ages of about 294 Ma (Hervé et al. 1988) and is isotopically enriched: εNd values are in the range of −2.5 and −7.5, and initial 87Sr/86Sr ratios vary within the range 0.705–0.715, indicating a mixing of at least two isotopically different lithospheric sources, with crust similar in composition to the exposed metasedimentary host rocks being the main contributor (Lucassen et al. 2001b).

Late Palaeozoic granitoids have also been identified (based on both radiometric and stratigraphic evidence control, in the Andean Cordillera at about 40°S (Martínez 1998) and, more widely, in the North Patagonian Massif (Cingolani et al. 1991), a morphological feature located in Argentina east of the North Patagonian Andes.

Triassic–Early Jurassic bimodal intrusions and associated volcanic rocks

A discontinuous belt of shallow-placed leucogranite and gabbro intrusions and associated mafic and felsic dykes is recognized along the Coastal Batholith. Between 30°30’S and 33°S these mildly peraluminous leucogranites and gabbros have been dated as between 209±4 and 191±4 Ma (Irwin et al. 1988; Parada et al. 1988) and assembled within the Limari Complex (Parada et al. 1999). The leucogranites are medium- to fine-grained with equigranular to porphyritic textures, and were emplaced passively into both a Late Palaeozoic subduction complex and Late Palaeozoic–Early Mesozoic marine formations (Rivano et al. 1985) under an extensional tectonic regime.

The volcanic equivalent of the Limari Complex is the Pichidangui Formation characterized by the presence of a bimodal association of tholeiitic basalts and rhyolites (Morata et al. 2000). Both bimodal associations, the Limari Complex and the Pichidangui Formation, have similarities and differences in geochemical and isotopic characteristics. The basic plutons have higher MnO contents (8–10%) than the basalts (4–5%), but both types of mafic rocks have εNd between +3 and +5 (Parada et al. 1999; Morata et al. 2000), probably related to an enriched mantle source. All the acidic rocks show compositional evidence of major crustal components in their source, although some isotopic differences between rhyolites and leucogranites are recognized. Initial 87Sr/86Sr ratios vary
from 0.7040 to 0.7116 in rhyolites, and two leucogranite samples have values of 0.7154 and 0.7121.

Evidence for Triassic–Early Jurassic plutonism south of 33°S is scattered, with isolated granite bodies distributed along the western margin of the Coastal Batholith. These include the epizonal Topocalma, La Estrella, Pichilemu, Constitución and Hualpén plutons that were emplaced during the 218–202 Ma interval (Hervé et al. 1988; Creixell et al. 2002) and coeval with a Late Triassic riftting episode in the Chilen continental margin (Charrier 1979). All these plutons are slightly peraluminous granites, although small gabbro intrusions have also been recognized. The Hualpén granite plu ton, the only body of the southern segment of the Coastal Batholith that has been studied isotopically, shows evidence of crustal contribution such as εNd values of about −2.5 and initial 87Sr/86Sr ratios of 0.708–0.712 (Lucassen et al. 2001b; Creixell et al. 2002).

Middle–Late Jurassic and Early Cretaceous plutonic belts

The Middle–Late Jurassic and Cretaceous belts respectively occupy a western and an eastern position and extend southward along the Coastal Cordillera to c. 34°S. South of this latitude, only the Cretaceous belt continues, occupying the Andean Cordillera. Numerous Jurassic plutons were emplaced into metasedimentary and metaoceanic rocks of the Late Palaeozoic subduction complex, as well as into Late Palaeozoic granitoids and Early Mesozoic felsic volcanic rocks (Triassic Pichidangui Formation and Early Jurassic Ajial Formation). The Early Cretaceous plutons intruded the Jurassic plutons, but were also emplaced into a thick Early Cretaceous volcanic–sedimentary succession deposited in a subsiding basin (Vergara et al. 1995). This succession includes the volcanic-dominated Lo Prado and Veta Negra formations, assembled in the Ocoite Group, and the sedimentary-dominated Las Chicas Formation.

The main lithologies of the Jurassic plutonic belt are medium-grained two-pyroxene diorites, gabbros and hornblende + biotite tonalites. The tonalites locally have abundant mafic enclaves and symplutonic dykes similar in composition to the diorite and gabbro bodies. Small plutons of leucogranites are also recognized. Whole-rock Rb–Sr isochron ages of 164±2 Ma, hornblende 40Ar/39Ar of 169±3.6 and 165±2.6 Ma, and zircon U–Pb ages of 157, 160 and 164 Ma are known from this belt (Parada et al. 1988; Irwin et al. 1987; Godoy & Loske 1988; Gana & Tosdal 1996). Hornblende crystallization pressures of 3–5 kbar were obtained in quartz-diortes and tonalites in localities near Valparaíso (Gana & Tosdal 1996). Despite their wide compositional range, most of the rocks of this belt can be classified as medium- to high-K series and have remarkably similar REE patterns characterized by a La3/Yb3 close to 7 and moderate negative Eu anomalies (Parada 1992; Parada et al. 1999).

The Cretaceous granitoids constitute the largest Mesozoic plutonic belt of central Chile. They include a wide lithological spectrum from gabbro to granite, defining a typical calcalkaline high-K suite. Abundant geochronological data exist on these granitoids obtained by different methods, with variable precisions (cf. Drake et al. 1982b; Parada et al. 1988, 2005a; Hervé et al. 1988, Gana & Gana 1996). Recent studies on the Calca pluton, a good example of the last Cretaceous plutonic event, indicate an age and pressure of emplacement of c. 95 Ma and c. 2 kbar, respectively (Parada et al. 2002, 2005b).

Jurassic and Early Cretaceous volcanism at a rifted continental margin

During the Middle Jurassic–Early Cretaceous interval episodic volcanism took place. At about 33°S these events are represented by the volcanic components of the Mid-Jurassic (Bajocian) Ajial Formation, Late Jurassic Horqueta Formation, and the Early Cretaceous (Valanginian–Barremian) Ocoite Group (Aguirre et al. 1989) and (Albian) Las Chicas Formation. The Ajial Formation includes a bimodal association of abundant dacitic ignimbrites and interbedded basalts and basaltic andesites. These rocks have a calcalkaline high-K affinity and exhibit poorly fractionated REE patterns (Vergara et al. 1995). The Horqueta Formation includes a suite of calcalkaline lavas from basic to acidic composition. The Ocoite Group is represented by a c. 15-km-thick succession that includes the Valanginian to Hauterivian Lo Prado Formation and the Hauterivian to Barremian Veta Negra Formation. A minimum effusion rate of about 500 km3/Ma has been calculated during the deposition of the 10-km-thick Veta Negra Formation (Vergara et al. 1995). At 30°S the volcanism is represented by the abundant volcanic rocks of the Hauterivian to Barremian Arqueros Formation and Barremian to Albian Quebrada Marquesa Formation.

The rocks of the Ocoite Group, as well as those of the Arqueros and Quebrada Marquesa formations, are subaerial porphyritic basalts and basaltic andesites with high-K to shoshonitic affinity (Levi & Aguirre 1981; Aberg et al. 1984; Levi et al. 1988; Vergara et al. 1995; Morata & Aguirre 2003). Remarkably similar flat REE patterns are recognized among the Early Cretaceous basic and intermediate volcanic rocks. They exhibit low (4–11) La0.6/Lu0.8 ratios which exclude the presence of garnet as a residual phase in the magma source. Additionally, the REE patterns show a slightly negative Eu anomaly, despite the commonly observed plagioclase-rich porphyritic textures (Morata & Aguirre 2003).

Volcanic rocks are scarce in Las Chicas Formation, which mainly consists of a lower section of limestone, sandstone and minor amount of andesite and rhyolitic tuffs, and an upper section of thick strata of volcanic–sedimentary breccias and conglomerates with coarse volcanic fragments (molasse-type deposit).

Rifting magmatism and associated subsiding basins are the most outstanding processes developed along the western margin of South America during Jurassic–Cretaceous times. They are probably the result of a trench retreat that led to crustal attenuation, asthenospheric upwelling to fill the gap, bimodal volcanism, burial metamorphism and plutonic activity (Aberg et al. 1984; Vergara et al. 1995; Aguirre et al. 1989). In Chile, between 25°S and 36°S, extensional bimodal volcanic activity developed during Jurassic–Early Cretaceous times, generating c. 1200-km-long belt with an average width of 30 km and thickness of 3 to 13 km. Associated with basin subsidence, very low-grade metamorphism developed, reaching its climax at 93–94 Ma in central Chile (Aguirre et al. 1999). These geochronological data show that the very low-grade metamorphism is coeval with the last Cretaceous plutonic event, suggesting that this metamorphism is not the result simply of burial, but also of enhanced regional thermal gradients related to the plutonism.

Unlike the observed temporal and compositional correlations between the Late Palaeozoic–Early Mesozoic plutonic complexes of the High Andes Batholith and its volcanic counterparts, the Middle–Late Jurassic and Cretaceous plutonic events are younger than the spatially related peak in volcanism (Table 4.2). Rhyolites of the Bajocian Ajial Formation are intruded by the Late Jurassic plutons. Similar stratigraphic relationships are observed between the 5–10-km-thick pile of basalts and basaltic andesites of the Early Cretaceous Veta Negra Formation (and equivalent units) hosting the extensive mid-Cretaceous plutonic belt.

The available geochronological data on the Mesozoic volcanism are restricted to the Veta Negra, Arqueros and Las Chicas formations. 40Ar/39Ar ages of about 119±2.4 Ma have been obtained on primary plagioclase in basaltic flows of the Veta Negra Formation at 33°30’S, and do not differ substantially from the 117±0.6 and 114±0.7 Ma obtained on primary Ca-plagioclase populations from the Arqueros Formation (30°S).
Zircon U–Pb ages in the range 109 ± 0.2 to 106.5 ± 0.4 Ma have been obtained in rhyolites of the lower section of the Las Chilcas Formation, whereas its upper section has whole-rock and plagioclase K–Ar ages in the interval 101 ± 3 to 95 ± 3 Ma (Wall et al. 1999).

A trend towards MORB isotopic signature with time
Sr–Nd data, obtained from rocks of the Coastal Batholith between 31°S and 34°S (Parada et al. 1999, 2002; Morata et al. 2001), are plotted in Figure 4.6 and show an abrupt change from an enriched Late Palaeozoic to a depleted Mesozoic source of the magma. This change coincides with a change in the tectonic regime from a compressional subduction associated with the formation of a subduction complex (Hervé et al. 1987; Mpodozis & Kay 1992), to extension attributed to arrested subduction related to an episode of slow sea-floor spreading (Larson & Pitman 1972). The absence of a magmatic event older than Late Palaeozoic in central Chile suggests the existence of an old and isotopically enriched lithospheric mantle source for the Late Palaeozoic granitoids and associated volcanic rocks.

Along the central Chilean continental margin, a progressive Mesozoic extension developed, culminating during Early Cretaceous times in an aborted marginal basin (Aberg et al. 1984) within which the thick Ocoite Group succession was deposited. This progressive Mesozoic extension correlates with isotopic (Sr–Nd) trends towards MORB signatures (Fig. 4.6) in both plutonic and volcanic rocks (31–34°S). This is attributed to a combination of: (i) a continuous asthenospheric upwelling and associated lithospheric removal that accompanied extension (Parada et al. 1999, 2005a); (ii) a decreasing participation of the crust due to its increasing refractory nature that resulted from successive extractions of felsic melts or subsolidus dehydration as extension proceeded.

It is interesting to note the more depleted isotopic signatures of the Mesozoic plutonic complex as compared with their host volcanic rocks (Fig. 4.6), confirming the independence between volcanism and plutonism. This situation differs from that described for the Coastal Batholith of Peru (Atherton 1990), which is considered to be a product of precursor volcanic rocks remelting within a rifted continental margin. The last plutonic event, during which the Caleu pluton formed, would have been generated from the most depleted source during the climax of Cretaceous extension.

Diachronous Mesozoic exhumations of the Coastal Batholith
Fission track dating has revealed an exhumation event in Palaeozoic–Early Cretaceous coastal granitoid rocks (32°30′S to 33°30′S) during the interval 106–90 Ma (Gana & Zentilli 2000; Parada et al. 2005a). The magnitude of this Cretaceous exhumation was larger along the present-day coastline at these latitudes where the once deep-seated (c. 3.0–5.0 km) or c. 11–18 km depth based on Al-in-hornblende geobarometry; Gana & Tosdal 1996; Sial et al. 1999) Palaeozoic to Jurassic plutonic–metamorphic belts are now exposed. In the case of the Cretaceous belt, the exhumation would have started at about 7 km depth.

The cause of the exhumation can be attributed to a well-recorded change in tectonic regime in the Coastal Cordillera and along the Chile–Argentina Andes during mid-Cretaceous times. At that time, extensional tectonics ended and a high-stress compressional regime started, giving rise to the Aconcagua fold-thrust belt located in the High Andes (Mpodozis & Ramos 1989; Ramos & Aleman 2000). The age range of the exhumation roughly coincides with the deposition of thick breccias and conglomerates (molasse-type deposits) of the upper section of the Las Chilcas Formation, suggesting a cause-and-effect relationship.

On the other hand, the Late Palaeozoic granitoids of the Nahuelbuta Complex (at about 38°S) were exhumed before the deposition of the Late Triassic sedimentary rocks of the Santa Juana Formation. In fact, basal conglomerates with abundant granitic fragments were deposited over Late Palaeozoic granitoids. Similar stratigraphic relationships have been observed between the pre-Triassic components of the High Andes Batholith and the overlying conglomerates of the Middle–Late Triassic Las Bresas Formation.

The Cenozoic plutonic belts: 30–38°S
These belts are poorly known, despite the close temporal and spatial relationships with porphyry copper deposits of this segment. The information given here comes mainly from regional mapping and a K–Ar geochronological programme carried out during the 1980s. Two north–south trending Cenozoic plutonic belts with eastward decreasing ages are clearly identified east of the Coastal Batholith, along the Andes of central Chile between 30°S and 38°S. The plutonic rocks of the segment between 30°S and 32°S are those most of the geological and radiometric data have been obtained (Rivano et al. 1985; Parada et al. 1988). Stratigraphic relationships and radiometric ages indicate that the western belt is a continuous Palaeogene belt that extends down to 32°S where it abruptly ends. The eastern belt is less continuous and has early and late Miocene ages. Further north of these latitudes the belts are poorly defined by isolated plutons with few radiometric data.

The jump in the locus of plutonism from the Coastal Batholith to the Palaeogene belt and then to the High Andes Neogene belt, corresponds essentially to non-magmatic intervals at 95–70 Ma and 40–26 Ma. During the development of the Neogene belt, a non-magmatic interval is also recognized between 24 and 17 Ma, although no jump in the locus of magmatism exists. The eastward propagation of the locus of plutonism with time correlates with a secular decrease in intensity of the plutonic activity, as suggested by the decreasing area of the respective belt exposures.

The Cenozoic plutons are epizonally emplaced into volcanioclastic successions, in the case of the Palaeogene intrusions, and into deformed Late Jurassic to Palaeogene sedimentary and volcanic formations, in the case of the Miocene bodies.
The Palaeogene belt in the 30–32°S segment includes medium-grained to porphyritic hornblende-pyroxene diorites and quartz-diorites and fine-grained leucogranites assembled in the Cogotí superunit. Eleven K–Ar mineral ages in the range 67–38 Ma have been reported for this superunit (Parada et al. 1988). In contrast, only two Miocene events have been identified based on 14 K–Ar mineral ages distributed in the intervals 26–24 Ma and 17–8 Ma. The early Miocene granitoids of this segment correspond to the Rio Grande superunit, exhibit a wide lithologic spectrum from gabbro to granite, and occur as shallow intrusions along the axis of the Andes. The late Miocene granitoids include hornblende–clinopyroxene–monzo-diorites tind dacitic porphyries assembled into the Rio Chicharra superunit (Parada et al. 1988). As with the early Miocene granitoids, most of the late Miocene granitoids occur as small epizonal bodies located along the highest Andean ranges.

Forming part of the Neogene belt, the late Miocene (c. 10 Ma) porphyritic intrusions of the Los Pelambres giant porphyry copper deposit show various features akin to the adakites. These intrusions, located at 31°43′S, 70°29′W, represent an oddity in the late Miocene magmatism of central Chile (Reich et al. 2003). In fact, they exhibit distinctly higher Na₂O contents and higher Sr/Y (100–300) and La/Yb ratios (25–60) ratios than coeval magmatic units such as La Ramada (c. 32°S) and Cerro Aconcagua volcanic centres (Kay & Mpodozis 2002) and La Gloria pluton (Cornejo & Mahood 1997), which have typical arc geochemistry with moderate La/Yb ratios and Eu. The adakite-like rocks from Los Pelambres, located in the modern flat-slab Andean segment, are closely related in time and space with the subduction of a west-east segment of the Juan Fernández Ridge. This locally restricted scenario has been considered favourable for the partial melting of young and hot rocks of the Juan Fernández Ridge under late Miocene flat-slab conditions and may explain why adakite rocks have not been found elsewhere along the Neogene belt (Reich et al. 2003). It is interesting to note that adakite-like intrusions have also been found associated with late Oligocene–early Oligocene porphyry copper systems of northern Chile (Oyarzún et al. 2001).

Neogene magmatism in the modern flat-slab Andean segment

The Neogene flat-slab Andean segment (28–33°S) forms a distinctive component of the Chilean Andes, and is located north of the active Andean Southern Volcanic Zone (SVZ; 33–46°S). The subducting slab in this segment presents a relatively smooth northern transition to the north, toward the Central Volcanic Zone (CVZ), and an abrupt southerly transition to segments with a steeper subduction angle (30°) (Cahill & Isacks 1992). Shallow subduction worldwide seems to be associated with the subduction of thickened oceanic crust, particularly seamounts or oceanic plateaux, and with the subduction of oceanic ridges. The latter phenomenon is thought to have been responsible for the Chilean flat-slab segment, beneath which has been subducted the aseismic Juan Fernández Ridge (Nur & Ben-Avraham 1981; Pilger 1981, 1984; Kirby et al. 1996; Gutscher et al. 2000a, b; Yañez et al. 2001, 2002).

The relationships between the magmatic history of the Chilean flat-slab, crustal thickening, and subduction of the Juan Fernandez Ridge requires an understanding of the changes in the chemical signatures of the magmas with time. These signatures can be related to the chemical components of the subducted slab, the overlying mantle wedge, and the crust. Based on a north to south geochemical change of the SVZ as a consequence of crustal thickness variations (Hildreth & Moores 1988), several studies have concluded that the temporal geochemical variation of the Cenozoic volcanic successions is related to crustal thickness variations in these three areas (Kay et al. 1987, 1988, 1991, 1999; López-Escobar et al. 1991; Kay & Kurtz 1995; Stern & Skewes 1995; Kay & Mpodozis 2001, 2002; Nyström et al. 2003). All these studies have used REE patterns to infer the presence or absence of garnet in the magma source. Accordingly, variations in La/Yb ratios have been used as a tool to determine crustal thickness variations with time. A compilation of the chronological characteristics of the volcanic units in the flat-slab segment is presented in Table 4.3.

Although this section deals primarily with the magmatism of central Chile between 28°S and 38°S, the igneous rocks of the Maricunga area in the High Andes between 26°S and 28°S are included because the influence of the flat-slab is recognized down to these latitudes. Accordingly, the following discussion offers a comparison between the magmatic histories of the Cenozoic volcanism in areas along the northern margin (26–28°S), the centre (28.5–32.5°S) and the southern margin (33.5–55°S) of the modern flat-slab segment (Fig. 4.7).

North: Maricunga area (26–28°S)

The magmatic history of the area in this region during late Oligocene–early Miocene times (24–20 Ma) is recorded in the Maricunga Belt, Andean Cordillera (Fig. 4.8), by the Cerros Bravos, La Coipa and Refugo lava dome complexes (Kay et al. 1994; Mpodozis et al. 1995). The volcanic rocks are mainly medium- to high-K calcalkaline andesites and dacites with La/Yb ratios ranging from 7 to 20. The best studied backarc volcanic rocks of this age come from the Segerstrom lavas (c. 23 Ma), which are medium-K basalts and basaltic andesites (Kay et al. 1999). The trace element characteristics of these are magmas indicate equilibration with amphibole-bearing residual assemblages. Comparison of the REE patterns of andesites and dacites with similar SVZ rocks suggests a crustal thickness of about 40–45 km during the interval 24–20 Ma.

Volcanic rocks with ages of 21 to 17 Ma are less voluminous than those of late Oligocene–early Miocene age, and correspond to small dacite dome complexes (21–17 Ma), pyroclastic deposits and dacitic ignimbrites (Kay et al. 1994; Mpodozis et al. 1995). This magmatic quiescence has been considered as a near-magmatic lull and interpreted as an episode of deformation and crustal thickening (Kay et al. 1999; Kay & Mpodozis 2001). After this lull, widespread volcanism developed and produced the Jotabeche Norte centre (27.5°S; c. 18–16 Ma) and the Ojos de Maricunga/Pastillos stratovolcanoes (27°S) (McKee et al. 1994; Kay et al. 1994; Mpodozis et al. 1995). The volcanic rocks here are medium- to high-K calcalkaline andesites and have La/Yb ratios up to 26–35 (Kay et al. 1994). Comparison of the REE patterns of these andesites with those of the SVZ rocks suggests a crustal thickness of over 50 km during the interval 18–16 Ma (Kay et al. 1999; Kay & Mpodozis 2001).

A new volcanic arc started erupting around 16–9 Ma, producing stratovolcano–dome complexes which have been subdivided into two groups (Kay & Mpodozis 2002): (1) Doña Inés volcano, Ojos de Maricunga/Pastillos centre and Cadillal complex, with ages between 16 and 13 Ma (Kay et al. 1994; Mpodozis et al. 1995); and (2) Copiapó ignimbrite/dome complex and centres near Nevado de Jotabeche, with ages from 11 to 9 Ma (Kay et al. 1994). Lavas of group (1) are mostly medium- to high-K calcalkaline andesites with La/Yb ratios between 15 and 26, suggesting a crustal thickness of about 55–60 km. Lavas of group (2) are high-K calc–alkaline andesites and dacites with La/Yb ratios larger than those from group (1). Backarc dacitic ignimbrites erupted at about 15–10 Ma and their geochemical features are similar to those of the western arc rocks (Kay & Mpodozis 2002). Finally, the arc magmatism migrated toward the east after 9 Ma, and high-K calcalkaline andesites were erupted (Mpodozis et al. 1997; Kay & Mpodozis 2000).

Late Miocene volcanism is represented by the Copiapó stratocone (8–7 Ma), the rhyodacitic Jotabeche caldera.
complex and the Pircas Negras mafic andesite flows. The last two units have ages between 8 and 5 Ma (Mpdodzi et al. 1995). All these units correspond to medium- to high-K calcalkaline andesites and dacites, and have $La/\text{Yb}$ ratios up to 57 (Kay et al. 1991), suggesting a crustal thickness of about 60–65 km (Kay et al. 1999; Kay & Abbruzzi 1996), the occurrence of high-angle thrusting in the Main Cordillera (Maksaev et al. 1984), the inception of deformation in the Precordillera (Kay et al. 1993), and the end of backarc volcanic rocks of the Doña Ana Group. Their trace element characteristics are OIB-like suggesting backarc extension over a steeply dipping subduction zone (Kay et al. 1991).

The initial shallowing of the subduction zone has been timed at c. 18 Ma based on the eastward broadening of the arc (Kay et al. 1987, 1988, 1991, 1999; Kay & Abbruzzi 1996), the occurrence of high-angle thrusting in the Main Cordillera (Maksaev et al. 1984), the inception of deformation in the Precordillera close to 18 Ma (Jordan et al. 1993), and the end of backarc volcanicism (Kay & Mpdodzi 2002). Thus, as in the Maricunga area, the magmatic activity corresponds to a near lull during deformation near 20 Ma (Kay & Mpdodzi 2002).

<table>
<thead>
<tr>
<th>Age (Ma)</th>
<th>26°–28°S</th>
<th>28.5°–32.5°S</th>
<th>33°–34.5°S</th>
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<tbody>
<tr>
<td>Maricunga Transect</td>
<td>Arc: Copiapó stratocone (8–7 Ma)–Jotabeché complex/Pircas Negras (8–5 Ma)</td>
<td>Arc: Cerros Bravos-La Crustal thickness: 40–45 km</td>
<td>Arc: Upper Sewell (Teniente Volcanic Complex: 9–7 Ma)</td>
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<td></td>
<td>Backarc: Farallón Negro (8–5 Ma)–Sierra de Famatina (5 Ma)</td>
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<td>Los Broncos/Río Blanco (7–5 Ma)–Bradencriscrimination 'Late Hornblende' dikes (4–3 Ma)</td>
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<td>Backarc: Sierra de San Luís (6–2 Ma)</td>
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<tr>
<td>Crustal thickness:</td>
<td>60–65 km</td>
<td>60 km</td>
<td>45 km</td>
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<tr>
<td>8–5</td>
<td>Arc: Copiapó complex (11–9 Ma) Valle Ancho (9 Ma)</td>
<td>Arc: Tambo Formation</td>
<td>Arc: Pirámide (11–9 Ma) Cerro Aconcagua (10–9 Ma) Lower Sewell (Teniente Volcanic Complex: 11–10 Ma)</td>
</tr>
<tr>
<td>Crustal thickness:</td>
<td>55–60 km</td>
<td>&gt; 45 km</td>
<td>30–35 km</td>
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<tr>
<td>11–9</td>
<td>Arc: Doña Inés volcano-Ojos de Maricunga/Pastillos center-Cadillal complex (16–13 Ma) Backarc: Valle Ancho (15–10 Ma)</td>
<td></td>
<td>Backarc: Huiñcán group (14–6 Ma)</td>
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<tr>
<td>Crustal thickness:</td>
<td>55–60 km</td>
<td>&gt;45 km</td>
<td>30–35 km</td>
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<tr>
<td>16–13</td>
<td>Arc: Jotabeché Norte center (18–16 Ma)–Ojos de Maricunga/Pastillos stratovolcanoes</td>
<td>Reduced volcanic or magmatic lull (20–18 Ma) Arc: Escabroso Formation (21–18 Ma) Tambo Formation (13–10 Ma) Backarc: Cerro Ulúlln (10–11 Ma)</td>
<td>Arc: Farellones Formation (25–7 Ma) Intrusives II</td>
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<td>Crustal thickness:</td>
<td>&gt;50 km</td>
<td>&gt;45 km</td>
<td>30–35 km</td>
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<td>Crustal thickness:</td>
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<td>30–35 km</td>
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<td>28–20</td>
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Centre: El Indio area (28°–32°S)

The late Oligocene–early Miocene volcanicism near 29°S includes the andesitic Cerro Pulido volcanic complex (23 Ma), the Cerro Chacay andesites (20 Ma), and the rhyolitic Cantarito ignimbrites (22–18 Ma) (Kay et al. 1991). The best studied volcanic rocks are from the Doña Ana Group (Maksaev et al. 1984; Kay et al. 1987, 1991) near 30°S, which comprises two formations: the Tilito Formation (Oligocene: 27–23 Ma) and the Escabroso Formation (Early Miocene: 21–18 Ma). These formations are separated by a mild discordance interpreted as a period of
20–18 Ma and is manifested by the Escabroso Formation (Early Miocene: 21–18 Ma) and the intrusion of subvolcanic plutons and porphyritic stocks from the Infiernillo unit (18–15 Ma; Maksaev et al. 1984; Kay et al. 1987). Lavas of the Escabroso Formation consist of pyroxene-bearing medium-K calcalkaline basaltic andesites and andesites with geochemical characteristics consistent with a crustal thickness of about 35–40 km. The Infiernillo unit is composed of high-K calcalkaline andesites and their geochemical features are poorly understood; their relationship with the crustal thickness is difficult to interpret (Kay et al. 1987, 1991).

Arc volcanism during middle Miocene times (17–14 Ma) is represented by the Cerro de Las Tórtolas Formation (Martin et al. 1997a). Lavas of this formation are mainly amphibole-bearing medium- to high-K calcalkaline andesites and dacites, and have La/Yb ratios up to 25, suggesting a crustal thickness of over 45 km (Kay et al. 1991, 1999; Kay & Mpodozis 2001). The Tambo Formation (middle–late Miocene: 12.7–10 Ma; Martin et al. 1997a) is composed of medium- to high-K calcalkaline intermediate to felsic tuffs with La/Yb ratios higher than those from the Cerro de Las Tórtolas Formation (Kay & Mpodozis 2002). At about 8 Ma the volcanism ended in this region and its final activity is recorded further east in Argentina, where andesites with La/Yb ratios up to 25 are found. Toward the east of these andesites, in the Argentine Precordillera, the ignimbrite of the Vallecito Formation (late Miocene: 7–5 Ma; Martin et al. 1997a) represents the only significant arc activity (Maksaev et al. 1984; Ramos et al. 1989; Kay et al. 1991). These rocks correspond to high-K calcalkaline rhyodacites and have La/Yb ratios as high as 35, suggesting a crustal thickness of about 60 km (Kay et al. 1999; Kay & Mpodozis 2001). In the Precordillera and the Sierras Pampeanas, in the backarc environment, the Cerro Blanco centre (31.5°S; c. 6 Ma; Kay et al. 1988; Kay & Abbruzzi 1996) and the Pocho volcanic field in the Sierra de Cordoba (32°S; 7.9–4.5 Ma; Kay & Gordillo 1994) are found.

South: Los Andes–El Teniente area (33–34.5°S)
Unlike volcanism in the Maricunga and El Indio areas, the late Oligocene–early Miocene magmatic history of this region is recorded in the Central Valley, which is a graben with eroded volcanic centres (López-Escobar & Vergara 1997), and in the Andean Cordillera (Fig. 4.9). Most of the studies carried out south of 32°S indicate that the Late Oligocene–Early Miocene volcanic and volcaniclastic deposits that crop out in the Central Valley can be included in the Abanico (37–15 Ma; Wyss et al. 1993; Rivera & Falcón 2000) and Farellones (25–7 Ma; Munizaga & Vicente 1982; Vergara et al. 1988) formations (Vergara & Drake 1979; Rivano et al. 1990; Charrier et al. 1996; Fuentes et al. 2002). In addition to the Abanico and Farellones formations, other Oligocene–Miocene magmatic units exist and correspond to intrusive bodies, mainly stocks, dykes, sills and volcanic necks, which occur within and to the east of the Central Valley. These bodies are called here Intrusives I (34–20 Ma) and correspond to a range of microgabbros, dolerites, basalts and pyroxene andesites (Drake et al. 1976; Gana & Wall 2000). Figure 4.7 shows the areas used to analyse the magmatic development in the modern flat-slab segment.
Volcanic rocks between 37 and 20 Ma define tholeiitic to calcalkaline series, whereas those between 20 and 16 Ma show mainly calcalkaline affinities. The rocks range between basalts and rhyolites, showing a broad compositional range. Trace element features are consistent with pyroxene-dominated mineral residues. La/Yb ratios of rocks from the Abanico Formation range between 2 and 7, whereas those of rocks from the Farellones Formation range between 4 and 9, suggesting a crustal thickness of about 30–35 km during the 37–20 Ma interval (Kay & Kurtz 1995; Kay et al. 1999; Fuentes et al. 2000; Nyström et al. 2003). It would appear that a backarc setting did not exist during this time range, but instead there was a wide volcanic arc comprising two volcanic chains.

During the interval 20–16 Ma, magmatic activity is mainly represented by the Farellones Formation and equivalents (Fig. 4.9) with lavas ranging from medium- to high-K basalts to rhyolites, and with La/Yb ratios from 5 to 14. Their REE patterns suggest a crustal thickness of about 30–35 km (Nyström et al. 2003). However, some lavas from this formation, as well as amphibole-bearing andesitic to rhyolitic porphyries called here Intrusives II (c. 21 and 14 Ma), show chemical features distinctive of adakites (Vatin-Perignon et al. 1996; Sellés 1999a, b; Fuentes et al. 2000) with La/Yb ratios up to 28. Two possible mechanisms and sources can be invoked for the generation of these rocks: (1) melting of mixtures of subducted metabasalt and sediments followed by extensive interaction of these melts with mantle wedge peridotite; and (2) melting of forearc continental crust tectonically removed by subduction erosion and incorporated into the oxidized, hydrated mantle wedge, followed by extensive interaction of these melts with mantle wedge peridotite.

To the east, in Argentina, amphibole-bearing andesitic lavas with ages of about 16 to 15 Ma, and La/Yb ratios of 19–22, are found at 32°40’S in the Cerro Aconcagua region (Kay et al. 1991; Ramos et al. 1996). Even further east, amphibole-bearing isolated subvolcanic bodies occur in the Uspallata Valley, Cerro Colorado (18.9–16.2 Ma; Kay et al. 1991) and other Argentinian localities. La/Yb ratios for these bodies are below 15 suggesting a crustal thickness of about 30–35 km, like that of the Farellones Formation. In summary, unlike the volcanism in the Maricunga and El Indio areas, no magmatic lull is recorded at c. 33°S during the interval 20–16 Ma. However, further south, at 34°S, a magmatic lull occurred at 19–16 Ma (Kay & Mpodozis 2002) and has been interpreted as a period of uplift and crustal thickening associated with compressional deformation (Kurtz et al. 1997).

During the interval 15–9 Ma, volcanic activity concentrated in Argentina, although the upper levels of the Farellones Formation are found in Chile, near the Argentinian border. Around 32–33°S, in Argentina, the La Ramada stratovolcano

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Fig. 4.8. Distribution of the Miocene volcanic rocks of the Maricunga Belt, high Andes Cordillera taken from Mapa Geológico de Chile (Sernageomin 2002).
complex (12.7–10.7 Ma), the Pirámide region centres (11.4–9 Ma) and the Cerro Aconcagua stratovolcano complex (10.3–9 Ma) occur. Lavas from these centres are generally amphibole-bearing medium- to high-K calcalkaline andesites and dacites with La/Yb ratios less than 15, suggesting a crustal thickness of 30–35 km (Kay et al. 1991). Near 34°S, the Teniente Volcanic Complex (15–7 Ma; Godoy et al. 1999) can be divided into three subgroups (Kay & Kurtz 1995; Kay et al. 1999; Kurtz et al. 1997; Kay & Mpodozis 2002): (1) the Maqui Chico subgroup, composed of medium- to high-K calcalkaline basalts to rhyolites with highly variable La/Yb ratios that range from 4 to 22, suggesting an amphibole (and titanite)-based residual mineralogy; (2) the Lower Sewell (11–9.5 Ma) subgroup; and (3) the Upper Sewell (9.3–7 Ma) subgroup. These last two subgroups correspond to medium- to high-K calcalkaline basalts to rhyolites with La/Yb ratios from 7 to 13, but with higher Sm/Yb ratios than those from subgroup (1). Kay et al. (1999) suggested that these subgroups originated in a progressively higher pressure magmatic environment in a thickening crust.

At about 34°S, in the Malargüe fold and thrust belt of Argentina, backarc volcanic rocks from the Huincán group (13.9–5.6 Ma) were erupted (Baldaf 1997; Baldaf et al. 1997). Their geochemical features correspond to typical backarc lavas and have similar REE patterns to those from the Maqui Chico and Lower Sewell subgroups.
At 33–34°S, the late Miocene magmatic activity in the Teniente Volcanic Complex waned at c. 7 Ma (Kay et al. 1999; Kurtz et al. 1997). The Los Bronces/Río Blanco tourmaline breccias (7–4.9 Ma; Warnaars et al. 1985; Stern & Skewes 1995), the El Teniente region Braden Breccia (4.7 Ma; Cuadra et al. 2005), and the “Late Hornblenden” dykes (6.2–2.8 Ma; Kay & Mpodozis 2002) represent the final phase of magmatism in the Main Cordillera. These units mostly comprise medium- to high-K calcalkaline andesites and dacites with La/Yb ratios as high as 77. Although these high ratios suggest a crustal thickness larger than 65 km, Kay et al. (1999) and Kay & Mpodozis (2001) have proposed a more conservative estimate of over 45 km. Additionally, source contamination resulting from forearc tectonic erosion has been invoked for the El Teniente rocks (Stern 1991a; Kay et al. 2005). In the Sierras Pampeanas, in the backarc, centres with ages of 6.4–1.9 Ma in the Sierra de San Luis (33°S; Ramos et al. 1991) are found.

**Tectonic controls on magmatic events in central Chile**

The spatial, temporal and geochemical patterns of Cenozoic magmatism over the flat-slab segment outlined above were presumably controlled by the composition and state of stress of the lithosphere and the crust through which the magmas erupted, by changes in the convergence parameters, and by changes in the physical nature of the subducted slab. In the flat-slab segment, this latter control corresponds to the subduction of the Juan Fernández Ridge, which arrived at this segment around 14 Ma (Yañez et al. 2001, 2002).

Tectonic reconstructions of the convergence between the Nazca (Farallon) oceanic and South American plates are given by Pilger (1984), Pardo-Casas & Molnar (1987) and Somoza (1998). Although these authors examine only the convergence area farther north than 32°S, other studies indicate that the convergence parameters of Somoza (1998) can be assigned to the entire zone from 30°S to 40°S (Jordon et al. 2001).

From about 38 to 28 Ma convergence between the plates was 6 cm/year and with a high degree of obliquity (55°) relative to the South American Plate margin. At about 28 Ma the rate of convergence increased to c. 9 cm/year, and by around 26 Ma it was even faster (15 cm/year), with the degree of obliquity dropping to c. 10° (Somoza 1998). This change in convergence rate at 28–26 Ma was due to the increase in the rate of motion of the subducting plate caused by the break-up of the Farallon Plate (Pardo-Casas & Molnar 1987). Consequently, no change in age or buoyancy of the oceanic lithosphere subducting in the Chile Trench occurred at 28–26 Ma. This period of fast, nearly orthogonal convergence continued until around 20 Ma (Pardo-Casas & Molnar 1987; Somoza 1998).

Along the flat-slab segment, magmatic and tectonic indicators suggest a more compressional Oligocene–Early Miocene (28–20 Ma) tectonic regime in the north than in the south (Kay & Mpodozis 2002). As shown above, calcalkaline arc and backarc magmas were associated with a thicker crust in the Marcungua area (Kay et al. 1994; Mpodozis et al. 1995) than in the Los Andes–El Teniente area (Kurtz et al. 1997; Kay et al. 1999; Godoy et al. 1999; Nystrom et al. 2003; Fuentes et al. 2000). Charrier et al. (2002b) have pointed out that during this period the South American active continental margin between 32°S and 36°S experienced a long-term extensional episode. Volcanic, volcanoclastic and sedimentary deposits accumulated in a north–south orientated, strongly subsident, extensive basin system that can be assigned to an intra–arc basin (Charrier et al. 2002b). Extensional basins have been suggested as far south as 42°S (Muñóz et al. 2000; Jordan et al. 2001).

Concerning the period between approximately 20 and 16 Ma, disagreements exist in the convergence models of the Nazca–South American plates, arising from uncertainties in sea-floor spreading histories. Three different convergence models have been proposed: (a) the rate of convergence increased from 15 to c. 18 cm/year and had only a small obliquity increase (Pardo-Casas & Molnar 1987); (b) the rate of convergence decreased from 15 to 13 cm/year and the degree of obliquity increased from c. 10° to c. 30° relative to the plate margin (Somoza 1998); and (c) the dextral obliquity of convergence slightly decreased (Godoy et al. 1999; Kay & Mpodozis 2002).

The change in the convergence parameters at about 20 Ma marked the beginning of the long period of Miocene compressional deformation and tectonic inversion along the flat-slab segment (Kay & Mpodozis 2002). The reduced magmatic activity during the interval 20–17 Ma recorded in some regions could be explained by a mechanical adjustment between the two converging plates due to the change in convergence parameters. Evidence for this compressional regime comes from: (1) tectonic inversion and uplift at 20–18 Ma at 34°S, suggested by Kurtz et al. (1997) and Godoy et al. (1999); (2) high-angle reverse faulting at 30°S (Maksaev et al. 1984) and inception of Precordillera thrusting (Jordan et al. 1993) at 18 Ma; (3) eastward expansion of the magmatic arc into the Calingasta/Usppallata Valley and western Precordillera (Kay et al. 1988). Additionally, from 32°S to 36°S during the interval 20–16 Ma, contractional conditions prevailed with the tectonic inversion of pre-depositional extensional faults taking place (Charrier et al. 2002b). Nevertheless, Nystrom et al. (2003) have indicated that, at least near 33°S, extensional conditions continued until c. 18 Ma, an age that in accordance with Kay et al. (1987, 1991, 1999) marks the initial shallowing of the subduction zone beneath the flat-slab segment. From 32°S to 34°S the frontal arc moved toward the east and an eastward broadening of the arc occurred, which widened the Central Depression as far as the western Precordillera in Argentina (Kay & Mpodozis 2002). Evidence for this eastward migration extends from 30°S to 36°S (Kay & Abruzzi 1996; Kay et al. 1987), but is not recorded in the Marcungua area where vigorous arc volcanism reinitiated there at 16 Ma (Kay et al. 1994).

High La/Yb ratios in some lavas erupted during this time, although indicative of high pressure mineral residuals, cannot be explained by crustal thickening (Sellés 1999b; Kay & Mpodozis 2002). The use of lower crustal contamination models like the MASH (melting, assimilation, storage, homogenization) model of Hildreth & Moorbath (1988) allows an explanation for long-term systematic REE changes in the Miocene flat-slab magmas (Kay et al. 1987, 1991; Kay & Abruzzi 1996), but transient variations require unreasonable changes in crustal thickness over a short period of time (Kay & Mpodozis 2002). Therefore, Kay & Mpodozis (2002) have related such high La/Yb ratios to the melting of forearc continental crust tectonically removed by subduction erosion and incorporated into the oxidized, hydrated mantle wedge (von Huene & Schlörl 1991; Stern 1991a; Kay & Mpodozis 1999). However, melting of mixtures of subducted metabasalt and sediments cannot be discarded.

At 14 Ma, the arrival of the Juan Fernández Ridge occurred. The subsequent period (14–9 Ma) is characterized by thrusting related to compressional shortening and andesitic arc volcanism along the entire flat-slab segment. All convergence models indicate a rate faster than 10 cm/year and a degree of obliquity of c. 12°. Kay & Mpodozis (2002), assuming that the Juan Fernández Ridge affected a region of about 150–200 km, suggested that the passage of the ridge through the Marcungua area at 14 Ma correlates with the transition from a stratovolcanic chain to an isolated ignimbrite complex at about 11 Ma, producing magmas with high La/Yb ratios (e.g. Copiapó Complex). Ignimbrites erupted in the backarc at 15–14 Ma when the ridge arrived, and other similarly felsic volcanic units were formed as the ridge migrated south. At 12–10 Ma, the ridge arrived at the backarc, coinciding with a change from the andesite-dominated Cerro de Las Tórtolas Formation to the dacite-dominated Tambo Formation. As in the Marcungua area, lavas of this latter formation have high La/Yb ratios.
Ridge subduction cannot explain coeval events in the Los Andes–El Teniente area, because this region is located south of the area affected by the ridge. The increase in La/Yb ratios from 14 to 10 Ma in this area coincides with crustal thickening associated with compressional shortening (Kurtz et al. 1997; Kay et al. 1999; Godoy et al. 1999). Deformation at 14–9 Ma in this region (Baldauf et al. 1997) is similar in age to that inferred for the Precordillera in the north (12–10 Ma) by Jordan et al. (1993). Therefore, crustal shortening and thickening have been associated with a regional compressional regime that extended beyond the region affected by the subduction of the Juan Fernández Ridge (Kay & Mpodozis 2002). Overall, this situation suggests that the Oligocene–early Miocene extensional basin between 33°S and 36°S could have been developed as far as 30°S and that processes of crustal thickening may be associated with tectonic inversion of the extensional basin, between 30°S and 36°S, in addition to the subduction of the Juan Fernández Ridge (Charrier et al. 2005a).

The subduction of the NE arm of the Juan Fernández Ridge changed around 10 Ma due to the subduction of the east–west segment of the ridge. In the Maricunga area, the arc front shifted eastward at 7–5 Ma, coinciding with the passage of the NE-trending arm. The magmas in this new arc (Copiapó stratocone and Volcan de Fuego complex) have very high La/Yb ratios, and the presence of residual mineralogy. The chemistry of these magmas has been interpreted as the result of an episode of forearc subduction erosion as the slab shallowed and the front migrated eastward (Kay & Mpodozis 1999, 2000, 2002). At 4 Ma, the NE-trending ridge segment had passed the Maricunga area, allowing the modern CVZ to be stabilized before 2 Ma.

In the El Indio area, arc volcanism ceased at 5 Ma and the eruption of the backarc Cerro Blanco centre at 7 Ma appears to be related to the passage of the bend in the ridge below the eastern Precordillera (Kay & Mpodozis 2002). In contrast in the Los Andes–El Teniente area, arc volcanism ceased at 9 Ma with the last eruptions of the La Ramada/Cerro Aconcagua/Pirámide centres, coinciding with the arrival of the bend in the ridge below the frontal arc (Kay & Mpodozis 2002). Further south, at 34°S, magmatic patterns cannot be explained by ridge subduction, because this region is outside of the influence of the ridge. However, again the magmatic and tectonic evolution in this region is similar to that of the north. The arc front migrated eastward at 7 Ma, coinciding with a major compressional deformation event (Kurtz et al. 1997; Godoy et al. 1999). Although the end of volcanism in the modern flat-slab frontal arc and the arc migration events at 8–4 Ma in the Maricunga and Los Andes–El Teniente areas coincide with the subduction of the kink of the ridge axis and the arrival of the east–west ridge segment, the ridge was only a perturbation in a much larger set of driving forces affecting the Andean margin (Kay & Mpodozis 2002). Such forces appear to be related to variations in the Nazca–South American plate convergence parameters.

The Central Depression and Coastal Cordillera late Oligocene–early Miocene volcanism (37°–44°S) (J.M.B., R.T.V. & C.R.S.)

In south-central Chile between 37°S and 44°S, late Oligocene–early Miocene volcanic and subvolcanic rocks are locally exposed both to the east and within the Main Andean Cordillera, as well as within the Central Depression and along the coast on the western slope of the Coastal Cordillera (Fig. 4.10). Particularly good exposures of these partially eroded Oligocene–Miocene volcanic complexes have been identified in the Los Angeles–Temuco segment within the Central Depression and along the coast on the western slope of the Coastal Cordillera at Bahía Capitanes, Caleta Estaquilla, Caleta Parga, Ancud and Guapi Quilan islands, the latter to the SW of Quellón town (Fig. 4.10). These exposures represent remains of individual volcanic complexes previously grouped together as the Coastal Cordillera Volcanic Belt (Vergara & Munizaga 1974), the Eocene–Miocene Central Depression Volcanic Belt (López-Escobar et al. 1976), the Central Depression Upper Oligocene–Miocene Volcanic Belt (Stern & Vergara 1992) and/or the Coastal Magmatic Belt (Muñoz et al. 2000). Both Central Depression and Coastal Cordillera volcanic complexes define a NNE-trending belt interpreted by Muñoz et al. (2000) as the late Oligocene–early Miocene volcanic front (Fig. 4.10).

Central Depression and Coastal Cordillera Oligocene–Miocene volcanic complexes were emplaced upon Palaeozoic–Triassic metamorphic basement and the effusive products underlie, cover or are interbedded with late Oligocene to early Miocene continental and/or marine sedimentary sequences. These relationships indicate that this magmatic activity was synchronous with the opening and/or subsidence of forearc and intra-arc continental and/or marine sedimentary basins (i.e. Temuco–Labranza, Valdivia, Osorno–Llanquihue and Chiloe basins; Fig. 4.10) (Cisternas & Frutos 1994; Martínez & Pino 1979; Muñoz et al. 1997, 2000; Elgueta et al. 2000b). Other Oligocene–Miocene intra-arc continental or marine sedimentary basins were also open along and to the east of the Main Cordillera between 37°S and 44°S. For example the Curahullán continental basin (Niemeyer & Muñoz 1983; Suárez et al. 1992; Suárez & Emparan 1995), Lago Ranco marine basin (Campos et al. 1998) and Ninilhuau marine and continental basin (Spalletti & Dalla Salda 1996). These Main Cordillera and extra-Andean sedimentary basins were also temporally related to Oligocene–Miocene volcanism, (i.e. Lago Ranco volcanics (Campos et al. 1998) and El Maitén Volcanic Belt east of the Main Cordillera (Rapela et al. 1988).

K–Ar geochronology

All available whole-rock K–Ar ages and cited localities for the Oligocene–Miocene volcanic complexes between 37°S and 44°S are shown in Figure 4.10. Most of K–Ar determinations in the Los Angeles–Temuco segment are in the range 29–20 Ma (Vergara & Munizaga 1974; Rubio 1993; Stern & Vergara 1992; Troncoso 1999; Elgueta et al. 2000b; Muñoz et al. 2000). Muñoz et al. (2000) reported K–Ar ages from 28 to 25 Ma for samples in the Los Angeles–Temuco segment and from 29 to 22 Ma in the Guapi Quilan Islands Complex, SW of Quellón. García et al. (1988) and Muñoz et al. (2000) obtained ages of 21 and 24 Ma for samples from the volcanic neck at Punta Polocué, and 20 to 23 Ma for samples in the vicinity of Ancud, both belonging to the Ancud Volcanic Complex. Stern & Vergara (1992) and Muñoz et al. (2000) determined ages of 25 and 23 Ma, respectively, for glassy compacted fragments separated from a rhyolitic pyroclastic flow cropping out within Ancud city. Also, a conventional U–Pb age determination for zircons separated from a 10 cm ashfall deposit separating two coal layers interbedded in a continental sedimentary sequence within the Catamutún coal mine in the western portion of the Central Depression, north of Osorno city, yielded an age at the Oligocene–Miocene boundary (23.5 ± 0.5 Ma; Elgueta et al. 2000b), in good agreement with the K–Ar data. Similar late Oligocene to early Miocene ages have been reported in the western slope of the Main Cordillera in the Colbún area (36°S; Vergara et al. 1988, 1999) and north, west, east and SE of Santiago (33–34°S; Wall & Lara 2001; Nyström et al. 2003; Fuentes et al. 2002; Muñoz et al. 2006), confirming the prolongation of the Oligocene–Miocene volcanism north of 37°S.

Two K–Ar ages for a basaltic andesite and a dacite from Bahía Capitanes Complex gave ages of 32.9 and 27.5 Ma (Muñoz et al. 2000), the former older than most of the other dated samples. Due to alteration assemblages indicative of
Fig. 4.10. Distribution and K–Ar ages for the Central Valley and Coastal Cordillera late Oligocene–early Miocene volcanic complexes, and related sedimentary basins between 37 and 44°S, including contemporaneous sedimentary basins and volcanic units within and east of the Main Cordillera (modified after Muñoz et al. 2000). Also shown are the approximate location of the Oligocene–Miocene and current active CSVZ volcanic fronts.
magma–seawater interaction, no viable K–Ar ages have been determined yet for samples from the Parga and Estaquillas complexes. Available K–Ar data between 37°S and 44°S suggest an older Coastal Cordillera Tertiary magmatic pulse in Eocene–early Oligocene times, represented by dacitic biotite-and quartz-bearing sills cropping out near Castro city (37.2 ± 1.2 Ma; Valdivia & Valenzuela 1988; Muñoz et al. 2000), altered dacitic porphyry at Río Futa (52.7 Ma in sercite; Peri & Rivera 1991), and tonalitic bodies at Metalqui locality south of the Ancud Volcanic Complex (39.6 ± 0.3 Ma; Arenas & Duhart 2003). Similar Eocene–early Oligocene ages have been reported in the western slope of the Main Cordillera at the lower part of the volcanic sequence in the Colbún area (36°S; Vergara et al. 1988, 1999) and north and SE of Santiago (Fuentes et al. 2002; Muñoz et al. 2006), confirming the existence of this older pulse. Whether the older early Oligocene age obtained at Capitanes belongs to this earlier pulse is still unresolved. Unfortunately, the alteration assemblage does not permit reliable ages for the Parga, Capitanes and Estaquillas complexes.

**Petrography**

Oligocene–Miocene Central Valley and Coastal Cordillera partially eroded volcanic complexes include porphyritic to aphanitic lava and pyroclastic flows, ashfall tuffs, hydrothermal and hyalopilitic breccias, glassy domes or bodies, columnar jointed volcanic necks, sills and dykes. Most of the lava flows, sills, necks and dykes range from basaltic to andesitic in composition, although dacitic and rhyolitic members are also represented at some localities (e.g. Los Angeles, Capitanes, Estaquillas and Ancud complexes). Although there is not a clear north–south petrographic or mineralogical trend, orthopyroxene phenocrysts are common in andesites of the Los Angeles–Temuco (37–39°S) segment and in the Guapi Quilan complex (43°30'S), but they are rare or absent in basalts and basaltic andesites in the Caleta Parga, Capitanes, Estaquilla and Ancud complexes (41–42°S), which show mostly clinopyroxene and olivine as phenocrysts and microcrysts. Plagioclase, clinopyroxene, orthopyroxene (+ hornblende) porphyritic subalkaline andesitic lava flows, necks, stocks and dykes, and quartz + K-feldspar dacites dominate in the Los Angeles–Temuco segment, with trachitic or glassy groundmass (Vergara 1982; Rubio 1993; López-Escobar & Vergara 1997; Troncoso 1999). No olivine basalts have been reported in this segment and red-coloured hornblende phenocrysts are common in the vicinity of Temuco.

Necks and lavas in the Capitanes, Estaquilla and Caleta Parga volcanic complexes are porphyritic to aphanitic plagioclase, clinopyroxene, olivine basalts and basaltic andesites, with trachytic, interstitial, ophitic or intergranular groundmass (Troncoso et al. 1994; Troncoso 1999; Muñoz et al. 2000; Vargas 2001). Dacitic domes and subvolcanic bodies, which occur east of the coastline and to the south of Estaquilla, have porphyritic textures and show trachytic groundmass. Pyroclastic rocks including accretionary lapilli and pyroclastic flows interbedded with volcaniclastic marine sandstones and polymictic conglomerates are also common in the Caleta Parga and Estaquillas complexes.

The Ancud Volcanic Complex (Vergara & Munizaga 1974; Valenzuela 1982, García et al. 1988; Stern & Vergara 1992; Muñoz et al. 1997, 2000) includes basaltic to andesitic lava flows and volcanic necks (Fig. 4.11), rhyolitic pyroclastic flows, and black-coloured obsidian bodies. Basaltic andesite necks and lava flows have fine-grained and porphyritic textures with labradorite plagioclase, clinopyroxene and olivine in a glassy groundmass. Pyroclastic rocks are abundant at Pumillahue bay and are mainly represented by white, glassy, lithic-rich rhyodacitic pyroclastic flows (Figs 4.12 & 4.13) with fresh plagioclase, orientated fragments of both white and compacted
black pumice, partially altered volcanic lithics, rounded clastic sedimentary fragments and carbonized wood, within a recrystallized perlite, silicic glass.

The secondary alteration assemblage (mainly calcite, silica, chlorite and/or zeolites, with silica, calcite and zeolites filling amygdales, veins and/or as disseminations) is well represented in all the volcanic complexes along the present-day coastline (Capitanes, Parga, Estaquillas and Ancud volcanic complexes) and has been suggested to represent magma–seawater interaction during eruptions (Muñoz et al. 2000). Also, hyalopilitic breccias in the Capitanes Complex have been interpreted as the centre of a submarine volcanic system (Alfaro et al. 1994). Clear evidence of hydrothermal alteration (e.g. infilled vesicles, chaledony, calcite and/or limonite veins, hydrothermal breccias and supergene clay minerals) and Au and As geochemical anomalies have been detected in most of the late Oligocene–early Miocene Central Valley and Coastal Cordillera volcanic complexes, but have been described in detail only in the Parga Complex (Vargas 2001).

Petrochemistry

Although most of the samples from the Oligocene–Miocene Central Valley and Coastal Cordillera volcanic complexes are arc-type subalkaline rocks in terms of silica, total alkalies, TiO₂, Ba, Nb and REE contents, an important number of samples from volcanic complexes along the coastline on the western slope of the Coastal Cordillera (Capitanes, Estaquillas, Parga and Ancud complexes) show alkali affinities and Ba/La, La/Nb and/or La/Yb ratios similar to oceanic island basalts (Muñoz et al. 2000).

In the Los Angeles–Temuco segment, low TiO₂ subalkaline orthopyroxene-bearing arc-type andesites show moderate LREE enrichment relative to HREE (La/Yb = 4–7.3), LIL enrichment relative to LREE (Ba/La = 24–31) and HFSE depletion relative to REE (La/Nb > 2.3), compared to ocean island basalts (López-Escobar et al. 1976; López-Escobar & Vergara 1997; Muñoz et al. 2000; see Fig. 4.14A & B). These ranges for La/Yb, Ba/La and La/Nb are all similar to those reported from stratovolcanoes and minor eruptive centres (MEC) along the current volcanic front at a similar latitude (37°S to 43°S, Hickey et al. 1986; Hickey-Vargas et al. 1989; López-Escobar & Vergara 1997), confirming the subalkaline affinities shown by petrographic and mineralogical observations.

As noted by Muñoz et al. (2000), not all basalts and basaltic andesites from the Capitanes, Parga, Estaquillas and Ancud volcanic complexes have clear geochemical signatures indicative of the incorporation of components derived from the dehydration of subducted oceanic lithosphere. Some basalts from Caleta Parga and some basaltic andesites from Bahía Capitanes have relatively high TiO₂ (≥ 2 wt%) compared to typical arc-type calcalkaline or tholeiitic mafic rocks, indicating a more alkaline composition. Also, REE contents in these complexes are higher than those in andesites from the Los Angeles–Temuco complexes, although their La/Yb ratios (4.5–5.2) are similar. In contrast, Ba/La (8.8–21.2) and La/Nb (1.4–2.1) ratios (Fig. 4.14A & B) of these mafic samples are more similar to oceanic island basalts and extend to significantly lower values than the samples from the Los Angeles–Temuco segment and the basalts from stratovolcanoes and MEC along the volcanic front of the active Central Southern Volcanic Zone (CSVZ) of the Andes.

Similarly, some basaltic andesite samples from the Ancud Complex have high TiO₂ (≥ 2 wt%) and more alkaline affinities, similar to the samples from Caleta Parga, whereas other basalts and basaltic andesites are clearly subalkaline. La/Yb ratios (4.6–5.9) of basalts and basaltic andesites from Ancud are similar to those from the Bahía Capitanes, Caleta Parga, and the Los Angeles–Temuco segment. In contrast, Ba/La ratios (12.5–19.2) are similar to the samples from Bahía Capitanes and Caleta Parga, but lower than those from Los Angeles–Temuco, and from stratovolcanoes or MEC basalts of the current volcanic front of the CSVZ, while La/Nb ratios (1.6 to > 3.5) are transitional between the values for samples from the Bahía Capitanes–Caleta Parga and the Los Angeles–Temuco segment. The silicic pyroclastic rocks from Ancud have high REE contents and a significant negative Eu anomaly, but similar La/Yb, Ba/La and La/Nb ratios to the associated mafic rocks.

![Image](https://via.placeholder.com/150)
**Sr, Nd and Pb isotopes**

In addition to petrochemical alkaline affinities and trace element similarities to ocean island basalts, basalts from Parga have initial Sr and Nd isotopic compositions (Fig. 4.14C) more primitive than most other samples determined for the Oligocene–Miocene Central Valley and Coastal Cordillera complexes (Muñoz et al. 2000) and the current volcanic front of the active CSVZ, including MEC basalts (López-Escobar & Vergara 1997). On the other hand, basaltic andesites, dacites and rhyolitic members from Capitanes and Estaquillas have higher initial ⁸⁷Sr/⁸⁶Sr ratios and lower initial ¹⁴⁳Nd/¹⁴⁴Nd isotopic ratios, with the higher Sr and lower Nd values represented by rhyolitic members, suggesting possible assimilation of Palaeozoic–Triassic crust. Some basaltic andesites from the Ancud Complex have primitive initial Sr and Nd isotopic compositions similar to basalts from Caleta Parga, while other mafic samples from this complex are similar to CSVZ and MEC along the present-day volcanic front (Fig. 4.14C). One basaltic sample from Los Angeles–Temuco complexes shows an Sr and Nd isotopic composition similar to that of the Caleta Parga complex, whereas other mafic samples from these complexes are similar to CSVZ and MEC along the present-day volcanic front (as is the Guapi Quilan complex).

As reported by Muñoz et al. (2000), Pb isotopic composition of the more mafic rocks at Parga are similar to magmas erupted from the active CSVZ, whereas dacites and rhyodacites show higher values of the Pb isotopic ratios, again possibly due to crustal assimilation. In contrast, the glassy rhyodacitic pyroclastic flow in the Ancud Complex is isotopically similar to more mafic rocks from the same complex.

**Petrogenesis**

The Central Valley and Coastal Cordillera volcanic complexes between 37°S and 44°S represent the late Oligocene–early Miocene volcanic front located between 80 and 100 km to the west of the current Andean volcanic front at a similar latitude (Fig. 4.10). Alteration assemblages, especially those in the Coastal Cordillera complexes, indicate magma–seawater interaction and submarine volcanism, suggesting that part of this volcanic front formed as an island arc built on the Palaeozoic–Triassic continental crust.

The Oligocene–early Miocene volcanism, the opening of sedimentary basins and the initiation of development of the present-day Central Valley in south-central Chile (37–44°S) all occurred during a widespread regional episode of crustal extension (Muñoz et al. 2000; Jordan et al. 2001; Muñoz & Stern 2003). This regional episode of late Oligocene to early Miocene extension, the related volcanism and the opening and subsidence of sedimentary basins have also been reported along and in the western slope of the Main Cordillera north of 37°S to the north, east and south of Santiago (Nyström et al. 1993, 2003; Thiele et al. 1991; Vergara et al. 1999; Godoy et al. 1999; Charrier et al. 2005a). Evidence for Oligocene–Miocene crustal extension includes low-angle normal faulting bounding the western edge of the present-day Central Valley (Muñoz et al. 2000) and the Cura Mallín sedimentary basin on the eastern flank of the Main Andean Cordillera (Spalletti & Dalla Salda 1996; Jordan et al. 2001). Also, regional negative gravity and positive magnetic anomalies (Muñoz & Araneda 2000) and interpretation of seismic reflection profiles (McDonough et al. 1997) suggest a thin crust (33 km) below the Central Valley and the Main Cordillera as a consequence of crustal thinning during late Oligocene–early Miocene extension. The late Oligocene (+29 Ma) initiation of the Central Valley and Coastal Cordillera volcanism and sedimentary basin opening were coeval with plate reorganization in the SE Pacific (Cande & Leslie 1986; Tebbens & Cande 1997), resulting in an increase in plate convergence rate below the southern Andes that changed the geometry of plate subduction from oblique to almost orthogonal (Pardo-Casas & Molnar 1987; Somoza 1998).

No obvious chronological, petrographic, chemical or isotopic north–south trend may be defined for the Oligocene– Miocene Central Valley and Coastal Cordillera volcanic rocks (Muñoz et al. 2000). However, subalkaline orthopyroxene andesites predominate in the Los Angeles–Temuco segment and the Guapi Quilan complexes, the northern and southern representatives of the Central Valley and the Coastal Cordillera volcanism between 37°S and 44°S. Primitive basalts and basaltic andesites with alkaline affinities are important in the Capitanes, Parga, Estaquillas and Ancud volcanic complexes along the Pacific coastlines of the Coastal Cordillera.

All Oligocene–Miocene mafic volcanic rocks have Pb isotopic compositions similar to the samples from the current volcanic front of the CSVZ at the same latitude, but this is not always true for trace element ratios and Sr and Nd isotopic compositions. The occurrence at some localities of mafic rocks with alkaline affinities, low La/Yb, Ba/La (<17) and La/Nb (<1.6) ratios and primitive Sr and Nd isotopic compositions (e.g. Parga and Ancud complexes), suggest local absence or at least lowered input of slab-derived hydrous fluids into the mantle source region. Also, Central Valley and Coastal Cordillera Oligocene–Miocene volcanism does not exhibit a negative correlation between either Ba/La or La/Nb and La/Yb, as do the current volcanic arc magmas at the same latitude. Trace element geochemical data suggest that dehydration of subducted slab may not have been the fundamental mechanism driving magma genesis for some Coastal Cordillera volcanic complexes. The fact that all the samples have similar Pb isotopic compositions and many have Ba/La and La/Nb ratios similar to those of the modern Andean arc magmas has been interpreted as the effects of source region contamination of the subcontinental mantle during earlier episodes of subduction of oceanic lithosphere below this portion of the southern South American continent (Muñoz et al. 2000). Geochemical data suggest that late Oligocene–early Miocene Central Valley and Coastal Cordillera magmatism was directly related to slab dehydration, but that chemically heterogeneous mantle sources were also involved, some resembling the sources of oceanic basalts and being apparently free of slab-derived components. Muñoz et al. (2000) suggested that some of these magmas (e.g. Parga, Capitanes and Ancud volcanic complexes) resulted from an asthenospheric slab window which developed in association with changes in convergence rate and direction, and with slab-rollback of the subducting Nazca Plate. This scenario is envisaged to have developed in response to a transient episode of invigorated asthenospheric wedge circulation caused by the three-fold increase in late Oligocene trench-normal convergence rates between the Nazca and South American plates. Thus, during the late Oligocene, more rapid (10 cm/year) subduction at a steeper angle resulted in the opening of a slab window and upwelling and melting of asthenospheric mantle uncontaminated by slab-derived components. This led in turn to interaction of these melts with subcontinental mantle lithosphere containing stored slab-derived components, and to crustal extension and subsidence (see model in Fig. 4.15).

The ending of Central Valley and Coastal Cordillera volcanism between 37°S and 44°S during the early Miocene (±20 Ma) did not coincide with any obvious change in plate convergence rates. It did, however, broadly coincide with closure of the Central Valley and Coastal Cordillera sedimentary basins (e.g. Osorno–Llanquihue basin) which occurred during early (Martínez & Pino 1979) or middle Miocene (Elgueta et al. 2000b) due to tectonic inversion that gently folded the sedimentary sequences. Early Miocene tectonic inversion also produced the closure and folding of the continental sedimentary basins and related sedimentary sequences along the Main Cordillera.
north and south of Santiago (Godoy et al. 1999; Charrier et al. 2005a). During the early to middle Miocene, westward extension of the continental margin and a decreasing angle of subduction returned the arc to its current position in the Main Cordillera and caused deformation, uplift and exposure of the late Oligocene–early Miocene volcanic and sedimentary rocks.

**Magmatism in the North Patagonian Andes (40–47°S)**

(M.A.P., D.M. & F.E.)

The North Patagonian Andes, developed between the North Patagonian Massif and the Coastal Cordillera, extend from about 40°S to 47°S, where the triple-junction between the Nazca, South America and Antarctic plates is located. The North Patagonian Andes is mainly the result of intensive subduction-related magmatism active since mid-Jurassic time. Evidence for this magmatism is provided by the Early Cretaceous–late Cenozoic North Patagonian Batholith (Pankhurst et al. 1999), the Jurassic–Eocene arc and backarc volcanism (Parada et al. 2001b), and the Holocene volcanoes of the southern segment of the Southern Volcanic Zone. One of the most striking structural features of the North Patagonian Andes is the north–south Liquiñe–Ofqui Fault Zone, a dextral strike-slip fault system recognized for c. 900 km from 39°S to 47°S (Thiele et al. 1986), that played a key role in the emplacement of Neogene–Holocene magmatic rocks.

The North Patagonian Batholith

The magmatic evolution of the North Patagonian Andes has been mainly documented in the arc domain, particularly in the North Patagonian Batholith, where distinct plutonic events have been identified (Fig. 4.16) by using different radiometric methods (Rb–Sr isochrons, K–Ar and 40Ar/39; Halpern & Fuenzalida 1978, Bartholomew 1984; Pankhurst et al. 1999; Suárez & De La Cruz 2001). This batholith occupies the axis of the North Patagonian Andes and constitutes, together with the South Patagonian Batholith, one of the largest batholiths on earth. Identification of individual plutons within the batholith is difficult due to their similar lithology and the dense forest cover. The western margin of the batholith is defined by intrusive contact with low-grade metasedimentary rocks of the Late Palaeozoic subduction complex, whereas the eastern margin is marked by an intrusive contact with the Jurassic Ibáñez Formation.

The North Patagonian Batholith was formed episodically from Early Cretaceous to late Tertiary times. Based on recently published geochronological data (Pankhurst et al. 1999; Suárez & De La Cruz 2001; Cembrano et al. 2002) it is possible to recognize three main plutonic events: Cretaceous, early Miocene and late Miocene–Pliocene. Unlike the Mesozoic–Cenozoic plutonic development in central Chile, the locus of the North Patagonian Batholith components does not change significantly with time. For example, along the northern (40–42°S) segment of the batholith, two north–south plutonic belts are recognized occupying the axis of the Cordillera: an eastern Early Cretaceous belt and a western early to late Cenozoic belt (Carrasco 1995). The boundary between the two belts roughly coincides with the Liquiñe–Ofqui Fault Zone. Further south (44–47°S) the batholith is, in a broad sense, formed by western plutons forming a poorly defined Early Cretaceous belt, a central Miocene to Pliocene belt of isolated plutons, and eastern Early to mid-Cretaceous intrusions forming the widest plutonic belt of the batholith (see Pankhurst et al. 1999; Suárez & De La Cruz 2001).

The batholith as a whole exhibits a wide lithological spectrum with a typical calcalkaline affinity. The Cretaceous granitoids include mainly metaluminous granodiorites and tonalites, whereas the Miocene–Pliocene granitoids are composed of different lithologies that cover a wide spectrum from gabbro to granite (Parada et al. 1987; Pankhurst et al. 1999). Peraluminous granites have been found as part of the late Miocene plutonic event closely related to the Liquiñe–Ofqui Fault Zone (Parada et al. 1987, 2000).

The North Patagonian Batholith shows progressive variation in Sr and Nd isotopes with time, from enriched Early Cretaceous granitoids to depleted late Cenozoic rocks (Pankhurst et al. 1999). This isotopic evolution, which is similar to that shown by the Late Palaozoic–Early Cretaceous evolution of the Coastal Batholith in central Chile, has been attributed to mixed sources, in different proportions, of mafic crustal underplate and lower crust (Pankhurst et al. 1999). An alternative explanation, however, based on thermochronological and geobarometric evidence, has been invoked for the origin of the late Cenozoic rocks, such as the Queulat Complex in the vicinity of the Liquiñe–Ofqui fault Zone at 44°30’S. This complex includes 18–16 Ma deep-seated (19–24 km) quartz-diorites and tonalites (Parada et al. 2000) and 10 Ma peraluminous shallow (<10 km) granitoids (Hervé et al. 1993), that would have been formed by partial melting of a lower crust that underwent modifications in its thermal structure. Such modifications would have been derived from rapid exhumation events along...
the Liquiñe–Ofqui Fault Zone (Hervé et al. 1993; Parada et al. 2000) during a contractional deformation generated from a nearly orthogonal subduction episode.

**The Mesozoic–Eocene backarc volcanism of the North Patagonian Andes**

The volcanic rocks located to the east of the North Patagonian Batholith are the result of a Meso-Cenozoic backarc extensional regime (Bartholomew & Tarney 1984). The area where these rocks are continuously exposed corresponds to the eastern flank of the North Patagonian Andes of the Aysén Region between 45°S and 47°S.

The products of the Mesozoic volcanism, erupted over highly deformed Late Palaeozoic metamorphic rocks cropping out south of 46°S, have been placed into a Jurassic Ibáñez Formation and a Cretaceous Divisadero–Ñirehuao Formation. The Eocene volcanism is represented by the Balmaceda basalts, the rhyolitic tuffs of the Chile Chico Formation (Niemeyer 1975), and the lower section of the Meseta Buenos Aires plateau basalts (Charrier et al. 1979; Petford et al. 1996).

The Ibáñez Formation includes felsic to intermediate volcanic rocks underlying an Early Cretaceous marine succession (Bell & Suárez 1997) of the Coyhaique Group, and is part of the Jurassic Large Igneous Province of Patagonia associated with the break-up of Gondwana (Féraud et al. 1999; Pankhurst et al. 2000). K-feldspar and biotite \(^{40}\)Ar/\(^{39}\)Ar ages and U–Pb zircon data indicate an age interval between 130 and 160 Ma for the deposition of this formation (Pankhurst et al. 2000; Parada et al. 2001b). The integration of field, radiometric, geochemical and scarce isotopic data available for these Jurassic volcanic rocks and related plutons indicates a significant crustal participation in the origin of these magmas (Parada et al. 1997; Pankhurst et al. 2000).

The Divisadero–Ñirehuao Formation, which represents the most intensive event of the backarc volcanism, overlies the Ibáñez Formation and marine rocks of the Coyhaique Group. It is composed mainly of felsic tuffs and rhyolites, although flat-lying basalts are also found in the eastern exposures of this unit. Available radiometric data for this formation indicate that Cretaceous volcanism in this region developed during the interval 115–75 Ma, slightly after the Cretaceous plutonic event of the North Patagonian Batholith (Parada et al. 2001b). Following this, after a period of magmatic quiescence lasting about 20 Ma, an Eocene volcanic event developed during the interval 46–55 Ma (Parada et al. 2001b) giving rise to olivine flood basalts and felsic tuff components of the volcanic-sedimentary Chile Chico Formation (Niemeyer 1975).

Rocks belonging to the Jurassic–Neogene successions in Chilean Patagonia between 43°S and 46°S have been affected by very-low to low-grade metamorphism (Aguirre et al. 1997). Differences in grade are related to the age of the rock successions, with the youngest metamorphosed to zeolite and the oldest to greenschist facies. In addition, in the case of the Jurassic Ibáñez Formation, a thermal input by Cretaceous granitoids has almost completely obliterated an earlier low-grade metamorphic event.

Based on geochemical and isotopic compositions, the backarc volcanic rocks between Coyhaique and Cochrane have been divided into two magmatic units (Parada et al. 2001b) referred to as the Northern and Southern magmatic domains. The Southern Magmatic Domain begins at about 46°30’S, and is composed of basalts and intermediate-composition volcanic rocks. They are typically calcalkaline and have enriched Sr–Nd

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**Fig. 4.16.** Distribution of the plutonic components of the Northern Patagonian Batholith.
isotopic values that are compatible with a lithospheric mantle source. In contrast, basalts from the Northern Magmatic Domain have alkaline affinities and depleted to slightly deplet ed Sr–Nd signatures, suggesting an asthenosphere-dominated source. With regard to the felsic rocks, those in the Southern Magmatic Domain are more isotopically (Sr–Nd) enriched than the equivalent rocks of the Northern Magmatic Domain. These isotopic distinctions between the two domains are attributable to: (i) the participation of Palaeozoic metamorphic basement in the origin of the volcanic rocks of the Southern Magmatic Domain; and (ii) a greater degree of tectonic extension in the Northern Magmatic Domain and, consequently, only minor participation of a thinned lithosphere (Parada et al. 2001b).

The Patagonian Plateau Basalts at about 47°S: the role of Chile Ridge subduction

The Cenozoic geodynamic evolution of the western margin of South America has been dominated by the subduction of different lithospheric plates and various oceanic spreading ridges (e.g. Cande & Leslie 1986). A particularly prominent event in the recent history of the margin has been the oblique subduction of the South Chile Spreading Ridge (SCR) beneath the South American Plate, a process that began at 14–15 Ma. At this time a segment of the ridge collided with the Chile Trench near Tierra del Fuego (c. 55°S; Cande & Leslie 1986), generating the Chile triple-junction. The northward migration of the Chile triple-junction involved the subduction of various fracture zone–ridge segments (orientated c. N160), the last of which started subducting at c. 0.3 Ma (SCR1; Cande & Leslie 1986; Bourgois et al. 2000) at the Taitao Peninsula (46°12'S; Guivel et al. 1999). According to palaeo-tectonic reconstructions (Cande & Leslie 1986), another active ridge, the Farallon–Aluk ridge, collided with the western border of South America during Palaeocene–Eocene times (c. 55–53 Ma). This triple-junction migrated southward reaching Patagonian latitudes at c. 50 Ma.

The subduction of divergent mid-ocean ridges below continental plates is thought to produce a gap between the two subducting plates under the continental backarc region, the so-called slab window, which allows the decompressional melting of upwelling asthenosphere from sub-slab regions (Dickinson & Snyder 1979; Thorkelson 1994, 1996) and subsequent generation of mafic plateau volcanism in the backarc domain. Magmas generated under these conditions are expected to reproduce the chemistry of the asthenospheric mantle beneath the subducting plate (Stern et al. 1990; Gorring et al. 1997; D'Orazio et al. 2000; Gorring & Kay 2001), together with the probable occurrence of contamination during their trip to the surface.

The Patagonian flood basalts form an extensive basaltic province east of the Andean Cordillera (Patagonian Plateau Lavas; Fig. 4.17A & B) and extend from approximately 34°S to 52°S (Baker et al. 1981). Around 46°S the flood basalts are mostly located south of Lago General Carrera (Fig. 4.17C) and on both sides of the Chile–Argentina international border. In this area, the flood basalts, which overlie Mesozoic–Cenozoic sedimentary and volcanic rocks, have been divided (Baker et al. 1981; Charrier et al. 1979) into four age and genetically related groups: (i) Late Cretaceous (c. 80 Ma) mainly tholeiitic basalts, with some calcalkaline affinities, related to subduction; (ii) Eocene (c. 57–43 Ma) olivine tholeiites to alkaline basalts emplaced in a backarc domain; (iii) Late Oligocene to Late Miocene (c. 25–9 Ma) dominantly alkali basalts; and (iv) Pliocene to Quaternary (c. 4–0.2 Ma) highly undersaturated basanites. The Tertiary magmatic events are of continental intraplate type and were probably developed in an extensional regime as an indirect response to the subduction of various active ridges, which would have induced the opening of different slab windows beneath the continent during Eocene and Mio-Pliocene times (Ramos & Kay 1992; Petford et al. 1996; Gorring & Kay 2001; Gorring et al. 1997, 2003, Espinoza et al. 2005).

The Chilean Patagonian flood basalts at this latitude form the Meseta Chico Chico, which is located south of Lago General Carrera (46°30′S to 47°S; Fig. 4.17C), c. 300 km east of the present-day position of the Chile triple-junction and slightly eastward of the inferred location of the subducted Chile Ridge Segment 1 (SCR–1; Fig. 4.17B) that collided with the South American Plate c. 6 million years ago. K–Ar ages define two main basaltic sequences in the Meseta Chico Chico: the Lower Basaltic Sequence (57–40 Ma) and the Upper Basaltic Sequence (16–3 Ma) (Charrier et al. 1979; Baker et al. 1981; Petford et al. 1996; Petford & Turner 1996; Flynn et al. 2002a; Espinoza 2003; Espinoza et al. 2003, 2005). Both sequences are separated from each other by late Oligocene–early Miocene calcareous sandstones of the Guadal Formation (Niemeier et al. 1984; Frassinetti & Covacevich 1999).

The Upper Basaltic Sequence comprises a 400-m-thick pile of basalts, with minor rhyolites, covering an area of about 300 km². This Mio-Pliocene upper series can be correlated with similar volcanic rocks at the Meseta del Lago Buenos Aires (Fig. 4.17B). Taking into account all the published K–Ar ages, a magmatic gap of 24 ± 4 Ma (between 38 ± 2 and 14 ± 2 Ma), is defined in this Patagonian sector during the Tertiary (Espinoza et al. 2005).

Petrology and geochemistry of the flood basalts

Olivine, clinopyroxene and plagioclase plus minor Fe–Ti oxides are the main phenocrysts in intergranular, pilotaxitic, rarely interstitial or ophtic to subophitic basalts from both the lower and upper sequences. A major petrographic difference is the presence in basalts of the upper sequence, of rounded quartz xenocrysts (1–3 mm) rimmed by clinopyroxene (<0.5 mm), the composition of which is similar to those of clinopyroxene xenocrysts of the host basalts (Espinoza et al. 2005).

Basalts of the lower sequence are mainly ne-normative olivine basalts and scarce hy-normative olivine tholeiites ranging in composition from basanites to trachybasalts, whereas those of the upper sequence classify as basalt, basanite, trachybasalt, basaltic trachyandesite (rocks with quartz xenocrysts) with alkaline affinities (mainly ne- and minor hy-normative olivine basalts) and high-K calcalkaline rhyolites. The basic and acidic terms of the upper sequence define a bimodal distribution with a distinctive lack of intermediate compositions between 54 and 72 wt% SiO₂. In both, the lower and upper series, the high [Mg#] number, together with high Ni, Cr and Co contents are consistent with those of mantle-derived primitive melts. Major, trace and rare earth element geochemistry of both lower and upper series basalts are indicative of an OIB-like signature of magmas.

According to Espinoza et al. (2005), the lower-series basalts have initial ⁸⁷Sr/⁸⁶Sr ratios of 0.70385 to 0.70311 and εNd values from +5.1 to +4.9. These values are generally consistent with previous analyses for the Eocene basalts (Hawkesworth et al. 1979; Baker et al. 1981; Parada et al. 2001). Moreover, the lower series basalts display similar initial ⁸⁷Sr/⁸⁶Sr ratios and εNd
values compared with the coeval Posadas Basalts (Ramos & Kay 1992; Kay et al. 2002). An andesitic sample from the upper series has an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.70414 and an $\epsilon_{\text{Nd}}$ value of +4.7. A Miocene rhyolite has an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.70449 and $\epsilon_{\text{Nd}}$ value of +0.7, very similar to those obtained in a coeval pluton (Paso de la Llaves granite; Pankhurst et al. 1999) located in a neighbouring locality.

Petrogenetic model and geodynamic implications
Using trace element ratios, REE and isotopic modelling, Espinoza et al. (2005) suggested that lower- and upper-series intermediate and primitive basaltic magmas were generated at approximately similar pressures by equivalent degrees of partial melting of similar mantle sources, characterized by the presence of garnet as a residual phase ($Y_b$, 6.6–9.5).
The genesis of the Patagonian flood basalts in the Meseta Chile Chico may be integrated within a model based on the opening of two different slab windows below South America during the Eocene and Pliocene epochs, respectively (Ramos & Kay 1992; Espinoza et al. 2005). This kind of model is based on the temporal evolution of ridge–trench collision and slab window development under southern South America and has been proposed for other Patagonian basaltic sequences deposited during Pliocene–Recent events (Goring et al. 1997, 2003; D’Orazio et al. 2000).

Despite several uncertainties concerning the related geodynamic evolution, a slab window model for Eocene magmatism is strongly supported by the deep mantle geochemistry (OIB-like signature) of the Eocene lavas (Ramos & Kay 1992; Kay et al. 2002). The geotectonic reconstruction for early Tertiary times proposed by Cande & Leslie (1986) shows that around 50 Ma one segment of the active Farallon–Aluk ridge would have collided with the Chile Trench at Patagonian latitudes (c. 46°S). This collision would have induced the opening of a slab window beneath the South America continental margin, with subsequent melting and ascent of subslab asthenosphere.

The basaltic episode of the upper series would have begun between 12 and 10 Ma, prior to the arrival of the slab window beneath the Lago General Carrera basalts. At c. 5 Ma, the partially opened slab window may have been located beneath this area. Consequently, as proposed for the Eocene period, together with a new extensional tectonic event (and uplift as a consequence of late Miocene compression; Ramos 1989; Flint et al. 1994; Morata et al. 2003), decompressional melting would occur in the subslab asthenosphere. The Meseta Chile Chico is characterized by the occurrence of long-lived magmatism since Eocene to Pliocene times, as a consequence of some kind of thermal and compositional anomaly responsible for the input of primitive OIB-like asthenospheric melt.

**Magmatism in the southernmost Andean segment (47–56°S) (M.C.)**

The present-day geodynamic situation of the southernmost segment of the Patagonian Andes, a 4000-m-high mountain belt with only rare volcanism and a complex fold and thrust belt to the east, is dominated by the subduction of the Antarctic oceanic lithosphere beneath the South American continental plate. Along the convergent margin several spreading-centre segments have been subducted during the Miocene epoch. This phenomenon is temporally and spatially related to pauses in arc volcanism and the eruption of large volumes of backarc plateau basaltic lavas (Ramos & Kay 1992; Goring et al. 1997). The northern limit of the convergent margin is given by the subduction of the Chile Ridge, resulting in the Pliocene obduction of the Taitao ophiolite (47°S) in the forearc region (Forsythe et al. 1986; Hervé et al. 2003b).

**South Patagonian Batholith**

The composite South Patagonian Batholith and the Fuegian Batholith, which extends to around 1200 km long and 50–100 km wide between 47° and 55°S, is flanked to the east and west by metamorphic complexes with protoliths of Palaeozoic ages (Fig. 4.18), and which record diachronous metamorphic events associated with the evolving continental margin of Gondwana and South America (Hervé et al. 2003a; see chapter 2). The composite Southern Patagonian Batholith consists of hypidiomorphic and medium- to coarse-grained mafic (diorite, gabbro), intermediate (tonalite, quartz-diorite, quartz-monzodiorite) and felsic (granite, monzogranite, granodiorite) lithologies. Layered gabbros containing pyroxene and olivine, offten with coronitic texture, are commonly found, whereas the dominant intermediate and felsic rocks contain hornblende and/or biotite. Accessory minerals in intermediate and felsic rocks include magmatic epidote (allanite), titanite and Fe–Ti oxides, among others (Weaver et al. 1990; Ureta 2000). The eastern side of the batholith is composed mainly of biotite granites (Nelson et al. 1988; Weaver et al. 1990) spatially related with migmatises and high-grade metasedimentary rocks and minor biotite–muscovite leucogranites with garnet and tourmaline (c. 49°S; Calderón et al. 2003). In general, between 48°S and 50°S intermediate and felsic compositions predominate in the western and eastern margin of the batholith, respectively, whereas mafic rocks are more abundant along the central part (Ureta 2000).

Hornblende geobarometry indicates that the components of the batholith were emplaced in the upper crust between 2 and 4 kbar (50–52°S; Dzogolyk et al. 2003), which is consistent with widespread andalusite–sillimanite in metapelitic rocks along a diachronous regional aureole at the eastern margin of the batholith (48–50°S; Calderón & Hervé 2000). Deeper levels of emplacement, equivalent to c. 5.5 kbar, have been estimated from the celadonite content of magmatic muscovite in garnet-bearing granites east of the Southern Patagonian Batholith (c. 51°40′S; Massonne et al. 2004).

The ages of the plutonic components of the Southern Patagonian Batholith range from 151 to 16 Ma (Bruce et al. 1991; Martin et al. 2001) and are distributed in a zoned pattern. Jurassic plutons occur along the eastern margin (c. 151–141 Ma), Early Cretaceous plutonic units have been identified in the west (c. 137 Ma), and Late Cretaceous to Tertiary granite intrusions have been concentrated near the central axis of the SPB (U–Pb conventional and SHRIMP zircon ages; Martin et al. 2001). This age distribution, which is also revealed by K–Ar and 40Ar/39Ar methods, has been considered as the result of an ‘inflating’ batholith, where younger intrusions are hosted in older granitoids within an extending magmatic arc (Bruce et al. 1991).

Low-temperature cooling ages of the batholith have been revealed by zircon and apatite fission track thermochronology along a southern NW–SE transect, at c. 50°S. Reset zircon fission track ages reveal a progressive decrease from 18 to 14 Ma (Thomson et al. 2001), which are similar to 40Ar/39Ar (in hornblende) and K–Ar (biotite) ages (c. 18 to c. 13 Ma) of Mesozoic satellite plutons located to the east of the Southern Patagonian Batholith (c. 51°S; Bruce et al. 1991).

Whole-rock major and trace element compositions of the Southern Patagonian Batholith plutons indicate calcalkaline metaluminous affinities (Stern & Sproul 1982; Weaver et al. 1990). The batholith has been divided into a tonalitic and a granodioritic series, suggesting that differentiation of two distinct parental magmas occurred during batholith formation (Nelson et al. 1988; Weaver et al. 1990). Initial 87Sr/86Sr ratios range from 0.7036 to 0.7074 and initial εNd values from +7 to –7 (Fig. 4.19). The older plutonic rocks have the highest initial 87Sr/86Sr ratios and lowest εNd values (48°S; Weaver et al. 1990). These isotope compositional variations have been interpreted as evidence for magma contamination by Palaeozoic metasedimentary rocks. An origin from mantle-derived magmas, with crustal melts that progressively decrease in importance with time, has been inferred (Weaver et al. 1990).

At Puerto Edén (c. 49°S), the easternmost biotite granites (assumed to be Jurassic) have initial 87Sr/86Sr ratios of c. 0.7075 and negative εNd of c. –7, which are considered to reflect a contribution from Palaeozoic metasedimentary rocks in their genesis, during low-pressure and high-temperature anatectic (Calderón et al. 2003; Hervé et al. 2003a). Western plutonic rocks of Early Cretaceous ages (c. 130–137 Ma) in the Archipiélago Madre de Dios (c. 50°S) have initial 87Sr/86Sr ratios ranging from 0.7046 to 0.7050 and low positive εNd from 0 to +1. The magma source of this pluton is thought to lie at the base of the continental crust (Duhart et al. 2003).
Mesozoic volcanic rocks

Jurassic felsic volcanic rocks

Along the eastern side of the Patagonian and Fuegian Andes, Jurassic and Cretaceous supracrustal and plutonic rocks (Fig. 4.18), formed during continental rifting, gave rise to the Rocas Verdes backarc marginal basin succession (Bruhn et al. 1978; Dalziel 1981; Stern & de Wit 2003). These units have subsequently been deformed by several thrusting episodes. The closure of this marginal basin began after 100–120 Ma (Dalziel 1981), with a later development of the Magallanes foreland basin after the Turonian (c. 92 Ma; Fildani et al. 2003).

Silicic volcanic rocks of the El Quemado and the Tobífera formations form the westernmost components of the Jurassic Chon Aike Large Igneous Province, which is considered to be the product of extensive crustal anatexis (Pankhurst & Rapela 1995; Pankhurst et al. 1998). The Tobífera and El Quemado formations have SHRIMP zircon ages of c. 172–173 Ma and c. 153 Ma, respectively (Pankhurst et al. 2000). The plutonic equivalent of these formations is the S-type biotite–garnet granite suite (of c. 160 Ma; U–Pb zircon age) of the Darwin Cordillera, interpreted as a product of the anatexis of upper crustal metasedimentary rocks during the earliest stages of backarc basin formation (Hervé et al. 1981c; Mukasa & Dalziel 1996).

Thrust sheets of foliated silicic rocks of the Tobífera Formation (Galaz et al. 2005), cropping out to the north and south of the southern Patagonian ice field, are correlated with subsurface silicic rocks in southeastern Patagonia, seen in boreholes in the Magallanes basin (Fuenzalida & Covacevich 1988; Pankhurst et al. 2000). However, this correlation is not supported by recent data of the Andean components which yield a Late Jurassic age of crystallization (c. 148 Ma; Calderón 2006) considerably younger than previously constrained. A subaquatic depositional environment has been proposed for at least part of the Tobífera Formation, preceding the formation of the ocean mafic floor of the Rocas Verdes marginal basin (Dalziel 1981; Fuenzalida & Covacevich 1988; Hanson & Wilson 1991).

Late Jurassic–Early Cretaceous mafic rocks

Discontinuous and incomplete ophiolite complexes crop out along the Sarmiento Cordillera (Sarmiento Complex; c. 52°S), Navarino Island (Tortuga Complex; c. 55°S) (Fig. 4.18) and the
island of South Georgia. They consist of submarine mafic intrusives, sheeted dykes, gabbros and local plagiogranites, flanked on both sides by successions of silicic volcanic rocks. They are interpreted as ophiolites formed at a mid-ocean-ridge-type spreading centre during Late Jurassic and Early Cretaceous times (Dalziel 1981; Stern et al. 1992; Mukasa & Dalziel 1996; Stern & de Wit 2003). The mafic rocks of the Tortuga Complex have tholeiitic trends, low initial ⁸⁷Sr/⁸⁶Sr ratios (0.70323 to 0.70429) and high εNd values (+6.8 to +7.6) (Fig. 4.19), indicating that they were derived from a depleted MOR-type asthenospheric mantle source (Stern 1991b). REE compositions and εNd values from +0.8 to +2 were obtained in the Sarmiento complex, suggesting a slightly enriched E-MORB-type magmatic source. The age of the ophiolite complexes has been inferred from U–Pb zircon ages of 136–142 Ma (Sarmiento Complex; Stern et al. 1992) and 150 ± 1 Ma (South Georgia island; Mukasa & Dalziel 1996) obtained in plagiogranites. The observed diachronism of the ophiolite emplacement is considered indicative of a northward ‘unzipping’ mode of marginal basin formation (e.g. Stern & de Wit 2003).

Early Cretaceous (c. 104 Ma) spilitized clinopyroxene–amphibole-bearing mafic dykes and lavas with mildly alkaline shoshonitic affinities, occur c. 20 km east of the Sarmiento Complex. These rocks have low initial ⁸⁷Sr/⁸⁶Sr ratios (c. 0.7030) and εNd values (c. +5.0), which are considered to reflect a contribution from the subcontinental lithospheric mantle in their magma genesis (Stern et al. 1991).

Pleistocene to Holocene volcanism in the Chilean Andes (L.L.-E., H.M. & O.F.)

The most recent Andean magmatism has produced three prominent volcanic belts in Chile, situated (i) in the far north, (ii) south from Santiago as far as 46°S, and (iii) in the far south. Volcanism in northern Chile extends into NW Argentina, SW Bolivia and southern Peru, north of which there is another major gap before a fourth major volcanic belt is reached (in the far NW of South America: see Fig. 5.1 in Chapter 5). Thus Andean volcanism in South America has been divided, from north to south, into four zones: Northern (NVZ), Central (CVZ), Southern (SVZ) and Austral (AVZ); only the last three of these crop out in Chile. These volcanic zones are separated by flat-slab segments where Plisse-Holocene volcanism is absent. The following description sets the scene for the more detailed examination of Chilean volcanism presented in Chapter 5.

Northern Chilean volcanoes

Miocene to Recent volcanic activity in the CVZ of northern Chile has produced stratovolcanoes, monogenetic centres and ignimbrites. The stratovolcanoes are the highest volcanic edifices (up to 2000 m above their base). Since they are built on the Altiplano–Puna, their altitude can reach over 6000 m a.s.l. (e.g. Ojos del Salado, 6887 m a.s.l.). They are mainly composed of andesitic lava flows (Déruelle 1979; Wörner et al. 1992a), and minor amounts of pyroclastic deposits, some with volcanic avalanche deposits generated by collapse of the central edifice. Domes comprise either dacites or rhyolites, and have either rounded forms and abrupt edges or consist of lateral flows, as illustrated by the Chao flow (22.1°S; de Silva et al. 1995). The monogenetic cones are products of single eruptions. They are small (about 100 m high) and their lavas are less differentiated (basaltic andesites) than those of the stratovolcanoes.

The ignimbrites cover most of the topography of this volcanic zone and, taken together, they constitute the world’s largest late Tertiary–Recent ignimbritic province. They are mainly early Miocene in age in the northernmost part of Chile (<21°S) and late Miocene–late Pliocene between latitudes 21°S and 24°S (Wörner et al. 2000b, and references therein). The abundant ignimbrites between 21°S and 24°S have been designated as the Altiplano–Puna Volcanic Complex (de Silva 1989a), which covers more than 50 000 km² and represents >30 000 km³ (Lindsay et al. 2001b). Most sources of ignimbrites correspond to calderas, identification of which has been revealed by satellite imagery. These calderas are generally located in the backarc zone, aligned parallel to the actual chain of stratovolcanoes. Some of them, such as the Pacana caldera, have huge dimensions (60 km × 35 km; Gardeweg & Ramirez 1987; Lindsay et al. 2001b).

Beneath this volcanic belt the crust–mantle boundary has a broad transitional character due to active processes such as hydration of mantle rocks, magmatic underplating and intraplate under and into the lowermost crust, and partial melting (ANCORP Working Group 2003). Seismic studies also show a pronounced low-velocity zone in the mid-crust (Wigger et al. 1994; Yuan et al. 2000; ANCORP Working Group 2003), which together with other geophysical observations (such as bright reflectivity, high conductivity, high heat flow values, negative anomaly in the residual gravity field), has been interpreted as a zone of partial melting (for a review see Babeyko et al. 2002; ANCORP Working Group 2003). The low velocities of the P waves (Vp = 6 km/s) and the anomalous low Poisson ratio (0.25) would imply that the crust has a felsic composition down to 50–55 km depth (Swenson et al. 2000, and references therein), being predominantly mafic at greater depths (Yuan et al. 2002).

Petrography and geochemistry

The stratovolcano lavas vary from basaltic andesite to rhyolite with a predominance of andesites and dacites. Plagioclase is the most abundant phenocryst, although olivine, orthopyroxene and clinopyroxene phenocrysts are also found in the basaltic andesites, whereas amphibole can be observed in andesites. The dacitic–rhyolitic domes are highly porphyritic (up to 50 vol%), presenting phenocrysts of amphibole, biotite, and two feldspars, with little or no quartz and pyroxenes. Dissequilibrium
Most of the rocks erupted from the CVZ centres have a medium-K calcalkaline character. Generally, TiO$_2$, Al$_2$O$_3$, FeO$^*$, MnO and CaO decrease, and K$_2$O increases, as SiO$_2$ increases; such patterns are consistent with the fractionation of plagioclase, pyroxenes and Fe–Ti oxides. In comparison with the lavas of SVZ, CVZ lavas are enriched in incompatible elements and have higher $^{87}$Sr/$^{86}$Sr and lower $^{143}$Nd/$^{144}$Nd, implying a greater amount of crustal contamination (for a review of Andean magma genesis see Stern 2004). The increase in the La/Yb ratio as the Yb contents decrease suggests that this contamination took place deep in the crust, where garnet is a stable phase (e.g. Hildreth & Moorhut 1988; Feeley & Davidson 1994). At this level, mantle-derived magmas evolve by contamination and crystal fractionation to basaltic andesitic composition, which decreases its density and permits the magmas to migrate to shallow crustal levels, where they accumulate at the bases of dominantly andesitic magma chambers. These chambers must be stratified and periodically refill with basaltic andesite magmas (Feeley & Davidson 1994; Matthews et al. 1999).

Based on isotopic composition, two types of magma differentiation are recognized (Davidson et al. 1991): (a) closed system, e.g. in Nevados de Payachata (Davidson et al. 1990, and references therein) and San Pedro–San Pablo (O’Callaghan & Francis 1986), where in spite of their wide variation in elemental composition, they exhibit only minor variations in Sr-, Nd- and Pb-isotopic composition; and (b) open system, e.g. Ollagüe (Feeley & Sharp 1995, and references therein), Licancabur (Figueroa 2001) and Lascar (Matthews et al. 1994), where isotopic compositions change with indices of magma differentiation, which is interpreted as the result of assimilation plus fractional crystallization (AFC).

Fractional crystallization played an important role in the origin of dacites from andesitic (or basaltic andesite) magmas. Magma mixing commonly occurred in addition to fractionation processes, as evidenced by disequilibrium textures (plagioclase resorption), mineralogical disequilibria (inverse zonations) and geothermometric considerations. The latter indicate, for example, higher equilibrium temperatures in the rims of hornblende phenocrysts than in their cores. In the specific case of Lascar volcano, such heterogeneities have been interpreted to result from the remobilization of crystallized or semicrystallized shallow intrusions of andesitic composition during the injection of basaltic andesite magma (Matthews et al. 1999).

The isotopic compositions of the large Altiplano–Puna Volcanic Complex ignimbrites ($^{87}$Sr/$^{86}$Sr > 0.709 and $^{143}$Nd/$^{144}$Nd < 0.5123) are considered to reflect large-scale mid-crustal melting (e.g. de Silva 1989a), which is supported by geophysical studies (see above). Therefore ignimbritic magmas are hybrid but the proportion of crustal to mantle-derived material is difficult to identify because of the heterogeneity of pre-Andean basement (Lindsay et al. 2001b). Mixing calculations indicate that compositions of Purico ignimbrites could be produced by adding c. 70% crustal components to c. 30% mantle-derived basaltic andesite (Schmitt et al. 2001), although small ignimbrites that erupted in the immediate vicinity of stratovolcanoes have less crustal influence (Déruelle et al. 2000, and references therein).

**Southern Chilean Volcanoes**

*The Southern Volcanic Zone (SVZ)*

Volcanic activity in Chile between 33°S and 46°S has been continuous and very active (one eruption per year on average) during post-glacial times (last 15 000 years). The Holocene volcanic front, whose axis is located about 280 km from the Chile–Peru Trench, has an average width of 40 km, although it is almost 80 km wide around 39°S. The volcanic activity is expressed as numerous composite stratovolcanoes, megacalderas (Diamante, Calabozos, Copahue), domes, and hundreds of minor eruptive centres consisting of scoria cones ± lava flows and maars. Lahars, ashfalls and lava flows have been the main volcanic hazards within historical times, although pyroclastic flows and surges, together with voluminous debris avalanches, have also occurred during the latest Pleistocene and Holocene. Among the 46 main stratovolcanoes between 33°S and 41.5°S, about 30 of them have erupted in post-glacial times and 18 of them have historical records. On the other hand, most minor eruptive centres are post-glacial, and three of them (Rininahue, Carrán and Mirador) erupted during the twentieth century. Individual stratovolcano elevations generally vary between 700 and 2100 m above the base, although further south they reach up to 2500 m.

Field and seismic evidence suggests that the tectonic regime within the SVZ arc has been dextral strike-slip for the last few million years (Cembrano 1992, and references therein). The fact that the direction of maximum horizontal stress (σ$_{\text{Hmax}}$) is roughly N50–70°E may reflect a transpressional tectonic regime, resulting from a combination of dextral strike-slip and shortening across the arc. Following the fracture propagation model of Shaw (1980), the ascent of magmas should be expected along inherited or newly created NE-trending tensile fractures and faults within the volcanic arc. This agrees with geochemical interpretations based on U–Th disequilibrium considerations (Tormey et al. 1991b).

Tectonism seems to control whether or not basaltic magmas either reach the surface or evolve to more differentiated products within the crust. In fact, most young volcanic centres (Late Pleistocene–Holocene), define NE-trending alignments erupting magmas of mainly basaltic to basaltic andesite composition (e.g. Osorno–Puntiagudo–Cordón Cenizos volcanic chain). This is consistent with an extensional regime that allows a short residence time of magmas in the crust, resulting in limited contamination and fractionation of mantle-derived magmas. On the other hand, volcanic edifices controlled by NW-trending fractures and faults (e.g. Villarrica–Quetrupillán–Lanín and Puyehue–Cordón Caulle volcanic chains) may be under a combination of shortening and strike-slip deformation, which will cause a longer intracrustal magma residence, yielding more differentiated compositions, including rhyolite.

Basaltic rocks erupted from either N50–60°W or N50–70°E transverse fractures tend to increase their incompatible elements abundances and La/Yb and $^{87}$Sr/$^{86}$Sr ratios and decrease their Ba/La, and $^{143}$Nd/$^{144}$Nd ratios from west to east. These geochemical characteristics seem to be independent of the orientation of fractures and faults where those centres are emplaced, suggesting that they are controlled mainly by subduction-related processes.

Even the most primitive basaltic of the Andean segment between 37°S and 46°S show evidence of having undergone fractional crystallization. They have MgO < 11%, and both Ni and Cr are lower than those expected in mantle-derived primary magmas. Olivine and plagioclase phenocrysts are present within the whole suite (basalts to rhyolites), with even a few rhyolites containing fayalitic olivine. Plagioclase phenocrysts vary in abundance from about 1–10% in minor eruptive centres volcanic rocks to 20–50% in stratovolcano products.

Isolated young andesitic stratovolcanoes, such as the Calbuco and Huequi volcanoes, lie on NW-trending lineaments tens of kilometres long. Lanín (Lara & Moreno 1994a) and Tronador (Mella et al. 2005) volcanoes are emplaced on an uplifted block, east of the LOFZ main trend. Calbuco was also emplaced on an uplifted block (López-Escobar et al. 1995a, and references therein). Unlike other centres, Mocho-Choshuenco, a basaltic andesite to dacitic stratovolcano, located at 40°S
(McMillan et al. 1989), is emplaced on a basement that, in addition to Jurassic igneous rocks and Triassic metasediments, contains Palaeozoic metasedimentary rocks of the coastal Western Series.

Two main types of basaltic rocks, one K-depleted (K$_2$O < 1%), also depleted in other incompatible elements such as Rb, La, Th) and the other K-enriched (K$_2$O=1–1.5%), have been distinguished in places (López-Escobar et al. 1995a, and references therein), as well as in strato-volcanoes and minor-eruptive-centre basalts. Only in the K-rich group are minor-eruptive-centre basaltic rocks significantly richer in MgO than strato-volcano basaltic rocks. K-poor basaltic rocks tend to have higher Ba/La and lower La/Yb ratios than K-rich ones. Some of the highest Ba/La and lowest La/Yb ratios are presented by basalts from the Pichihuinco maar (one of the westernmost minor eruptive centres of this Andean region). In contrast, one of the lowest Ba/La and highest La/Yb ratios are exhibited by the Puyuhuapi minor-eruptive-centre basalts (Demant et al. 1994; Lahsen et al. 1994; López-Escobar & Moreno 1994). In fact, the latter basalts have La/Yb ratios as high as those presented by andesites from the northernmost part of the SVZ (33–34.5°S) (López-Escobar 1984; López-Escobar et al. 1995a), where the thickness of the continental crust is 55–60 km. The relatively low Ba/La ratios of the Puyuhuapi and, in general, of K-rich basaltic rocks have been explained by subduction-related processes (López-Escobar et al. 1993). These processes could also explain the enrichment of these magmas in incompatible elements and their relatively high La/Yb ratios. The extremely high La/Yb ratio of the Puyuhuapi basalts, which is mainly due to their low Yb contents, also suggests that their parental magmas were in equilibrium with garnet at their mantle source or at lower crustal levels. In both groups of basalts high 87Sr/86Sr ratios are generally accompanied by low 143Nd/144Nd ratios, a trend commonly interpreted as evidence of contamination with crustal material.

The Austral Volcanic Zone (AVZ)

The southernmost Andean volcanic segment, located between latitudes 49°S and 55°S, has only six volcanic centres (Chapter 5), and these have erupted andesites and dacites, lacking the basalts and basaltic andesites commonly seen elsewhere in Chile (Kilian et al. 1991; Stern & Kilian 1996; Stern 2004). Unlike the CVZ and SVZ volcanoes, the AVZ volcanic centres are the result of subduction of the oceanic Antarctic plate below the South American plate. Relatively low base and summit elevations, and the presence of garnet-free granulitic lower crust xenoliths, indicate a relatively thin crust (Stern & Kilian 1996).

The essential mineralogy of all the AVZ andesites and dacites involves pyroxenes, amphibole and plagioclase (Stern 1984; Kilian et al. 1991; Stern & Kilian 1996). Both a greater proportion of crustal xenoliths and larger, more complexly zoned phenocrysts are observed from south to north, suggesting increasing importance of near-surface magma chamber processes such as slow cooling and crystallization, magma mixing, wall-rock assimilation, and volatile degassing.

In comparison with typical orogenic andesites and dacites, the AVZ eruptive rocks have high AI, low Ti and relatively high [Mg#]. They also show adakitic chemical features (Stern & Kilian 1996), with SiO$_2$ > 56 wt%, Al$_2$O$_3$ > 15 wt%, low HREE (Yb < 1.9 ppm) and Y < 18 ppm, high Sr > 400 ppm and Sr/Y > 40, and positive Sr and Eu anomalies. These characteristics have been interpreted to reflect the presence of residual garnet, amphibole and pyroxene, but with little or no olivine and plagioclase, in the source region.

AVZ andesites and dacites also have low concentrations of HFSE, which is a common feature of both adakites and typical convergent plate boundary magmas. However, other chemical characteristics, such as LREE content, LREE/HREE, LILE/LREE, LREE/HFSE and isotopic composition, which may reflect source composition rather than source mineralogy, vary significantly between, but not within, individual centres (Stern & Kilian 1996).

Some thermal models indicate that partial melting of the subducted oceanic crust is probable below the AVZ due to the slow subduction rate (2–3 cm/year) and the young age (< 24 Ma) of the subducted oceanic lithosphere. The source of AVZ adakites is thus likely to be subducted oceanic basalt, recrystallized to garnet-amphibolite or eclogite (Stern 1984; Ramos & Kay 1992; Stern & Kilian 1996; Kilian & López-Escobar 2000). To explain the origin and evolution of these adakitic andesites and dacites, Stern & Kilian (1996) propose a multi-stage, four-component (MORB, subducted sediment, mantle wedge and crust) model in which the proportions of the different source materials involved in the genesis of AVZ magmas vary significantly from south to north. Thus, andesites from Cook Island volcano, where subduction is more oblique, have MORB-like isotopic and trace-element ratios. These can be modelled by small degrees (2–4%) of partial melting of eucrotic MORB, yielding a tonalitic primary magma, followed by limited interaction of this magma with the overlying mantle (c. 90% MORB melt, c. 10% mantle), and very little (c. 1%) or no participation of either subducted sediments or continental crust. In contrast, models for the magmatic evolution of AVZ andesites and dacites located between 49°S and 52°S require melting of a mixture of MORB (35–90% MORB-derived mass) and subducted sediment (c. 4% sediment-derived mass), followed by interaction of this primary melt with the overlying mantle (10–50% mass contribution) and the continental crust (0–30% mass contribution). AFC processes and the mass contribution from the continental crust become more significant northwards in the AVZ as the angle of convergence becomes more orthogonal.

Concluding remarks on recent magmatism

The main sources that are involved in the geochemical characteristics of the Andean Pleistocene to Holocene volcanism are as follows. (1) The subducted oceanic crust which in the CVA and SVZ provides aqueous fluid solutions, enriched in alkaline and alkaline-earth elements, that contaminate the mantle wedge. In the AVZ, melts from the Antarctic plate seem to be the main source of magmas. (2) The asthenospheric part of the mantle wedge is the main source of magmas in the CVZ as well as in the SVZ. (3) The lithospheric part of the mantle wedge has been proposed as a source of contaminants of magmas coming from below. (4) The lower crust, especially in those places where crust is thick, seems to play an important role in modifying the geochemical characteristics of mantle-generated magmas. (5) The upper crust is an important source of meteoric water and is a place where important geochemical processes take place. Many different processes have been proposed to explain the geochemical characteristics of Andean magmas: dehydration of the subducted slab (CVZ and SVZ), melting of the subducted slab (AVZ), fractional melting of the asthenosphere, contamination at mantle lithosphere and lower crust, magma mixing at crustal level, assimilation of crustal material, fractional crystallization at low and high pressures, differentiation along the conduits, and various combinations of the above. These processes, their eruptive products, and the hazards associated with specific volcanic centres, are examined further in Chapter 5.

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There are over 200 Pleistocene and Holocene Andean arc volcanoes along western South America, occurring in four distinct segments (Fig. 5.1) called the Northern (NVZ; 2°N–5°S), Central (CVZ; 14°S–28°S), Southern (SVZ; 33°S–46°S) and Austral (AVZ; 49°S–55°S) volcanic zones. In the Andes of Chile alone there are more than 100 Pleistocene and Holocene stratovolcanoes, as well as a number of large volcanic fields and giant caldera complexes, of which 60 have documented Holocene eruptive activity (Simkin & Siebert 1994; González-Ferrán 1995). These are located in the CVZ of northern Chile, the SVZ of central-south Chile, and the AVZ of southernmost Chile. Pleistocene and Holocene backarc volcanic centres, the westernmost part of the Patagonian plateau basalts, also occur in southern Chile along the border with Argentina. In addition, intraplate oceanic volcanoes form Chilean islands in the Pacific Ocean, submarine volcanism takes place along the Chile Ridge, and slab-window volcanic activity occurs above the region where the Chile Ridge is currently being subducted, both along the west coast of Chile and in the submarine environment near the trench.

Pleistocene and Holocene volcanoes of the Chilean Andes provide a natural laboratory for the study of volcanoism, magma genesis and volcanic hazards in the context of oceanic–continental plate collision. Andean volcanic activity results from subduction of the Nazca and Antarctic oceanic plates below the continental lithosphere of western South America (Fig. 5.1). Volcanoes in the CVZ of northern Chile and the SVZ of central-south Chile occur where the angle of subduction of the Nazca Plate is relatively steep (>25°) at depths >90 km. These two active arc segments are separated by the Pampean flat-slab segment, below which subduction angle decreases and becomes relatively flat (<10°) at depths >90 km. The northern boundary of the SVZ corresponds to the locus of subduction of the Juan Fernández Ridge of volcanic oceanic islands. At the southern end of the SVZ, the eastern extension of the Chile Ridge, which separates the Nazca from the Antarctic plate, is being subducted below South America, and the Patagonian gap in volcanism is observed between the SVZ and AVZ. These relations imply a clear genetic connection between plate subduction and volcanism that is confirmed by geochemical studies indicating that generation of Andean magmas is initiated by dehydration and/or melting of subducting oceanic lithosphere and interaction of these slab-derived fluids/melts with the overlying mantle wedge. Continental crust is Incorporated into Andean magmas by a combination of subduction of crust into the subarc mantle and the assimilation of crust into mantle-derived magma.

Detailed studies of individual Andean volcanoes in Chile indicate that there is no unique ‘Andean-type’ volcanic arc (Stern 2004). The diversity of Pleistocene and Holocene Andean volcanoes and volcanic rocks in Chile, and the corresponding potential for different types of volcanic hazards, reflect a variety of magma genesis processes that need to be evaluated in the context of both along-arc and across-arc variations in the geological and tectonic character of the continental South American Plate, the oceanic plates it is colliding with, and aspects of the collision process. Characteristics of the continental lithosphere, including pre-Andean basement ages, Andean structural evolution, crustal thickness and composition, and Neogene continental tectonic regimes vary significantly from north to south along the length of Chile, and these influence the genesis and fine structure of the Andean Pleistocene and Holocene volcanic arc in Chile. Both rates and directions of convergence differ north and south of the Chile Ridge triple-junction (Fig. 5.1). The age of the oceanic lithosphere being subducted also varies from north to south along-strike of the Chilean continental margin, as does depth of the trench. The amount and type of sediment in the trench changes as the result of north to south differences in climate, which have affected sediment supply to the trench since Miocene times (Alpers & Brimhall 1988). This, in turn, influences rates of sediment subduction and subduction erosion of the continental margin (Scholl et al. 1980), degree of hydration and consequently shear stress in the subduction zone (Ruff 1989), dynamics of mountain building (Giese et al. 1999; Lamb & Davis 2003), as well as Andean magma chemistry in different segments of the Chilean Andes (Stern 1991b, 2004).

This chapter describes Pleistocene and Holocene Andean volcanoes, volcanic hazards and magma genesis in Chile in the context of these tectonic and geologic variables, as well as other volcanic activity in Chile that occurs in a variety of tectonic settings.

**CVZ volcanoes of northern Chile (J.E.C., J.A.N. & C.R.S.)**

Volcanoes of the Central Volcanic Zone (CVZ; Figs 5.1 & 5.2) are located in the Central Andes of southeastern Peru, western Bolivia, northern Chile and northwestern Argentina. In total, more than 1100 volcanic vents and/or edifices have been identified in the CVZ of the Andes (de Silva & Francis 1991). However, due to the hyperarid conditions that have prevailed in the region during the Pleistocene and Holocene, erosion has been minimal and some of these are most likely pre-Pleistocene in age. The hyperarid conditions result in excellent exposures of extremely well preserved volcanic features, and for these reasons studies of the evolution of CVZ centres in northern Chile have proved especially useful for understanding volcanic processes.

In northern Chile, Pleistocene and Holocene CVZ volcanoes form an essentially continuous chain between Tacora volcano, located near the Peru–Chile border at 17.5°S, to the Nevado Ojos del Salado volcanic complex at 27°S, the world’s highest active volcano summit (6886 m; Simkin & Siebert 1994).
Geological and tectonic setting

Volcanoes of the CVZ result from subduction of the Nazca Plate below the Central Andes (Figs 5.1 & 5.2). The northern end of the CVZ is located in Peru at 14°S, where subduction of the Nazca Ridge results in a dramatic decrease in subduction angle below the Peruvian flat-slab segment to the north. The southern margin of the CVZ, where the Tres Cruces–Nevado Ojos del Salado volcanic chain is located, coincides with a seismic discontinuity at 27°S (González-Ferrán et al. 1985), superimposed on a gradual decrease in subduction angle associated with the Pampean flat-slab segment to the south, where Pleistocene and Holocene magmatism is absent (Kay & Mollo-Christensen 2002). Convergence velocity between the Nazca and South American plates in northern Chile is 7.8–9.4 cm/year (DeMets et al. 1990). Obliquity of the convergence angle varies between 0° at the latitude of Arica (18°S; Fig. 5.2) to c. 24° west of the southern margin of the CVZ. The age of the oceanic Nazca Plate being subducted below northern Chile is <60 million years (Fig. 5.1). Subduction angle as defined by the Wadati–Benioff zone is c. 25° (Cahill & Isacks 1992), and below northern Chile the subducted plate descends to a depth of >400 km (Dorbath et al. 1996). The volcanic front of the CVZ is located approximately 120–150 km above the subducted slab and 260–340 km east of the Peru–Chile Trench, which is >7000 m deep everywhere west of the CVZ and reaches a maximum depth of 8055 m below sea level at 23°S. The trench is devoid of sediment (Thornburg & Kulm 1987a) because of the hyperarid conditions in northern Chile (Hartley et al. 2000; Hartley 2003).

Most of the Pleistocene and Holocene volcanic centres in the CVZ of northern Chile occur along the western boundary of two morphotectonic regions of the Central Andes referred to as the Altiplano (15–23°S) and Puna (23–28°S). Both these regions include a high central plateau, at an elevation of 3700 to 4200 m above sea level, bounded by Eastern and Western Cordilleras, with numerous peaks reaching >6000 m elevation. They occur above the thickest crust on Earth: 60–65 km thick below the Altiplano and Puna plateaux and >70 km thick below the bounding mountain belts (James 1971; Beck et al. 1996). The present-day topography of the Altiplano and Puna is essentially the result of the last Andean orogenic or deformation episode, known in the literature as the ‘Quechua Phase’ (Noble et al. 1997). This widely recognized episode has been dated as beginning at c. 27 Ma in the Eastern Cordillera (Semperé et al. 1990; Hérail et al. 1996) and more recently in the Western Cordillera (Muñoz & Charrier 1996; García et al. 1999a; Wörner et al. 2000c). This deformation resulted from an increase in the rate of subduction below the Central Andes at c. 27 Ma, which also produced the generation of large volumes of magmas of different compositions (Sébrier & Soler 1991), although the onset of volcanism from this event was not simultaneous along the whole Central Andes. Very thick crust below the Central Andes may be due to either crustal shortening (Isacks 1988; Beck et al. 1996; Allmendinger et al. 1997; Kley et al. 1999), magmatic underplating (Tosdal et al. 1984; Schmütz et al. 1997, 1999), or a combination of the two, with more crustal shortening in the Eastern Cordillera (Cordillera Oriental) and more magmatic underplating and lithospheric hydration below the Western Cordillera (Cordillera Occidental; James 1971; James & Sacks 1999; Giese et al. 1999; Victor et al. 2004).

The CVZ also occurs above some of the oldest crust along the west coast of South America, exposed as the Palaeoproterozoic (2.0–1.8 Ga) metamorphic and igneous rocks of the Arequipa terrane of southern Peru. Metamorphic rocks with similar protolith ages have also been reported in the Antofalla basement exposed at Belén, Chile, implying that such rocks may underlie a large part of the CVZ of northern Chile (Wörner et al. 2000c). Protoliths in these terranes may be either parautochthonous (Perigondwanan) and related to the Brazilian craton (Tosdal 1996), or allochthonous and part of the Grenvillian province of Laurentia (Dalziel 1994, 1997; Wasteney et al.

Fig. 5.1. Schematic map, modified from Stern (2004), of the four volcanically active zones in the Andes, subduction geometry as indicated by depth (in km) to the Benioff zone, oceanic ridges, ages of oceanic plates close to the trench, and convergence rates and directions along the length of the Andes. Andean volcanoes in Chile occur in the Central, Southern and Austral volcanic zones.
1995; Loewy et al. 2004). Rocks with ‘Grenvillian’ (1.3–1.0 Ga) protolith ages also occur at Choja in northern Chile (Loewy et al. 2004). These Proterozoic rocks were subsequently affected by the Neoproterozoic to Early Palaeozoic (680–510 Ma) Pampean igneous and metamorphic event, first identified in the basement further south in the Sierras Pampeanas ranges in Argentina (Rapela et al. 1998a, b), as well as younger Palaeozoic (470–340 Ma) events (Wörner et al. 2000c; Jaillard et al. 2000; Ramos & Alemán, 2000; Lucassen et al. 2001). South of the Antofalla basement, below the southern part of the CVZ, accretion and amalgamation of the Oaxaquia, Famatina, Cuyania and Chilenia allochthonous terranes took place during the Palaeozoic (Ramos et al. 1986; Pankhurst & Rapela 1998; Ramos & Alemán 2000; Astini & Dávila 2002, 2004; Loewy et al. 2004).

A notable and significant aspect of the western margin of northern Chile is the total lack of Mesozoic and Cenozoic arc–trench gap accretionary complexes, and the occurrence of pre-Andean Proterozoic to Early Mesozoic exotic and parautochthonous terranes along the coast (Rutland 1971; Kulm et al. 1977; Schweller & Kulm 1978). This is interpreted to result from subduction erosion, or rasping off of the continental margin by a combination of both continental slope retreat and erosion of the underside of the upper plate (Rutland 1971; Scholl et al. 1980; Ziegler et al. 1981; von Huene & Lallemand 1990; Stern 1991b; von Huene & Scholl 1991, 1993; Lallemand et al. 1992a, b; Lallemand 1995; von Huene et al. 1999, 2004; von Huene & Ranero 2003; Sallarés & Ranero 2005). Other geological evidence for subduction erosion along the west coast of northern Chile includes the c. 250 km eastward migration of the Andean Jurassic to Cenozoic magmatic belts (Ziegler et al. 1981; Peterson 1999; Giese et al. 1999), the NW strike and almost complete disappearance north of 27°S of the Late Palaeozoic Gondwanan subduction accretionary complexes that form the Coastal Cordillera in central Chile (Stern 1991b; Stern & Mpodozis 1991), and shortages in the amount of crustal shortening that can be accounted for by crustal area balance calculations (Schmitz 1994).

The suggested mechanism for slope retreat is continental margin collapse towards the trench caused by extensional faults, generating debris that are pushed into the horst and graben structures on the subducting oceanic plate, which resemble the teeth of a chain saw (Hilde 1983; von Huene et al. 2004). Such structures and processes have been well understood in similar tectonic settings (von Huene et al. 2004).
documented along the continental margin of southern Peru (Lallemand et al. 1992a, b; Hampsell et al. 2004a, b) and northern Chile (Delouis et al. 1996, 1998; von Huene et al. 1999; Hartlett et al. 2000; von Huene & Ranero 2003; González et al. 2003; Lovelace et al. 2003; Sallarés & Ranero 2005). Rates of subduction erosion depend on factors such as subduction angle, sediment supply to the trench, and subduction of buoyant features such as ridges (Scholl et al. 1980; Shreve & Cloos 1986; Ruff 1989; Bangs & Cande 1997; Lamb & Davis 2003). Both low angle subduction and low sediment supply to the trench, features of the plate boundary west of northern Chile, are factors that enhance the potential for increased rates of subduction erosion. Here the Peru–Chile trench receives almost no sediment from the few mountain glacier-fed drainage systems that cross the Atacama Desert (Thornberg & Kulm 1987a; Hartley 2003).

Long-term average rates of subduction erosion west of the Central Andes may be as high as 40 km$^3$ of crust per million years per kilometre of trench in order to have removed an estimated 250 km of 40-km-thick crust in the last 25 million years of near-orthogonal convergence (Ziegler et al. 1981; Lallemand 1995). This is equivalent to 4% of the volume of subducted oceanic crust in the same time period, which is well within the carrying capacity of the underthrusting grabens entering the subduction zone below northern Chile (von Huene & Ranero 2003). This subducted crust may have an important influence on the extent to which the subducted slab is hydrated (Giese et al. 1999; Ranero & Sallarés 2004), the dynamics of the interaction between the subducting oceanic and overlying continental plate (Ruff 1989; Lamb & Davis 2003), and on the chemistry of CVZ magmas in northern Chile (Stern 1991b, 2004; Kay et al. 1999; Macfarlane 1999).

Subduction erosion and decreasing subduction angle have caused the Andean magmatic arc to migrate into its current position in the Altiplano–Puna region. Large volumes of rhyolitic magmas started to erupt here at c. 27 Ma, most of them probably associated with large calderas, forming extensive ignimbritic plateaux to the west of the Altiplano. During the Miocene epoch a series of medium to large andesitic to dacitic stratovolcanoes was formed. Magmatic activity increased dramatically approximately 10 million years ago (Allmendinger et al. 1997). In the late Miocene and Pliocene, new stratovolcanoes of the same compositions were formed, as well as large-volume rhyodacite ignimbrites, such as the Laucar, Purico, Atana and Toconce ignimbrites (Gardeweg & Ramírez 1987; Kött et al. 1995; Wörner et al. 2000b; Lindsay et al. 2001a, b; Schmitt et al. 2001; García et al. 2004). Due to the hyperarid conditions in northern Chile, many of these volcanic centres and ignimbrites are extremely well preserved. The recent CVZ volcanic chain was formed during Late Pleistocene to Holocene times, with magmas mainly of the same compositions as in the Miocene and Pliocene.

**CVZ volcanoes and hazards**

Late Pleistocene to Holocene volcanism in northern Chile has been characterized by the formation of both andesitic stratovolcanoes and dacitic dome complexes, with associated pyroclastic flow, tephra fallout, debris avalanche and block-and-ash flow deposits. Less common are small monogenetic andesitic to basaltic andesite eruptive centres, such as the Ajata flank centres of Parinacota volcano, La Porrúa pyroclastic cone on the lower western flank of Ollague volcano and La Porrúa pyroclastic cone and associated lava on the western flank of San Pedro volcano. A small number of isolated high-silica dacitic to rhyolitic flows, some forming ‘torta’-type domes, have been recognized locally, such as the Chao (de Silva et al. 1994) and Tocorpuri (de Silva & Francis 1991) domes in Chile, and others in Bolivia (Watts et al. 1999). Large caldera complexes, including the giant La Pacana system in Chile and others in southwestern Bolivia and northwestern Argentina, have produced the Pliocene and Pleistocene ignimbrites of the Altiplano–Puna Volcanic Complex that crop out in northern Chile.

Although most of the active volcanoes of northern Chile are located in remote areas, with few people living nearby, future eruptions of some of them could affect inhabited areas located in southwestern Bolivia and/or northwestern Argentina, as was the case during the 1993 eruption of Láscar volcano (Gardeweg & Medina 1994). Also, although low-altitude prevailing winds in the area blow generally from west to east, high-altitude winds may blow from east to west in some seasons. Therefore, tephra falls from large eruptions could disperse to the west into Chile and affect both the local population and wetlands (bofedales), which are the main feeding area of altiplano camelids (llamas and alpacas), and therefore damage the local economy (Castellaro et al. 2004).

Some of the largest, most active and better studied of the many Late Pleistocene to Holocene CVZ volcanic centres in northern Chile (Fig. 5.2) include Tacora (Clavero et al. 2006; Parinacota (Wörner et al. 1988, 1992a; Clavero et al. 2002, 2004a), Taapaca (5850 m; Clavero et al. 2004b), Guallatiri (García et al. 2004), Arintica (5597 m), Isluga (5501 m), Irupuñuncu (Clavero et al. 2005), Oca–Parumá (5450 m), Aucanquilcha (6176 m; Grunder et al. 2004), Ollagüe (Fleeley et al. 1993; Fleeley & Davidson 1994; Clavero et al. 2004c), Azufre (5846 m), San Pedro (Francis et al. 1974; O’Callaghan & Francis 1986), Putana (5890 m), Escalante (5971 m), Lican-cabar (5916 m; Figueroa 2001), Guayayques (5584 m), Colachi (5631 m), Acamarachi (6046 m), Aguas Calientes (5924 m), Láscar (Matthews et al. 1994, 1997, 1999; Gardeweg et al. 1998b), Chiliques (5778 m), Punta Negra (5852 m), Pular (6233 m), Socoma (Francis et al. 1985; van Wyk de Vries et al. 2001), Lullaillaco (Gardeweg et al. 1984; Richards & Villeneuve 2001), Escorial (5447 m), Lastarria (Naranjo, 1988, 1992; Naranjo & Francis 1987), Bayo (5401 m; Naranjo 1988), Sierra Nevada (6127 m; Clavero et al. 1997), Nevado Ojos del Salado (González-Ferrán et al. 1985; Baker et al. 1987b; Gardeweg et al. 1998a) and Nevado Tres Cruces (Gardeweg et al. 2000).

We briefly describe below a selection of the most important CVZ volcanoes, chosen according to criteria including their historic activity and/or prominent features such as the occurrence of debris avalanches or notable pyroclastic fans.

**Tacora volcano (17.7°S, 69.8°W, 5980 m)**

Located close to the Peru–Chile border, this is the northernmost active volcano in Chile. It has permanent fumarolic activity on its western flank. Although there are no reliable records of historic eruptive activity, unconfirmed reports suggest two eruptive events in 1930 and 1937 (Simkin & Siebert 1994). The volcanic edifice is formed by a series of lavas and domes, mainly of dacitic composition, although many andesite flows also occur, associated with block-and-ash flow deposits, located mainly on its southwestern flank. It has been active at least since the Middle Pleistocene (c. 720 ka; Clavero et al. 2006), with its more recent eruptive activity located on its upper western flank (c. 50 ka). It exhibits a permanent degassing activity, which mainly consists of CO$_2$ and SO$_2$, also accompanied by seismic activity (Clavero et al. 2005). The edifice has suffered strong glacial erosion and is partially covered with moraines on its lower flanks. It also suffered a partial sector collapse of its southern flank, directed towards the south. This generated a debris avalanche deposit that partially inundated the Caracarani basin (Clavero et al. 2006).

**Parinacota volcano (18.2°S, 69.1°W, 6345 m)**

Located on the Chile–Bolivia border, this is a prominent young stratovolcano (Fig. 5.3) that has evolved in three main stages beginning in the late Pleistocene (Wörner et al. 1988, 2000b; Davidson et al. 1990; Clavero et al. 2002; 2004a; Hora et al. 2006).
A dacitic volcano (35 km³; Fig. 5.4) which has been active since 2004. Although there are no records of historical eruptive activity, the whole of the new, almost perfect, volcanic cone has been built in the last 8000 years, and Aymara legends talk about hot clouds coming down from the volcano shortly before the Spanish arrived in the area. During its earlier stage (Parinacota 1, late Pleistocene, 300–40 ka) rhyolitic to andesitic magmas were erupted, forming a voluminous lava-dome complex with associated pyroclastic deposits (mainly block-and-ash flow deposits), deposited in the Upper Lauca basin. It later evolved to a steep-sided composite stratocone (Parinacota 2, late Pleistocene–Holocene, 40–8 ka), mainly formed by andesitic and dacitic lava flows and scoria tephra fallout deposits. Around 8000 years ago this ancestral Parinacota volcano partially collapsed towards the west, inundating the upper Lauca basin in a single and catastrophic event that produced the Parinacota Debris Avalanche deposit (140 km² and 6 km³; Clavero et al. 2002, 2004a). Soon after the collapse a new stratocone started to build, with the emission of andesitic lava flows and pyroclastic flows and their associated fallout deposits (Holocene, <8 ka). Some lahars, mainly shed towards the south, west and east, were generated in Holocene times. Contemporaneously with the formation of the central cone, a series of flank cones and their associated basaltic andesite to andesitic lava flows were formed (Ajata centres, 6–1.4 ka; Entenmann 1994; Wörner et al. 2000b; Clavero et al. 2004a). The new cone (Parinacota 3 unit) has an estimated minimum volume of 18 km³, which gives a minimum eruption rate of 2.25 km³/ka for the last 8000 years, making Parinacota volcano one of the most active volcanoes in northern Chile during the Holocene.

Taapaca Volcanic Complex (18.1°S, 69.5°W, 5850 m)
Located on the western margin of the western Cordillera, this is a dacitic volcano (35 km³; Fig. 5.4) which has been active since at least the lower Pleistocene (Wörner et al. 2000b; Clavero et al. 2004b). Although it has no record of historical eruptions, it has had persistent effusive and explosive eruptive activities during both the late Pleistocene and Holocene (Clavero et al. 2004b). Volcanism has built a substantial volcanic complex, including a large partially eroded stratocone and a large dome complex. Four stages of evolution are recognized, with volcanism occurring in short bursts between much longer periods of dormancy. Apart from early poorly preserved silicic andesites in its first stage, Taapaca has generated remarkably similar porphyritic hornblende–biotite dacites with distinctive sanidine megacrysts for at least 1.5 Ma. The main products of the volcano are dacite lavas and domes with associated pyroclastic deposits, of which the most common ones are block-and-ash flow deposits. There have also been several sector collapses that have generated debris avalanches. These were closely associated with volcanic blasts and episodes of dome growth. The main focus of volcanic activity has migrated 4–5 km towards the SW with time. Late Pleistocene to Holocene activity has involved at least three sector collapses of hydrothermally altered domes along the flanks of the stratocone. Volcanic blasts, block-and-ash flows, debris avalanches and lahars have been distributed down the southwestern flanks of the complex. These areas correspond to the main populated part of the Chilean Altiplano, including Putre, the largest town of the Chilean Precordillera (Fig. 5.4), and also the location of the main road between Bolivia and the Pacific coast. For this reason Taapaca is the most hazardous volcano in northern Chile, since a future eruption would certainly produce several of the recurrent volcanic hazards mentioned above and therefore threaten these areas.

Guallatiri volcano (18.4°S, 69.2°W, 6071 m)
Together with the Acotango, Humarara and Capuruta centres, this forms a southward-younging north–south volcanic chain (García et al. 2004). Guallatiri is the southwesternmost and youngest volcano of this chain. It exhibits permanent degassing activity on its upper western and southern flanks. According to historic records it has been one of the most active volcanoes of the northern Chilean Andes during the nineteenth and twentieth centuries, with several small explosive eruptions which have generated thin tephra fallout deposits (Simkin & Siebert 1994). Guallatiri volcano consists essentially of a dome complex that has evolved in two main stages of middle to late Pleistocene and late Pleistocene to Holocene ages (García et al. 2004). It has generated a series of silicic andesite to dacite domes and lava-domes and their associated block-and-ash flow fans and tephra fallout deposits, with ages ranging from 710 to 5 ka. Block-and-ash flow fans are developed on its south and southwestern flanks, whereas the tephra fallout deposits are mainly located on the southern and eastern flanks of the volcano. A future eruption of Guallatiri could threaten the village of Guallatiri, located on its lower southern flank, and tephra could affect small villages in western Bolivia.

Irruputuncu volcano (20.7°S, 68.6°W, 5163 m) (Fig. 5.5)
This is located on the Chile–Bolivia border close to the large Collahuasi porphyry copper mine. Irruputuncu volcano shows permanent degassing activity, mainly through its central summit. The main gas released to the atmosphere corresponds to SO₂ (Clavero et al. 2005). Despite this permanent degassing, there are no reliable records of historical eruptive activity,
although unconfirmed press reports suggest a small eruptive event in 1989 (Simkin &Siebert 1994). Irruputuncu consists mainly of silicic andesite to dacite domes and lava-domes and their associated block-and-ash flow fans, especially developed towards the western and southern flanks. It has developed two craters, the southwestern one currently active. Wörner et al. (2000b) reported one reliable K–Ar age of a young dome on the upper western flank of c. 140 ka. Clavero (unpublished data) obtained a young 1570 ± 90 years BP 14C date for a very fresh-looking block-and-ash flow deposit on the southwestern flank, demonstrating that Irruputuncu volcano has had explosive activity related to dome collapse in the Late Holocene. Although located in a very remote area, renewed activity could affect the main access road from Iquique to Collahuasi copper mine.

Ollagüe volcano (21.3ºS, 68.2ºW, 5868 m) (Fig. 5.6)
Located on the Chile–Bolivia border SW of the large Uyuni Salar, this volcano exhibits permanent fumaroles on its upper western and southern side, the latter being recently more active. The main gases are SO2 and H2O, with minor fluxes of CO2 close to the fumarole vent (Clavero et al. 2005). Despite this fumarolic activity, there are no reliable records of historic eruptive activity, although unconfirmed reports suggest that an eruption may have occurred in 1903 (Simkin &Siebert 1994), and that an increase in fumarole ‘glowing’ may have occurred in November 2005 (A. Amigo and V. Solear, pers. comm.). Ollagüe volcano is a long-lived stratovolcano, active since at least 1.2 Ma, that has produced rhodacitic domes, silicic andesite lavas and domes, and dacitic domes and lava-domes, with their associated pyroclastic flow, surge and block-and-ash flow deposits (Feeley et al. 1993; Feeley & Davidson 1994; Clavero et al. 2004c). The volcano has evolved mainly in three stages (Clavero et al. 2004c). Initially, at 0.9–1.2 Ma, a rhodacitic dome complex formed along with its associated pyroclastic flow fan, including column-collapse pyroclastic flows and dome-collapse block-and-ash flows, which were directed mainly towards the northern and western flanks. It later evolved (600–900 ka) to a stratocone formed by andesitic to dacitic lavas, domes and associated block-and-ash flow deposits. During this stage the parasitic La Poruñita pyroclastic cone was also erupted on the lower western flank of the volcano. At some point between 400 and 600 ka, a large sector collapse partially destroyed the ancestral edifice, generating a volcanic debris avalanche towards the west, covering a c. 50-km²-area of the Carcote Salar with an estimated debris volume of 1 km³, and involving salar deposits in the avalanche flow. Since the collapse (<400 ka), a series of dacitic domes and andesitic lavas have formed and partially filled the avalanche amphitheatre on the western flank of the volcano. Tephra fallout as well as block-and-ash flow deposits have also been generated, the latter forming the youngest pyroclastic fan towards the lower western flank of the volcano.

San Pedro (21.9ºS, 68.4ºW, 6145 m)
This is a very large composite volcano, its peak rising 2500 m above its base. It forms the largest centre in a 16 km east–west volcanic lineament which also includes San Pablo volcano.
an estimated volume of 30–40 km$^3$, active since 230 ka, whose volcano consists of an almost permanent degassing through the of Talabre, the closest Chilean town to the volcano, 12 km to volumes of ash, which was dispersed towards northwestern occurred in May 2005 when a small explosion generated limited (before the time of writing: December 2005) of Láscar volcano Medina 1994; Calder Lindsay 2004). These flows were directed towards the north- formed when the column partially collapsed (Gardeweg & Lindsay 2004) developed on an (Matthews 1974). The amphitheatre generated by this collapse was filled by viscous lavas and pumice flows and block-and-ash flows, originally described as ‘hot avalanche deposits’ (Francis et al. 1974).

Láscar volcano (23.4°S, 67.7°W, 5592 m) Located east of the Salar de Atacama, Láscar is the most active volcano of the northern Chilean Andes in historic times (Simkin & Siebert 1994), with more than 15 eruptive events of different magnitude in the twentieth century. The largest eruption (volcanic explosive index, VEI = 3–4) in the last 9000 years occurred in April 1993 (Fig. 5.7; Gardeweg et al. 1998b), when a subplinian eruptive column more than 20 km high rose above the crater, marking the end of an eruptive cycle that started in 1984 (Matthews et al. 1997). Small-volume pumice flows were formed when the column partially collapsed (Gardeweg & Lindsay 2004). These flows were directed towards the northwestern and southern flanks of the volcano, reaching up to 9 km from their source (Guarinos & Guarinos 1993; Gardeweg & Medina 1994; Calder et al. 2000). The most recent eruption (before the time of writing: December 2005) of Láscar volcano occurred in May 2005 when a small explosion generated limited volumes of ash, which was dispersed towards northwestern Argentina. This event was not even noticed by the inhabitants of Talabre, the closest Chilean town to the volcano, 12 km to the west (Naranjo et al. 2006). The current activity of Láscar volcano consists of an almost permanent degassing through the active vent.

Láscar is a young, elongate, composite stratovolcano, with an estimated volume of 30–40 km$^3$, active since 230 ka, whose structure is formed by two cones, on which five craters are developed, aligned in an ENE direction. Its eruptive activity has been divided into four stages (Gardeweg et al. 1998b). Stage I (230 to >26 ka; Gardeweg & Lindsay 2004) developed on an older stratocone formed mainly by andesitic lavas developed on the eastern edge of the present edifice. This stage culminated with the eruption of andesitic pyroclastic flows. During Stage II (>26–22 ka) the magma composition changed to silicic andesite and dacite, and the eruptive activity migrated to the west, probably related to the formation of a dome complex. This stage ended with the Plinian eruption of a major ignimbrite: the Soncor ignimbrite (Gardeweg et al. 1998b; Matthews et al. 1999), which shows strong evidence for magma zonation, with juvenile material ranging from andesite to rhyodacite. A debris avalanche and the formation of a fluvioglacial fan followed this large eruption between 22 and 19 ka. During Stage III (19.2–9.2 ka) a new stratocone was formed in the vent area of the Soncor eruption, consisting mainly of silicic andesite to dacite lavas, together with some andesitic pyroclastic flow deposits. This stage ended with a major explosive event, the Tumbres Eruption, which generated thin pumice fall deposits and a series of scoriaceous andesitic pyroclastic flow deposits. After the Tumbres event (Stage IV, 9.2 ka to the present) the activity shifted again to the east. Three aligned craters were formed and a series of thick lavas were erupted. The historical activity has been mainly characterized by periods of dome growth, crater subsidence and Vulcanian explosions. Since 1984, the activity has consisted of dome growth, crater subsidence, Vulcanian explosions and explosive eruptions, including the April 1993 Subplinian eruption, which generated small-volume pumice flow deposits.

Although located in a remote zone, far from populated areas, the 1993 eruption generated important effects in northwestern Argentina, demonstrating that the major hazard related to Chilean CVZ volcanoes corresponds to the effects of tephra dispersal reaching populated areas to the east in northwestern Argentina or southwestern Bolivia due to the predominant low-altitude wind direction in the Altiplano of northern Chile (Amigo & Gallardo 2004). However, larger eruptions could reach higher altitudes, where the wind direction changes, and tephra could be directed towards populated areas to the west in Chile (Amigo et al. 2006), as occurred several times between the Miocene and the Pleistocene in northern Chile (Basso et al. 2004).

Socompa volcano (24.4°S, 68.3°W, 6051 m) Located along the Chile–Argentina border, this is a large stratovolcano without historical records of eruptive activity (de Silva & Francis 1991). Socompa volcano has mainly evolved in two stages, separated by a large sector collapse that affected the northwestern part of the edifice: The initial stratocone, which was higher than the current one, was formed mainly by andesitic to dacitic lavas, coulées, domes and subordinate pyroclastic flow and tephra fallout deposits (Déruelle 1978; Wadge et al. 1995). At c. 7,200 years BP (Francis & Wells 1988), a large volcanic debris avalanche, the largest known in the CVZ of the Andes, was formed, involving the partial collapse of the volcanic edifice and large volumes of its ductile basement (Francis et al. 1985; Wadge et al. 1995, van Wyk de Vries et al. 2001). The avalanche flow, directed towards the NW, travelled as far as 40 km from its source, inundating a total area of more than 500 km$^2$, and generating a deposit with a minimum volume of 36 km$^3$ which was originally interpreted as the result of a ‘nuée ardente’ (Déruelle 1978). A pyroclastic flow, directed towards the NE, has been interpreted to have immediately followed the sector collapse (Francis et al. 1985). After this catastrophic event, a series of post-collapse domes, coulées and lavas, mainly dacitic in composition, partially filled the amphitheatre created by the collapse. A series of explosion craters within the collapse amphitheatre are likely to represent the most recent eruptive activity of Socompa volcano (Wadge et al. 1995).

Llullaillaco volcano (24.7°S, 68.5°W, 6739 m) Located on the Chile–Argentina border, this is considered to be the second highest active volcano summit in the world.

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Fig. 5.7. April 1993 Subplinian (VEI = 3–4) eruption of the Lascar volcano. Photo courtesy of M. Gardeweg.
Although it shows no signs of current fumarolic activity, there are records of at least three eruptions during the nineteenth century (Simkin & Siebert 1994). The evolution of Lullaillaco volcano has been divided into two main stages according to geological mapping both in Chile (Gardeweg et al. 1984) and Argentina (Zapettini & Blasco 1998) and geochronological data (Richards & Villeneuve 2001). The first constructive stage (early Pleistocene) generated extensive dacitic lavas from two centres. These lavas have been strongly affected by hydrothermal alteration and glacial erosion. The second stage (late Pleistocene) generated a series of viscous dacite lavas, coulées and domes and associated block-and-ash flow deposits (de Silva & Francis 1991), forming a steep edifice built on top of the older eroded one. At some point during this late Pleistocene constructive phase (at c. 150 ka or later; Richards & Villeneuve 2001), a partial sector collapse affected the volcanic edifice, generating a large debris avalanche flow directed towards the east, which travelled for at least 25 km, resulting in a deposit that covers an area of c. 165 km² and has an estimated minimum volume of 1–2 km³.

**Lastarria (25.2°S, 68.5°W, 5697 m)**

This is a late Pleistocene to Holocene composite volcano which includes three morphostructural components: the Espolon Sur (Southern Spur), Lastarria volcano *sensu stricto*, whose activity has migrated northward with time along four seminested craters, and Negriales, a nearby 5.4 km² lava field (Naranjo 1988, 1992). PETROGRAPHICALLY, the Lastarria complex consists of pyroxene andesites and pyroxene–amphibole dacites (Naranjo 1992). Lastarria has notably active fumaroles, but no reported record of historic eruptions. The stratigraphic succession suggests that the evolution of the volcanic complex started at Espolon Sur, and that Lastarria and Negriales evolved contemporaneously. No visible break occurs between the older and younger units. The modern Lastarria cone consists of well preserved lava flows and pyroclastic flows, including block-and-ash flows derived from the northernmost dome. A debris avalanche interpreted as a notably high-velocity flow was produced on the southeastern flank of the volcano after the collapse of the tephra-fall layers that form the entire upper part of the slightly concave slope of the cone (Naranjo & Francis 1987). In addition, 220–350-m-long sulphur flows, exhibiting well preserved pahoehoe structures, resulted from melting of hydrothermal sulphur deposits on the northwestern flanks of the cone due to enhanced thermal activity (Naranjo 1988).

**Nevado Ojos del Salado Volcanic Complex (27.1°S, 68.5°W, 6687 m)**

Located on the southern edge of the CVZ (Fig. 5.2), this is the highest active volcano summit in the world. It forms a volcanic group, together with the Pleistocene volcanic centres Nevado Tres Cruces, Volcán Solo, El Fraile and San Francisco (Gardeweg et al. 1997), oblique to the main chain of Pleistocene and Holocene CVZ volcanoes. Despite the fresh-looking aspect of the youngest lavas and domes, there are no reliable records of historical eruptive activity (Simkin & Siebert 1994), although unconfirmed reports by mountaineer guides suggest there was fumarolic activity in the 1980’s (González-Ferrán et al. 1985). The Nevado Ojos del Salado volcanic complex corresponds to an extensive (c. 160 km²), long-lived field of andesitic, dacitic and minor rhyodacitic lavas, coulées and domes (Baker et al. 1987b; Gardeweg et al. 1997, 1998a) that has been active at least from early Pleistocene (c. 1.5 Ma, based on an age for a dacite lava of the lower northern flank) to late Pleistocene (c. 35 ka, based on the age for an upper dome of the northern flank) times, according to geochronological data (Gardeweg et al. 1997, 1999). This implies that its last eruptive activity was older than 30 ka. Despite the high silica content of most of the products of the Nevado Ojos del Salado volcanic complex, no important pyroclastic deposits have been generated associated with this activity. The extensive pyroclastic cover of the volcano and its surroundings corresponds to a large explosive eruption of its neighbour Nevado Tres Cruces volcano, which occurred c. 67 ka ago.

**Nevado Tres Cruces volcano (27.1°S, 68.8°W, 6748 m)**

Located 20 km to the west of the Ojos del Salado volcanic complex, this eruptive centre also forms part of the oblique Pleistocene volcanic chain mentioned above. There are no records of historical activity, and so far the centre has not been listed as an active volcano (Simkin & Siebert 1994). Nevado Tres Cruces, which consists of at least three coalescent cones aligned in a north–south direction, is a long-lived stratovolcano, active since at least 1.5 Ma (Gardeweg et al. 2000). It is formed by a series of dacitic to rhyodacitic lavas, coulées, domes, explosive craters, and small-volume pyroclastic flow, surge and tephra fallout deposits (Gardeweg et al. 1997, 1999, 2000). It has produced at least two significant Pleistocene explosive eruptions, which have generated pyroclastic flows and tephra fallout deposits. The oldest of these explosive events took place at c. 1.5 Ma, generating a small-volume pyroclastic flow directed towards the west, whereas the youngest one took place at c. 67 ka, producing medium-volume pyroclastic flows, including 15-m-thick basal surge deposits, directed towards the east and SE. This eruption was associated with extensive tephra fallout deposits, which partially cover most of the surrounding areas towards the east, and were originally interpreted as being related to the Nevado Ojos del Salado volcanic complex (González-Ferrán et al. 1985). The youngest product of the volcano so far dated corresponds to a dacitic lava flow on its upper southern flank (c. 28 ka; Gardeweg et al. 2000), which implies volcanic activity in very late Pleistocene times.


Volcanoes of the Southern Volcanic Zone (SVZ; Figs 5.1 & 5.8) occur both in Chile and Argentina. In Chile alone, the SVZ includes >70 Pleistocene and Holocene composite stratovolcanoes and large volcanic fields, and at least nine caldera complexes, as well as hundreds of minor eruptive centres (MEC) formed by scoria cones ± lava flows and maars. These form a continuous volcanic arc segment 1400 km long, extending from 33.3°S to 46°S. Between 33.3°S and 34.4°S the SVZ is a narrow chain of volcanoes located along the Chile–Argentina border, but between 34.4°S and 39.5°S the arc widens to >200 km and occurs in both Chile and Argentina (Fig. 5.8). South of 39.5°S the SVZ again consists of a relatively narrow chain of volcanoes with the volcanic front in Chile, and south of 42°S all the arc volcanoes are located in Chile.

The SVZ has more historic volcanic activity, with about one eruption per year on average, than either of the other two volcanic zones in Chile. Villarrica and Llaima volcanoes are two of the most active volcanoes in the entire Andean arc, with together more than 80 reported episodes of activity since 1558 (Simkin & Siebert 1994; Petit-Breuilh 1994; Petit-Breuilh & Lobato 1994; Naranjo & Moreno 2005). The largest historic eruption (c. 9.5 km³) of any Chilean volcano, and the largest of any Andean volcano in the last 100 years, was generated by Quizapu in 1932 (Hildreth & Drake 1992). Among the >70 stratovolcanoes complexes and volcanic fields between 33.3°S and 46°S, at least 40 have erupted in the Holocene and 20 have had historic eruptions. Most of the hundreds of MEC cones and maars are also Holocene, and three of them (Riñahue, Carrán and Mirador) erupted during the last century. Six of the nine calderas have erupted in the past <1 Ma. Large volumes of middle Pleistocene rhyolite pyroclastic flows derived from the Diamante and Calabozos calderas (Fig. 5.8; Stern et al.
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Fig. 5.8. Schematic map of the location of some of the volcanoes and larger caldera systems in the Southern Volcanic Zone. Dark cross-hatched line indicates the eastern limit of Cenozoic arc volcanism, which divides the area of transitional (to the west) from cratonic (to the east) alkali olivine backarc basalts of the Patagonian plateau lavas (Stern et al. 1990). The approximately north-south dashed line labelled LOFS is the Liquiñe–Ofqui Fault System.

1984a; Hildreth et al. 1984) have travelled more than 100 km from their sources in the High Andes and crossed the Central Valley between Santiago and Puerto Montt where the largest part of the population of Chile lives. Although such large eruptions are infrequent, Hildreth et al. (1984) noted that the ‘long record of voluminous silicic pyroclastic activity’ associated with these giant systems ‘has important implications for the safety of several major cities near the mouths of Andean canyons’. Smaller, but still aerially extensive pyroclastic flows, derived from explosive eruptions of other SVZ volcanoes, as well as lahars and debris flows, have also covered what are today highly populated regions of the Central Valley.
**Geological and tectonic setting**

The Southern Volcanic Zone (SVZ) of central-south Chile results from the subduction of the oceanic Nazca Plate under the South American Plate (Figs 5.1 & 5.8). Its northern end coincides with the impingement of the Juan Fernández Ridge (JFR) upon the Chile–Peru trench, which is accompanied by a dramatic change in the subduction angle of the Nazca Plate below South America (Fig. 5.1). The southern boundary of the SVZ is the intersection of the Chile Ridge with the trench. Convergence direction between the Nazca and South American plates is oblique (c. 20–30°; Jarrard 1986; Dewey & Lamb 1992) and convergence velocity is 7–9 cm/year (DeMets et al. 1990). The age of the Nazca Plate at the trench decreases from 45 Ma in the north to 0 Ma at the southern end of the SVZ. Subduction angle increases from c. 20° at the northern end of the SVZ to >25° further south. As a consequence, the distance from the trench to the volcanic front decreases from >290 km in the north to <270 km in the south. In part because of the change in the distance of the volcanic front from the trench, the volcanoes along the front at the northern end of the SVZ occur in the High Andes along the drainage divide between Chile and Argentina, while to the south they occur in Chile along the western margin of the Andean Cordillera with the Central Valley, an extensional basin formed at the beginning of late Oligocene times (Muñoz et al. 2000). Crustal thickness also decreases from 55–60 km at the northern end of the SVZ to <35 km south of 37°S. The crust below the SVZ consists of Palaeozoic and Mesozoic pre-Andean basement and Mesozoic–Cenozoic igneous rocks.

In southern Chile (39–47°S), oblique convergence results in slip along the arc-parallel Liquiñe–Ofqui Fault System (LOFS in Fig. 5.8; Cembrano et al. 1996, 2000; Lavenu & Cembrano 1999a, b; Arancibia et al. 1999). This fault system controls the location of many of the larger volcanic centres as well as MEC monogenetic cones in the southern SVZ (López-Escobar et al. 1995a). In contrast, in northern and central Chile (20–34°S), crustal shortening occurs with only minor arc-parallel slip (Dewey & Lamb 1992). A transitional zone between these two areas occurs in central-south Chile (34–39°S), where the SVZ is widest and includes extensional intra-arc basins with small monogenetic basaltic cones and alkali basalt lava fields (Fig. 5.8; Muñoz & Stern 1988, 1989; Folguera et al. 2002). In this part of the arc the location of stratovolcanoes and calderas is controlled chiefly by NW–SE and NE–SW structures.

**SVZ volcanoes and volcanic hazards**

Post-glacial eruptions of SVZ volcanoes include the full range of Hawaiian, Strombolian, Subplinian, Plinian, Vulcanian and phreatomagmatic types, with VEI (Newhall & Self 1982) ranging from 0 to 6. There appears to be a relation between magma composition and eruption style. Basaltic to basaltic andesitic magmas have had mainly Strombolian eruptions, which normally can also include Hawaiian phases, such as the typical historic eruptions of Antuco, Llaima, Villarrica and Osorno. High-silica andesitic and dacitic magmas commonly have had Subplinian eruptions, as for example the historic eruptions of Descabezado Grande (1932), Cordón Caulle (1922 and 1960) and Calbuco (1961). High-silica dacites to rhyolitic magmas have had Subplinian and Plinian eruptions, such as the eruptions of Quizapu in 1932 and Hudson in 1991. Important exceptions are the pre-historic eruptions of the 24 km³ basaltic to basaltic andesite Curacautín Ignimbrite from the Llaima volcano at c. 13.5 ka (Naranjo & Moreno 1991), and the c. 6.7 ka Hudson explosive eruption of >18 km³ of low-silica andesite (Stern 1991a, Naranjo & Stern 1998).

Lahars, lava flows and ashfalls have been the main volcanic hazards within historic times. Lahars produced by the eruptions of Villarrica volcano in 1948–49, 1963–64 and 1971–72 together killed more than 75 people (Lara & Clavero 2004). A giant lahar generated in 1971 by melting of glacial ice in the Hudson caldera flowed over 40 km from its source to the sea (Best 1992). Lahars may also be generated by lava flows temporarily obstructing drainage systems. Ashfall produced by both the 1932 eruption of Quizapu and the 1991 eruption of Hudson adversely affected very large regions to the east in Argentina. Another introduced by explosive eruptions in the SVZ has the potential to significantly disrupt both domestic and international air traffic across the southern part of South America (Hauser & Naranjo 2006).

Pyroclastic flows and surges, together with voluminous debris avalanches, have also occurred during latest Pleistocene and Holocene times. Three late Pleistocene large-volume pyroclastic flows have crossed the Central Valley between Santiago and Puerto Montt, travelling more than 100 km from their sources. Lateral blasts followed by small surges took place in the 1864 eruption of Mocho-Choshuenco, and the 1893–95 and 1929 eruptions of Calbuco volcanoes. San Pedro, Antuco and Calbuco had Bezymianian- and Bandai-types cone sector collapses between 14000 and 6200 years BP, generating large debris avalanches (6–15 km³).

On the basis of the tectonic controls affecting the distribution of volcanic centres, as well as petrologic and geochemical considerations, the Pleistocene and Holocene volcanoes of the SVZ are divided into four main provinces or segments (Fig. 5.8; Tormey et al. 1991a; López-Escobar et al. 1995a; Stern 2004). These are the Northern (NSVZ; 33.3–34.4°S), Transitional (TSVZ; 34.4–37°S), Central (CSVZ; 37–42°S) and Southern (SSVZ; 42–46°S) segments. Volcanoes and volcanic hazards for each of these provinces are discussed separately below.

**The NSVZ**

The NSVZ is formed by eight composite stratovolcanoes and the giant Diamante caldera, aligned in a north–south trend along the High Andes drainage divide between Chile and Argentina (Fig. 5.8). An older early to mid-Pleistocene volcanic chain forms the base for younger late Pleistocene to Holocene volcanoes. The volcanic front in the Pliocene was further to the west, but migrated to its current position as a result of both decreasing subduction angle and subduction erosion below the northern end of the SVZ (Stern 1989, 1991b; Kay et al. 2005), possibly due to the southward migration of the locus of subduction of the Juan Fernández Ridge (Fig. 5.1; Stern & Skewes 1995, 2005; Yahre et al. 2001, 2002).

Summit elevations of volcano edifices in the NSVZ are <2000 m above their base, and basal elevations decrease in height from north to south. Thus Tupungato volcano (6570 m), the northernmost and highest volcano in the SVZ, has its base over Mesozoic strata at 4600 m, while the Maipo volcano (5290 m), the southernmost in the NSVZ, located inside the Diamante Caldera (Fig. 5.9), overlies basement with an average elevation of 3600 m. The Diamante caldera erupted an estimated 450 km³ of rhyolite pyroclastic material in the middle Pleistocene (Stern et al. 1984a). The largest stratovolcano in the region is the lower to middle Pleistocene Marmolejo centre (Fig. 5.10), but the volumes of other NSVZ volcanoes are relatively small, from 5 to 55 km³ (Hildreth & Mooar 1988). This reflects a relatively low rate of magma extrusion from the stratovolcanoes in the NSVZ during the late Pleistocene and Holocene (Stern et al. 1984a).

From north to south in the NSVZ, the older group of centres includes the Nevado de Huéqueo (6019 m) and Marmolejo (6108 m; López-Escobar et al. 1985) centres, and further south the isolated Castillo (5485 m) volcano, the northernmost and highest volcano in the SVZ, has its base over Mesozoic strata at 4600 m, while the Maipo volcano (5290 m), the southernmost in the NSVZ, located inside the Diamante caldera (Fig. 5.9), overlies basement with an average elevation of 3600 m. The Diamante caldera erupted an estimated 450 km³ of rhyolite pyroclastic material in the middle Pleistocene (Stern et al. 1984a). The largest stratovolcano in the region is the lower to middle Pleistocene Marmolejo centre (Fig. 5.10), but the volumes of other NSVZ volcanoes are relatively small, from 5 to 55 km³ (Hildreth & Moorbath 1988). This reflects a relatively low rate of magma extrusion from the stratovolcanoes in the NSVZ during the late Pleistocene and Holocene (Stern et al. 1984a).
the San José volcanic complex (5856 m) overlying Marmolejo (Fig. 5.10), and Maipo volcano inside the Diamante caldera (Fig. 5.9).

Among these volcanoes, Tupungatito, San José and Maipo have had historic eruptions, most recently in 1987, 1960 and 1912, respectively, and both Tupungatito and San José show permanent solfataric activity. These three historically active volcanic centres all have high summit elevations (> 5290 m), are snow- and ice-covered, and occur at the source of relatively narrow mountain drainage systems that ultimately feed into the Maipo River, which flows through the southern edge of Santiago, the largest city in Chile. Thus, although these centres are not visible from Santiago, and are located relatively far from the city, they pose a potential sector-collapse, debris-flow hazard to the > 6 000 000 inhabitants of the city, as well as tephra-fall hazard that might impact the Province of Mendoza to the east in Argentina and disturb air traffic in both countries.

Rhyolite pyroclastic material from the middle Pleistocene eruption that formed the giant Diamante caldera flowed into the area of the northern Central Valley now occupied by the cities of Santiago and Rancagua, forming the > 30-m-thick Pudahuel ignimbrite, as well as out to the east onto the plains south of Mendoza, Argentina (Stern et al. 1984a). However, such very large eruptions are extremely infrequent, and no such giant eruption has occurred anywhere in the Andes during Holocene times.

The TSVZ
From 34.4°S to 37.0°S, the Pleistocene and Holocene volcanic front trends N20°E. The arc also widens considerably eastward into Argentina (Fig. 5.8), with arc volcanic centres located along NW–SE trending uplifted blocks separated by intra-arc extensional basins containing small basaltic cones and lava flows (Muñoz & Stern 1988, 1989). Within the arc, transverse alignments control the location of several volcanic vents south of the Planchón–Peteroa–Azufre volcanic complex. Thus, Descabezado Chico and the San Pedro–Tatara–Pellado volcanic complexes are arranged along a NE–SW alignment and a NW–SE trend controls both the Nevado de Longaví–Loma Seca–Resago volcanoes and the Nevados de Chillán volcanic complex.

The TSVZ is characterized by compound stratovolcanoes and large volcanic complexes that overlie older basaltic shield volcanoes, as well as volcanic fields and giant calderas. Most of the latter are related to late Pliocene to Pleistocene volcanic activity. The heights of the volcanic edifices in the TSVZ are < 1900 m above their bases, and most of them are only between 1400 and 1700 m. Their volumes are relatively small, from 3 to 60 km³ (Hildreth & Moorbath 1988), but some of the caldera systems have erupted large volumes of Pleistocene rhyolites (Hildreth et al. 1984, 1999). Basal elevations along the volcanic front decrease from north to south, both because the Andes are not as high towards the south, and also because the location of the volcanic front moves westward from along the drainage divide between Chile and Argentina towards the western edge of the Cordillera (Hildreth & Moorbath 1988). Thus the small Palomo volcano (4850 m) overlies Mesozoic and Cenozoic rocks that have a height of about 3400 m, while the larger Nevados de Chillán volcanic complex (3216 m) overlies a basement with an average height of only 1500 m. Numerous small cities and rural population centres in the Central Valley occur along rivers that have their origin on the flanks of large TSVZ volcanoes.

From north to south, the 12 most important middle–late Pleistocene to Holocene TSVZ volcanic centres and volcanic fields in Chile are Palomo (4850 m), the Tinguiririca cluster (Arcos et al. 1988), Planchón–Peteroa–Azufre (Tormey et al. 1995; Naranjo et al. 1999a; Naranjo & Haller 2002), the Descabezado Grande–Quizapu–Azul volcanic complex (Hildreth & Drake 1992), the Descabezado Chico volcanic complex (3250 m), the Calabozos caldera (Hildreth et al. 1984), the Del Medio volcanic complex (3508 m), the San Pedro–Tatara–Pellado volcanic complex (Dungan et al. 2001), the Laguna del Maule volcanic field (3175 m; Frey et al. 1984; Hildreth et al. 2004), which has a longer history that began in Pliocene times, the Nevado de Longaví (3242 m; Sellés et al. 2004), Loma Blanca (2230 m), Resago (1890 m) and the Nevados de Chillán volcanic complex (Dixon et al., 1999; Naranjo and Lara, 2004). Older highly eroded stratovolcanoes and calderas, probably of lower to middle Pleistocene age, include the western margin of the giant 35 × 45 km wide Del Atuel caldera, Alto del Padre–Sordo Lucas (3548 m) and Campanario (4020 m; Hildreth et al. 1998).

The largest volcanic areas in the TSVZ of Chile are the ones formed by the Descabezados, Calabozos and Laguna del Maule volcanic fields, which together contain more than 200 vents, including stratovolcanoes, craters, scoria cones, domes and maars. The main historically active and potentially hazardous volcanic complexes, each summarized briefly below, are Tinguiririca, Planchón, Descabezado Grande–Quizapu–Azul, San Pedro–Tatara–Pellado and Nevados de Chillán.

**Tinguiririca (34.48°S, 70.3°W, 4300 m).** This is a Holocene cluster of about ten scoria cones, the southernmost one called Fray Carlos. The cluster overlies a lower to middle Pleistocene plateau of andesitic lavas (Arcos et al. 1988). The last eruption was in 1917 and this centre has permanent solfataric activity.
Planchón–Peteroa–Azufre volcanic complex (35.2°S, 75.6°W). This comprises the Pleistocene–Holocene Planchón stratovolcano (3977 m), which is developed on top of the older eroded volcanic centres of Peteroa (4107 m; located 2.5 km south of Planchón) and Azufre stratovolcano (3603 m). The whole volcanic field covers an area of c. 78 km². The evolution of Planchón is characterised by the emission in its earlier stage of a series of basaltic lavas. A large-volume (> 10 km³) sector collapse affected this structure during the last glaciation (c. 90–20 ka), generating a giant debris avalanche, directed towards the west, which travelled downvalley for up to 95 km (Naranjo et al. 1999a; Naranjo & Haller 2002). The Holocene activity is mainly explosive, with the formation of five explosive craters, and the emission of more evolved magmas, such as the c. 1000 years bp Los Baños Subplinian eruption which generated a dacite pumice fallout deposit. In 1837 a small lava flow and a scoria cone were formed within the Planchón collapse caldera and both phreatic as well as low-magnitude phreatomagmatic eruptions have taken place from these craters in 1991 and 1998. Based on its most recent activity, the major hazards related to Planchón volcano are small phreatic eruptions with the emission of tephra fallout, which could involve acid fluids, causing contamination of the surrounding drainages.

Descabezado Grande–Quizapu–Azul volcanic complex (35.6°S, 70.8°W). This complex has more than 40 vents, covering an area of about 500 km². Descabezado Grande (3953 m) and Azul (3788 m) are stratovolcanoes, and Quizapu (3050 m) is a flank vent of the latter. The volcanic complex has generated large Pleistocene ignimbrites (0.8–0.3 and 0.15 Ma; Hildreth et al. 1984) and a large volume of both tephra and lavas during Holocene times, including historic eruptions. In April 1932, Quizapu had the largest Plinian eruption of any Andean volcano in the 20th century, releasing c. 9.5 km³ of dacitic pumice (Hildreth & Drake 1992), although sustained Plinian activity lasted less than 20 hours and produced an eruptive column only c. 30 km high. Nevertheless, tephra dispersal toward the east covered one-third of Argentinian territory, an area of more than 2 x 10⁶ km², with pumice-fall causing heavy damage to agriculture and cattle (Fig. 5.11; Larsson 1937). A similar eruption to that of 1932 would be catastrophic at the present time. Descabezado Grande and Quizapu have an eruptive record since the nineteenth century and a new eruption is likely to take place in the near future. Subplinian to Plinian pumice columns, widespread tephra dispersal and pyroclastic flows can be expected, and a very large area in both Chile and Argentina could be under a heavy ashfall impact. Also, if an eruption takes place during winter, lahars could be generated and affect the nearby Maule river hydroelectric power plants in Chile.

San Pedro–Tatara–Pellado volcanic complex (36°S, 70.7°W). This has been active from the Pliocene to the Holocene. A middle Pleistocene 6 x 12 km caldera and the eroded Pellado stratovolcano (3121 m) underlie the Tatara basaltic–andesite shield volcano (Dungan et al. 2001). The mainly andesitic San Pedro is a late Pleistocene to Holocene stratovolcano that rises above the Tatara edifice and has a small scoria cone nested within its summit crater. Active fumaroles have been found SE of Pellado.

Nevados de Chillán volcanic complex (36.9°S, 71.4°W). This forms an elongate chain comprising two major subcomplexes, Cerro Blanco and Las Termas, located 6 km apart and separated by a series of cones and explosive craters aligned

Fig. 5.11. Isopachs of tephra fall from the 1932 eruption of Quizapu (Larsson 1937; Hildreth & Drake 1992).
The CSVZ Late Pleistocene to Holocene volcanic front includes, from north to south, the following centres: Antuco (López-Escobar et al. 1981), Callaqui (Polanco 1998), Tolguaca (2806 m; Polanco 1998), Lonquimay (2865 m; Moreno & Gardeweg 1989; Naranjo et al. 1992), Llaima (Naranjo & Moreno 1991, 2005; Moreno & Naranjo 2003), Sollipulli caldera (2282 m; Naranjo et al. 1993a; Gilbert et al. 1996), Caburgua cluster (998 m), Villarrica (Clavero & Moreno 1994, 2004; Petit-Breuilh 1994; Moreno 2000; Lara & Clavero 2004), Mocho-Choshuenco complex (Echebarray 2004), Carrán–Los Venados cluster (Campos et al. 1998; Lara et al. 2006b), Cordillera Nevada caldera (1799 m), Cordón Caulle fissural sequence (Lara et al. 2004b, 2006a, b) and Puyehue caldera (Gerlach et al. 1988), Antillanca (or Casablanca) cluster (1990 m), Cenizos (1668 m), Osorno (López-Escobar et al. 1992) and Calbuco (Hickey-Vargas et al. 1995; López-Escobar et al. 1995b). Among them are lower and/or middle Pleistocene eroded stratovolcanoes such as Sierra Velluda (3585 m), Lancú (2296 m), Sierra Nevada (2554 m), Llallicupe (1915 m), Quinquilil (2002 m), Quinchila caldera (1632 m), Mencueca (1840 m), Sarnoso (1630 m), La Picada (1715 m) and Hueñuhueñu (1207 m).

Between 38°S and 39.5°S, the eastern part of the arc in Chile includes the middle Pleistocene to Holocene Copahue stratovolcano (Fig. 5.12; Varekamp et al. 2001; Polanco 2003; Naranjo & Polanco 2004) located on the western edge of the giant El Agrio (or Caviahue) caldera (Muñoz & Stern 1988, 1989; Folguera & Ramos 2000; Melnick & Folguera 2001; Melnick et al. 2006). Southeast of Copahue, small Holocene scoria cones and maars overlie a Pliocene and Pleistocene volcanic plateau and old eroded stratovolcanoes that include the Lower Pleistocene Pino Hachado caldera (Fig. 5.12; Muñoz & Stern 1988, 1989; Tunstall & Ramos 2005). Behind the volcanic front south of 39.5°S, volcanic centres are mostly eroded lower to middle Pleistocene stratovolcanoes, but a few are late Pleistocene to Holocene stratocone. These include, from north to south, Quetrupillán complex (Pavez 1997; Pavez et al. 2004), Llaima (Naranjo & Polanco 1994b; Lara et al. 2004b), Huanquihue complex (2139 m), Chihuio (Lara and Moreno 2004; Lara & Folguera 2006), Pirihueico (Lara & Moreno 2004; Lara & Folguera 2006), Mirador (1795 m), Pantoja (2024 m), Puntiagudo (2493 m), Tronador (3491 m; Mella et al. 2005), Cerro Volcánico (1930 m) and Cuernos del Diablo (1862 m).

This CSVZ is the most volcanically active segment along the whole Andes and includes two of the most active volcanoes in South America: Llaima and Villarrica. Llaima, Villarrica and Tronador are the largest individual volcanoes in the CSVZ, with volumes up to ~400 km³. Growth rates of 4 km³ in less than 10000 years have been determined for Antuco volcano. The largest volcanic fields in the CSVZ are the ones formed by the Cordillera Nevada caldera, Cordón Caulle fissural range, Puyehue, Mencueca and Carrán–Los Venados clusters that together cover an area of more than 1500 km², with almost 200 vents including stratovolcanoes, calderas, craters, scoria cones, domes, fissures and maars. Numerous minor eruptive centres such as Rucapillán, Caburgua, La Barda, Redondo, Relicura, Huelemolle, Rihinahue, Carrán and Mirador, among many others, also occur in the vicinity of the larger stratovolcanoes and along the LOSF (López-Escobar et al. 1995a).

The numerous historically active volcanoes, their large size and proximity to the eastern edge of the Central Valley make the CSVZ an area of high volcanic risk. Risk has been enhanced in this region by the construction of thousands of small drainage channels originating on the flanks of volcanoes. In 1996 the Servicio Nacional de Geología y Minería (SERNAGEOMIN) created the Observatorio Volcanológico de los Andes del Sur (OVDAS) in Temuco to monitor the more hazardous volcanoes along the SZV, particularly Llaima and Villarrica volcanoes. Currently, Lonquimay, Mocho–Choshuenco,
Osorno and Calbuco volcanoes are also seismically monitored through OVDAS.

The historically most active and potentially most hazardous CSVZ volcanic complexes, briefly described below, include Antuco, Callaqui, Copahue, Lonquimay, Llaima, Villarrica, Quetrupillán, Carrán–Los Venados, Puyehue–Cordón Caulle fissural range, Osorno and Calbuco.

Antuco (37.4°S, 71.4°W, 2979 m). This is a compound stratovolcano, active since late Pleistocene times, which overlaps the larger eroded Sierra Velluda stratovolcano (3385 m). An early edifice collapsed laterally to the west after a 6.2 ka Bandai-type eruption, leaving a 2-km-wide horse-shoe-shaped crater and generating a >5 km$^3$ debris avalanche that reached the Central Valley. The current stratocone grew inside the avalanche crater. Antuco has had at least 17 historic eruptions, with the last big eruption occurring in 1853 and the last recorded event being in 1911. Currently it has weak fumarolic activity on the summit.

Callaqui stratovolcano (37.9°S, 71.4°W, 3100 m). This eruptive centre is strongly controlled by a NE–SW fissure with an 11-km-long row of craters and a main summit crater filled with ice. It is a dominantly basaltic–andesite to andesite volcano and has been active since the Late Pleistocene. At least five historic eruptions have been recorded for this volcano. The last was a small event that took place in 1980. Currently it has strong solfataric activity on the south flank near the summit. Based on morphostructural features and both K–Ar and $^{14}$C dates, six chronostratigraphic units, with ages ranging from 171 ka to Holocene, have been mapped (Naranjo et al. 1999b). Small-volume pyroclastic flow deposits have been dated between 9950 and 320 BP. One of the most important hazards associated with Callaqui volcano is the generation of large lahar flows that are widely distributed in the surrounding drainages.

Copahue volcano (37.8°S, 71.2°W, 2965 m). This is a Pleistocene to Holocene stratovolcano, built on the southwestern edge of the Pliocene El Agrio (or Caviahue) caldera (Fig. 5.12; Muñoz & Stern 1989; Pesce 1989; Linares et al. 1999; Folguera & Ramos 2000; Melnick & Folguera 2001; Melnick et al. 2006). Copahue volcano has an elongated shape (22 km $\times$ 8 km), with nine summit craters aligned at N60°E. The easternmost crater (2800 m) is currently active. Since its earlier stages at c. 1.2 Ma (Polanco 2003), Copahue has produced mainly andesite to basaltic andesite lavas, although subordinate rhyolite domes and pyroclastic deposits also occur. Holocene explosive eruptions have generated several pyroclastic deposits (c. 8–2 ka BP), including pyroclastic flow, surge and tephra fallout (Polanco 1998, 2003; Polanco et al. 2000). There are records of at least 12 eruptive events in the last 250 years, most of them of phreatic origin (Petit-Breuilh 1996; Martini et al. 1997). One of the best-documented eruptions took place between July and August 1992, when both phreatic and phreatomagmatic explosions occurred (Delpino & Bermúdez 1993). The last important eruption of Copahue volcano took place from July to October 2000 (Naranjo & Polanco 2004). The main volcanic hazards related to a future eruption of Copahue volcano are the generation of pyroclastic flows, tephra fallout, lavas and lahars (Delpino & Bermúdez 1994; Naranjo et al. 2000a). However, the presence of an acid lake within the active crater (Varekamp et al. 2001) represents one of the most hazardous aspects that could threaten Caviahue, the closest village to the volcano.
Lonquimay volcano (38.3°S; 71.5°W, 2865 m) (Fig. 5.14). This is a stratovolcano that has been active mainly in Holocene times (Moreno & Gardeweg 1989). It has a truncated cone due to an ENE fracture control and an estimated edifice volume of c. 20 km$^3$. Its eruptive activity has been divided into six lava units, including the most recent eruptive products generated in the 1988–89 eruption. Lavas are andesite to basaltic andesite in composition and vary from aa to blocky in morphology, with minor basalts and dacites. A succession of pyroclastic deposits, which include both pyroclastic flow and tephra fallout, have been recognized. Although they correspond to small-volume eruptions (Strombolian to Subplinian), some pyroclastic flows reached areas located more than 10 km from their source, with the last of these being erupted only c. 200 years ago (Naranjo et al. 2000a, b). There are no reliable records of historic eruptions occurring through the main vent. Instead, seven documented eruptive events have been attributed to flank eruptions, including the last long-lasting eruption of 1988–89. This eruption generated a flank pyroclastic cone (Navidad cone; Fig. 5.14) and an associated lava flow (Moreno & Gardeweg 1989; Gardeweg et al. 1990; Naranjo et al. 1992). The maximum eruptive column height was estimated at 9 km above the vent with an estimated VEI of 1–2, and the lava flow reached up to 10.2 km from the source. The high fluoride content produced widespread pollution of land, air and water which strongly affected both humans and grazing animals. According to the historical and post-glacial activity, the main hazards related to Lonquimay volcano are the formation of new flank vents, the emission of volcanic products with high fluoride content and the consequent contamination of water, grass and atmosphere, and tephra fallout (Naranjo et al. 2000a, b). Another important hazard, although less recurrent, is small-volume pyroclastic flows, which can reach populated areas.

Llaima volcano (38.6°S, 71.6°W, 3179 m) (Fig. 5.15). This is one of the most historically active volcanoes in the entire Andean chain, with more than 49 reported eruptions since 1640 (Petit-Breuilh & Lobato 1994; Naranjo & Moreno 2005). It is a large compound stratovolcano, one of the largest in the Chilean Andes (up to 400 km$^3$ in volume), whose products cover an area of more than 500 km$^2$. It consists of a composite stratocone that developed on the eastern side of a NE fissural range, both built on top of an older shield volcano formed by a lava succession erupted from the central and flank vents. The evolution of Llaima volcano has been divided into three main stages (Naranjo & Moreno 2005). During its early stage it generated a shield-like volcano mainly formed by basaltic andesite lavas, with minor basaltic andesites, subsequently glacially eroded during the last glaciation. This old edifice suffered a huge collapse at c. 13.7 ka during a large explosive eruption (c. 24 km$^3$), that generated the Curacautín Ignimbrite, a series of pyroclastic flows, basaltic andesite in composition, reaching up to more than 70 km from their source (Moreno & Naranjo 1989, 1991, 2003). This explosive stage continued until about 7 ka with a series of andesitic to dacitic pyroclastic flows, surges and Plinian tephra fallout deposits (Fig. 5.16). The second stage of evolution was built on top of the western and northern slopes of the old volcano, with many parasitic centres. Lavas and pyroclasts have basaltic andesite to andesitic composition. The present composite stratovolcano began to form after 3 ka and is largely composed of lavas and pyroclastic deposits of basaltic and basaltic andesitic composition. There are almost 40 parasitic centres located on its western and northeastern flanks, including pyroclastic cones and minor fissural centres, mostly aligned along a 25-km-long SW–NE arc structure.

The main hazard related to Llaima volcano corresponds to the generation of lahars due to ice/snow melting of its ice-cap during an eruption from either the central or a flank vent (Moreno & Naranjo 2003). Emission of lava flows (up to 30 km long) and tephra fallout (mainly distributed towards the east) are also two important hazards that could damage villages located on the lower flanks of the volcano. Large explosive eruptions have not occurred at Llaima since the generation of the Curacautín Ignimbrite at c. 13.7 ka, and another complex Plinian eruption at c. 9 ka (Naranjo & Moreno 1991, 2005).
Nevertheless, the remote possibility of the generation of pyroclastic flows has to be considered during severe Strombolian to Subplinian eruptions (Moreno & Naranjo 1989, 1991, 2003).

Villarrica (39.5°S, 71.9°W, 2847 m). This is a large (c. 250 km³) Middle Pleistocene–Holocene compound volcano (Fig. 5.17), located on the western edge of the Villarrica–Quetrupillán–Lanín volcanic chain, which is oblique to the main volcanic chain of the SVZ (Fig. 5.8). Villarrica has had more than 30 reported eruptions in historic times (Petit-Breuilh 1994; Petit-Breuilh & Lobato 1994; Simkin & Siebert 1994). The most important eruptive events in recent times occurred in 1948–49, 1963–64, 1971–72 and 1984.

The evolution of Villarrica volcano has been divided into three stages according to geochronological, stratigraphical and morphostructural criteria (Clavero & Moreno 2004; Moreno & Clavero 2006). In its early evolution (Villarrica 1 unit, middle to upper Pleistocene) an ancestral stratocone was built, mainly with basaltic to basaltic andesite lavas and fallout deposits. At c. 100 ka this ancestral edifice partially collapsed, generating an elliptical caldera structure, 6.5 km by 4.2 km in diameter. Subglacial basaltic andesite lavas and dacitic domes and dykes were generated during the Llanquihue Glaciation (95–14 ka), as well as a series of pyroclastic flows and both basic and acid tephra fallout deposits, which are mainly preserved on the lower western flank of the volcano (Clavero & Moreno 2004; Gaytán et al. 2005), although it is not clear if an edifice was formed during this period. A second caldera collapse, nested in the previous structure, occurred at c. 13.7 ka (caldera 2), with the generation of a series of pyroclastic flows, including the large-volume Licán Ignimbrite (c. 10 km³; Clavero 1996; Clavero & Moreno 1994, 2004), marking the beginning of an explosive phase, which seems to have continued to the present day. A new stratocone located on the northwestern rim of calderas 1 and 2 was formed through successive effusive and explosive eruptions (Villarrica 2 unit, 14–3.7 ka). At c. 3.7 ka this stratocone partially collapsed, forming a smaller summit caldera (caldera 3) associated with the Pucón Ignimbrite eruption (Clavero 1996; Clavero & Moreno 2004; Silva et al. 2004), which marked the end of Villarrica 2 unit. Soon after the collapse a new stratocone started to build within the summit caldera. This new cone has been built through successive effusive and explosive eruptions (Villarrica 3 unit, 3.7 ka to present). The last large explosive event at Villarrica volcano occurred c. 530 years ago (Moreno & Clavero 2006), with the formation of a pyroclastic flow directed towards the northeastern flank of the volcano. In total, the post-glacial eruptive history of Villarrica volcano (13.7 ka to present) has produced at least 15 small to large-volume pyroclastic flows and surges.

Historic eruptions that have occurred at Villarrica volcano have been mainly Hawaiian to Strombolian (Petit-Breuilh 1994; Petit-Breuilh & Lobato 1994; Moreno 1993; González-Ferrán 1995; Clavero & Moreno 2004). The main related hazards are generation of lahars, lava flows, tephra fallout and small-volume pyroclastic flows. Lahars, which are the only hazard that have produced casualties in Villarrica’s history, can be generated by ice/snow melting by lava or pyroclastic flows on top of the permanent ice-cap (Rivera et al. 2006), and could

Fig. 5.16. Late Pleistocene to Holocene pyroclastic deposits derived from Llaima volcano, including the large-volume mafic Curacautín ignimbrite at the base. Photo by J. A. Naranjo.

Fig. 5.17. (A) Villarrica volcano looking SW from Pucón. Note the gas plume dispersed towards the east. (B) The partially covered lava lake, in April 2005, below a small cone inside the Villarrica crater. The crater lava lake is a feature that apparently has persisted since the 1971–72 eruption (Calder et al. 2004; Witter et al. 2004). Photos by J Clavero.
be directed all around the volcano (Moreno 2000), reaching extensively populated areas, such as the towns Pucon and Villarrica on its northern foot. Tephra fallout, produced during Strombolian eruptions, could affect areas located mainly towards the east, according to the prevailing low to medium-altitude winds in the area, although it could be directed towards different zones when unusual winds blow. Lava flows, as lahars, can be directed almost all around the volcano and reach distances of up to 25 km. Finally, but not least important, Villarrica volcano has generated abundant pyroclastic flows ranging from small to large ignimbrites. The last of these highly explosive eruptions occurred only around 530 years ago (Moreno & Clavero 2006), indicating that such an explosive eruption could be generated at any time in the future, threatening the highly populated areas near the volcano, which increase their population up to three-fold in the summer. Small-volume pyroclastic flows, less than 5 km long, were generated during the 1948–49 eruption. Larger explosive eruptions may produce flows that reach more distal areas.

A lava lake in Villarrica’s crater (Fig. 5.17B; Calder et al. 2004; Witter & Delmelle 2004; Witter et al. 2004) also generates a hazard to the numerous tourists who climb to the crater almost every day. The occurrence of small explosions, and emissions of SO2 and HCl exceed recommended limits, threaten the safety and health of the visitors (Witter & Delmelle 2004).

**Quetrupillán (39.5ºS, 71.7ºW, 2360 m).** This is a Pleistocene to Holocene compound stratovolcano formed by two nested calderas, and a series of domes and pyroclastic cones (Pavez 1997; Naranjo 2004; Pavez et al. 2004). An extensive post-glacial (<15 ka in the area) pyroclastic sequence, which includes pyroclastic flow, surge and tephra fallout deposits, has been recognized (Naranjo 2004; Lara & Moreno 2004). Dacitic domes and coulées surround the caldera structures, while the pyroclastic cones are aligned NE–SW (Pavez 1997). At least one historic eruption has been reported, which occurred in 1872 (Simkin & Siebert 1994). Although no large populated areas surround the volcano, a large explosive eruption, such as those that occurred after the last glaciation, could generate pyroclastic flows that would reach inhabited areas.

**Mocho–Choshuenco volcanic complex (39.9ºS, 72.0ºW, 2430 m).** This is a Pleistocene–Holocene complex, formed by the remnants of an old edifice (Choshuenco peak; Fig. 5.18) and a small stratocone (Mocho) built within a caldera structure. The evolution of the complex probably started in the late Pleistocene with the emission of a series of basaltic andesite to andesitic lavas. In the early post-glacial period (c. 11 ka) an explosive phase started with the generation of large-volume Plinian fallout: the Pihuehueco (Naranjo et al. 2001a, b) and Neltume (Echegaray et al. 1994; Echegaray 2004; Naranjo et al. 2001a, b) Plinian tephra-fall deposits, of dacitic to andesitic composition. The Neltume eruption tephra-fall deposits covered an area of more than 4000 km2; reaching distances over 50 km to the north. One of these eruptions probably produced the partial collapse of the ancestral edifice forming a caldera. The explosive activity continued in the Holocene with the generation of a series of pumice- and scoria-rich pyroclastic-flow, dry and wet surges, and tephra fallout deposits both of andesitic and dacitic composition (Echegaray 2004; Pérez 2005). Simultaneously, a series of flank cones (Fuy series) were formed on the lower northeastern flank, and a small stratocone with lavas and tephra fallout layers started to infill the caldera depression. The last explosive eruption of Mocho–Choshuenco volcano occurred in 1864, with the formation of a highly destructive pyroclastic surge towards the western flank of the volcano. According to its post-glacial and recent eruptive history, Mocho–Choshuenco volcano must be considered one of the most hazardous volcanoes of the SVZ, and one of the most highly explosive (Naranjo et al. 2001a, b).

The major hazards related to Mocho–Choshuenco are the generation of highly explosive eruptions with the formation of pyroclastic flows and/or surges as have occurred several times in its very recent history. Tephra fallout could reach populated areas in Argentina (to the east) affecting both tourism and agriculture. Another hazard is related to the permanent ice-cap that fills the caldera depression (Rivera et al. 2005), as it is a source for lahar generation. Finally, it is likely that new flank cones will form with their associated lava flows, especially on its northeastern lower flank, as has been recurrent in the last 6000 years.

**Carrán–Los Venados (40.4ºS, 72.1ºW, 709 m).** This forms a cluster of mainly Holocene basaltic and basaltic andesitic pyroclastic cones and maars which form a NE-trending alignment (Moreno 1977; Rodríguez et al. 1997; Campos et al. 1998; Lara et al. 2006b). Eruptions observed during the twentieth century occurred in 1907, 1955 (when fine ash reached the city of Santiago, more than 900 km to the north) and 1979.

**Cordón Cautle–Puyehue (40.5ºS, 72.2ºW).** This middle Pleistocene–Holocene volcanic complex comprises the Cordillera Nevada caldera, Cordón Cautle fissure system and the Puyehue volcano (2240 m) in a NW-trending alignment (Moreno 1977; Lara et al. 2004b) that has emitted voluminous Holocene andesitic to dacitic pyroclastic ejecta. A remarkable fissure eruption of ryhodacites followed the high-magnitude (9.5) earthquake in 1960 (Lara et al. 2004b). This complex has an active geothermal field, which has been recently studied as a geothermal prospect (Sepúlveda et al. 2004).

**Osorno (41.1ºS, 72.5ºW, 2652 m).** His is a Pleistocene and Holocene stratovolcano (Moreno et al. 1979, 1985) with a field of Holocene basaltic pyroclastic cones on its flanks (Fig. 5.19). At least ten historical eruptions have been reported, the last in 1835 when a NE-trending system of fissures and cones formed (López-Escobar & Parada 1991; Petit-Breuilh 1999).

Osorno has the potential to produce lava flows that could possibly affect almost any of the areas all around the volcano. Lava flows might be erupted from either the central cone or the parasitic scoria cones on the volcano’s flanks and could flow up to 12 km from the vent. However, lahars are the most important volcanic hazard associated with Osorno volcano, based on their recurrence both in the historic and geological record. Scenarios for the generation of lahars vary according to the time of the year because of changes in the seasonal snow-covering thickness. Thus, in summer the Petrohué valley could be affected by lahars reaching about 80 × 106 m3 in size, but in winter the volume could increase to about 200 × 106 m3.
Although pyroclastic flows are not the most common volcanic product in Osorno’s recent eruptive history, their widespread distribution together with their highly destructive character make them an important hazard for the areas surrounding the volcano. Thus although the possibility of generation of a pyroclastic flow is very low, it nevertheless represents the most dangerous type of threat posed by Osorno volcano. According to the prevailing high-altitude wind directions (from NW to SE) tephra-fall hazards could affect areas located mainly to the ESE of the volcano. Although there are no important human settlements towards the east in Chile, a highly explosive eruption could produce tephra fallout deposits that could reach more populated areas such as San Carlos de Bariloche in Argentina.

Calbuco (41.3°S, 72.6°W, 2015 m). This is a Pleistocene–Holocene compound andesitic volcano (López-Escobar et al. 1992, 1995b; Hickey-Vargas et al. 1995). A sector collapse occurred during the early post-glacial period, generating a large avalanche which flowed to the north (Moreno et al. 1985), and an andesitic dome grew inside the collapse amphitheatre. Eleven historic eruptions have been recorded, the last in 1961 (Petit-Breuilh 1999).

Calbuco has produced blocky lava flows in both pre-historic and historic eruptions. Similar eruptions could affect areas located mainly toward the NE and SE of the volcano. The lava flows could be erupted from the central dome, but because of their high viscosity would not be expected to reach >9 km from the vent. Lava flows could potentially block the Caliente river valley forming a dam and an unstable and hazardous lake that could produce violent flooding downstream.

As Calbuco volcano is highly explosive, ballistic pyroclasts can reach long distances around the volcano, and pumice bombs up to 10 cm in diameter have been found in deposits 24 km east of the volcano. During the 1893–95 eruptive cycle, bombs up to 30 cm in diameter were thrown out to 8 km from the vent, killing cattle and causing forest fires. Dispersal of ash plumes would cause tephra fall affecting areas located mostly on the east and southeastern side of the volcano. Tephra fallout deposits could also reach more populated areas further to the east such as El Bolsón, San Carlos de Bariloche and Villa La Angostura in Argentina.

Historic eruptions of Calbuco, like that during 1893–95, have been explosive, but have produced only small-volume pyroclastic flows/surges and block-and-ash flows. However, the geological record indicates that the average recurrence of larger pyroclastic flows is less than 500 years, and the current evolution of the volcano suggests that the main central dome is still growing. For this reason, the possible future generation of a large pyroclastic flow (≥ 1 km³) that could cover an area greater than 1000 km² represents the most dangerous type of threat from Calbuco volcano.

Lahars have also been one of the most dangerous and destructive processes associated with eruptions of Calbuco. Lahars from Calbuco are generated in two different ways: by sudden melting of ice and seasonal snow or by mixing of block-and-ash flows with stream waters. The eruptive historic record for Calbuco includes the formation of ‘hot lahars’ toward the NE, SE and south of the volcano, that rapidly indurate, suggesting that they are generated by mixing of pyroclastic flows/surges and block-and-ash flows with river waters (Moreno & Naranjo 2004). Lahars generated by melting of ice and snow have also travelled down through the main valleys on the volcano slopes, and spread over wider areas when reaching lowlands, especially close to the shores of Llanquihue and Chapo lakes. As with Osorno, the scenario for an eventual lahar generation during an eruption varies significantly according to the time of the year because of variations in the seasonal snow-covering thickness. For example, in summer the Blanco and Hueñuhueñu river valleys could be affected by lahars reaching a volume of about 3 x 10⁶ m³, although in winter the volume could increase drastically to about 82 x 10⁶ m³.

The SSVZ

The SSVZ consists of 13 Quaternary volcanoes forming a narrow chain that crops out completely in Chile (Fig. 5.8; Stern et al. 1976b; Futa & Stern 1988; López-Escobar et al. 1993; D’Orazio et al. 2003; Naranjo & Stern 2004). Numerous minor eruptive centres also occur in the SSVZ along the LOFS and subparallel faults (Demant et al. 1994). The larger composite stratovolcanoes of the SSVZ include Yate, Hornopirén, Michinmahuida, Corcovado, Yanteles, Melimoyu, Mentolat, Cay and Macá (Fig. 5.20). Hudson volcano is a 10-km-diameter caldera. Huequi is a relatively small cinder cone, but may be part of larger eroded complexes, and Hualaihué–Cordón Cabrera is also a small pyroclastic cone aligned along a fissure system which includes eroded volcanic necks, domes and cones. Chaitén is formed by a rhyolite dome inside an explosion crater or small caldera (Fig. 5.21) located immediately to the west of the western flanks of the Michinmahuida complex, the largest volcano in the SSVZ.

Michinmahuida had historic eruptions in 1742 and 1834–35, and Hudson had an event that produced a large lahar in 1971.

Fig. 5.20. Maca volcano viewed from the west across the Moraleda channel. Photo by C.R. Stern.
(Best 1992), followed by a large explosive eruption in 1991 (> 3.6 km³; Naranjo et al. 1993b). All the other SSVZ volcanoes have had Holocene eruptions, except perhaps Cay volcano (Naranjo & Stern 2004). At least ten of the SSVZ centres have had medium to large explosive eruptions during the Holocene, namely Yate, Huequi, Chaitén, Michimahuida, Corcovado, Yanteles, Melimoyu, Maca, Mentolat and Hudson (Naranjo & Stern 1998, 2004). The towns of Chaitén, Futaleufú, Puerto Aisén, Coihaique, Puerto Ibáñez and Chile Chico in Chile, as well as Esquel in Argentina, occur within the > 10 cm tephrafall isopach of at least one of these eruptions. Puerto Ibáñez was significantly impacted by both ashfall and volcanic ash washed into General Carrera lake by the Ibáñez river after the 1991 eruption of Hudson volcano (Fig. 5.22; Naranjo et al. 1993b), and the port remains shut down as a result of this eruption. Hudson also had two very large pre-historic Holocene explosive eruptions, including one at c. 6700 years BP that may be among the largest Holocene explosive eruptions in the SVZ (> 18 km³; Naranjo & Stern 1998). This eruption deposited > 15-cm-thick tephra layers as far south as Tierra del Fuego (Fig. 5.23; Stern 1991d, 2007). Hudson is anomalous in the SSVZ with regard to the number of large-volume explosive eruptions it has produced, which may relate to its proximity to the Chile triple-junction.

Because of the high precipitation and cool climatic conditions in southern Chile, the volcanic centres of the SSVZ in general have a great deal of snow cover, which increases lahar potential, as illustrated by lahars generated by a thermal event that caused the melting of the glacial fill inside the caldera of Hudson volcano in 1971 and 1991 (Best 1992; Naranjo et al. 1993b). Although the SSVZ volcanoes are generally remote from local population centres, the town of Chaitén is located directly at the base of Chaitén volcano (Fig. 5.21), which has had at least one Holocene explosive eruption, and drainage systems from Michimahuida volcano also pass through Chaitén.

**AVZ volcanoes of southernmost Chile (C.R.S.)**

The AVZ volcanic segment in southernmost Chile occurs between latitudes 49°S and 55°S (Fig. 5.24). It includes only six volcanic centres: Lautaro (49°S), Viedma (49.4°S), Aguilera (50.2°S), Reclus (51°S), Monte Burney (52.3°S) and Cook Island (54.9°S). The AVZ was first recognized as a separate Andean volcanic segment in 1976 (Stern et al. 1976b), while the Cook Island volcanic complex, the southernmost in the Andean Quaternary arc, was not discovered until 1978 (Puig et al. 1984; Martinic 1988) and Reclus volcano was accurately located only in 1987 (Harambour 1988). Lautaro, Viedma and Aguilera volcanoes occur within the region of the southern Patagonian
ice-cap, where the border between Chile and Argentina is as yet unresolved, but the other three centres occur in Chile.

**Geological and tectonic setting**

The AVZ results from the subduction of the Antarctic Plate below the South American Plate. It is separated from the SVZ by the Patagonian gap in active volcanism just south of the Chile Ridge–Trench triple-junction (46–49°S; Figs 5.1 & 5.24), where very young oceanic lithosphere (<6 Ma) is being subducted. The age of the Antarctic Plate at the trench changes from 12 Ma below the northern part of the AVZ to 24 Ma in the south. Convergence velocities between the Antarctic and South American plates are relatively low (2–3 cm/year), and there is no Benioff zone of seismic activity associated with subduction of this oceanic plate, possibly because the slow rate of convergence and young age of the oceanic plate may cause the subducted slab to be several hundreds of degrees hotter than in more typical subduction environments (Peacock et al. 1994). Because of the lack of seismicity, the geometry of the subducted slab and mantle wedge below the AVZ cannot be determined. The convergence direction changes from nearly orthogonal in the north to oblique in the south. Strike-slip plate motion takes place at the latitude of Cook Island, which occurs on the Scotia microplate (Forsyth 1975), south of the Magellanes fault system (Fig. 5.24).

The southernmost Andes consist of Late Palaeozoic to Early Mesozoic metamorphic basement. Mesozoic and Cenozoic mafic, intermediate and silicic igneous rocks, as well as sedimentary rocks, were deformed and uplifted beginning in mid- to Late Cretaceous times (Bruhn & Dalziel 1977). Cook Island volcano directly overlies plutons of the Patagonian batholith, and Mt Burney, Reclus and Aguilera volcanoes occur along the eastern margin of the batholith. Viedma and Lautaro volcanoes occur within the area of the Patagonian ice-cap, which impedes knowledge of their bedrock geology. Base elevations of AVZ volcanoes are less than a few hundred metres above sea level, and summit elevations are less than 3600 m. This suggests that crustal thickness is <35 km. Lower crustal xenoliths found in the Patagonian plateau basalts are pyroxene + plagioclase granulites, without garnet, consistent with a relatively thin crust below southernmost South America (Selverstone & Stern 1983).

The Chile Ridge was subducted below the southernmost Andes during late Miocene times, resulting in the development of a slab window and extensive basaltic plateau volcanism in southern Patagonia prior to the reinitiation of subduction and AVZ arc magmatism (Ramos & Kay 1992; D’Orazio et al. 2000, 2001; Gorring & Kay 2001). Ridge subduction may have also caused subduction erosion along the continental margin (Bourgois et al. 1996, 2000; Polonia et al. 1999; Guivel et al. 2003, 2006), and the western edge of the Patagonian batholith lies much closer to the trench south of the Chile Ridge–Trench triple-junction than north of it (Cande & Leslie 1986). These authors also suggest that active subduction erosion is occurring in conjunction with the subduction of the Chile Ridge at 46°S, and presumably also occurred west of the AVZ as the locus of subduction of this spreading ridge migrated northwards during the Miocene.

**AVZ volcanoes and volcanic hazards**

The Cook Island volcanic complex consists of a group of post-glacial domes (Fig. 5.25; Puig et al. 1984; Heusser et al. 1989–90), some of which also occur on nearby Londonderry and Kelvin Islands (Dreher et al. 2005). All the other AVZ volcanoes are glaciated stratovolcanoes. Monte Burney (Fig. 5.26) has a small summit crater and a steep vertical northern wall, which may have formed from glacial erosion, a lateral blast or sector collapse. A broad plain to the NE, east and SE of Mt Burney is formed by a >5-m-thick section of Holocene pyroclastic flows. Small isolated outcrops of volcanic rocks...
engulfed by these flows (Fig. 5.26B) may represent highly eroded pre-Holocene Mt Burney lavas. However, these outcrops have not been dated. Other evidence for pre-Holocene volcanic activity in the AVZ is provided by 0.17 Ma K–Ar dates from Lautaro volcano (Orihashi et al. 2004). Lautaro is also the only centre in the AVZ with a documented historic eruption, which occurred in 1959 (Martinic 1988). However, tephra records indicate that each of the five AVZ stratovolcanoes has had Holocene explosive activity (Stern 1990b, 1992, 2000, 2007; Kilian et al. 2003; Motoki et al. 2003, 2006). Mt Burney, Recluse and Aguilera have each had large (> 2.5 km³) late glacial and/or Holocene explosive eruptions, as well as a number of smaller eruptions. Although all the AVZ volcanoes are remote from any population centre in southernmost Patagonia, such as Punta Arenas, Puerto Natales and Calafate, recurrence of an eruption as large as one of the larger Holocene events could produce significant tephra-fall in the area of these cities, as well as affecting the livestock ranching industry, as did the 1991 eruption of Hudson further north in Patagonia. Nevertheless, the presence of only six volcanic centres in a region extending over six degrees of latitude, and the relatively few documented explosive eruptions, suggests that magma production rates and the potential for explosive eruptions in the AVZ are less than in the SVZ, presumably as a result of the lower plate convergence and subduction rates below the AVZ (Stern 1990b, 2000, 2007).

Petrogenesis of the Quaternary Andean volcanoes of Chile (C.R.S., L.L.-E., H.M. & M.A.P.)

Andean magmatism is initiated by the dehydration and/or melting of the subducted oceanic lithosphere resulting in the addition of subducted components into, and melting of, the overlying mantle wedge (Thorpe 1984; Stern 2004). In this sense the initial subcrustal stages of generation of Andean magmas are similar to magma-generation processes in oceanic convergent plate-boundary island arcs. In fact, tholeiitic and high-Al basalts from the volcanic front of the Central SVZ of the Andes, where the crust is < 35 km thick, have isotope and trace-element ratios similar to oceanic island arc basalts, and they have not obviously assimilated continental crust (Hickey-Vargas et al. 1984, 1989; Hickey et al. 1986; Futa & Stern 1988).

However, differences in isotopic composition of volcanic rocks from the Northern SVZ and CVZ of the Andes, compared with those erupted in the Central SVZ and oceanic island arcs, do indicate the participation of continental crust and/or subcontinental mantle lithosphere in the formation and evolution of the NSVZ and CVZ Andean magmas. This may occur during interaction of magmas derived in the subarc asthenospheric mantle wedge with continental lithosphere (Rogers & Hawkesworth 1989; Stern & Kilian 1996; Hickey-Vargas et al. 2002), by intracrustal assimilation (AFC or MASH processes; James 1984; Hildreth & Moorbath 1988; Davidson et al. 1991), and/or by source region contamination of subarc mantle by subducted continental components (Stern et al. 1984b; Stern 1988, 1990a, 1991b; Stern & Skewes 1995; Kay et al. 1999, 2005). Relative roles of each process differ in the different Andean volcanic zones as a function of variations in the thickness and composition of the continental crust and rates of subduction erosion of the continental margin.

**SVZ Petrography, Geochemistry and Petrogenesis**

Along-arc petrochemical variations in the SVZ are discussed for each zone, beginning with the CSVZ and SSVZ, since volcanic rocks and magma-generation processes in these two zones are most similar to oceanic island arcs.

**The CSVZ and SSVZ**

In the Central and Southern SVZ, where the crust is relatively thin (< 35 km), tholeiitic and high-Al basalts and basaltic andesites are the dominant rock types erupted from both stratovolcanoes and many minor eruptive centres (López-Escobar...
Orthopyroxenes appear in intermediate to silicic rocks (their modal percentage ranges from only about 1–10% in MEC rhyolites. Plagioclase phenocrysts are also ubiquitous, although in the entire suite, including iron-rich olivine in some of the few these two SVZ segments is that olivine phenocrysts are present derived primary magmas. A remarkable feature of the rocks in and Cr abundances both lower than those expected in mantle-crystallization, as their MgO contents are <11%, and their Ni and Cr abundances both lower than those expected in mantle-derived primary magmas. A remarkable feature of the rocks in these two SVZ segments is that olivine phenocrysts are present in the entire suite, including iron-rich olivine in some of the few rhyolites. Plagioclase phenocrysts are also ubiquitous, although their modal percentage ranges from only about 1–10% in MEC volcanic rocks to 20–50% in composite stratovolcano products. Orthopyroxenes appear in intermediate to silicic rocks (>57% SiO2). An important feature of the CSVZ volcanic segment is that most silicic rocks lack hydrous minerals such as amphibole and mica. Indeed, a key difference that distinguishes the SSVZ from the CSVZ is the presence, in the most silicic SSVZ rocks, of hydrous minerals like hornblende (in rocks with >59% SiO2) and scarce biotite, the latter only in the Chaitén rhyolites.

Tectonism seems to control whether basaltic magmas reach the surface or evolve to more differentiated products at crustal levels. Most late Pleistocene and Holocene volcanic centres defining NE-trending alignments, including both composite stratovolcanoes ± MEC, contain mainly basaltic to basaltic andesite lithologies, such as seen in the Osorno–Puntiagudo–Cordón Cenizos volcanic chain. This is consistent with a short residence time of magmas in the crust (Tormey et al. 1991b), resulting in limited contamination and fractionation of mantle-derived magmas. On the other hand, volcanic edifices controlled by NW-trending fractures and faults, such as in the Villarrica–Quetrupillán–Lanín and Puyehue–Cordón Caulle volcanic chains, have more differentiated compositions, including rhyolites.

Sr, Nd, Pb and O isotopic data for CSVZ and SSVZ basalts preclude any significant assimilation of continental crust. Detailed studies of these basalts therefore provide information on their genesis and chemical composition of mafic mantle-derived Andean magmas that have not interacted extensively with continental crust, and thus a ‘baseline’ for evaluating the extent of crustal interaction in other segments of the Andean arc. These studies have concluded that CSVZ and SSVZ basalts form by melting of the subarc mantle contaminated by fluids derived from the dehydration of the subducted oceanic lithosphere, including sediments. This conclusion is based on: (1) Be isotope data which imply a subducted sediment component in CSVZ basalts (Morris et al. 1990; Sigmarsen et al. 1990; Hickey-Vargas et al. 2002); (2) excess 226Ra over 230Th and 238U over 236Th which both imply addition of slab-derived fluids to the mantle magma source (Sigmarsen et al. 2002); (3) Pb isotopic data that suggest Pb is derived from a mixture of mantle and subducted Nazca Plate sediment (Barreiro 1984; Macfarlane 1999); (4) high ratios, compared to oceanic island basalts (OIB), of large-ion lithophile elements (LILE) to rare earth elements (REE; for example Ba/La; Fig. 5.27A), REE to high-field-strength elements (HFSE; for example La/Nb; Fig. 5.27B), and very high ratios of LILE to HFSE (for example Ba/Nb; Fig. 5.27C), which reflect the fact that LILE are more soluble than REE, and REE more soluble than HFSE, and therefore relatively enriched in slab-derived fluids (Hickey-Vargas et al. 1984, 1989; Hickey et al. 1986); and (5) Sr (Fig. 5.27C). Nd and 18O isotopic compositions of these slabs are small relative to the mass of the mantle source of these basalts (Hickey-Vargas et al. 1984, 1989; Hickey et al. 1986; Stern et al. 1990).
Decreasing Ba/La, La/Nb and Ba/Nb ratios (Fig. 5.27) in magmas erupted progressively east of the volcanic front in the CSVZ, suggest decreasing input of slab-derived fluids into the mantle source of volcanoes behind the volcanic front as a result of progressive dehydration of the descending slab (Futa & Stern 1988; Stern et al. 1990). Decreasing Ba/La and La/Nb across the CSVZ arc are associated with increasing light REE (La; Fig. 5.27A) content and light REE to heavy REE ratios, both of which are interpreted as a measure of the degree of partial melting of the mantle. Thus, as the input of slab-derived fluids into the subarc mantle decreases, so does the percentage of mantle partial melting (Hickey et al. 1986; Hickey-Vargas et al. 1989; Muñoz & Stern 1989; López-Escobar et al. 1995a).

Further to the east in the backarc region, alkali basalts derived by relatively low degrees of partial mantle melting exhibit even lower Ba/La, La/Nb and Ba/Nb ratios, similar to OIB (Fig. 5.27), and therefore little or no evidence for input of slab-derived components (Skewes & Stern 1979; Stern et al. 1990; Kay et al. 1993; Gorring et al. 1997). Also, samples of the South American continental mantle lithosphere, which occur as peridotite xenoliths in backarc alkali basalts, present evidence of metasomatism by slab-derived fluids only when their host basalts were erupted in areas of Cenozoic arc volcanism, such as at Cerro del Fraile, Argentina (Figs 5.24 & 5.35; Kilian & Stern 2002). However, xenoliths in alkali basalts erupted further to the east, in regions with no evidence of Cenozoic arc volcanism such as Pali-Aike, Chile, and Lota 17, Argentina (Fig. 5.24), have been metasomatized by fluids without any slab-derived isotopic or trace-element signature (Stern et al. 1986, 1989, 1999; Gorring & Kay 2001).

Andesites, dacites and rhyolites in CSVZ and SSVZ volcanoes generally have the same isotopic composition as basalts and basaltic andesites, indicating that they formed either by crystal-liquid fractionation without assimilation (Gerlach et al. 1988), or assimilated young, isotopically similar crust such as Miocene plutonic rocks (McMillian et al. 1989).

The NSVZ
Rocks from the NSVZ segment range in composition from basaltic andesite to dacites, with a variable mineralogy that commonly includes plagioclase+clinoxyroxene+orthopyroxene+biotite+hornblende (in rocks with SiO₂ > 57%) + olivine (in rocks with SiO₂ > 60%) + olivine (Hildreth & Moorbath 1988; Sruoga et al. 2005). Basalts have not been yet reported, and they are either very rare or simply absent. Rhyolites are scarce, but a large rhyolite pyroclastic flow was erupted in Pleistocene times from the Diamante caldera (Stern et al. 1984a).

Geochemically, these have relatively high K, Rb, Sr, Ba, La, Th and U contents and higher Rb/Cs, La/Yb, K/La, Rb/La, Ba/La, Hf/Lu and ⁸⁷Sr/⁸⁶Sr isotope ratios (Fig. 5.28), but lower K/Rb and ¹⁴⁳Nd/¹⁴⁴Nd (Fig. 5.28) isotope ratios, than either basalts or rocks of similar SiO₂ content in the CSVZ, indicating incorporation of continental crust in the magmas erupted from these volcanoes (Stern et al. 1984b; López-Escobar et al. 1985; Futa & Stern 1988; Hildreth & Moorbath 1988; Stern 1988, 1989, 1991b). How this crust is incorporated in these rocks is still under debate. Hildreth & Moorbath (1988) outlined a model of mixing, assimilation, storage and homogenization (MASH) within intracrustal magma chambers. Stern et al. (1984b) and Stern (1988, 1991b, c) noted that Sr and Nd isotopic ratios were independent of SiO₂ in the range from basaltic andesites to dacites, and suggested that isotopic differences between the NSVZ and CSVZ were caused by a northward increase in mantle source region contamination by subducted continental components, due to either (1) a smaller volume of mantle wedge associated with the northward decrease in subduction angle and increase in crustal thickness below the NSVZ; and/or (2) increased subduction erosion caused by the southward migration of the locus of subduction of the Juan Fernández ridge. Stern (1991b) proposed that a small increase, of from 1% to 2% subducted components, would be sufficient to cause the isotopic differences observed in the NSVZ compared to the CSVZ (Fig. 5.29), while intracrustal assimilation of a mass of crust equal to the mass of mantle-derived basalt would be required to generate the same isotopic change. NSVZ volcanic rocks also have higher La/Yb and lower Yb than CSVZ volcanic rocks, suggesting garnet in their source, which could be in either the mantle (López-Escobar et al. 1977; Stern et al. 1984b; Stern 1991b) or deep crust (Hildreth & Moorbath 1988).

Stern & Skewes (1995, 2005), Nystroém et al. (2003) and Kay et al. (2005) demonstrated that the isotopic difference between the NSVZ and CSVZ developed during late Miocene and Pleiocene times, prior to the migration of the volcanic front to its current position in the Main Cordillera above relatively thick (>45 km) crust. At the latitude of the Maipo volcano (34°S), progressive temporal changes between the early Miocene and Pleiocene occurred in the isotopic composition of mantle-derived olivine basalts, implying changes in the isotopic composition of the mantle source region (Stern & Skewes 1995). Similar progressive changes in isotopic composition occurred between the early and late Miocene further north at latitude 32°S, and between the early Miocene and Pleiocene at latitude 33°S. The diachronous southward migration of these changes closely follows the southward migration of the locus of subduction of the Juan Fernández Ridge. This suggests that an important part of these changes resulted from increased contamination by crustal components of the mantle source region below the arc, caused by either increased subduction erosion or decreasing subduction angle and mantle volume in association with ridge subduction. In contrast, no isotopic changes occurred in conjunction with the 40 km eastward migration of the front in the Pleiocene, and therefore these changes are not correlated with crustal thickness (Stern & Skewes 1995).

The TSVZ
The volcanic rocks of this segment vary considerably from one centre to the other, and can even be very different in centres belonging to related group of volcanoes, as seen, for example, in the Planchón–Peteroa–Azufre group (Tormey et al. 1991a, 1995). Azufre is a bimodal stratovolcano with basaltic andesites of tholeiitic affinities and dacite, Planchón has erupted only tholeiitic basalts and basaltic andesites, while Peteroa lavas are mostly calcalkaline basaltic andesites with abundant mixed lavas of andesites and dacites composition. The Tatara–San Pedro–Pellado volcanic complex (Dungan et al. 2001) ranges in composition from basaltic andesite to rhyolite, but more than half of the volume of erupted material comprises porphyritic dacite flows. The oldest units of Nevado de Longavi (Sellés et al. 2004) are formed by basalts to andesites, while the Holocene ones range from basaltic andesites to dacites. Nevados de Chillán late Pleistocene to Holocene lavas are calcalkaline basaltic andesites to rhyolites (Déruelle & López-Escobar 1999). Finally, Laguna del Maule is also bimodal: the most abundant rock types are basaltic andesites and high-K rhyolites (Frey et al. 1984). Large volumes of rhyolites were also erupted in the Chilean TSVZ from the Pleistocene to Recent Calabozos caldera (Hildreth et al. 1984; Grunder 1987; Gruner & Mahood 1988) and Puelche volcanic field (Hildreth et al. 1999). As noted by Hildreth et al. (1999), these TSVZ centres are all located over the fold-and-thrust belt developed along the eastern edge of the Main Cordillera (Folgueira et al. 2002), although this region of the arc is also characterized by the occurrence of intra-arc extensional basins (Muñoz & Stern 1988). Isotopic data indicate that the generation of these rhyolites involved significant contributions from crustal partial melts.
The Sr and Nd isotopic ratios of the Nevados de Chillán Holocene lavas are among the most primitive within the SVZ arc of the Andes (Deruelle & López-Escobar 1999). Model calculations suggest that the two sources involved in the generation of the most mafic lavas are the asthenospheric mantle, which is the main source, and fluids coming from the subducted slab that enrich the mantle source in some incompatible elements such as K and Rb. The fact that rocks in the basaltic andesite to rhyolite range have similar values of Sr and Nd isotopic ratios, in conjunction with the behaviour of major and trace elements and incompatible element ratios, suggests that the Nevados de Chillán post-caldera magmas evolved through a closed-system fractional crystallization process, and that assimilation of crustal material was limited to <2 %.

At Azufre–Planchón–Peteroa (Tormey et al. 1991a, 1995), geochemical data suggest that two types of contamination are evident: a lower crustal component which caused increased La/Yb but had no effect on $^{87}$Sr/$^{86}$Sr, and an upper crustal component which increased $^{87}$Sr/$^{86}$Sr and $^{6}$H O. Peteroa andesite and dacite lavas formed by magma mixing, indicating that the volcanic system evolved over the past 500 ka from bimodal tholeiitic basalts and dacites (Azufre) to calcalkaline mixed andesite (Peteroa). Banded pumice and preservation of textural disequilibrium suggest that mixing occurred at shallow crustal levels. Thickness of the crust, magma supply rate and crustal temperature are the three main physical controls on compositional variation in this volcanic complex.

In the Tatara–San Pedro–Pellado volcanic complex (Feeley & Dungan 1996; Feeley et al. 1998; Dungan et al. 2001), three source components are recognized as contributing to parental magmas. These include the subarc asthenospheric mantle (mid-ocean ridge basalt (MORB) source), slab-derived fluids which would be the major source of LILE, and the continental lithosphere including the mantle and/or lower crust. Incompatible element variability in the most mafic rocks is interpreted as a result of source heterogeneity and deep-level processes. Many basaltic andesites to andesite lavas show evidence for magma mixing, and for the underlying magma chamber having been subject to repeated mafic magma replenishment and incorporation of silicic melts. The study of the Tatara–San Pedro complex indicates that open-system processes involving mixing newly arrived magma batches with evolved magmas in conduit–reservoir systems played a more significant role than differentiation in large, long-lived reservoirs.
Dungan et al. (2001) suggest that these processes may account for a large part of the trace element variability observed northward along the arc from the CSVZ to the NSVZ. However, Tormey et al. (1991a, 1995) suggest instead that spatial variations in trace element ratios of basalts northward from the CSVZ through the TSVZ are regional trends caused by a combination of both decreasing degree of mantle partial melting and an increase in the influx of slab-derived fluids as the volume of the subarc mantle decreases northwards. It is clear that the Tatara–San Pedro complex does not exhibit the isotopic variability observed between the CSVZ and NSVZ (Fig. 5.28; Davidson et al. 1987, 1988; Dungan & Davidson 2004), and thus the results from this one centre do not actually address directly the problem of the regional along-strike variations in the quantity of crustal components in SVZ magmas. The restricted ranges in isotope ratios of Sr, Nd and Pb of the Tatara–San Pedro–Pellado volcanic complex is interpreted as due to the small isotopic contrast between magma and wall rock.

**Petrography and petrogenesis of CVZ volcanoes**

Andesites, dacites and rhyodacites are the dominant rock types erupted in the CVZ, although basaltic andesites, rhyolites and occasional basalts also occur. Plagioclase is usually the most abundant phenocryst, along with olivine, orthopyroxene and clinopyroxene phenocrysts in the basaltic andesites, and amphibole in andesites. The dacitic–rhyolitic domes are highly porphyritic (up to 50 vol%), presenting phenocrysts of amphibole, biotite and two feldspars, with little or no quartz or pyroxene. Disequilibrium textures, such as sieve textures in plagioclase, and magma-mixing textures are common.

CVZ magmas have elevated $^{87}$Sr/$^{86}$Sr and $^{6}$O (Fig. 5.30; James 1982, 1984), and lower $^{144}$Nd/$^{143}$Nd (Fig. 5.31; Wörner et al. 1988; Davidson et al. 1990, 1991) than SVZ magmas, as well as distinctly different Pb isotopic compositions (Fig. 5.32; Barreiro 1984; Barreiro & Clark, 1984; Wörner et al. 1992b; Aitcheson et al. 1995). These data imply that a greater amount of continental crust has been incorporated in CVZ than in SVZ magmas (Davidson et al. 1991; Francis & Hawkesworth 1994; Lindsay et al. 2001b; Schmitt et al. 2001, 2002), as might be expected in this region of both extremely thick crust and high rates of subduction erosion.

Correlated variations of Sr, Nd and O isotopes with increasing SiO$_2$ content in some volcanic suites (Francis et al. 1974; Feeley & Davidson 1994), and regional correlations of Pb isotopic compositions and Andean basement age (Wörner et al. 1992a, b, 1994; Aitcheson et al. 1995), indicate that an important part of the process of crustal contamination of CVZ magmas occurred by intracrustal assimilation combined with crystallization of these magmas, and/or crustal anatexis (Hawkesworth et al. 1982; Schneider 1987; Davidson & de Silva 1995). Geophysical studies interpret crustal structure as having been affected by partial melting (Schmitt et al. 1997). Temporal variations indicate that the extent of crustal contamination of volcanic rocks in northern Chile has increased through time, from prior to the Miocene to Recent, as the crust has thickened to $>70$ km (Rogers & Hawkesworth 1989; Miller & Harris 1989; Trumbull et al. 1999; Lucassen et al. 2001a).

However, not all CVZ volcanoes have correlated variations of Sr, Nd and O isotopes with increasing SiO$_2$ content. For example, the Nevados de Payachata volcanic group (18ºS) exhibits wide variation in elemental composition, but only minor variations in Sr, Nd and Pb isotopic composition. Davidson et al. (1990, 1991) explain these chemical and isotopic features through a complex petrogenetic evolution involving primary mantle-derived magmas which are modified by deep crustal interactions to produce magmas with ‘baseline’ isotopic compositions (Figs 5.30 & 5.31) that are parental to those erupted at the surface. They suggest that these ‘baseline’
parental magmas subsequently evolve at shallower levels through assimilation–crystallization processes involving upper crust and magma mixing, which in both cases were restricted to end-members of low isotopic contrast. However, lack of correlated variations in isotopic compositions and increasing SiO2 content argues against such combined assimilation–crystallization processes, and is consistent instead with the generation of CVZ parental magmas with enriched ‘baseline’ isotopic compositions resulting from modification of the subcrustal mantle source by subducted crust and sediment (Figs 5.29 & 5.30; Stern 1988, 1990a, 1991b, c).

Unfortunately, because of the lack of basalts in this segment of the Andean arc, which may in itself be an indication of extensive subarc source-region contamination below the CVZ, the degree to which the baseline composition of magmas generated in the subarc mantle below the CVZ has been affected by source-region contamination is difficult to evaluate. Nevertheless, a number of different lines of evidence indicate that subduction erosion and source-region contamination play an important role in determining the baseline isotopic and trace-element composition of mantle-derived mafic magmas below the CVZ of northern Chile. Most significant is the fact that some primitive mantle-derived mafic CVZ magmas with enriched isotopic signatures also have high Mg numbers and Cr and Ni contents, implying that they erupted through the thick Andean crust without significant intracrustal assimilation (Kay

Fig. 5.30. Sr versus O isotopes of Andean magmas, modified after (A) Davidson (1991) and (B) Stern (1991c). Each figure illustrates the different trends expected from mantle source contamination, which affects Sr but not O isotopic composition significantly, and intracrustal assimilation, which affects both isotopic systems. As interpreted by Davidson (1991), isotopic values of SVZ versus CVZ volcanic rocks suggested greater intracrustal contamination involving older crust in the CVZ. As interpreted by Stern (1991c), the data suggest changes in the ‘baseline’ isotopic composition of CVZ relative to SVZ magmas due to greater source-region contamination, followed by intracrustal assimilation in each area.

Fig. 5.31. Sr versus Nd isotopic compositions for magmas erupted from volcanoes in different Andean arc segments (Davidson et al. 1991). The figure illustrates the isotopic differences between SVZ and CVZ volcanoes, compared to MORB and oceanic island arc volcanoes, and indicates the enriched Sr and Nd isotopic composition of ‘baseline’ parental magmas for volcanic rocks in the CVZ. This ‘baseline’ composition is produced either by lower crustal MASH processes (Davidson et al. 1991) or subcrustal mantle source region contamination (Stern 1991b).

Fig. 5.32. Pb isotopes for volcanic rocks from different segments of the Andean arc (Davidson et al. 1991). The different Pb isotopic compositions of north and south CVZ volcanic rocks depend in part on the isotopic composition of the basement they have erupted through.
mantle-derived clinopyroxene from these volcanoes (Stern 1990; Kilian 1996). The AVZ andesites and dacites include pyroxenes, plagioclase. Biotite occurs only in the andesites and dacites of the three northernmost AVZ volcanoes (Stern 1990b). Cook Island and Mt. Burney andesites contain mantle-derived clinopyroxene + olivine xenocrysts (Stern & Kilian 1996). The presence of both a greater proportion of crustal xenoliths and larger, more complexly zoned phenocrysts is observed in AVZ volcanic rocks from south to north, suggesting an increasing importance of near-surface magma chamber processes such as slow cooling and crystallization, magma mixing, wall-rock assimilation, and volatile degassing. The AVZ andesites and dacites have adakitic chemical features (Stern et al. 1984; Futa & Stern 1988; Stern & Kilian 1996), with SiO₂ > 56 wt% and Al₂O₃ > 15 wt%, relatively high Mg#, low HREE (Yb < 1.9 ppm) and Y < 18 ppm, high Sr > 400 ppm and Sr/Y > 40 (Fig. 5.33), and positive Sr and Eu anomalies. These characteristics have been interpreted to reflect the presence of residual garnet, amphibole and pyroxene, but little or no olivine and plagioclase in the source region. AVZ andesites and dacites also have low concentrations of HFSE, including TiO₂, which is a common feature of both adakites and typical convergent plate boundary magmas. The AVZ occurs in an area of relatively thin crust (< 35 km), and the source of AVZ adakites is thus likely to be subducted oceanic basalt, recrystallized to garnet-amphibolite or eclogite (Stern & Kilian 1996; Kilian & López-Escobar 2000). Geothermal models indicate that partial melting of the subducted oceanic crust is probable below the AVZ due to the slow subduction rate and the young age of the subducted oceanic lithosphere. However, other chemical characteristics, such as LILE (K₂O, Rb, Ba and Cs) and LREE content, LREE/HREE and LILE/LREE ratios, and Sr, Nd (Fig. 5.34), Pb, Hf and O isotopic compositions, which may reflect source composition rather than source mineralogy, vary significantly between, but not within, individual centres (Stern et al. 1984; Futa & Stern 1988; Stern & Kilian 1996; Harttori et al. 2005). Although some Cook Island adakites contain xenoliths of granite from the Patagonian batholith (Dreher et al. 2005), others are xenolith-free and have MORB-like isotopic compositions (Stern & Kilian 1996; Sigmarsson et al. 1998; Bindeman et al. 2005; Harttori et al. 2005). The other AVZ centres to the north have isotopic compositions suggesting a greater proportion of subducted sediment or assimilated continental crust involved in their genesis.

To explain the origin and evolution of these AVZ adakitic andesites and dacites, Stern & Kilian (1996) proposed a multistage four-component model involving subducted MORB metamorphosed to garnet-amphibolite or eclogite, subducted...
sediment, mantle wedge and continental crust. According to this model, the proportions of the different source materials involved in the genesis of AVZ magmas vary significantly from south to north. The adakitic andesites from Cook Island volcano, where subduction is most oblique (Fig. 5.24), can be modelled by small degrees (2–4%) of partial melting of eclogitic MORB with little participation of subducted sediments, yielding a primary magma that interacts to only a very limited extent with the overlying mantle and continental crust. In contrast, models for the magmatic evolution of other AVZ andesites and dacites located between 49°S and 52°S, where subduction is more orthogonal (Fig. 5.24), require melting of a mixture of MORB and subducted sediment, followed by interaction of this primary melt with the overlying mantle and the continental crust. Based on geochemical data for Chile Trench sediments (Kilian & Behrmann 2003), these models suggest that the input of subducted sediments is variable, between 0 and 20% for the AVZ volcanoes, and that intracrustal AFC processes and the mass contribution from the continental crust become more significant northwards in the AVZ as the angle of convergence becomes more orthogonal.

**Backarc alkali plateau basalts (M.A.S. & C.R.S.)**

Pleistocene and Holocene backarc volcanism occurs to the east of the Andean arc between latitudes 34°S and 52°S (Skewes & Stern 1979; Baker et al. 1981; Stern et al. 1990). Pleistocene and Holocene backarc volcanic activity, which generally consists of alkali basalt lava flows erupted from small monogenetic spatter and scoria cones, forms the youngest part of the Patagonian plateau basalts. The Patagonian plateau basalts are located dominantly in Argentina, but the western parts of this volcanic province, including Pleistocene and Holocene cones and flows, also occur in southernmost Chile (Fig. 5.24), specifically along the western edge of Meseta Vizcachas exposed in Cordillera Baguales (50.5°S, 72.5°W; Muñoz 1981; Kilian & Stern 2002), and in the Pali-Aike volcanic field (51–52°S, 69–71°W; Skewes & Stern 1979; D’Orazio et al. 2000), which straddles the Chile-Argentina border close to the Atlantic coast just north of the Magellan Straits.

**Geological and tectonic setting**

Pleistocene and Holocene backarc volcanic activity in southern Chile forms the youngest part of the Patagonian plateau lavas located south of the locus of subduction of the Chile Rise at 46°S (Fig. 5.24). The eastward extension of the Chile Ridge, which separates the Nazca and Antarctic plates, was subducted below southernmost Patagonia beginning in the Miocene (Ramos & Kay 1992; Gorring et al. 1997; D’Orazio et al. 2000, 2001; Gorring & Kay 2001; Kilian & Stern 2002; Espinoza et al. 2005; Guivel et al. 2006). Neogene backarc volcanic activity in southernmost Patagonia has been attributed to the subduction of the Chile Ridge, which resulted in the opening of an asthenosphere slab-window between the relatively rapidly down-going Nazca Plate and the more slowly subducting Antarctic Plate (Fig. 5.35).

The Pali-Aike volcanic field occurs in the area of the Magellan basin, which began subsiding in the Jurassic in conjunction with the opening of the southern Atlantic Ocean. The continental crust underlying the Pali-Aike basalts consists of Jurassic silicic volcanic rocks of the Tobifera Formation (Bruhn et al. 1978), and Jurassic to Miocene marine and continental sedimentary rocks deposited on pre-Andean
metamorphic basement. Meseta Vizcachas basalts occur along the western margin of this basin, above Cenozoic continental sediments uplifted along the eastern margin of the Austral Basin fold-and-thrust belt.

**Pali-Aike volcanic field**

The Pali-Aike volcanic field is >50 km wide and >150 km long. It consists of basaltic lava flows extruded from hundreds of monogenetic spatter and scoria cones (Figs. 5.36 & 5.37), and tuff rings associate with numerous maars (Fig. 5.38). Cones and maars are orientated along NE–SW and NW–SE lineations. The field takes its name from the Pali-Aike spatter cone, an archaeological site within which a cave formed along the inner wall of the cone was found to contain evidence of human occupation dating back >10,000 ¹⁴C years BP (Bird 1938, 1988).

The Pali-Aike field formed between 3.8 Ma and the Holocene, but has not had historic activity. Three different units are recognized in the field (Skewes 1978; Skewes & Stern 1979; D’Orazio et al. 2000). The oldest unit forms an extensive area of plateau-like basaltic lavas approximately 100 m thick. The two younger units, which are spatially more restricted, are both post-glacial (<1 million years). The partially eroded and aeolian sediment-covered Pali-Aike cone (Fig. 5.36) and Laguna Ana maar (Fig. 5.38) both belong to the second unit in the Pali-Aike field, estimated to have formed between 130–17 ka based on K–Ar ages (Meglioli 1992). The youngest lava flows are associated with the Cerro Diablo scoria cone (Fig. 5.37). Based on their fresh glassy appearance and lack of any aeolian soil cover, these flows are certainly <10,000 years old.

Pali-Aike basalts are nepheline normative alkali olivine basalts, with notably high TiO₂ content and intraplate geochemical characteristics similar to some oceanic islands and lacking any evidence for the incorporation of subducted components into their source (Skewes & Stern 1979; Stern et al. 1986, 1989, 1999). Their chemical compositions are consistent with derivation by low degrees of melting of mantle asthenosphere, either due to subduction-induced thermal and mechanical perturbation (Stern et al. 1990), or rising through a slab-window (Fig. 5.35; D’Orazio et al. 2000; Kilian & Stern 2002).

Pali-Aike basalts contain ultramafic xenoliths of mantle origin, including a unique suite of garnet-bearing peridotite xenoliths, which generally are not found in alkali basalts (Skewes & Stern 1979; Stern et al. 1986, 1989, 1999). Mineral chemistry geothermometry and geobarometry for these xenoliths indicate that the continental lithosphere below Pali-Aike is relatively thin (<100 km; Figs. 5.35 & 5.39). Major and trace-element, and isotope chemistry of Pali-Aike garnet peridotites indicates that the deeper portions of this lithosphere consist of fertile garnet lherzolites with trace-element and isotopic compositions similar to the mantle source of mid-oceanic ridge basalts (Fig. 5.35). The data imply the absence of a deep, low-density, olivine-rich mantle root below the Palaeozoic–Mesozoic crust of southernmost South America such as has been inferred below Archaen cratons (Stern et al. 1999). Pali-Aike garnet peridotites have been both modally and cryptically metasomatized by fluids genetically related to the alkali host basalts. These xenoliths, most of which occur in tuff rings associated with maars, have been transported to the surface at velocities estimated to be between 1 and 6 m/s (Selverstone & Stern 1983; Demouchy et al. 2006), similar to velocities estimated for the emplacement of kimberlite magmas.
Quaternary volcanism in Cordillera Baguales

The deeply eroded, c. 1000-m-thick western edge of Meseta Vizcachas (Fig. 5.40) forms Cordillera Baguales along the border between Chile and Argentina at 50.5°S latitude (Fig. 5.25; Muñoz 1981). Lavas from the western margin of Meseta Vizcachas at Cerro del Fraile in Argentina are interbedded with fluvial–glacial sediments and have been dated to 1–2 Ma (Fleck et al. 1972). In Argentina, Meseta Vizcachas is covered by numerous small spatter and scoria cones, presumably younger than the eroded lavas dated at Cerro del Fraile. Sills and dykes are exposed within the deeply eroded lava cliffs of Cordillera Baguales.

Chile Ridge and Chile triple-junction volcanism (L.E.L. & C.R.S.)

Several submarine volcanic centres are located along the NW–SE trending Chile Ridge, which is formed by relatively short active ocean-floor spreading segments bounded by NE–SW fracture zones. The Chile Ridge is being subducted beneath the South American Plate west of the Taitao Peninsula (46ºS; Figs 5.1, 5.8 & 5.41). Dredged rocks from the Chile Ridge are mainly MORB-type basalts, but they show a wide spectrum of geochemical signatures, in particular trace-element ratios (Klein & Karsten 1995; Karsten et al. 1996). Far from the trench, MORB-like signatures predominate, but near the margin arc-like trace-element signatures are observed. These arc-like geochemical characteristics in Chile Ridge basalts are considered to reflect contamination of a depleted N-MORB source mantle with slab-derived components due to slab break-up or shearing in conjunction with subduction of this young buoyant lithosphere, and subsequent entrainment of these slab components into the subridge mantle (Klein & Karsten 1995; Karsten et al. 1996; Guivel et al. 2003, 2006).
Moreover, in the area referred to as the synsubduction segment of the Chile margin (Bourgois et al. 2000), Pleistocene–Holocene (70 ka to 2 Ma) calcalkaline basalts and dacites, with subduction-related geochemical characteristics, have been dredged from both the Chile Trench and the Taitao ridge (TR, Fig. 5.41; Guivel et al. 2003). Chemically similar Late Miocene, Pliocene and Pleistocene igneous rocks also occur in the Cabo Raper plutonic suite (4.8–5.1 Ma; Guivel et al. 1999) and the volcanic–sedimentary sequence of the Chile Margin Unit (1.5–5.3 Ma; Bourgois et al. 1993, 1996) on the forearc of Peninsula Taitao. The age of these volcanic rocks gets younger towards the present-day triple-junction.

**Petrology and petrogenesis**

Among the dacites, both low- and high-silica varieties occur. They are both interpreted to be derived from relatively shallow hydrous melting of subducting oceanic basalts and associated sediments under high thermal conditions (Fig. 5.42). Low-silica dacites show chemical characteristics of adakitic slab melts, with depleted HREE abundances suggesting the occurrence of residual garnet in their source at depths of 25–45 km. High-silica dacites are less depleted in HREE and are interpreted to represent melts of MORB and sediment at even shallower depths of <25 km. Basaltic andesites have predominantly mantle-derived geochemical characteristics and are thought to result by mixing of dacite magmas with MORB-type liquids derived from a buried spreading ridge (Guivel et al. 2003). The high temperatures required to produce melting at shallow depths result from what has been referred to as a ‘blowtorch’ effect (Lagabrielle et al. 2000, 2004), caused by the back (south) and forth (north) migration of the triple junction due to the successive subduction of the short NW–SE Chile Ridge segments separated by NE–SW fracture zones (Fig. 5.41).

Various models have been suggested to explain the spatial proximity of these young volcanic rocks to the current trench axis. These include: (1) rise of their parent magmas from their source to the surface along the subduction thrust plane; (2) rapid tectonic erosion caused by ridge subduction (Fig. 5.42); or (3) changes in the geometry of the subducted slab, causing a steepening of subduction angle in response to mantle counterflow moving from the back arc towards the trench in the mantle wedge (Guivel et al. 2003).

**Intraplate oceanic volcanism: the Pacific volcanic islands of Chile (L.E.L. & C.R.S.)**

Three separate clusters of islands in the southeastern Pacific are formed by intraplate oceanic volcanoes occurring along linear chains of seamounts on the Nazca Plate. These three clusters include (1) Easter, Salas and Gómez islands along the western edge of Salas and Gómez seamount chain, (2) San Félix and San Ambrosio islands along the eastern edge of this chain, and (3) Robinson Crusoe, Santa Clara and Alejandro Selkirk islands along the Juan Fernández seamount chain. These islands are the subaerial upper parts of large, nested and partially eroded stratovolcanoes.

Most of these islands had early Pliocene (c. 4–3 Ma) stages of submarine growth, Surtseyan pyroclastic deposits and dyke swarms that fed later subaerial volcanoes, some of them active in the Holocene. They also show advanced erosion, mainly related to sector-collapse processes. On a larger, tectonic-plate scale, they could represent the long-lived manifestations of mantle hot-spot plumes, the temporal evolution and petrophysical behaviour of which are still not well understood in the area of the southeastern Pacific. What is clear is that subduction of the aseismic ridges of which these islands form a part has important effects on the South American margin. For example, subduction of the Juan Fernández Ridge, from the Eocene to the present, may have produced changes in subduction angle (Yañez et al. 2001, 2002), affected coastal tectonics and the present pattern of seismicity of both outer-rise and forearc domains in central Chile (von Huene et al. 1997), and resulted in the genesis of giant Andean copper deposits in central Chile (Stern & Skewes 2005).

**Easter and Salas and Gómez islands**

Easter Island (27º09’S, 109º23’W, 560 m) occurs 350 km east of the East Rift spreading ridge on the eastern margin of the Easter microplate, and near the western edge of a long chain of seamounts that also includes both Salas and Gómez Island (26º28’S, 105º22’W, 30 m) as well as more than 553 other cone-shaped or flat submersed seamounts recognized from high-resolution bathymetry (Rappaport et al. 1997).

Easter Island volcanic complex is formed by three coalescent volcanic centres (Rano-Kao, Poike and Terevaka-Tangaroa) and several superimposed pyroclastic cones and lava flows.
The nearly coeval Rano-Kao and Poike stratocones were built beginning in the Late Pliocene (c. 2.6 Ma). They are eroded, exposing thick volcanic successions with interbedded pyroclastic ejecta produced by phreatomagmatic eruptions. Late extrusion of rhyolitic domes occurred on the flanks of both Rano-Kao and Poike. K–Ar ages for Maunga Orito and Te Manavai domes associated with the Rano-Kao volcano are, respectively, 230 ka and 180 ka. Jet-black rhyolite obsidian occurs associated with a rhyolitic dome in Rano-Kao caldera.

The youngest volcanic centre is the Terevaka-Tangaroa volcano, a shield complex active since approximately 1.5 Ma. A thick pile of basaltic lavas was emitted from several fissure vents on this volcano. Possibly Holocene lavas erupted from several flank vents are the youngest evidence of activity at Easter Island. Historical reports and local myths suggest eruptions as recently as 2000 years ago.

San Félix and San Ambrosio islands
San Félix (26°17'S, 80°07'W, 193 m) and San Ambrosio (26°20'S, 79°58', 479 m) islands, together with other small emerged lands such as the Peterborough Cathedral, are parts of huge shield volcanoes and perihedral centres located in the easternmost segment of the Salas and Gómez submarine volcanic chain. San Félix Island corresponds to a tuff-ring related to phreatomagmatic eruptions, which truncated an older alkaline shield volcano. Lava flows and pyroclastic deposits with fresh morphology cover the island. San Ambrosio island is also a remnant of a shield volcano where two unconformable sequences are cut by vertical dykes dated to 2.93 ± 0.15 Ma (Bonatti et al. 1977).

Alejandro Selkirk and Robinson Crusoe islands
Robinson Crusoe (33°3'S, 78°53'W, 915 m) and Alejandro Selkirk (33°45'S, 80°45'W, 1650 m) islands occur along the Juan Fernández Ridge (Fig. 5.1), a c. 900-km-long, east–west-trending volcanic chain of mostly submarine volcanoes. Alejandro Selkirk and Robinson Crusoe (Fig. 5.43) islands are remnants of ancient shield volcanoes. Alejandro Selkirk consists of a homoclinal lava sequence that yields K–Ar ages of 0.85 ± 0.30 and 1.3 ± 0.3 Ma (Booker et al. 1967). Robinson Crusoe island, in turn, has a heavily eroded profile where two unconformable sequences can be recognized. They are cut by coalescent vertical dykes and some laccoliths around which has developed intense hydrothermal alteration. Marine sedimentary deposits partially cover the island and record sea level rising during the Pleistocene (Valenzuela 1978). Booker et al. (1967) dated the lower lava sequence of Robinson Crusoe to between 3.5 ± 0.8 and 3.1 ± 0.9 Ma. Baker et al. (1987a) obtained 4.0 ± 0.2 Ma for the same unit. Possible historical renewal of volcanic activity was suggested by Sutcliffe (1839) during Charles Darwin’s expedition to the South Pacific area.

The Juan Fernández Ridge is supposed to be related to a mantle hot-spot, the present location of which would be west of Alejandro Selkirk island, below the vicinity of the Domingo and Friday seamounts (Farley et al. 1993; Devey et al. 2000), but these have not been dated. The present-day nearest-trench seamount (Monte O’Higgins) was dated as c. 8.5 Ma by 40Ar/39Ar (von Huene et al. 1997). Although the subduction of the eastern extension of this ridge played an important long-term role in the Cenozoic evolution of the South American margin (Yañez et al. 2001, 2002; Stern & Skewes 2005), the precise track of the supposed plume-fed hot-spot volcanism remains unknown.

Petrogenesis of Chilean oceanic islands
Geochemical signatures of the volcanic rocks from Easter Island are mainly tholeiitic with scarce silica-poor alkaline varieties (Clark & Dymond 1977). Hékinian et al. (1996) have described a mixing of N-MORB and E-MORB characteristics for Easter Island, suggesting a mixing between Pacific mid-oceanic ridge and hot-spot sources, as might be expected from the proximity of Easter Island to the Pacific Ridge. Other young seamounts and submarine volcanic fields in the area, such as Ahu and Tupa, which may be related to the Easter Island hot-spot, exhibit similar chemical characteristics (Haase & Devey 1996).

Baker et al. (1987a) studied the transitions from tholeiitic to alkaline affinities in volcanic rocks from Robinson Crusoe and Alejandro Selkirk islands. They concluded that the chemistry of the early tholeiitic stage of formation of the islands resulted from a high degree of partial mantle melting, consistent with a steep geothermal gradient likely to be associated with a hot-spot. Later-stage alkaline rocks were derived by remelting of a source region previously subjected to small but variable degrees of partial melting and melt migration over a prolonged period. More recently, Devey et al. (2000), as a result of a study of the younger Domingo and Friday seamounts along the Juan Fernández Ridge, suggested melting of a MORB-like suboceanic harzburgite source modified by metasomatic reactions with CO2-rich plume-melts, similar to the metasomatism related to the generation of kimberlitic magmas. They proposed that these plume-melts were derived from a source that included ancient recycled oceanic crust.

Gerlach et al. (1986) described the mantle heterogeneities beneath the Nazca Plate from Sr–Nd systematics, recognizing both EM1 and EMII components as parental composition of some hot-spot volcanoes that follow distinct geochemical trends. In contrast, Devey et al. (2000) suggested that mixing of PREMA and HIMU mantle sources produced the isotopic variability observed among the various islands and seamounts along the Juan Fernández Ridge.

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This chapter describes the metallic ore deposits of Chile, their mineralized host rocks and the processes involved in ore formation, and provides a brief overview of the mining history of this Andean copper-rich country. The ore deposits are ordered according to their respective economic importance. Thus, after mining history and a general introduction, Chilean porphyry copper–molybdenum deposits are described first, with subsequent sections dealing with epithermal precious metals, iron oxide copper–gold and iron oxide–apatite deposits, stratabound copper–silver ores, precious metal veins, sedimentary-hosted gold and porphyry gold deposits, skarn rock ores and, finally, an overview of metallogenetic evolution.

Mining history

About 40% of the known copper resources of the world occur in Chile, with the native populations using the red metal at least since 500 BC. Bracelets, earrings and weapons that have been found in archaeological sites in northern Chile were made of either native copper or copper-rich minerals that were melted in small quantities and subsequently hammered. Copper production during Spanish colonial times (1541–1810) amounted to some 80 000–85 000 tons, with high-grade oxidized copper minerals being exploited and melted with charcoal. Despite this mining activity, however, Spaniards regarded copper as ‘plebeian metal’ because of its relatively low value, and it was used mostly as ballast for ships returning to Spain, rather than for technological or industrial purposes. The colonial Spaniards were much more interested in gold and silver, and mining activities were consequently mostly orientated towards precious metals.

Prior to Spanish conquest the Incas dominated northern Chile and had already exploited some placer gold deposits. During the second half of the sixteenth century placer deposits were the only source of gold, which was extracted by the Spaniards with intensive use of native labour, and annual gold production was in the range of 1 to 2 metric tonnes (t), declining by the end of the sixteenth century as the shallow, highest grade placer deposits were exhausted (Cuadra & Dunkerley 1991). During the seventeenth century overall Chilean metal production was meagre, with the entire gold production amounting to only 350 kg (Millán 2001), suppressed by harsh battles between the colonizers and the natives of southern Chile, destructive earthquakes, and frequent pirate assaults. In contrast, through-out the eighteenth century gold production increased progressively and the exploitation of precious metal from hard rock deposits began. In 1749 official gold coin production started in Chile, mining laws were published in 1754, and a mining court was established to resolve innumerable title disputes. However, mining development overall remained hampered by the economic monopoly imposed by the Spanish Crown.

Chile declared independence in 1810, but conflict with the Spanish colonial powers continued until 1826. The war of independence seriously affected Chilean mining activities, and gold production fell dramatically (by 1829, for example, no gold coins were stamped), although several bonanza-type silver-bearing vein deposits were discovered and developed (e.g. Cachinal de la Sierra in 1822, and Arqueros in 1825). Mining became revitalized once the political turmoil ended, and so in the decade from 1832 a total of 418 hard rock mines and alluvial placer exploitations were active in central Chile (Cuadra & Dunkerley 1991). This mining revival was mainly focused on copper, although precious metal exploitation continued to be important. Further bonanza-type silver deposits were mined (e.g. Chañarcillo, 1832; Tres Puntas, 1848; Caracoles, 1871), mostly hosted by Lower Cretaceous carbonate sedimentary rocks. Total nineteenth century silver production was 7544 t, which placed Chile as a major world silver producer. In contrast, gold production declined during this century, with no new gold mines having been discovered (Cuadra & Dunkerley 1991).

Thus post-colonial metallic ore mining in Chile became dominated by copper, with 100 000 t of the metal being produced from 1841 to 1850, placing Chile as the second largest copper producer in the world after Great Britain (Camus 2003). The increase in Chilean copper production during the nineteenth century coincided with a rise in demand caused by the Industrial Revolution, mainly for weapons and shipyard uses. In 1851 Chile rose to become the number one copper producing country, a position it retained until 1880. Total Chilean copper production during the nineteenth century amounted to 1 876 000 t, all of which came from the high-grade (>6–8% Cu) oxidized and supergene enriched zones in vein deposits of the Atacama and Coquimbo regions. More specifically, it was the Carrizal Alto, Dulcinea, La Higuera, Andacollo, Brillador and Tamaya mining districts of northern Chile that were the main contributors to copper production at that time. Although the occurrence of large porphyry copper deposits such as Chuquicamata and El Teniente had been known since the second half of the nineteenth century, these were originally uneconomic due to the lack of proper technology for their exploitation, except for their richest parts (e.g. the enargite and chalcopyrite veins at Chuquicamata).

The use of copper for electrical applications by the end of the nineteenth century further increased the demand for the metal. However, the shallow and high-grade supergene enriched parts of vein deposits were largely exhausted, as a result of which Chilean copper production declined from 1876 to 1891. Production increased again by the end of the nineteenth century, really taking off at the beginning of the twentieth century when the introduction of novel ore drilling, blasting, loading and transport technologies allowed bulk mining of low-grade and high-tonnage porphyry copper deposits. From 1906 to 1927 large-scale copper production started at El Teniente, Chuquicamata (Fig. 6.1a, b), and Potrerillos mines, which were developed by North American interests. Since then the huge, world-class Chilean porphyry copper deposits have continuously been the main source of Chilean copper and molybdenum production, which is still on the rise. Chile regained its number one position in copper production in the late twentieth century (1982), and it remains current world leader (2005).
Finally, to return to precious metal mining, from the beginning of the twentieth century to 1979 gold production was, for the most part, a by-product of the exploitation of porphyry copper deposits, copper-bearing veins and disseminated deposits. Annual gold production ranged from about 2 to 10 t. Chilean gold production was subsequently boosted by the exploitation of El Indio epithermal deposit in 1980 and further discoveries and development of 20 other precious-metal-bearing deposits. Gold production peaked during the years 1996 and 2000 with 53.1 and 54.1 t per year, respectively. However, since then gold production has been dropping, and currently only four gold mines remain operating on a regular basis (La Coipa, Alhué, El Peñón and Pullalli), with 39.9 t of gold being produced in 2004.

**Metallic ore deposits: an introduction**

Some of the world’s largest and richest porphyry Cu–Mo deposits occur in the Andes of northern Chile, and these have allowed Chile to become not only the leading copper-producing country of the world (5 418 800 t of copper were produced in 2004, 37% of annual global copper production), but also the second producer of molybdenum (41 883 tonnes in 2004), which is a by-product of the copper exploitation. A total of 14 porphyry deposits are currently exploited, and these contribute most of the Chilean copper production. Despite the fact that this overall huge richness in mineral resources is mainly due to the large size of these copper–molybdenum deposits, there are other significant metallic ore occurrences. These latter deposits include precious metal epithermal, iron oxide copper–gold, volcanic-hosted stratabound copper–silver, Fe oxide–apatite, mesothermal copper–gold-bearing veins, and minor skarns. Gold-rich porphyry deposits also occur, but none is currently exploited.

The Chilean Andes provide a typical example of a mountain belt developed along a convergent plate margin. This is a non-collisional orogen formed over a long-lived, currently active subduction system. The geological evolution of this orogen was characterized by the development of successive north–south trending magmatic arcs on the active continental margin. The magmatic foci of the arc migrated inland stepwise after stages of compressive deformation and associated crustal thickening. The abundance of calcalkaline igneous rocks is a distinctive characteristic of the Andes in Chile, and metallic ore deposits are mostly hydrothermal in origin, with an inherent and temporal relationship to the arc-related magmatic activity: they are believed to derive most of their metallic content from underlying subduction processes.

Most of the Chilean metallic mineralization occurs in the arid and semi-arid northern section of the country (18–34°S), an area which probably contains the richest ore deposits in the entire Andean mountain belt. In contrast, only sparse and less significant metallic deposits occur south of latitude 34°S, in the humid Andes of southern Chile. This difference arises in part from the climatic changes along the Andes, which control erosion rates and the efficiency of supergene enrichment.
Porphyry copper–molybdenum deposits

Porphyry copper deposits are the most abundant type of mineralization in the Chilean Andes, and a detailed review has been published (in Spanish) by Camus (2003). Porphyry Cu–Mo deposits occur in six longitudinal belts along the Andes of northern Chile, each representing a discrete metallogenic period. The porphyry belts are: Late Palaeozoic–Triassic (298–230 Ma), Early Cretaceous (132–97 Ma), Palaeocene–Early Eocene (60–50 Ma), late Eocene–early Oligocene (43–31 Ma), and late Miocene–early Pliocene (12–4.3 Ma) (Camus 2002, 2003). The two youngest porphyry belts are unquestionably the Late Eocene–early Oligocene porphyry Cu–Mo belt (Fig. 6.2). This arc-parallel belt includes 30 porphyry Cu–Mo deposits and prospects with the highest amount of copper resources. Overall it constitutes the largest copper concentration in the world, totalling about 220 million tonnes (Mt) of copper (resources plus production; Camus 2002). The most distinctive geological characteristic of this porphyry belt is its spatial relationship with the Domeyko Fault System (Fig. 6.2) which is an arc-parallel regional fault zone that follows an uplifted crustal block (Domeyko Cordillera) cored by Upper Palaeozoic basement rocks (Maksaev & Zentilli 1988, 1999; Boric et al 1990; Reutter et al 1991, 1996). The copper-bearing stocks occur along-strike of master faults in this regional system (e.g. Copaquire, Quebrada Blanca, Ujina, Radomiro Tomic, Chuquicamata, La Escondida and Cerro Zaldívar), as well as along NW-trending subsidiary faults (e.g. Río Blanco of Collahuasi (Fig. 6.1d), El Abra, Potrerillos and La Fortuna). An apparent exception to this close link between faults and mineralization is provided by El Salvador, one of the oldest deposits of this group, which is located some 9 km west of the

Late Eocene–early Oligocene porphyry Cu–Mo belt

This is the most significant Chilean porphyry belt. It extends for more than 1400 km along the Domeyko Cordillera and can be traced from the border with Peru (18°S) to latitude 31°S (Fig. 6.2). This arc-parallel belt includes 30 porphyry Cu–Mo deposits and prospects with the highest amount of copper resources. Overall it constitutes the largest copper concentration in the world, totalling about 220 million tonnes (Mt) of copper (resources plus production; Camus 2002). The most distinctive geological characteristic of this porphyry belt is its spatial relationship with the Domeyko Fault System (Fig. 6.2) which is an arc-parallel regional fault zone that follows an uplifted crustal block (Domeyko Cordillera) cored by Upper Palaeozoic basement rocks (Maksaev & Zentilli 1988, 1999; Boric et al 1990; Reutter et al 1991, 1996). The copper-bearing stocks occur along-strike of master faults in this regional system (e.g. Copaquire, Quebrada Blanca, Ujina, Radomiro Tomic, Chuquicamata, La Escondida and Cerro Zaldívar), as well as along NW-trending subsidiary faults (e.g. Río Blanco of Collahuasi (Fig. 6.1d), El Abra, Potrerillos and La Fortuna). An apparent exception to this close link between faults and mineralization is provided by El Salvador, one of the oldest deposits of this group, which is located some 9 km west of the
major north–south trending Cerro Castillo fault. However, El Salvador and another six mineralized centres are aligned for 4 km along a NNE trend which suggests there is also a structural control, albeit more arcane than is normally the case (Gustafson et al. 2001). A number of authors have suggested a direct genetic relationship between the Domeyko Fault System and copper mineralization (e.g. Sillitoe 1981; Hunt et al. 1983; Baker & Guilbert 1987; Maksaev & Zentilli 1999; Richards 2000, Richards & Villeneuve 2002), and this association has successfully been used for exploration (e.g. Lowell 1991; Ortiz 1995).

The late Eocene–early Oligocene porphyry Cu–Mo deposits are not regularly distributed along this north–south trending belt, but rather form local clusters within areas of <200 km². The clusters typically approximate alignments rather than equidimensional groupings of deposits, with the individual centres being strung out either parallel (e.g. Chuquicamata district) or transversely (e.g. Copaquire, Quebrada Blanca, Rosario de Collahuasi, and Ujina group of deposits at 21ºS; Sillitoe 2004) (Fig. 6.2). The copper-bearing stocks cluster in areas where late Eocene–early Oligocene pre-mineralization equigranular dioritic to granodioritic plutons occur (‘plutonic precursors’ of Sillitoe 1988). Some of these plutons have radiometric ages that are very close to, or overlap with, the age of mineralized porphyry copper deposits. These plutons locally host stocks of mineralized stocks, which, in some cases, appear to be associated with porphyritic facies of otherwise more equigranular plutons. The Southern Granodiorite at El Abra (Ambrus 1977) and the Fortuna Granodioritic Complex at Chuquicamata are the most conspicuous examples of these plutonic precursors (Sillitoe 1988; Dilles et al. 1997).

The concurrence of the locus of unmineralized and mineralized late Eocene–early Oligocene plutons suggests that several batches of magma followed a similar ascent path through the crust, and were emplaced at distinct centres of igneous activity. This is reminiscent of the separation of modern volcanic centres along the Southern Volcanic Zone of the Andes. The separation of the igneous centres may have resulted from long-term rheological contrast of the crust along the magmatic arc, with porphyry Cu–Mo clusters/signeous centres localized at zones of diapiric uprise of magmas as envisaged by Yanëz & Maksaev (1994) for the Chilean porphyry belt. However, Richards (2000) and Richards & Villeneuve (2002) have suggested instead a structural magma focusing. In addition, Behn & Camus (1997) and Behn et al. (2001) showed that the porphyry clusters match with regional magnetic anomalies. These lie transverse to the Cordillera, and have been attributed to hypothetical batholiths at depth, which are thought to be common parental intrusive complexes for the mineralized stocks. Minor post-mineralization dykes and stocks commonly occur, which according to available geochronological data were emplaced shortly after the mineralization (e.g. latite dykes at El Salvador (Gustafson & Hunt 1975), and the rhyolite dome and rhyolite dyke at La Escandida (Padilla et al. 2001)).

The late Eocene–early Oligocene porphyry Cu–Mo deposits are mostly associated with granodioritic to quartz monzonite porphyritic stocks that intruded unrelated country rocks of diverse nature. Comagmatic volcanic piles, if they ever existed, are not preserved. The plutonic complexes are typically characterized by successive, small (0.5–2 km²), roughly cylindrical porphyritic intrusions, and include many pre-, syn- and post-mineralization porphyry dykes or epizonal stocks (e.g. El Salvador deposit; Gustafson & Hunt 1975). Copper and subordinately molybdenum sulphides typically occur in disseminated grains and stockwork veins throughout a large volume of intrusive rock, indicating significant brittle fracturing and volume increase of the igneous rock mass during mineralization (e.g. Burnham 1985). Breccia bodies with a tourmaline matrix occur in Quebrada Blanca and El Salvador, and breccias with a biotite matrix occur in El Abra (Hunt et al. 1983; Gustafson & Hunt 1975; Ambrus 1977). However, hydrothermal breccias are only a minor part of these large porphyry systems, whose dominant copper-bearing stockwork ores are largely hosted by intrusive rocks. At Chuquicamata local mechanical brecciation related to faulting also occurs. Pervasive hydrothermal alteration zones of potassic, phyllic, argillic and propylitic types typically form annular shells around intrusions similar to the model of Lowell & Guilbert (1970), although significant distortions due to local structural conditions or multiple intrusive pulses occur (e.g. El Abra (Ambrus 1977) and Chuquicamata (Alvarez et al. 1980)). A late advanced argillic alteration cap occurs at El Salvador, La Escondida and Rosario of Collahuasi (Gustafson & Hunt 1975, Gustafson et al. 2001, Padilla et al. 2001; Bisso et al. 1998). The hypogene mineralization of chalcopyrite, bornite, pyrite, molybdenite and locally enargite is primarily associated with the potassic cores and phyllic alteration zones of the deposits. In most cases these zones were exhumed and exposed at the surface. The ore bodies are irregular, but conform to roughly domal cappings that are restricted to the exposed apical section of the host intrusive complexes and their immediate host rocks. The hypogene shape of the ore bodies has been in almost all cases modified by supergene processes, so that the richest copper ores commonly underlie partially leached, oxidized rocks forming irregular blankets containing the supergene copper sulphaes (chalcoite, dıutrite, covellite and anilite), which have replaced hypogene sulphide minerals.

The Chuquicamata deposit is the largest porphyry copper deposit of this belt, and has produced a total of 31 Mt of Cu metal from 1915 to 1997, with an equivalent amount of contained Cu still remaining in the main orebody. Chuquicamata had an intimate and complex relationship with an active regional fault during its mineralization, which included the superposition of at least two distinct periods of copper introduction related to an early stage of potassic alteration, followed by a later structurally controlled phyllic overprint (Ossandón et al. 2001; Lindsay et al. 1995). Despite its uniqueness this enormous copper deposit may be catalogued as a classic porphyry copper deposit following the classification of McMillan & Panteleyev (1980) and the same applies to most other porphyry Cu–Mo deposits of this belt. An exception is the Fortuna cluster, where Au-rich porphyry deposits occur (La Fortuna, El Negro and El Morro prospects; Perelló et al. 1996). These are relatively small porphyry systems that occur along a local NNE structural trend in the High Andes at 28º38’S, between major NNE-trending reverse faults that post-date porphyry mineralization. These porphyries have a potassic alteration core (dominated by an assemblage of biotite and magnetite) that grades to a narrow outer zone of propylitic alteration. Phyllic alteration is poorly developed, being only significant in the uppermost sections of El Morro and La Fortuna prospects. An epithermal advanced argillic alteration assemblage is overprinted on previous potassic and phyllic alteration zones in the uppermost section of La Fortuna prospect, extending to the SE into the Cantarito epithermal prospect. The alteration zoning is largely vertical rather than lateral and these systems seem to be vertically elongated. Stockworks of polydirectional quartz and sulphide veiners locally occur, but most of the mineralization of these porphyry Cu–Au–Mo deposits is structurally controlled, and dominated by sheeted-vein zones formed of subparallel sets of near-vertical quartz and sulphide veiners. Sheeted-vein orientations match the NW and NNE dominant fault trends of the district.

Abundant K–Ar, 40Ar/39Ar, Rb–Sr and U–Pb radiometric ages have been reported for the porphyry Cu–Mo deposits of northern Chile (Camus 2002, 2003). The isotopic ages demonstrate that the copper mineralization took place within a discrete interval of geological time from 41 to 31 Ma. In fact, according to the available radiometric dates, most major deposits were formed within the period from 38 to 36 Ma (Copaqueir,
Rosario de Collahuasi, Ujina, Quebrada Blanca, El Abra, Toki, Apache, Genoveva, Chimborazo, Cerro Zaldívar, La Escondida and Potrerillos), El Salvador and Gaby were formed at 42–41 Ma, and Fortuna and El Negro Au-rich porphyry deposits at about 35 Ma. However, the largest and richest deposits are those that are the youngest in this belt, i.e. Chuquicamata, Radomiro Tomic and MM, formed at 34–31 Ma (Camus 2003).

**Late Miocene–early Pliocene porphyry Cu–Mo belt**

The late Miocene–early Pliocene porphyry belt in the Andes of central Chile (32–34°S) is the second in economic importance in the country (Fig. 6.3). This belt includes world-class porphyry Cu–Mo deposits such as El Teniente, Los Bronces-Rio Blanco and Los Pelambres. Overall, these deposits total 183 Mt of contained copper (resources plus production; Camus 2002). According to recent geochronological data El Teniente and Rio Blanco–Los Bronces were formed at the same time, from 6.46 to 4.37 Ma (Deckart et al. 2003; Maksaev et al. 2004), whereas mineralization at Los Pelambres developed earlier, from about 13 to 10 Ma (Bertens et al. 2003). These Cu–Mo deposits are typically related to multiphase porphyritic stocks and associated magmatic-hydrothermal breccia complexes.

The composition of the porphyritic stocks varies from quartz diorite to quartz monzonite. The porphyritic bodies were emplaced within country rocks of diverse nature, including Lower Cretaceous volcanic and sedimentary rocks at Los Pelambres (Rivano and Sepulveda 1991), a Middle Miocene granodioritic batholith at Rio Blanco–Los Bronces (San Francisco Batholith) and minor Miocene volcanic rocks (Serrano et al. 1996), and Miocene mafic rocks at El Teniente (Cuadra 1986; Skewes et al. 2002).

Pervasive alteration zones of potassic, phyllic, propylitic and minor argillic types in general form annular shells around the porphyry intrusions similar to the model of Lowell & Guilbert (1970), but they are overprinted by successive intrusive phases with associated high-temperature hydrothermal events, and particularly by the development of late breccia bodies and associated phyllic alteration. Copper and molybdenum sulphides occur within a stockwork of veinlets throughout a large volume of rocks related to the potassic and phyllic alteration zones, but the occurrence of large, mineralized, biotite-bearing and tourmaline-rich breccia pipes is characteristic of these deposits (Skewes & Stern 1995; Skewes et al. 2002). About 50% of the copper ore occurs within the matrix of hydrothermal breccias at Rio Blanco–Los Bronces (Serrano et al. 1996). Ore-grade material is largely restricted to the intrusive bodies at Los Pelambres and Los Bronces, but at Rio Blanco it also occurs within volcanic country rocks, and at El Teniente more than 70% of the ore is contained within biotitized andesitic–basaltic–gabbroic Miocene country rocks. Diatreme breccia pipes occur at El Teniente and Rio Blanco; the conspicuous funnel-shaped Braden breccia pipe occurs roughly in the centre of El Teniente orebody and is 1200 m in diameter at the surface, tapering down to 600 m at 1800 m below the surface. This diatreme is poorly mineralized (c. 0.3% Cu), but a rim of tourmaline-rich breccia (Marginal breccia) with high copper grades surrounds it. At Rio Blanco a post-mineralization early Pliocene volcanic dacitic diatreme occurs and its tuffaceous ejecta cover part of the deposit (La Copa volcanic complex; Serrano et al. 1996).

Large-scale tectonic and local structural control is of utmost importance in focusing magma injection and hydrothermal fluid flow for porphyry systems. The late Miocene–early Pliocene porphyry Cu–Mo deposits of the High Andes are north–south aligned overall, but they appear to be located at NE- and NW-trending fault intersections. Recent explorations by Codelco at El Teniente and Rio Blanco districts have identified first-order NE-trending lineaments that appear to have controlled precursor tonalitic intrusions, the location of porphyry deposits, and other prospects at these districts. NE-trending faults are also the dominant structures that controlled hypogene sulphide mineralization at Los Pelambres orebody. At all three of these porphyry systems intersecting NW-trending faults also occur, which are regarded as second order structures that controlled the dacite hypogene mineralization at El Teniente, the hydrothermal breccia development at Rio Blanco–Los Bronces, and subsidiary hypogene mineralization at Los Pelambres. The occurrence of NW-trending faults appears to be related to inherited basement structures that occur as major NW-trending lineaments in the Argentinean Andes. These major lineaments coincide with the location of
modern volcanic centres of the Southern Volcanic Zone of the Andes as well as with the porphyry deposits.

**Palaeocene–early Eocene porphyry Cu–Mo belt**

A belt of Palaeocene–early Eocene porphyry Cu–Mo deposits extends for over 1300 km from 16°S latitude in southern Peru to 29°30′S in northern Chile. This belt comprises the porphyry Cu–Mo deposits of Mocha, Cerro Colorado, Spence, Sierra Gorda, Centinela, Polo Sur, Lomas Bayas, Fortuna del Cobre, Relincho and Las Pascualas in northern Chile (Fig. 6.4). The same belt includes the most significant Cu–Mo porphyry deposits of Peru (Cerro Verde–Santa Rosa, Cuajone, Quellaveco and Toquepala) and shows K–Ar ages ranging between 58 and 51 Ma (Sillitoe 1980). The Peruvian porphyries represent about 39 Mt of contained copper as a whole (resources plus production; Camus 2003). The Chilean porphyries of this belt total 12.7 Mt of contained copper (resources plus production; Camus 2003). Owing to the overall poorer mineralization of the Chilean Palaeocene–early Eocene porphyries, only those with well-developed supergene enrichment are currently economic. This is the case of Cerro Colorado that is currently exploited, and Spence that is under development.

Cerro Colorado is located 95 km ENE of Iquique, and its proven and probable reserves were 228 Mt at 1.0% Cu (0.5% cut-off; Bouzari & Clark 2002). The copper mineralization at Cerro Colorado is related to tonalitic and monzonitic porphyries (40Ar/39Ar 52.8 ± 0.5 Ma (Bouzari & Clark 2002); molybdenite Re–Os 55.5 ± 0.3 Ma (Cotton 2003)) that were emplaced within a sequence of Cretaceous andesites, and developed apical breccias. A wide zone of propylitic alteration surrounds a core of potassic alteration, and both are overprinted by a pervasive quartz-sericitic alteration that masks the original texture and composition of the host rocks. The mineralized rocks are andesites (79%), breccias and porphyries (21%) within an area 2000 m long (east–west) by 1000 m wide (north–south). The upper 70 to 120 m of the deposit is leached and barren, followed at depth by a 60–120-m-thick oxidized ore blanket (chrysocolla, atacamite), and deeper by a supergene chalcocite-enriched blanket 50 to 70 m thick, reaching locally up to 200 m (Bouzari & Clark 2002).

The Spence porphyry copper deposit was discovered in June 1996 by Rio Algom mining company and currently a mine is being developed. The global geological resource is 330 Mt at 1.18% copper (cut-off grade of 0.6%). This enriched Cu–Mo porphyry deposit occurs under an alluvial plain of the Atacama Desert and is completely covered by gravel with an average thickness of 70 m. The porphyry copper is related to three Palaeocene dacite porphyry stocks and tourmaline-bearing hydrothermal breccias aligned NE–SW. Laser 40Ar/39Ar plateau ages for magmatic biotite in pre- and late-mineralization stocks are 57.0 ± 0.7 and 56.6 ± 0.6 Ma, whereas hypogene alunite from 46 to 113 m below the base of the enrichment blanket yielded 40Ar/39Ar ages of 33.2 ± 1.9, 33.4 ± 1.1 and 35.6 ± 1.4 Ma (Rowland & Clark 2001). Under the gravel cover and leached rocks a 20-m-thick copper oxide zone occurs (atacamite), which is followed by a well-developed supergene enrichment blanket (chalcocite) with an average thickness of 60 m. Supergene alunite and natroalunite from leached and chalcocite enrichment zones range from 44.4 ± 0.5 to 27.7 ± 5.4 Ma (13 inverse isochron 40Ar/39Ar dates; Rowland & Clark 2001).

The belt of Palaeocene–early Eocene porphyry copper deposits extends farther south at least to 29°S latitude as attested by the occurrence of the Relincho and Las Pascualas Prospects; Relincho was explored by Outokumpu reporting resources of 130 Mt at 0.8% Cu. Tourmaline breccias are commonly associated with the Palaeocene–early Eocene porphyry copper deposits (e.g. Copucha breccia pipe at Sierra Gorda; Boric et al. 1990). In addition, Palaeocene granodioritic plutons in the Copiapó and Inca de Oro areas of northern Chile (about 27°S) host a great number of vertical tourmaline-bearing breccia pipes; these are mostly smaller than 100 m in diameter. Some breccia pipes are mineralized with Cu–Mo and a number also contain Au and/or W. The main examples are the San Pedro de Cachiayuyo, Cachiayuyo de Llampos, Los Azules and Cabeza de Vaca districts (Sillitoe & Sawkins 1971; Colley et al. 1991; Sillitoe 1985). Cu–Mo mineralization is restricted to the matrix of the breccia bodies and no disseminated or stockwork mineralization has been found to date within the host intrusives at these districts.

**Early Cretaceous porphyry copper belt**

Early Cretaceous porphyry copper deposits occur along a belt that extends along the Coastal Cordillera of northern Chile and comprise Galenosa, Puntillas, Antucoya–Buey Muerto, Dos Amigos at Domeyko, Pajonales, Andacollo and Colliguay (Fig. 6.5). Most of them are low-grade and only the chalcocite...
supergene enrichment blanket of Andacollo and Dos Amigos are currently mined. These porphyries were early recognized as forming a distinct belt that was referred as the ‘Pacific Belt of porphyry copper and hydrothermal systems’ along the Coastal Cordillera (Llaumet et al. 1975) and K–Ar dates ranging from 132 to 97 Ma indicated an Early Cretaceous age (Munizaga et al. 1975; Reyes 1991). Andacollo is economically the most important porphyry deposit of this belt; its resources total 300 Mt at 0.70% Cu, 0.015% Mo and 0.23 g/t Au (Llaumet et al. 1975; Reyes 1991).

Fig. 6.5. Early Cretaceous porphyry Cu–Mo belt of northern Chile.

Late Palaeozoic–Triassic porphyry Cu–Mo belt

Pre-Mesozoic mineralization is poor in the Chilean Andes, but a belt of porphyry Cu–Mo prospects is distributed along the northern extension of the Late Carboniferous to Early Triassic magmatic arc which extends for over 2500 km from northern Chile to southern Argentina (Mpodozis & Ramos 1989; Camus 2003; Fig. 6.6). This Late Palaeozoic–Triassic belt of porphyry Cu–Mo deposits was first reported by Sillitoe (1977) in the Argentinean Andes, but recent exploration and K–Ar dates ranging from 298 to 195 Ma show that this belt extends to the Andes of northern Chile (Camus 2003). The Late Palaeozoic–Triassic copper occurrences are interspersed with Cenozoic porphyry Cu–Mo deposits of northern Chile (Camus 2003). However, most of the Palaeozoic systems are poorly mineralized and despite systematic exploration none has been proven to be economically viable. The poor mineralization of these porphyry systems might be a result of significant unroofing of Upper Palaeozoic to Lower Triassic igneous rocks, leaving only the roots of the porphyry systems. However, a detailed study of the San Jorge deposit (2 Mt Cu) revealed that high levels of this porphyry Cu–Mo deposit are preserved in the Argentinean Andes (Williams et al. 1999) and also shallow epithermal alteration is preserved at the Lila prospect in northern Chile. This may indicate that erosion of pre-Andean rocks probably plays only a minor metallogenic role and thus suggests that conditions for significant copper concentration were not optimal during Late Palaeozoic to Triassic times.

Epithermal precious metal deposits

The Neogene volcanic belt of the Central Andes contains hundreds of hydrothermal alteration zones of epithermal affiliation, and some of which have world-class epithermal precious metal deposits (Maksaev et al. 1984; Camus 1990b; Sillitoe 1991; Ericksen & Cunningham 1993). Neogene precious metal epithermal deposits occur in the easternmost Andes of northern Chile (Fig. 6.7), but also are present within the Andes and the Puna high plateau of Peru, Bolivia and Argentina. Rich precious metal epithermal deposits occur within the local Maricunga and El Indio belts of northern Chile (Sillitoe 1991; Bissig et al. 2002). In addition, a number of mesothermal to epithermal silver-bearing veins and gold-bearing veins of high-sulphidation and low-sulphidation affiliation are associated with Palaeocene–early Eocene volcanic rocks in northern Chile. Some of these precious metal deposits are peripheral to Palaeocene–early Eocene porphyry Cu–Mo deposits.
Miocene precious metal epithermal deposits

The most significant precious metal deposits associated with Miocene volcanic rocks occur in the High Andes of Chile and Argentina from 27°S to 30°S (Fig. 6.7), and from north to south are: La Coipa (c. 6 million ounces (Moz) of Au equivalent), Pascua (c. 12 Moz Au) and El Indio (c. 12 Moz Au). The Chilean Neogene precious metal epithermal deposits are predominantly of the high-sulphidation type according to the classification of Hedenquist (1987). The deposits basically correspond to complex fault-controlled vein systems (e.g. El Indio, La Pepa), but some also include breccia-hosted orebodies (Choquelimpie, Esperanza, El Tambo) and disseminated ores (La Coipa, Pascua). The rich El Indio vein system includes an early high-sulphidation copper mineralization dominated by massive enargite-pyrite-alunite veins with advanced argillic and quartz-sericite halos, and later quartz-gold bonanza-type veins with alteration halos of quartz-ilite. The gold-bearing veins of El Indio show that tennantite associated with Au mineralization replaced enargite, indicating an evolution to relatively lower sulphidation states during gold mineralization (Jannas et al. 1990, 1999). In addition, the Río del Medio banded quartz-rhodochrosite vein located 4.5 km

Fig. 6.6. Late Palaeozoic–Triassic porphyry Cu–Mo belt of northern Chile (ages after Camus 2003).

Fig. 6.7. Miocene precious metal epithermal deposits and Au-rich porphyry deposits of northern Chile (ages after Gröpper et al. 1991; Moscoso et al. 1993; McKee et al. 1994; Bissig et al. 2001, 2002; Muntean & Einaudi 2001).
north along-strike of El Indio is akin to low-sulphidation Au–Ag–base metal systems (Jannas et al. 1999).

The host rocks to these epithermal precious metal ore deposits are Miocene volcanic rocks, commonly associated with conspicuous bleached areas with advanced argillic alteration. El Indio is hosted by a Miocene rhyolitic to dacitic tuff unit (Amiga tuff, 27–21 Ma; Jannas et al. 1999), Choquelimpie occurs within a dacite dome complex in the core of an andesitic stratovolcano dated at 6.6 ± 0.2 Ma (Gröpper et al. 1991), whereas La Coipa and Esperanza are associated with 22–24 Ma dacite dome complexes (Oviedo et al. 1991; Vila 1991; Moscoso et al. 1993). In contrast, the Miocene gold dissemination and gold-bearing stockworks of Pascua are hosted by Late Palaeozoic granite and rhyolite (Bissig et al. 2000), and at La Coipa the early Miocene Ag–Au mineralization hosted by Miocene dacite continues in the underlying basement formed by Triassic shale beds (Oviedo et al. 1991). The conspicuous and extensive hydrothermal alteration zones that are associated with the epithermal deposits are largely of high-sulphidation type, and residual vuggy silica ledges flanked by quartz–alunite zones that grade outwards to kaolinite–dickite and illite–smectite to chloritized rocks are a common feature. Both structural and lithological control of hydrothermal alteration occurs, and contrasting intensity of hydrothermal alteration is common within the altered areas. Hypogene and epigenetic alunite are widespread in the altered rocks, and abundant within some of the ore deposits, namely Choquelimpie, La Coipa, Pascua-Lama, El Tambo and El Indio. However, quartz–illite alteration halos are related to the gold-bearing quartz veins at El Indio, which are overprinted on previous quartz–alunite alteration (Jannas et al. 1999). In contrast, the low-sulphidation Rio del Medio vein is hosted by dark chloritized Miocene andesites.

Since 1980, mines at El Indio and El Tambo have produced 126 t Au, 630 t Ag and 0.32 Mt Cu, with much of the Au coming from high grade ore (El Indio produced 191 000 t of direct shipping ore that averaged 209 g/t Au), but mining in the district ended in 2002. The early massive enargite–pyrite veins of El Indio are banded veins comprising an alteration of dark enargite–pyrite with light alunite crystalline bands; the late rich gold–silver-bearing quartz veins cross-cut these. Enargite also occurs at depth in Choquelimpie (Gröpper et al. 1991) La Coipa (Oviedo et al. 1991) and in at least one vein at La Peca (Sillitoe 1991), but this mineral phase is not a significant ore mineral in these deposits.

Porphyry-type stockwork Cu–Mo mineralization occurs about 2 km south of El Indio epithermal deposit and at a lower topographic level. This porphyry occurrence appears to be part of the same overall metallogenic event (Aranda et al. 1982; Walthier et al. 1985; Siddeley & Aranda 1986, 1990). Some other deposits (Choquelimpie, La Coipa and Pascua) are also thought likely to possess genetically related porphyry-type mineralization, but erosion has been insufficient to expose it (Sillitoe 1991). Although marked canyon incision is common in the High Andes, particularly in the Maricunga and El Indio areas, steam-heated alunite–quartz alteration with native sulphur occurs in the upper topographic sections of El Tambo, El Indio, La Peca, Pascua and La Coipa, indicating that only limited unroofing of these deposits (probably in the order of 100 to 200 m) has taken place since their formation (Bissig et al. 2002). Limited fluid inclusion data from one vein at Choquelimpie suggest that mineralization was associated with boiling hydrothermal fluids at a minimum depth of 300 m below the palaeo-surface: this deposit is located at 4860 m altitude within the core of an eroded volcanic edifice that has a maximum elevation of 5300 m (Gröpper et al. 1991). Despite abundant hydrothermal alteration zones between latitudes 19°S to 26°S, no epithermal deposits are known in the Chilean Andes along the Miocene to Recent volcanic chain. This latitude is coincident with the hyperarid Atacama Desert where Miocene volcanoes and domes preserve their original landforms, and it appears that erosion has been insufficient to expose potential epithermal mineralization associated with the volcanic chain in this part of the Chilean Andes.

The epithermal deposits of El Indio and Maricunga districts occur along the flat-slab subduction segment of the active continental margin (27–33°S) where active volcanism is absent (e.g. Thorpe et al. 1984). A progressive shallowing of the Benioff zone under the continental border has developed since around 20 to 17 Ma (Jordan et al. 1983; Isacks 1988; Allmendinger et al. 1983; Kay et al. 1987, 1988). This flattening of the subducting oceanic plate was accompanied by compressive deformation and crustal thickening, eastward migration of subduction-related magmatism, waning of main arc volcanism by c. 10 Ma, and complete termination of volcanic activity within this segment by c. 5 Ma when the modern flat-slab was completely established (Kay et al. 1999). According to Kay et al. (1999), Au and Cu mineralization at El Indio and Maricunga districts formed during the waning stages of arc volcanism, preceding either the eastward migration of the arc front or the actual end of igneous activity. In this interpretation, precious metal mineralization was ascribed to the 13–10 Ma period when hydrous, hornblende-based, residual assemblages that were in equilibrium with the erupted magmas dehydrated by amphibole breakdown to yield high-pressure, garnet-bearing assemblages in the crust. However, Bissig et al. (2000, 2001) reported new 40Ar/39Ar ages for ore-related minerals at El Indio (6.21 ± 0.26 and 7.62 ± 0.29 Ma), Rio del Medio (7.08 ± 0.19 Ma), El Tambo (7.08 ± 0.19, 7.97 ± 0.37 and 8.24 ± 0.29 Ma), Pascua (8.10 ± 0.19 and 8.73 ± 0.23 Ma) and Lama (9.00 ± 0.22, 9.40 ± 0.16 Ma). Based on these time constraints Bissig et al. (2000, 2001) argue that epithermal mineralization took place during the late Miocene in El Indio district during an episode of crustal thickening and uplift, in association with magmas largely derived from a deep crustal source anatectic environment in which garnet was a residual phase. This interpretation contrasts with the metallogenic model proposed by Kay et al. (1999) and Kay & Mpodozis (2001), who argued that no mineralization took place under these conditions.

At Choquelimpie only oxidized ore was mined (open pit and heap leaching), which in general extends to depths of 10 to 40 m, but locally reaches down to 130 m in deep oxide troughs along fracture-controlled breccia bodies. In contrast, at La Coipa silver-bearing jasperite, Ag-halides and gold dissemination resulted from oxidation within the host Miocene dacitic rocks. Thus, supergene oxidation was essential to produce economic ore in these low-grade, bulk-mining deposits. Oxidized ore also occurs at Pascua, especially in the section previously known as Nevada, but most of the Pascua orebody is hypogene. Similarly, although local oxidation occurs at El Indio (e.g. Indio Norte; Walthier et al. 1985; Jannas et al. 1990), most of this Au (–Ag, Cu) ore is hypogene.

**Palaeocene–early Eocene precious metal deposits**

The Palaeocene–early Eocene metallogenic belt includes a great number of mesothermal to epithermal silver-bearing veins (Challacollo, Huanchaqutia, El Inca, Caracoles, Cachinal de la Sierra, Vaquillas, Sierra Juncal, Tres Puntas–Chimberos and Lomas Bayas districts; Fig. 6.8) and epithermal gold-bearing veins of both high-sulphidation affiliation (El Guanaco, Cachiyuyo de Oro districts; Fig. 6.8) and low-sulphidation ore also occurs at Pascua, especially in the section previously known as Nevada, but most of the Pascua orebody is hypogene. Similarly, although local oxidation occurs at El Indio (e.g. Indio Norte; Walthier et al. 1985; Jannas et al. 1990), most of this Au (–Ag, Cu) ore is hypogene.
Iron oxide copper–gold deposits

Copper–gold deposits associated with iron oxides have been known since the seventeenth century in the Punta del Cobre district some 15 km south of Copiapó in northern Chile (Fig. 6.9). This mining district includes a number of veins, tabular or lens-like breccia bodies, as well as stockwork and stratabound Cu–Au deposits with Fe oxide gangue (Marschik & Fonboté 1996; Marschik et al. 1997; Sillitoe 2003). However, this type of deposit has only become really significant in Chile since the discovery of the Candelaria deposit in 1986 (which has resources of 479 Mt at 0.95% Cu, 0.22 g/t Au, and 3.1 g/t Ag; Marschik 2001) and Manto Verde in 1988 (which has resources of 130 Mt at 0.76% Cu at mine start-up in 1995, plus additional 180 Mt at 0.5% Cu further discovered; Zamora & Castillo 2001). A detailed review of this type of Andean deposit has been published by Sillitoe (2003).

Candelaria is the western extension of the Punta del Cobre district and lies within the contact metamorphic aureole of a granodioritic batholith with 40Ar/39Ar dates ranging from 117.2 ± 1.0 to 110 ± 1.7 Ma, which is exposed 1 km west of the deposit (Marschik 2001; Marschik & Fonboté 2001a). The bulk sulphides that carry the Cu–Au mineralization at Candelaria are related to pervasive Na–Ca metasomatism (actinolite–albite–scapolite–quartz–K-feldspar–hornblende–hedenbergite) superimposed on a previous pervasive biotitization and magnetite metasomatism that affected Lower Cretaceous andesites and andesitic tuff-breccia strata (Punta del Cobre Formation; Arévalo 1994). The stage of Ca–Na metasomatism remobilized and introduced new magnetite that locally forms up to 20% of the host rocks. The metasomatic ore bodies at Candelaria are roughly tabular, indicating
lithological control of the mineralization, although they occur immediately adjacent to regional faults in which occur magnetite-bearing mylonites. The dominant sulphide is chalcopyrite accompanied by minor pyrite (chalcopyrite/pyrite = S/I) and pyrrhotite, and the gold occurs as micrometre-sized inclusions studies revealed relatively high-grade hydrothermal alteration mineral assemblage related to mineralized rocks at Candelaria indicates high temperatures (>2450°C), an immediate igneous source for mineralizing fluids has not been identified. Re/Os molybdenite dates of 114.2 ± 0.6 and 115.2 ± 0.6 Ma were reported by Mathur et al. (2002) and were interpreted as mineralization ages. These ages are coincident with the 115.14 ± 0.18 Ma 40Ar/39Ar plateau age of biotite associated with chalcopyrite–pyrite reported by Marschik & Fontboté (2001a) and the biotite 40Ar/39Ar plateau ages of 114.2 ± 0.8 and 114.1 ± 0.7 Ma of Ullrich & Clark (1999). A younger amphibole 40Ar/39Ar plateau age of 111.7 ± 0.8 Ma (Ullrich & Clark 1999) and similar biotite 40Ar/39Ar ages of 111.0 ± 1.7 and 110.7 ± 1.6 Ma (Arévalo et al. 2000) probably represent a later event of alteration or mineralization (Mathur et al. 2002). The timing of the mineralization at Candelaria indicates that the deposit was formed after the deposition of about 2000 m of carbonate marine sediments during Early Cretaceous times over the host volcanic strata (Chañarcillo Group; Segerstrom 1967; Arévalo 1994). Therefore, Candelaria was originally formed under this c. 2000-m-thick carbonate sedimentary cover.

At Manto Verde Cu–Au mineralized breccias with an iron oxide matrix occur within transensional domains along the Manto Verde fault, a subsidiary branch of the regional strike-slip Atacama Fault System (Vila et al. 1996; Zamora & Castillo 2001). Breccias with specularite matrix occur at the Manto Verde North orebody, whereas breccias with a magnetite matrix are present at Manto Verde South pit. Only oxidized copper ores are currently mined, with oxidation extending down to 200 m and the two near-surface oxidized ore bodies being connected to a larger underlying hypogene sulphide orebody. Hypogene sulphides are dominated by chalcopyrite and pyrite (chalcopyrite/pyrite = S/I) that are disseminated within the specularite and magnetite matrix of breccias, with a magnetite matrix dominating at depth. A pervasive potassic alteration characterized by intergrowths of K-feldspar and chlorite and minor quartz and hematite affected the volcanic and intrusive host rocks at Manto Verde. Preliminary fluid inclusion studies of magnetite indicated relatively low homogenization temperatures (180–250°C), and high saline content (30–50% NaCl equivalent), and trapping of boiling hydrothermal fluids is suggested. Whole-rock K–Ar ages of altered andesites and dykes of 117 ± 3 and 121 ± 3 Ma were initially taken to indicate the age of the primary mineralization process (Vila et al. 1996), but a more precise zircon U–Pb age of 128.9 ± 0.6 and a titanite U–Pb age of 126.4 ± 0.5 Ma for a quartz monzonite to granodioritic dyke with potassic alteration, led Gelcich et al. (2003) to conclude that mineralization at Manto Verde is most probably related to the cooling and differentiation of the Sierra Dieciocho pluton, which forms the eastern margin of the deposit.

The Chilean iron oxide copper–gold deposits belong to a class of deposits characterized by abundant Ti-poor oxide mineralization referred to as the iron oxide (Cu–U–Au–REE) class or categorized as low-sulphur Cu–Au deposits (e.g. Hitzman et al. 1992; Barton & Johnson 1996). The genetic relationships with contemporaneous plutonic rocks are controversial (see review of Sillitoe 2003). Some authors favour a model where metal-bearing fluids exsolve from crystallizing magma and deposit metals in the adjacent country rocks (e.g. Gow et al. 1994; Rotherham et al. 1998; Williams 1998; Williams et al. 1999). An alternative suggestion is a model of evaportie-derived, thermally driven fluids to leach and redeposit metals (Battes & Barton 1995; Barton & Johnson 1996; Ullrich & Clark 1999). Both of these models have been proposed for Candelaria (e.g. Ullrich & Clark 1999; Marschik & Fontboté 1996). Potassic and Na–Ca high-temperature alteration assemblages related to copper sulphide deposition at Candelaria and Agustina mine at Punta del Cobre indicate that the Cu–Au mineralization processes took place at temperatures well above 450°C (Ryan et al. 1995; Hopf 1990). Hence, the origin of these deposits is likely to have a magma connection, as supported by Rb–Sr isotopic data, even though a hypothetical evaporite fluid source cannot be completely discarded (Mathur et al. 2002). A magmatic–hydrothermal origin for the copper–gold mineralization of Manto Verde has been postulated by Vila et al. (1996); fluid inclusion data indicate formation at c. 180–320°C with boiling of hydrothermal fluids during mineralization. Manto Verde appears to be a relatively shallow representative of this type of copper–gold mineralization.

**Iron oxide–apatite deposits**

Chile is currently (2005) only a modest iron ore producer (8.0 Mt of iron oxide ore were produced in 2004), although the so-called ‘Chilean Iron Belt’ comprises more than 40 Fe oxide–apatite (Ti-poor) deposits and prospects contain resources exceeding 1000 Mt of ore at 60% Fe. This belt of iron deposits extends some 600 km along the Coastal Cordillera from 25°30’S to 32°00’S (Ménard 1995; Espinoza 1990), but only Los Colorados, El Algarrobo and Romeral deposits are currently mined (Fig. 6.9). These iron deposits occur within the domain of the southernmost segment of the Atacama Fault System. This is a major regional sinistral strike-slip fault system that extends more than 1000 km along the Coastal Cordillera of northern Chile, and was developed through Jurassic–Early Cretaceous times as an intra-arc fault related to oblique subduction of the Aluk plate (Boric et al. 1990; Scheuber & Andriessen 1990). The main iron-bearing ore bodies are irregular, formed of massive magnetite with minor specularite and local traces of pyrite and chalcopyrite. Host rocks are mostly Lower Cretaceous meta-andesites adjacent to c. 100–120 Ma (K–Ar) dioritic to granodioritic plutons or meta-andesitic roof pendants on these intrusions, although massive magnetite veins hosted by diorites also occur. The economically relevant ore bodies are 100 to 1000 m long, up to 200 m wide and up to 650 m high, although more minor magnetite veins and lenticular ore bodies are also common along the belt. Martite occurs as a product of supergene oxidation of the hypogene magnetite, and locally magnetite has been completely transformed to hematite. Although gangue minerals are generally rare, the iron ores contain apatite (chlorapatite), actinolite and chlorite. The ores are massive, black in colour, and hard, but to some extent granular as they are formed of octahedral crystals of magnetite (martite). Minor pyrite and chalcopyrite locally occur.

The andesitic country rocks of the iron oxide ores show a strong alteration that consists of a mineral assemblage of actinolite–scapolite–biotite–tourmaline–chlorite–chlorapatite–sphe–minor garnet and pyrite. This high-temperature alteration mineral assemblage commonly obliterates the original porphyritic texture of the host andesites in the vicinity of the iron ore bodies, passing outward into quartz–albite±K-feldspar±biotite alteration and silicified±tourmalinized±argillized rocks (Naslund et al. 2002). K–Ar ages from the exploited Romeral, Algarbo and Los Colorados deposits range between 112 and 108 Ma (Munizaga et al. 1985), but the K–Ar data for the iron belt range from 128 to 100 Ma (Oyarzún et al. 2003). Recent U–Pb dating of apatite from a magnetite–apatite vein 500 m south of the main Carmen pit yielded 128.9 ± 3.0 Ma, coincident with the zircon U–Pb age of 130.6 ± 0.3 Ma for a quartz dioritic stock in the Carmen northern pit, and with the U–Pb ages in the Sierra Aspera pluton immediately west of Carmen (Gelcich et al. 2003).
Another belt of Fe oxide deposits of Cenozoic age (El Laco, Incahuasi and Magnetita Pedernales; Fig. 6.9) occurs in the High Andes and is related to late Miocene and Pliocene volcanic edifices. The El Laco deposit is located at elevations of between 4600 and 5200 m a.s.l. in a Pliocene stratovolcano, and comprises seven discrete bodies of massive magnetite within an 8 × 4 km area. The magnetite bodies aggregate some 500 Mt of ore at about 60% Fe. Park (1961) proposed that the El Laco magnetite is consolidated lava, occurring as a series of partly preserved flows and related feeder dykes. Several subsequent investigators have supported this concept based on textural features displayed by the magnetite (Naslund et al. 2002, and references therein). Others support a metasomatic replacement origin for El Laco based on detailed geological and laboratory studies (Rhodes & Oreskes 1999; Sillitoe & Burrows 2002).

A longstanding controversy persists about the origin of the Chilean iron oxide–apatite (Ti-poor) ore deposits. Some authors (e.g. Nyström & Henríquez 1994, 1995; Naslund et al. 2002) contend that many such deposits are derived by melt crystallization, compare them to the Kiruna deposits of Sweden, and relate the mineralization to injections of iron-rich magmas at subvolcanic and volcanic levels (Vivallo et al. 1994; Nyström & Henríquez 1994, 1995; Gelcich 1999; Naslund et al. 2002). On the other hand, others support the proposal that these iron deposits have a metasomatic replacement origin (e.g. Bookstrom 1995; Sillitoe 2003). The hypothetical iron metasomatism was accompanied by pervasive actinolitization of the volcanic host rocks in the temperature range 475–550°C and at a pressure of about 2 kbar (Ruiz et al. 1965, 1968; Bookstrom 1977, 1995; Ménard 1995; Hritzman et al. 1992; Sillitoe & Burrows 2002). Ménard (1992, 1995) proposed that the Chilean magnetite–apatite mineralization was produced by the emplacement of deep-seated (>4 km), subduction-related, anhydrous (pyroxene-bearing) gabbros and diorites. These mafic intrusions evolved under relatively high oxygen fugacities and developed the iron oxide mineralization and associated sodic alteration at high temperature (600–450°C).

**Stratabound copper–(silver) deposits**

Stratabound copper deposits with subordinate silver, long known as 'Chilean Manto-type', occurs along the Coastal Cordillera of northern Chile hosted by Jurassic and Lower Cretaceous volcanic and volcanosedimentary rocks. These Mesozoic stratabound deposits historically were the second source of Chilean copper production after Cenozoic porphyry copper deposits, though have recently been displaced to third place by the exploitation of the large Lower Cretaceous iron oxide copper–gold deposits at Candelaria and Manto Verde (Marschik & Fontboté 2001a; Vila et al. 1996; Zamora & Castillo 2001).

Two groups of significant stratabound Cu–(Ag) occur along the Coastal Cordillera of northern Chile. The first of these is seen between 21º30’S and 26ºS, and is hosted by a Jurassic volcanic sequence (La Negra Formation; Kojima et al. 2002). The second group occurs from 30ºS to 34ºS and is hosted by a Jurassic bimodal suite of rhyolitic and andesitic rocks, but the largest deposit of this group, Mantos Blancos, is hosted by a bimodal suite of rhyolitic and andesitic rocks, which according to ongoing investigations led by C. Palacios are mineralized sills, dykes and intrusive breccia pipes.
Stratabound Cu–(Ag) deposits occur near gabbroic, dioritic or andesitic subvolcanic intrusive bodies, such as dykes, sills, stocks or volcanic necks, but these intrusives in most cases are unmineralized (e.g. Buena Esperanza, Susana, Santo Domingo; Palacios 1986; Espinoza et al. 1996). The intrusives have been interpreted as feeder conduits of the Jurassic volcanism (Palacios & Definis 1981a, b; Espinoza 1981; Espinoza et al. 1996). Mantos Blancos includes at least four main lenticular disseminated and fracture-filling sulphide ore bodies (Sorpesra, Aida, Nora and Marina) forming an overall, slightly unconformable, subhorizontal tabular ore deposit (Chavez 1985). The thickness of mineralized levels at Mantos Blancos ranges from 150 to 350 m and the deposit extends irregularly over an area of 2.6 by 1.2 km. At Buena Esperanza and Susana copper sulphides occur within the matrix of a central breccia pipe and are disseminated in a number of conformable stratabound orebodies (‘mantos’) around the breccia pipe (Palacios 1990; Espinoza et al. 1996). The stratiform orebodies (2 to 25 m thick) are commonly restricted to the amygdaloidal and brecciated sections of the Jurassic lava flows and minor veins occur along local faults and fractures (Palacios & Definis 1981a, b; Dreyer & Soto 1985).

The main hypogene sulphides are chalcocite and bornite, minor chalcopyrite, and locally covellite and digenite. Gangue minerals are quartz, hematite, pyrite, chlorite and calcite. Minor magnetite occurs as dissemination within the mineralized rocks, but is mostly replaced by hematite or maghemite. At Mantos Blancos specularite occurs; this crystallized early in the paragenesis and is mostly concentrated within a barren andesite flow that overlies the orebody, unlike fine reddish hematite dissemination, which occurs within mineralized rocks. A lateral hypogene zonation has been described at Mantos Blancos and Santo Domingo. This includes copper-rich cores dominated by chalcocite–bornite–digenite surrounded by a rim of bornite–chalcopyrite or sole chalcopyrite, and an external, mostly uneconomic halo of chalcopyrite–pyrite ore (Chavez 1985; Definis 1985). The hydrothermal alteration assemblage of albite–chlorite–hematite–quartz–calcite–epidote–sphene–scapolite–anatase–minor sercite is associated with ore minerals in these stratabound copper deposits, but the primary textures of the volcanic rocks are preserved (Losert 1973; Chavez 1985; Palacios 1986, 1990). This local alteration is thought to be superimposed upon a regional alteration/metamorphism (prehnite–pumpellylite facies) of the volcanic sequence, but little or no alteration contrast occurs between mineralized and barren volcanic country rocks. The hydrothermal alteration is particularly pervasive in Mantos Blancos, where lithochemistry shows significant metasomatism of the host rocks with the addition of Na, Fe and Mg (Chavez 1985), although deeper parts of the orebody within intrusive breccias shows potassic alteration dominated by K-feldspar.

Most of the stratabound copper deposits have an upper oxidized zone that extends to a maximum depth of 250 m. The degree of supergene oxidation is variable: some deposits include almost exclusively copper oxides (e.g. Mantos de la Luna, Juarez), whereas others are formed exclusively of hypogene sulphides (e.g. Buena Esperanza). The boundary between the supergene and hypogene ores is gradual, so that some deposits include a zone with a mixture of copper sulphides and oxides (e.g. Susana, Santo Domingo). The oxidized ores include mainly atacamite, minor chrysocolla, malachite, copper sulphates, and rare cuprite and native Cu. Copper grades are similar in the oxidized upper section and the underlying hypogene sulphide zone, so that no significant Cu transport occurred during the supergene oxidation (in situ oxidation). Only the largest deposits such as Mantos Blancos and Susana have local enriched pockets with supergene chalcocite-group minerals and covellite (Chavez 1985; Wolf et al. 1990). The poor development of supergene enrichment within the stratabound copper deposits could be explained by the insufficiency of hypogene pyrite to release supergene acid solutions under oxidation, and the profuse occurrence of calcite gangue that may have readily neutralized any supergene acid solutions, so precluding the leaching of metallic cations from the oxidized zone.

Sparse and imprecise K–Ar and Rb–Sr minimum radiometric dates indicate that the main stratabound deposits hosted by Jurassic volcanics rock of northern Chile were formed at about 150–140 Ma (Boric et al. 1990; Venegas et al. 1991; Tassinari et al. 1993; Vivallo & Henriquez 1998). This radiometric age range is younger than the volcanic host rocks that have been dated by Rb–Sr and K–Ar methods in the 186–165 Ma range, but overlapping with 40Ar/39Ar, Rb–Sr and K–Ar ages of granodioritic batholiths that intrude the volcanic sequence along the coastal area (age compilation in Vivallo & Henriquez 1998). Minor stock and dykes spatially related to the stratabound Cu–(Ag) deposits have minimum K–Ar ages ranging between 154 and 133 Ma, except for the gabbroic stock at Buena Esperanza that has yielded a K–Ar age of 168 ± 5 Ma in plagioclase (Boric et al. 1990).

Stratabound copper deposits hosted by Lower Cretaceous volcanic rocks occur in central Chile (30–34°S). The largest deposit is El Soldado, whose production plus resources totals over 200 Mt at 1.4% Cu (Boric et al. 2002). The second largest in volume is the now-exhausted Lo Aguirre deposit where 19 Mt at 66% total Cu (0.9% soluble Cu) were exploited (Saric et al. 2003). El Soldado copper deposit is located about 120 km north of Santiago, in the Coastal Cordillera of central Chile (Fig. 6.10). It is hosted by a sequence of felsic and basaltic units dipping c. 30° to the east of the Upper Member of the Lower Cretaceous Lo Prado Formation (Rivano et al. 1993). The deposit consists of numerous isolated orebodies with intervening barren zones, distributed in about a dozen individual orebody clusters or blocks. These blocks are spatially distributed within a volume about 2 km long by 1 km wide and 600 m in vertical extent (Wilson & Zentilli 1999). Within the blocks, individual subvertical orebodies are extremely variable in size, from very small to 450 m long, 150 m wide and 450 m in vertical extent (Boric 1997; Boric et al. 2002). Although described as stratabound on a regional scale, in detail the El Soldado orebodies are distinctly discordant, displaying a strong structural control (Ruge 1985). The orebodies are preferentially developed within a generally north–south to NNW regional fracture system, especially where north–south, east–west and NW faults intersect. Away from these zones of structural permeability, orebodies can be best described as veins (Boric 1997). Lithological control is exerted by the relatively more brittle felsic and intermediate flows (dominant) and their finer and more matrix-richly mineralized in comparison to basaltic flows, tuffs, and underlying sedimentary rocks (Ruge 1985). Andesitic–basaltic dykes are generally barren (Boric 1997).

Hypogene ore minerals are chalcopyrite, bornite and chalcocite and occur as dissemination and veins, largely filling primary and secondary porosity of the volcanic host rocks. According to Boric (1997) many individual orebodies in a block show mineralogical zoning: a core of chalcocite–hematite or chalcocite–bornite–hematite is followed outwards by approximately concentric zones of bornite–chalcocite pyrite chalcopyrite, and pyrite in the most external zone. The deeper roots of the orebodies contain relatively more pyrite than their upward terminations and are surrounded by a halo of pyrite dissemination. This pyrite halo and the deep pyrite bodies are early, low-temperature, diagenetic and probably biogenic in origin, related to basinal, often petroleum-rich fluids. Most of these deposits have formed in two stages: one low temperature, near the age of the host rocks, and another hydrothermal episode later. Although oxidized copper zones exist near the surface and enriched ores were exploited, supergene enrichment is not significant. Common waste or gangue minerals are pyrite, hematite, calcite, chlorite, albite, microcline, bitumen and minor amounts of sphalerite, galena and arsenopyrite. Copper grade is extremely variable (Klohn et al. 1990). Lateral limits of the
orebodies are characterized by abrupt changes of Cu grade from a nucleus with about 2% Cu to outside zones containing 1.2–0.5% Cu within a few metres (Ruge 1985; Klohn et al. 1990). The wall rock between the orebodies is generally barren (<0.15% Cu; Klohn et al. 1990). Hydrothermal alteration consists of abundant calcite, chlorite, albite, microcline, epidote, opaline silica, titanite and rutile–anatase, and some sericite and clay minerals, but primary rock textures are largely preserved (Holmgren 1987; Boric 1997).

Highly saline (c. 30–40% NaCl equivalent) three-phase fluid inclusions in quartz related to sulphide mineralization at El Soldado yielded homogenization temperatures of 200–257°C (without pressure correction) and no evidence for boiling was observed (Holmgren 1987). The same study found that fluid inclusions within late barren calcite at El Soldado are also highly saline but with lower homogenization temperatures of 82 to 104°C. On the other hand, three-phase fluid inclusions in quartz from quartz–bornite-filled amygdales at El Salado consisted of abundant calcite, chlorite, hematite, epidote, zeolites and albite, with copper sulphides at Talcuna orebodies yielded homogenization temperatures of 250 to 430°C, and some vapour-rich inclusions suggest trapping of a boiling hydrothermal fluid (Nisterenko et al. 1973). El Salado is a minor volcanic-hosted stratiform Cu–(Ag) deposit located 18 km SE from El Soldado and close to the western border of a granodioritic host located 18 km SE from El Salado deposit yielded homogenization temperatures of 250 to 430°C, and some vapour-rich inclusions suggest trapping of a boiling fluid (Nisterenko et al. 1973). El Salado is a minor volcanic-hosted stratiform Cu–(Ag) deposit located 18 km SE from El Soldado and close to the western border of a granodioritic intrusion with a biotite K–Ar age of 118 ± 3 Ma (Rivano et al. 1993).

Many radiometric dates (K–Ar, 40Ar/39Ar, Rb–Sr) reported from El Soldado range from 131 to 96 Ma, supporting the Early Cretaceous age for this copper deposit (Boric & Munizaga 1994; Boric 1997). The oldest radiometric dates at El Soldado (from 131 to 118 Ma) are coeval with the stratigraphic Neocomian age of the host volcanics. A group of younger ages from 113 to 96 Ma are considered to represent the alteration/mineralization event: this younger group of radiometric ages includes dates of K-feldspar and albite veinlets associated with copper sulphides that are mostly concentrated between 105 and 101 Ma (Boric 1997). Wilson et al. (2003a) have dated ten samples of K-feldspar (adularia) from El Soldado by the step-heating 40Ar/39Ar method. For hydrothermal K-feldspar in close association with copper sulphide precipitation, the ages range from 100.5 ± 1.5 Ma to 106 ± 1.1 Ma, with a mean of 103 ± 1.3 Ma, which these authors interpret to be the main age of copper mineralization at El Soldado. Fission track dating of apatite in the host rocks yields an age of c. 90 Ma, indicating post-mineralization fast cooling of the system (Wilson et al. 2003a). According to the radiometric data, mineralization at El Soldado coincides temporally with the K–Ar age range (118–96 Ma, as determined by Wilson et al. 1998). A distinct group of 40Ar/39Ar ages from El Soldado comprises a subhorizontal bituminous lapilli tuff level with veins: rich pockets are restricted to a specific stratum and interconnected by poorly mineralized sections. Copper sulphides concentrate in the upper few metres of a specific sedimentary or pyroclastic level that normally underlies either a massive volcanic or mudstone impervious stratum. Typical examples occur at Talcuna (Boric 1985) and Cerro Negro (Elgueta et al. 1990). Hypogene minerals are bornite, chalcopyrite, chalcolite, pyrite, and minor sphalerite and galena. Gangue minerals are calcite, chlorite, hematite, epidote, zeolites and local magnetite. At Talcuna stratiform Cu–(Ag) mineralization occurs mostly along a subhorizontal bituminous lapilli tuff level 10 to 15 m thick, but highest-grade ore occurs along the intersection between the lapilli tuff horizon and subvertical NNW-trending veins. These veins have open-space filling textures, evidence of hydraulic fracturing, and have been interpreted as feeders for hydrothermal fluids that mineralized the bituminous lapilli tuff level under an impervious manganerich level of tuffaceous sandstones and mudstones (Boric 1985). Present exploitation in the Talcuna district concentrates on NW-longate ore shoots that developed along the intersection of the above-mentioned lapilli tuff level with veins: rich pockets coincide with the occurrence of abundant bitumen. Coexisting liquid-rich and vapour-rich fluid inclusions in calcite (suggesting trapping of a boiling hydrothermal fluid) associated with copper sulphides at Talcuna orebodies yielded homogenization temperatures of 70 to 170°C and salinities from 5 to 27 wt% NaCl equivalent (Oyarzún et al. 1998). These authors interpreted the wide range of salinity variation as the result of complex interaction between boiling of the hydrothermal fluid and mixing with non-saline waters during mineralization at Talcuna. In addition, fluid inclusions in calcite from cavity fillings in andesites in the area of the deposits yielded homogenization temperatures in the range 120–205°C and salinities from 11 to 19 wt% NaCl equivalent. These were taken to represent an earlier stage of mineralization at Talcuna (Oyarzún et al. 1998).

The stratabound deposits of this group have been formed within the Lower Cretaceous sequences deposited in shallow marine to intermontane lacustrine sedimentary basins with coeval volcanoclastic deposits. Stratiform Cu–(Ag) deposits of this type, found in intravolcanic depressions probably were pull-apart basins related to active sinistral strike-slip faults during Early Cretaceous times. In addition, Zentilli et al. (1997) and Wilson & Zentilli (1999) documented that bitumen commonly occurs within these stratabound Cu–(Ag) deposits hosted by Lower Cretaceous volcanosedimentary strata (Uchumí, Talcuna, Cerro Negro), suggesting that copper mineralization in these deposits was deposited within degraded petroleum reservoirs.

The origin of the copper stratiform deposits of northern Chile has long been a matter of controversy. Stratiform orebodies were first regarded as syngenetic volcanic exhalative (Ruiz et al. 1965, 1971; Stoll 1965). However, their epigenetic origin has now been widely demonstrated, due to the subsequent discovery of unconformable orebodies, the spatial relationship of copper mineralization around Upper Jurassic intrusive stocks and sills, and significant hydrothermal alteration (albite, chlorite, quartz, sericite, calcite, sphene, scapolite and anatase) associated with copper-rich sulphide dissemination (chalocite, bornite) within the volcanic host rocks (Palacios & Definis 1981a, b; Dreyer & Soto 1985; Espinoza et al. 1996). The most widely accepted hypothesis is the hydrothermal derivation of these volcanic-hosted copper deposits, related to subvolcanic intrusive bodies (Espinoza 1981, 1982;
Chavez 1985; Palacios 1990; Espinoza et al. 1996), although some authors have also suggested a diagenetic–metamorphic origin (Sato 1984; Sillitoe 1990) or a genetic connection with underlying batholiths (Losert 1974; Vivallo & Henríquez 1998; Maksaev & Zentilli 2002).

There has been some debate over the origin of the stratabound Cu–(Ag) deposits that occur peripheral to Upper Jurassic batholiths emplaced within transensional sinistral regional faults within the Mesozoic volcanic pile. The prevalent view that these Cu–(Ag) deposits have an inherent genetic relationship with hydrothermal fluid derivation from subvolcanic stocks and dykes was contested by Maksaev & Zentilli (2002), as these minor intrusions are largely unmineralized and this hypothesis does not fit well with Sr, Os and Pb isotopic data that call for contribution of these elements from the country rocks (e.g. Maksaev & Zentilli 2002; Wilson et al. 2003b). The Cu–(Ag) stratabound mineralization appears to be the product of hydrothermal fluids of mixed origin that were mobilized during the emplacement and cooling of Upper Jurassic batholiths within the Mesozoic sequence and deposited copper where cooled and mixed with meteoric fluids away from their igneous heat sources.

Copper-, silver- and gold-bearing veins

Numerous copper-bearing veins with iron oxide gangue occur along the Coastal Cordillera of central and northern Chile, especially between latitudes 21ºS and 26ºS. Upper Jurassic dioritic to granodioritic batholiths host these veins, the most significant of which occur in the Tocopilla, Gatico, Naguayán–Desesperado, Julia and Monte Cristo districts (Fig. 6.11). These vein districts of the Coastal Cordillera were of primary economic importance in the second half of the nineteenth century and early twentieth century when most of the Chilean copper production came from the exploitation of copper-bearing veins, but they are now long abandoned. The largest veins are Minita–Despreciada (Tocopilla District), Toldo–Velarde (Gatico District), and Julia–Reventón (Julia District). These are mostly NE-trending, steeply dipping veins (though some veins are striking WNW, east–west and north–south), from 750 to 2000 m long, 1 to 12 m wide and with about 370 to 670 m of known vertical extent. Copper ores concentrate in rich pockets along these structures, separated by low-grade or barren sections. The hypogene paragenetic sequence of the largest veins is tourmaline–actinolite–quartz–magnetite–hematite–pyrite–chalcopyrite–bornite–calcite (Boric et al. 1990). Strong silicification, argillic alteration and chloritization occur within these copper-bearing veins, and extend some metres into their intrusive wall rocks. The hypogene minerals fill fractures and openings, either as irregular and discontinuous veinlets or massive pockets with banded textures, or as fine dissemination. The structure of the Cu-bearing veins is regular and continuous within intrusive bodies, but quite irregular and discontinuous when veins extend within the intruded volcanic rocks, as in the Naguayán–El Desesperado district (Boric et al. 1990). Cu-bearing veins from the Tocopilla and Guanillos districts contain hypersaline fluid inclusions (48–68 wt% NaCl equivalent) with homogenization temperatures of 320 to 540ºC, mostly between 380 and 420ºC. The characteristics of this type of magnetite- and copper-bearing vein indicate that they are high- to moderate-temperature mesothermal deposits (Ruiz et al. 1965; Boric et al. 1990; Vivallo & Henríquez 1998). They are genetically related to the emplacement and cooling of Upper Jurassic batholiths (K–Ar and Ar/Ar ages ranging from 167 to 140 Ma; Maksaev 1990).

The apparent time and space relationships between the Jurassic volcanic-hosted Cu–(Ag) stratabound deposits and the mesothermal Cu-bearing veins hosted by plutonic rocks, plus geochemical and isotopic comparison, led Vivallo & Henríquez (1998) to postulate that the batholith-hosted, iron-rich, Cu-bearing veins were the deep feeder structures for the hydrothermal fluids that produced the stratabound deposits within the Jurassic volcanic strata. Although this hypothesis cannot be discarded, these two types of deposits were formed under rather different pressure and temperature conditions, their timing of formation is poorly constrained, and there are no direct field relationships between them.

Mesothermal gold-bearing veins occur in faults and fractures within Jurassic batholiths and their country rocks near the contact with the intrusive masses along the Coastal Cordillera in northern and central Chile (from 20ºS to 34ºS; e.g. Carmen, Filomena, Carizalillo, Talca, Hornillos, Las Vacas, Pullalli and Curacavi districts). Individual veins are generally less than 350 m long, 0.5 to 1 m wide, and most have been exploited only within the oxidized zone down to depths of less than 100 m. The
Upper exploited section of these deposits includes fine dissemination and stringers of native gold and minor chrysocolla, malachite and atacamite with gangue of quartz, hematite, limonite, calcite and rare magnetite, sericite and tourmaline. The hypogene ore minerals at depth are mostly specularite, auriferous pyrite and variable amounts of chalcopyrite, bornite and atacamite. Although these are minor, largely abandoned gold deposits, they appear to be the primary source of alluvial gold that has been exploited in placer deposits within the Coastal Cordillera of central Chile. A good example is the Marga-Marga placer deposits 120 km NW from Santiago city (Fig. 6.11), where about 2 Moz of alluvial gold have been extracted since colonial times.

Silver-bearing veins occur in the old district of Santa Rosa Huantajaya in the Coastal Cordillera of Iquique (latitude 20°15′S; Fig. 6.11). This district has been exploited intermittently since colonial times, becoming a significant silver producer by the end of the nineteenth century and early years of the twentieth century, but has been abandoned since 1942. The veins are hosted by Middle to Late Jurassic limestone strata that were intruded by dioritic and andesitic dykes and sills with local skarn development. Veins are orientated N60°E/60°S, N60°W/60°S and east–west/subvertical, extend up to 700 m long, are 0.5 to 2 m wide, and have rich pockets at vein intersections, which were exploited down to 500 m. The exploited ores came mainly from the oxidation zone, which extends down to 100 m and includes an assemblage of chlorargyrite, malachite, azurite, and limonite, but also from a well-developed supergene enrichment sulphide zone, from 100 to 400 m deep, with native silver, acanthite, chalcocite and covellite. The hypogene ores at depth include arsenopyrite, pyrite, chalcocopyrite, enargite, argentiferous tetrahedrite, galena and sphalerite.

Various mesothermal gold deposits are associated with Lower Cretaceous volcanic and intrusive rocks (e.g. Henriquez et al. 1991; Fribla 1991; Camus 1990b). These deposits are widespread along the Coastal Cordillera between latitudes 27°S and 36°S (Fig. 6.11) and some districts were once significant gold and copper producers, although all are now inactive.

About 100 NW-trending gold-bearing veins of various sizes, along with stratabound deposits associated with K-feldspar–hematite or chloritic alteration haloes occur at Andacollo Gold (Oyarzún et al. 1996). Gold is contained in gold-bearing pyrite, which is disseminated along pink-coloured, pervasively altered strata, and accompanied by sphalerite, galena and chalcocite within veins. Cinnabar and tennantite also occur within the most distal veins (3 km away from the Andacollo porphyry copper deposit). A whole-rock K–Ar age of 91 ± 6 Ma obtained from a gold-bearing sample is regarded as a minimum age for the mineralization (Munizaga et al. 1985; Reyes 1991).

In the Mantos de Punitaqui district a number of gold-bearing quartz–sulphide veins and carbonate–sulphide breccia pockets occur (Los Mantos, Delirio, Milagro, Azogues, La Culebra and Farellón mines). These deposits are distributed along the Punitaqui NNE-trending sinistral shear zone and on echelon NNW-trending subsidiary faults. The Punitaqui fault has juxtaposed a sequence of Lower Cretaceous andesites and skarnified limestone intercalations with a granodioritic pluton that has yielded a biotite K–Ar age of 114 ± 3 Ma. Hypogene sulphides are pyrite, chalcopyrite, tetrahedrite, bornite, cinnabar, schwatziite and realgar; specular hematite and magnetite are also common. Highest gold and copper grades occur within breccia bodies with a calcite matrix and abundant chalcocopyrite and hematite. At Colliguay gold-bearing quartz–pyrite veins (Tirillenta) and gold-bearing silicified breccia bodies (Vizcaino) occur peripheral to Lower Cretaceous granodioritic porphyry pervasively altered to quartz-sericite-chlorite and containing a stockwork of pyrite veins (Hernandez et al. 1999; Townley et al. 2000). A sericite K–Ar age of 129 ± 3 Ma was obtained for gold-bearing breccias in this district.

Gold-bearing veins hosted by Lower Cretaceous plutons occur at El Capote, Ojancos and La Higuera districts. The hypogene section of the veins consists mainly of quartz, pyrite, chalcocopyrite and arsenopyrite, with significant specularite occurring locally. Limited sericitic hydrothermal alteration haloes extend some metres in the host rocks. Similar gold-bearing quartz–sulphide veins, but hosted by Lower Cretaceous volcanic rocks, occur at Las Palmas and El Tigre districts, whereas gold-bearing breccia bodies with quartz–sulphide matrix hosted by a Lower Cretaceous granodiorite occur at El Chivato (Camus 1990b).

Numerous mesothermal to epithermal Au–Cu-bearing vein systems are related to Late Cretaceous igneous rocks. Along the Coastal Cordillera of north-central Chile the districts of Inca de Oro, El Espino, Farellón Sanchez, El Bronce de Petorca, Alhuelo and Lo Chancón (Fig. 6.11) are all representative of this type of mineralization. Moreover, silver-bearing mesothermal veins that are hosted by Jurassic and Lower Cretaceous limestone strata at Caracoles, Chimbberos–Tres Puntas, Chañarcillo and Arqueros districts probably were also formed in the Late Cretaceous. Although these districts have long mining histories, most of the mines became uneconomic and are now abandoned. Mechanized mining activities are at present limited to Alhuelo and Farellón Sanchez districts in central Chile.

Complex structurally controlled vein systems characterize the Late Cretaceous metallic ore deposits. Individual veins are formed of quartz, sulphides and sulphosalts (pyrite, arsenopyrite, with variable amounts of chalcopyrite, sphalerite, galena and tetrahedrite), and common but minor barite and carbonates. In addition, some veins include significant amounts of magnetite and hematite (Alhuelo). The veins are hosted by Cretaceous volcanic rocks, but are commonly close to Upper Cretaceous stocks (Inca de Oro, El Espino, Farellón Sanchez, El Bronce de Petorca, Alhuelo and Lo Chancón districts). Most veins were primarily exploited down to depths of 20–60 m in their richest upper oxidized sections formed by quartz, limonite and hematite with grades of 8 to 30 g/t Au, but quartz–sulphide ores have also been mined from the largest veins and richest ore shoots down to 100 to 200 m depth, and exceptionally to 400 m (El Bronce de Petorca) with grades of 3 to 6 g/t Au; copper has commonly been recovered from these sulphide ores.

K–Ar dates at El Bronce de Petorca bracket the precious metal mineralization between 86 and 79 ± 3 Ma (Camus et al. 1991). K–Ar and 40Ar/39Ar dating of the Maqui vein indicates that mineralization occurred between 86 and 82 Ma (Cotton et al. 1999). At Inca de Oro intrusive bodies with K–Ar ages between 80 and 77 Ma have been taken to indicate that precious metal mineralization occurred in Late Cretaceous times (Palacios et al. 1993). Although no radiometric data are available for other districts, the Upper Cretaceous stratigraphic age of the host volcanic rocks provides a maximum age for the mineralization.

Silver-bearing veins hosted by Jurassic and Lower Cretaceous limestone strata occur in the old Caracoles, Chimbberos–Tres Puntas, Chañarcillo and Arqueros deposits. These silver deposits were intensively exploited during the nineteenth century, but are long abandoned. The Chañarcillo district, renowned for its rich bonanzas, is located 50 km south of Copiapó city and includes a large number of subvertical veins with NE, NW and north–south strikes (Whitehead 1919; Segerstrom 1962; Diaz et al. 1998).

Sedimentary rock-hosted gold deposits

A Jurassic limestone sequence cropping out east of the old Potrillillos copper mine hosts disseminated gold deposits such as El Hueso, Agua de la Falda and Jerónimo. El Hueso is a sedimentary rock-hosted epithermal gold deposit located 3 km east of the Potrillillos porphyry copper. The original mineral
reserves at El Hueso were 12 Mt at 1.5 g/t Au (0.4 g/t Au cut-off) that were mined out by Homestake Company from 1988 to 1994. El Hueso occurs within a 150-m-thick section of Jurassic limestone strata intruded by stocks and dykes of dacitic porphyry. The sedimentary rocks at the mine pit are pervasively silicified, and show siliceous–argillic alteration. Intermediate argillic alteration encompasses the mineralized rocks. Diseminated gold occurs along some stratigraphic levels forming an overall stratobound orebody (0.4 to 2.0 g/t Au) striking north–south and dipping 30°W. In addition, the stratobound gold dissemination is intersected by a number of mineralized east–west to N70°W trending faults that dip 70°N. Irregular, massive siliceous ledges occur along these faults (Esperanza, Hueso, Tunnel 3 and Central veins). The highest gold grades occur within these siliceous veins assaying locally up to 30 g/t Au, but averaging 4 g/t Au. Ore minerals consist of native gold, electrum, rare nacogait, scorodite, cervantite, stibnite, enargite, pyrite, cinnabar, realgar, orpiment, arsenopyrite, galena, chenexvite, chalcantite, malachite and chrysocolla; gangue minerals are quartz, alunite, goethite, jarosite and minor barite. Gold occurs as 3 to 25 µm sized grains as native ore being dominated by pyrite with low gold content (Ilanes 1991).

El Hueso was customarily regarded as a high-sulphidation epithermal gold deposit with ‘Carlin-like’ affinities and a genetic link to the Potrerillos porphyry copper deposit (Colley et al. 1989; Sillitoe 1991; Davidson & Mpodozis 1991). However, the Au/Ar geochronological data and field observations of Marsh et al. (1997) indicate that El Hueso was formed from 40.2 to 40.8 Ma as a low-sulphidation epithermal gold deposit pre-dating by some 5 Ma the nearby Potrerillos porphyry copper deposit. An advanced argillic alteration assemblage pre-dating by some 5 Ma the nearby Potrerillos porphyry copper deposit. An advanced argillic alteration assemblage present at Cerro Silica at El Hueso, consisting of alunite, pyrophyllite, diaspora, zunyite, dickite, rutile and pyrite, was dated at 36.2 Ma (alunite 40Ar/39Ar plateau). According to Marsh et al. (1997) this is a late event lacking significant gold mineralization and also pre-dating the Potrerillos porphyry copper.

The Agua de la Falda mine located 2 km east of El Hueso is another oxidized sedimentary-hosted gold deposit that was exploited by Homestake mining company. Silicified and Fe–Mn-stained Jurassic limestone beds host gold dissemination. The oxidized mineral resources at Agua de la Falda amount to 1.3 Mt at 7.4 g/t Au (2.0 g/t Au cut-off). Down-dip in the same stratigraphic level the Jerónimo sulphide orebody occurs, though both deposits are separated by a post-mineralization normal fault. Homestake has delineated a resource of 15.05 Mt at 5.79 g/t Au at Jerónimo (Homestake, unpublished data). Jerónimo is a stratobound disseminated gold deposit within a section of Jurassic limestone strata of El Asiento Formation located 5 km east of Potrerillos porphyry copper deposit. The mineralized level (N20°E/12–15°W) is about 12 m thick and extends at least 200 m down-dip. Mineralization is also structurally controlled as ore minerals are found in subvertical fractures and joints.

The ore mineral assemblage at Jerónimo consists of pyrite, arsenopyrite, sphalerite, lead sulphosalts, orpiment and realgar, with minor coloradoite, aluita, cinnabar and cassiterite. Although minor visible gold is present, it generally occurs as 5 to 30 µm grains that are encapsulated in pyrite, arsenopyrite, quartz and realgar, and gold also occurs within vugs in the silicified matrix of host rocks. Alteration phases include: (1) strong, pervasive replacement-style silicification; (2) vug-filling by manganese carbonate and calcium carbonate (rhodocrosite, kuttunohrite); and (3) argilization consisting of widespread disseminated and veinlet illite, and vug-filling kaolinite in the centre of the deposit. Other common alteration minerals include apatite, rutile, monazite and barite. Pb and O isotopic data indicate ore fluid derivation mostly from a magmatic source with some input from wall-rock sources (Gale 2000). Some affinity to Carlin-type sedimentary rock-hosted gold deposits has been suggested for Jerónimo (e.g. Sillitoe 1991). However, according to Gale (2000) Jerónimo is a gold-rich carbonate replacement deposit that shows distinct differences from Carlin-type gold deposits. These differences include enrichment in base metals, carbonate alteration, the presence of gold as visible grains, occurrence in a district containing porphyry and related styles of mineralization, and evidence for a migmatic contribution to metals and hydrothermal fluids. Although no geochronological data are available from Agua de la Falda and Jerónimo gold-bearing orebodies, a number of upper Eocene stocks intrude the Jurassic limestone sequence in the region (e.g. Marsh et al. 1997; Mpodozis et al. 1994) suggesting that the hypogene mineralization at these deposits was formed in late Eocene times, probably close in age to the nearby El Hueso epithermal gold deposit.

### Gold-rich porphyry deposits

A number of gold-rich porphyry deposits occur in the High Andes between latitudes 27°S and 28°S (Maricunga belt; Muntean & Einaudi 2001) and include Refugio, Cerro Casale, Marte, Lobo, Santa Cecilia, and other prospects (Fig. 6.7). This group of deposits probably forms the largest gold concentration in the Andes as their resources exceed 40 Moz of gold, but they are contained within low-grade deposits (Table 6.1). Only Refugio was recently under exploitation, but closed mineral extraction in May 2001 due to a low gold price, although it is expected to start up again in May 2005, with expanded crushing facilities. Marte was also exploited from 1989 to 1991, but the mine closed due to deficient gold recovery from heap leaching. The rest of these Au porphyry deposits are essentially low-grade, large-tonnage prospects (Table 6.1), except for Cerro Casale, which contains a core of hydrothermal breccia with high Au and Cu grades (reverse circulation hole 176 intersected 42 m of breccia at 17.7 g/t Au and 1.5% Cu; Bema Gold Corporation, unpublished data, 1999).

### Table 6.1. Tonnage and Au grade of the main Au porphyry deposits of the Maricunga area, northern Chile (data from Muntean & Einaudi 2000)

<table>
<thead>
<tr>
<th>Deposit</th>
<th>Ore (Mt)</th>
<th>Au grade (g/t)</th>
<th>Cu grade (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Marte</td>
<td>46</td>
<td>1.44</td>
<td>0.5</td>
</tr>
<tr>
<td>Lobo</td>
<td>53</td>
<td>1.73</td>
<td>1.0</td>
</tr>
<tr>
<td>Refugio (Panco)</td>
<td>216</td>
<td>0.88</td>
<td>0.5</td>
</tr>
<tr>
<td>Verde (Refugio)</td>
<td>81</td>
<td>0.85</td>
<td>0.5</td>
</tr>
<tr>
<td>Cerro Casale</td>
<td>1183</td>
<td>0.71</td>
<td>26.9</td>
</tr>
</tbody>
</table>

Mioocene volcanic edifices of the southernmost Central Volcanic Zone of the Andes (Mpodozis et al. 1995) host the Au porphyry deposits of the Maricunga area. An example is the Marte deposit which occurs in the core of Pastillitos volcano and is a parasitic cone on the eastern foothills of the large compound Copiapó stratovolcano. Geochronological data indicate that a group of deposits was formed from 24 to 22 Ma (Refugio, Santa Cecilia, Cavancha at La Pepa) and another group was formed from 14 to 13 Ma (Cerro Casale, Marte, Lobo) (Sillitoe et al. 1991; McKee et al. 1994; Mpodozis et al. 1995). The Mioocene gold-rich porphyries occur within a section of the Chilean Andes with a NNE structural trend (deflection) between latitudes 27°S and 29°S and dominated by major reverse faults. Despite this structural disposition, the porphyry stocks, intrusive breccia bodies, gold-bearing sheeted veins,
local faults, dykes and silica ledges related to these deposits are preferably aligned along NW trends. Thus, on a local scale the Au porphyry mineralization appears to be controlled by subsidiary NW-trending faults that are especially common in the Maricunga area. These NW faults accommodate structural shortening and have sinistral strike-slip displacements at 27°S latitude (Tomlinson et al. 1994; Mpodozis et al. 1994). A similar structural setting is shared by the older Lower Oligocene gold-rich porphyry copper deposits at the La Fortuna cluster located some 50 km SSW from the Maricunga Au porphyries. Whether this specific structural setting was a controlling factor for Au-rich porphyry mineralization or just the effect of large-scale tectonic processes unrelated to mineralization cannot be resolved at this time, although gold porphyries are notably absent outside of the mentioned NNE-trending deflection of the Chilean Andes.

The Au porphyry systems of the Maricunga area are related to dioritic–quartz dioritic multiphase stocks intruded into Miocene andesitic to dacitic volcanic strata. The systems tend to be vertically elongated with a core of potassic alteration characterized by widespread biotite as dissemination, veinlets and fine aggregates forming amphibole pseudomorphs. K-feldspar occurs in veinlets and veinlet haloes, and both biotite and K-feldspar are accompanied by profuse magnetite and hematite that form up to 5 to 10% of the mineralized zone. A marginal propylitic zone occurs, mostly developed within the host volcanic rocks. The upper sections of the Au porphyries include irregular zones with assemblages of quartz–sericite–chlorite–clays or pyrite–albite–clays that may be interspersed with patches of propylitic or potassic alteration. The upper section of some deposits (Marte, La Pepa and Santa Cecilia) includes advanced argillic alteration overprinting. This juxtaposition of early deep porphyry-type hydrothermal and late shallow epithermal alteration appears to be the result of active erosion and collapse of the volcanoes that host the porphyries during active hydrothermal circulation (Sillitoe 1994). Gold-rich porphyry orebodies are characterized by the occurrence of multiphase quartz stockworks, which are either multidirectional or frequently occur as closely spaced subparallel sets or sheeted veins with a preferred NW strike. Sulphides are dominated by pyrite, with minor chalcopyrite, and traces of bornite and molybdenite. Copper content is variable, from 0.03% Cu in the Verde orebody at Refugio, to 0.26% Cu at Cerro Casale (Table 6.1). Volcanic reconstruction and fluid inclusion data indicated to Vila & Sillitoe (1991) that the gold-bearing stockwork and sheeted-vein ores were generated some 600 to 1000 m beneath palaeo-surfaces. Further fluid inclusion studies of Muntean (1998) revealed that early quartz–magnetite–biotite–chalcopyrite veinlets were formed at temperatures >500°C and at pressures estimated as between 200 and 400 bars, suggesting depths of 800 to 1600 m under lithostatic loads. In contrast, late banded quartz veinlets indicate temperatures <350°C and estimated pressures <200 bars, suggesting depths of 200 to 1500 m under hydrostatic conditions. Gold crystal composition at Cerro Casale indicates that gold was introduced both during the early potassic alteration and a late white-mica-dominated alteration (Palacios et al. 2001). Hydrothermal breccias are common within the deposits, but the only significant richly mineralized breccia occurs underground at Cerro Casale. It is a funnel-shaped fragment-supported breccia body with a vertical extent of some 300 m and an elliptical plan section up to 100 × 150 m in size. The breccia comprises subrounded intrusive fragments 5–10 mm in diameter within a matrix of anhydrite, barite, tournamite, rhodochrosite, dolomite, chalcopyrite, pyrite, galena, sphalerite, quartz and sericite.

Depths of supergene oxidation in the Maricunga Au-rich porphyries range from a few metres to a hundred metres with Cu leaching in the zone of oxidation, but constant Au. Locally, supergene enrichment blankets occur, e.g. Bema Gold reported a 5–35-m-thick chalcocite blanket at Cerro Casale averaging 1–1.3% Cu (Northern Miner, 22 April 1996), whereas at Cavancha porphyry a supergene chalcocite blanket ranging from 15 to 70 m in thickness and containing 0.1 to 0.5% Cu occurs (Muntean 1998). However, supergene copper-enriched zones are lacking in the other gold-rich porphyries.

Transitions between gold-bearing porphyry-type stockworks and shallower zones of advanced argillic alteration containing precious metal mineralization occur in Maricunga (Vila & Sillitoe 1991). Silica ledges that are replacement veins of quartz–alunite alteration with local core zones of vuggy silica host high-sulphidation epithermal mineralization. The most obvious example occurs at La Pepa where gold-bearing bonanza-type veins that strike N10°–40°W overlie and are peripheral to the low-grade Cavancha porphyry deposit (Muntean 1998). The advanced argillic alteration partly overprints the Cavancha porphyry, and geochronologic data of Muntean (1998) indicate that there is a 140 000 to 900 000 year gap (40Ar/39Ar at 95% confidence level) between the Au-rich Cavancha porphyry alteration and later advanced argillic overprint and precious-metal-bearing vein formation at La Pepa. Another example is Cerro Casale where a N50°–80°W striking vein system containing Pb, Zn, Ag, Sb and minor Au occurs immediately west of the porphyry (Vila & Sillitoe 1991). 40Ar/39Ar geochronological data from alunite in silica ledges at the Vein Zone, as well as biotite from the Cerro Casale porphyry, indicate that both were formed synchronously at 13.9 Ma (Muntean 1998).

Skarn deposits

At least 13 skarn deposits have been documented in Chile, but only a few are currently being exploited (e.g. Cabildo and El Toqui districts). Cu skarns are by far the dominant type, but local examples of Zn–Pb and Fe skarns also occur (Fig. 6.12). The most prominent are Cu skarn deposits that occur along the Coastal Cordillera of northern Chile hosted by skarnified units of Lower Cretaceous carbonate rocks intercalated within a volcanosedimentary sequence. A few examples also occur in the High Andes and are again hosted by skarnified Lower Cretaceous carbonate rocks, which were deposited in a shallow backarc marine basin. In the Patagonia area of southern Chile the El Toqui Zn–Pb skarn is the only Chilean producer of Zn, and is also hosted by a metamorphosed Lower Cretaceous marine sedimentary sequence. One unusual skarn with lazurite-rich pockets, within an outer marble facies, occurs in the High Andes of northern Chile. It provides lapis lazuli widely used as a gemstone in the country. This gemstone was extracted by native Andean inhabitants long before Spanish colonization (Rivano & Sepúlveda 1991).

The Cabildo copper district (Fig. 6.12) is one of the largest Cu skarn mineralization systems in the Coastal Cordillera, and the only one currently exploited by underground mining. The copper deposits are hosted by sedimentary intercalations within the mostly volcanic Lower Cretaceous Lo Prado Formation, immediately west of a batholith composed of diorite, tonalite and granodiorite, with K–Ar dates ranging from 106 ± 3 to 96 ± 3 Ma (Rivano et al. 1993). The limestone, sandstone, marl, tuff and conglomerate beds form a 400–500-m-thick intercalation within Lower Cretaceous volcanic rocks. The calcareous units are altered to skarn in the central and northern section of the district and comprise garnet (grossularite–andradite), clinopyroxene (diopside–heedenbergite), amphibole (tremolite–actinolite), epidote, calcite, scapolite, quartz, chlorite and titanite. Clastic and pyroclastic units are normally transformed to siliceous hornfels. North–south trending normal faults dipping 60–70° west displace the skarn units (with a sinistral lateral component) and some fault zones are mineralized. A number of stratiform, irregular and vein-type copper-bearing orebodies are hosted by the skarnified units. Stratiform orebodies are 2
orebodies occur within the garnet zone; these are 20 to 50 m thick, extending up to 160 m along-strike, and 60 m along-dip. Chalcopyrite is the main ore mineral, but pyrrhotite, pyrite, sphalerite, arsenopyrite, galena, magnetite, hematite and minor molybdenite occur in the interstices of skarn minerals. Ore deposition appears to have taken place at 115 ± 3 Ma (K–Ar in phlogopite; Ardila 1993), although Sugaki et al. (2000) reported a K–Ar date of 132 ± 7 Ma for a dioritic stock located at the east of the Panulcillo mine. Fractures and faults filled with hypogene minerals in the Panulcillo deposit are compatible with a north–south sinistral shear zone active at the time of mineralization and focusing fluid flow and ore deposition (Ardila 1993).

The La Campana district is located in the Coastal Cordillera 70 km north of Santiago (Fig. 6.12) in the southern and southwestern slopes of the Campaña hills. The Opositora, Veta Grande, Felicidad, Pronosticada, Guanaco and Mina de Hierro deposits occur within the metamorphic contact aureole of the Caleu pluton. Stratiform Cu and Fe deposits are hosted by skarnified limestone intercalations of the Lower Cretaceous Lo Prado Formation near the contact of a gabbro that is a western facies of the Caleu Pluton, located within La Campana National Park. The Opositora mine indicated resources of 6.25 million t at 1.5% total Cu, 25 g/t Ag and 8 g/t Au, which increase to 197 000 t if inferred resources are included. The Caleu pluton shows variations in lithology, mineralogy, texture and chemical composition that allow identification of three north–south trending plutonic units: Gabbro/Diorite, Tonalite and Grano-diorite zones. Its age according to 40Ar/39Ar dating ranges from 117 to 94 Ma (Parada et al. 2002), but the La Campana gabbro stock has a plagioclase 40Ar/39Ar date of 130 ± 1.5 Ma (Parada & Larrondo 1999). The skarn units are composed of garnet (gossan–andradite), pyroxene (diopside–salite), calcite and quartz. The orebodies are up to 20 m thick, extending up to 200 m along the NNW strike of the host rocks, and 100 m along the dip. They comprise disseminated chalcopyrite, pyrite, pyrrhotite, bornite, molybdenite, sphalerite, magnetite and minor galena arsenopyrite, marcasite and hematite. Gold and silver are associated with the sulphides.

The Maria Cristina Zn–Pb skarn occurs in Quebrada Galena, about 80 km south of Copiapó, and is hosted by carbonate rocks of the Lower Cretaceous Chañarcillo Group. The alteration of the carbonate host rocks at Maria Cristina involved recrystallization of the limestone to marble and the formation of coarse crystalline skarn lenses. The prograde assemblage comprises garnet (andradite Ad₃₀Gr₇₀), diopside pyroxene, magnetite and epidote. The occurrence of former pyrrhotite is evidenced by its pseudomorphic replacement by marcasite, pyrite and calcite. The mineralization consists of coarse-grained pyrite-rich massive sulphides within the carbonate rocks, and massive and semi-massive sulphides in retrograde skarn and within the matrix of coarse-grained volcanioclastic rocks. It occurs at the contact of potassically altered diorite porphyry of mid-Cretaceous age (93.6 ± 0.4 Ma, amphibole 40Ar/39Ar plateau date; Lieben et al. 2000). The ore minerals comprise sphalerite, pyrite and galena, minor marcasite and magnetite, and trace amounts of chalcopyrite and tetrahedrite. The deposit occurs at the intersection of two NE and NNE trending faults, which controlled the mineralization within folded carbonate rocks (Lieben et al. 2000).

A stratiform iron oxide deposit occurs at Cerro Bandurrias, 55 km south of Copiapó (Fig. 6.12), hosted by the Lower Cretaceous Chañarcillo Group. The ore horizon is 7 m thick, extending 1.7 km across the western slope of Bandurrias hill. Dioritic intrusives crop out in the area, and a skarn assemblage (garnet, scapolite) associated with these intrusions occurs at the same stratigraphic level that hosts the iron oxide horizon. The ore is formed of idio- to hypidiomorphic magnetite crystals, partially martitized, with local garnet bands. Although the setting of this iron deposit is strongly suggestive of skarn affiliation (e.g.
Oyarzún 2000), it was formerly interpreted to be of primary volcanicogenic exhalative origin with superposed contact metamorphism (Cisternas 1990; Espinoza 1990). Magnetite is again the dominant ore mineral at the San Cristobal skarn deposit in the Coastal Cordillera south of La Serena, although the skarn has actually been exploited for copper. Iron oxide is also abundant in the copper skarns of the San Antonio district 24 km NE of La Serena.

The Zn–Pb (Au) deposits of El Toqui are located in Chilean Patagonia about 1700 km south of Santiago, and provide the only source of zinc production in Chile. The district covers about 25 km² and contains stratiform (Concordia, San Antonio, Mallín–Mónica, Doña Rosa–Mallín Sur, Aserradero and Estatuas orebodies) and discordant vein deposits (Antolín and Zúñiga orebodies). It is hosted by skarnified Lower Cretaceous marine sedimentary rocks with volcanic intercalations, which are intruded by a mid-Cretaceous sill of quartzfeldspar porphyry altered as an assemblage of quartz, albite, K-feldspar and sericite. Whole-rock K–Ar and ⁴⁰Ar/³⁹Ar ages of the porphyry range from 100 ± 2 to 108 ± 4 Ma (Townley & Godwin 2001). North of El Toqui River, a tonalite stock also occurs. The Cretaceous succession strikes north–south to NNW–SSE, dips ENE from 5° to 25°, and shows propylitic alteration. A number of carbonate units, varying from 0.5 to 10 m in thickness, are skarnified to garnet, pyroxene, amphibole, epidote, chlorite and ilvaite. Ore minerals include marmatite (iron-rich sphalerite), galena, pyrrhotite, pyrite, arsenopyrite, tetrahedrite, chalcocite, electrum, native gold, magnetite, silver sulphosalts, native bismuth, hessite and maldonite. Veins up to 0.6 m thick and 60 m in length are exposed in the district and comprise a stockwork of veinlets of spalerite, galena, pyrite, chalcocite and pyrite, quartz and calcite. The mineralization was formerly interpreted as exhalative volcanicogenic (Wellmer et al. 1983; Wellmer & Reeve 1990), but subsequent studies have shown its skarn affiliation (Palacios et al. 1994), and the skarn model has successfully been applied in exploration in the district (Kakarieka 2002). Fluid inclusion studies have shown that the deposits were formed by two stages of hot fluid circulation: an early stage of saline fluids (11% NaCl equivalent) under temperatures of 350 to 300°C, and a later stage of more dilute fluids (4–8% NaCl equivalent) at 200 to 160°C (Palacios et al. 1994).

The only example of skarn development related to Chilean porphyry copper deposits occurs at Potrerillos (26°30′S, 69°22.9′W; Camus 1985) where the mineralized Cobre porphyry stock dated at 35.87 ± 0.21 Ma (⁴⁰Ar/³⁹Ar in hornblende; Marsh et al. 1997) intrudes a Jurassic–Lower Cretaceous carbonate sedimentary sequence. Surrounding the porphyry stock is a 150-m-wide halo, in which metamorphic and metasomatic assemblages and minor skarn are developed. Within this aureole occurs an assemblage of garnet, wollastonite, enstatite, diopside, epidote, zoisite, clinozoisite and tremolite, with pyrite and specular hematite (Marsh 1935).

The Mina Lar, a small (< 200 000 t) oxidized Cu–Au skarn deposit, occurs 20 km south of Copiapó in the Coastal Cordillera of Atacama region, northern Chile. Phelps Dodge Mining Company explored this deposit, which is hosted by skarnified Lower Cretaceous limestone strata of the Chañarcillo Group, and discovered in 1986 the underlying large, volcanic-hosted Candelaria iron oxide copper–gold deposit (Ryan et al. 1995; Marschik 2001).

At the Mantos Grandes Cu–Au mine (30°51′4.1″S, 70°33.2′W) a sequence of Lower Cretaceous marine carbonate rocks crops out. These are folded, repeated by a reverse fault (Tulahuencito fault), and intruded by diorite to gabbro stocks ascribed to the Tertiary (Mpodozis & Cornejo 1988). In a north-south elongated zone of 500 × 250 m the carbonate rocks are altered to a skarn of andradite–hematite, wollastonite-bearing marble, and silicified rocks. A number of en echelon subhorizontal orebodies (N20°E/30°W) occur within the skarn, controlled by tensional subsidiary fractures related to the Tulahuencito reverse fault. Individual orebodies are from 1.8 to 3 m thick, extend 100 m along-strike and 20 m down-dip. The mineralized rock has abundant andradite crystals 2 to 4 mm in diameter that form aggregates within calcite, and specular hematite forming discoidal crystals up to 5 cm in diameter. The main ore mineral is partly oxidized coarse-grained chalcopyrite.

The Escalones deposit is located in the High Andes of central Chile, east of Santiago (34°07.2′S, 69°57.9′W). The copper deposit is hosted by a Lower Cretaceous sequence of marine carbonate rocks and gyspum of the Lo Valdes Formation (striking NW and dipping 60° NE) that is intruded by Neogene andesite porphyry and north–south andesitic, vertical dykes 2–3 m thick. Skarn alteration occurs within an area of 800 × 100 m, with garnet, actinolite, tourmaline, quartz, calcite, zoisite, quartz, chlorite and minor sericite. Ore minerals are chalcopyrite, bornite, magnetite, hematite, pyrite, pyrrhotite and galena, and hypogene grade is 1.7% Cu (Flores 1943). Drill and channel sampling by General Minerals Corporation reported high-grade copper–gold skarn mineralization at Escalones Alto, including an 81-m channel sampled in a road cut in the ‘core area’ that yielded 1.54% Cu, 0.74 g/t Au and 9.0 g/t Ag (GNM 2000), but no further work was subsequently undertaken on the deposit.

**Metallogenic evolution**

Hydrothermal processes related to suprasubduction igneous activity (mostly felsic to intermediate plutonism) are the source of the metallic wealth of Chile. These hydrothermal ore deposits are genetically associated with a succession of north–south trending magmatic arcs that developed since Jurassic times over a poorly mineralized basement that corresponds to an accretionary prism of Late Paleozoic to Triassic age (Mpodozis & Ramos 1989). The arc-parallel ore deposit belts young progressively from Jurassic in the west to late Miocene–Pliocene in the east.

Jurassic to Early Cretaceous Andean evolution took place under an overall extensional tectonic setting characterized by a subsiding volcanic arc with structurally controlled emplacement of shallow batholiths within the nearly coeval volcanic succession. Two other important components of the tectonic setting were the development of the intra-arc, sinistral, strike-slip Atacama Fault System along the present Coastal Cordillera, and the formation of backarc marine carbonate sedimentary basins. This early period of the evolution of the Chilean Andes was strongly dominated by copper mineralization, characterized by a number of Late Jurassic and Early Cretaceous volcanic-hosted Cu(Ag) stratabound deposits and many Cu-bearing vein deposits, as well as Early Cretaceous iron oxide copper–gold deposits. In addition, Fe-oxide apatite deposits were formed during the Early Cretaceous within the domain of the Atacama Fault System, related to contact zones of Neocomian volcanic rocks with Early Cretaceous intrusions. Porphyry copper deposits were unimportant during this initial stage, except for some low hypogene grade porphyry Cu–Au deposits that were formed in the latest Early Cretaceous and related to shoshonitic porphyry intrusions (e.g. Andacollo; Reyes 1991).

A compressive tectonic setting, with basin inversion and episodic uplift, has characterized the evolution of the Chilean Andes since Late Cretaceous times. The magmatic arc migrated stepwise east ( inland), and the new geological setting along the active continental margin led to a strong dominance of calcalkaline porphyry Cu–Mo deposits which continued during Tertiary times. A belt of Paleogene–early Eocene porphyry Cu–Mo deposits extends from southern Peru down to 29°30′ latitude in northern Chile, totalling 12.7 Mt of contained copper in Chile (resources plus production; Camus 2003). However, only those porphyries with a well-developed supergene
METALLIC ORE DEPOSITS

enrichment blanket are currently economic, such as the Cerro Colorado and Spence deposits. In addition, a number of precious metal epithermal deposits, also related to volcanic rocks, occur along the Palaeocene–early Eocene outcrop in northernmost Chile, which have been preserved due to the limited extent of denudation in the Atacama Desert.

A major metallogenic episode took place during late Eocene–early Oligocene times when enormous porphyry Cu–Mo deposits were formed along the Domeyko Cordillera of northern Chile. These constitute the largest copper concentration in the world, totalling about 220.5 Mt of copper (resources plus production; Camus 2002). The mineralizing period extended from 43 to 31 Ma, though about 60% of the copper resources of the belt were accumulated from 34 to 31 Ma (Camus 2002), and hydrothermal ore precipitation was genetically associated with the closing igneous activity along this range, prior to a 30-km eastward arc migration in response to plate tectonic interaction. The late Eocene–early Oligocene porphyry Cu–Mo deposits are also spatially associated with a major intra-arc strike-slip shear system, the Domeyko Fault System, that developed during a period of NE-directed oblique convergence, transpression, crustal thickening and denudation of the Domeyko Cordillera (Maksaev & Zentilli 1999).

The main precious metal epithermal deposits of Chile are related to later Miocene volcanism that developed in the easternmost section of the Chilean Andes (extending to the east into Argentinean and Bolivian territories). Most deposits are of the high-sulphidation type with structural control of the orebodies, and occur within extensive zones of altered, bleached volcanic rocks. Miocene porphyry Au mineralization occurs in close spatial and temporal association with high-sulphidation epithermal deposits in the Maricunga area (27–28°S), particularly in the core of eroded volcanic edifices (Vila & Sillitoe 1991).

The last major metallogenic episode of the Chilean Andes was the formation of enormous porphyry Cu–Mo deposits in the Andes of Central Chile (i.e. Los Pelambres, Río Blanco–Los Bronces and El Teniente) which taken together have an estimated yield of 183 Mt of copper (resources plus production; Camus 2002). This youngest porphyry Cu–Mo mineralization followed stages of compressive deformation, crustal thickening and denudation during Miocene and early Pliocene times. Hydrothermal ore precipitation was once again genetically associated with the cessation of igneous activity along the western slope of the main Andes, prior to a 50-km eastward arc migration.

The economic viability of many Cu and Ag deposits in arid northern Chile (north of c. 30°S), especially porphyry copper deposits, is dependent on the size and quality of their supergene enrichment blanket. Supergene enrichment processes have been active there, episodically or continuously, from Eocene to Miocene times (Hartley & Rice 2005), a significant supergene metallogenic epoch in Chile.

The strong dominance and recurrence of copper deposits in the Chilean Andes probably results from a relatively homogeneous source of the metal, and the existence of a long-lived subduction-related system of magma generation. However, the distinct and relatively short duration of the major mineralization episodes within this subduction setting implies that a number of geological conditions have to combine to create conditions suitable for the generation of such anomalous metal concentrations along the Andean Cordillera.
## Geology of Chile

### Chapter 7 – Author queries

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This chapter overviews the origin, exploration and exploitation of Chilean industrial minerals (IM), the main such resources being evaporites, brines and other chemical products located in the northern part of the country. Following a general introduction to the subject, the geological processes relevant to IM generation are described, followed by more detailed accounts dealing with the evaporite deposits of the northern Chile salt flats, especially the nitrate and iodine ores that have provided the most valuable IM contribution to the Chilean economy.

Chile is administratively divided into ‘Regiones’ numbered I to XII from north to south, as well as the Metropolitan Region around Santiago, and the description of mineral deposits and mines is based on these geographical subdivisions (Fig. 7.1). Chilean industrial mineral production is concentrated in northern Chile, accounting for about 3–7% of Chilean mining exports in 2004, with a value of about US$500 million.

The exploitation of industrial minerals has been closely related to Chilean economic development, with the modern mining history of the country beginning during the late nineteenth and early twentieth centuries when an outstanding income was derived from the nitrate industry. During their heyday, these resources were in high demand because of their use in the manufacture of fertilizers and gunpowder. For a brief period in the nineteenth century Chile was also an important producer of borax.

The importance of IM resources in Chile is based on their variety and quality and also the huge nature of some of the reserves. Variety is provided by the wide geological setting and geographical diversity of the country that includes regions with climatic conditions ranging from arid and hyperarid desert environments to temperate areas with abundant rainfall. With regard to overall reserves, Chile possesses more than one-third of world lithium and iodine, as well as vast amounts of natural nitrates and gypsum/anhydrite. The exceptionally large salt deposit of Salar Grande has reserves that could supply world demand for centuries (Fig. 7.2). Chile is therefore today a world leader in the production and export of lithium, iodine and nitrate ores as well as the main producer of borates and salt (Figs 7.3 & 7.4).

Notwithstanding this important mining tradition, and major potential contribution to world demand, the development of the IM industry in Chile has been erratic and always subordinate to metallic mineral production. This is exactly the opposite of mining development in the more industrialized countries where the main investments are in industrial minerals. In the USA, for example, industrial minerals represent some 75% of total mining investment, whereas in Chile the investment is 91% in metals (mostly copper) and only 9% in industrial minerals. The reasons for this include the low industrialization levels of Chile and neighbouring countries, which tend to export only raw materials instead of more sophisticated mining products of greater economic value. Similarly, the lack of research and development in Chile does not encourage the creation of final products that require the integration of different ores (e.g. fertilizers). Geographical location, a lack of investment and little knowledge of the international markets are also important reasons for the slow development of the Chilean IM industry.

IM resources are separated into five groups according to their geological origins, reserve volumes and export potential (Gajardo 2000) (see Table 7.1).

**Group I.** This includes industrial minerals with very good to excellent geological and export potential. Lithium and potassium brines, nitrates, iodine, sodium sulphate, borates, salt and diatomite are part of this group, and these dominate the Chilean IM export market.

**Group II.** This group comprises IM resources with good to very good geological potential, combined with the ability to exploit large volumes, although with only limited export demand. They are mostly used for supplying the domestic market, especially in the construction industry and in related activities such as glass, ceramics and paint manufacture. This group includes sand and gravel, different types of clays including calcium bentonite, kaolin, quartz, feldspar, pumice and carbonate rocks.

**Group III.** In this group are industrial minerals with reasonably good geological potential but that, even when relatively abundant, can only be exploited on a small scale with minor export possibilities. Classed in this group are alunite, andalusite, apatite, sulphur, barite, calcium bentonite, phyllophyllite, ‘guano’, dimension stone, perlite and talc. Recently important titanium deposits have been discovered and have begun to be exploited.

**Group IV.** These include industrial minerals with poor geological potential, typically restricted to just one interesting deposit and with limited exploitation possibilities. With the notable exception of lapis lazuli, the minerals within this group have no export potential and are used only in domestic consumption. In this group are included dolomite, fluorite, phosphatic rocks, garnets, lapis lazuli, mica, iron oxides, wollastonite and zeolites.

**Group V.** This group comprise industrial minerals without economic potential according to present-day knowledge. They are known from the literature, and may have been sporadically exploited on, at best, a very small scale. In this group are included asbestos, calcium bentonite, chromite, graphite, magnesite, zirconium, volcanic glasses such as obsidian and perlite, and rare earth elements.
 geological processes producing Chilean industrial mineral ores (G.C.D. & A.G.C.)

The geological processes that have produced the vast majority of IM in Chile are either sedimentary or magmatic.

Sedimentary processes

These mineral deposits are produced either directly or indirectly by water and its weathering action, associated erosion, transportation and sedimentation, which together provide the main agents of Chilean industrial mineral ore formation. This general category includes mechanical (clastic), biogenic, evaporitic and chemical processes.

Ores resulting from mechanical sedimentary processes are sand and gravel, plastic and common clays, silica sands, andalusite and garnet. The main transport agents are fluvial, lacustrine and coastal marine processes, and such deposits are distributed throughout Chile, with ages ranging from Neogene to Quaternary. Sand and gravel are found as fluvial, alluvial and pediment deposits across the whole country. Common clays of lacustrine type are found in Regions I and III and between Regions Metropolitana and VIII. The main deposits of silica sands and plastic clays, of both marine and lacustrine

Fig. 7.1. Distribution of ‘Regiones’ in Chile.
origin, are located between Regions V and IX. Sedimentary deposits of garnets and andalusite can be found as placers in fluvial deposits in Region VIII.

Biogenic processes in a wide range of sedimentary environments have generated several resources. The more important of these are limestone, including shell deposits, diatomite and phosphatic rocks in shallow-marine and lacustrine environments, together with guano in supralittoral to continental environments. Limestone deposits range in age between Palaeozoic and Tertiary and are distributed across the whole country with the exception of Regions X and XI. The more important deposits of shells, diatomite, phosphatic rocks and guano, of Neogene to Quaternary age, are distributed between Regions II and VIII (shells), I and X (diatomite), II and IV (phosphoric rocks), and I and II (guano).

Evaporitic sedimentary deposits are located in the northern part of the country (Regions I, II and III) and make an important contribution to the export market in Chilean industrial minerals. In addition to these evaporites are chemical sedimentary deposits which depend for their formation on specific changes in pH, Eh, temperature, pressure and salinity rather than evaporation. Such deposits include nitrate ores and some dolomites, with the former being of especial interest because they are the only such deposits in the world. The dolomite ore deposits are Lower Cretaceous marine carbonates in Regions II and IV, as well as Neogene volcanosedimentary deposits in Region I. There are also some Triassic dolomites in the Precordillera of Region II.

Another group of IM deposits in Chile owes its origin to weathering processes such as lateritization and kaolinization: these kinds of ores require very specific climatic conditions for their formation. Generation of laterites is favoured where highly permeable and porous rocks containing water-soluble minerals crop out in areas where the climate is wet (rainfall > 2000 mm/year), vegetation cover is abundant, temperature averages over 20°C, and the topography is generally flat. These are ideal conditions under which to promote the formation of bauxite, bauxitic clay, kaolin and iron oxides. In Chile such lateritization processes occurred during Late Cretaceous to Palaeogene times, forming a particularly useful and unique clay deposit up to 10 m thick. This kaolin is used in ceramics, paints, paper and different types of rubbers.
MAGMATIC PROCESSES

Both intrusive and extrusive magmatic processes, as well as metamorphic and hydrothermal activities, have been responsible for producing a range of different IM ore deposits in Chile. In the case of intrusive activity, there are many potentially valuable outcrops of dimension stone varying from granite to gabbro in composition. They are traded as granite sensu lato and were produced during five main intrusion events (Palaeozoic, Jurassic, Jurassic–Lower Cretaceous, Upper Cretaceous–Palaeogene and Neogene). The main currently exploited deposits are located in Regions II, IV and V, most of which occur within batholithic intrusive bodies of Palaeozoic and Late Cretaceous age. Quartz, sodium and potassium feldspar and mica pegmatite deposits of various shapes and sizes are also associated with batholiths or stocks of granitic composition, ranging in age from Late Palaeozoic and Jurassic to Late Cretaceous–Palaeogene. Reserves of these different bodies vary in size from just tens to hundreds of thousands of tons, typically showing rather homogeneous compositions, especially in the distribution of silica in quartz, feldspar mineralogy and absence of impurities like iron oxides. Deposits of quartz and feldspar are known between Regions II and VIII, whereas mica deposits are hosted in Palaeozoic rocks, mainly in Regions VI to VIII.

Geothermal springs and solfataric activity in the volcanic region. These phenomena include the peculiar dripping fogs (camanchacas) that, although essentially coastal in nature, can extend inland. Geothermal springs are characterized by high temperatures and mineral content, often hot enough to boil water. They are typically found along active or recently active volcanic zones. These springs are important sources of thermal energy for geothermal power plants, which convert the heat from the Earth's interior into electricity. Geothermal energy is a sustainable and renewable energy source that can help reduce greenhouse gas emissions and combat climate change. Geothermal activity in Chile is concentrated in certain areas, such as the Atacama Desert and the Andean Cordillera, where the Earth's crust is thin and hot, allowing heat to more easily escape to the surface.

GROUP I
Very good to excellent geological potential

GROUP II
Good to very good geological potential

GROUP III
Regular to good geological potential

GROUP IV
Scarce to regular geological potential

GROUP V
No geological potential

Borates
Lithium chloride
Potassium chloride
Sodium chloride
Diatomites
Nitrates
Sodium sulphate
Iodine
Sand and gravels
Common clays
Plastic clays
Silica sands
Limestones
Kaolin
White calcium carbonate
Shells
Quartz
Feldspar
Pumice
Tuffs and ignimbrites
Gypsum or anhydrite
Alunite
Andalucite
Apatite
Bauxite clays
Sulphur
Barite
Calcite bentonite
Combarbalte
Granite
Guanu
Marble
Perlite
Pyrophyllite
Talc
Travertine

ALUMBRES (Al/Mg sulphates)
Dolomite
Fluorite
Phosphoric rocks
Garnets
Lapis lazuli
Mica
Iron oxides
Wollastonite
Zoisites

GROUP VI
No geological potential

Asbestos
Sodium bentonite
Chromite
Graphite
Magnesite
Tin oxide
Titanium
Zirconium

NORTHERN CHILE SALT FLATS (G.C.D. & A.H.)

The western slope of the Andean Orogen of northern Chile (latitude 18°–27°S) is characterized by a north–south alignment of morphotectonic blocks. Its altitude decreases from east to west, from the volcanic summit (in places over 6000 m a.s.l.) down to what are still some of the highest coastal cliffs in the world (up to 2000 m a.s.l.), dropping abruptly into the Pacific Ocean. Geomorphological units across this region are mostly enclosed (endorheic) basins separated by ranges, a geomorphology that has broadly existed at least from Oligocene times. Thus today these units define distinct tectonostratigraphic belts which from west to east comprise the Coastal Cordillera (extensional), the Central Depression (extensional and/or transpressional), the Preandean Range or Precordillera (transpressional and/or compressional), the Preandean basins, and the High Andes (Western Cordillera) or Altiplano (compressional) (Fig. 7.5).

This region is known as Norte Grande de Chile and has distinctive climatic and geological features. It is a tectonically active coastal desert, the climate of which is strongly influenced by the Humboldt current and Pacific anticyclone (Chong 1984; Hartley & Chong 2002; Houston & Hartley 2003; Hartley 2003). Extreme aridity in the west (Coastal Cordillera and Central Depression) changes eastwards to arid and semi-arid conditions (Preandean Range), with humidity increasing with altitude in the High Andes. The boundaries between these climatic belts are not only irregular but also affected by other important climatic phenomena superimposed over the whole region. These phenomena include the peculiar dripping fogs (camanchacas) that, although essentially coastal in nature, can reach more than 100 km inland. Another climatic phenomenon is the generation of winter storms (the Invierno Altiplanico; December–March) which are generated in the High Andes but, controlled by the regional slope, sometimes reach westwards as far as the coast (Chong 1984, 1988).
Although the regional base level is the Pacific Ocean, important local basins define their own base levels as they are essentially enclosed, with their waters rarely, if ever, reaching the sea. General aridity, interspersed with episodes of hyperaridity alternating with wetter periods, has been the climatic pattern for millions of years, as evidenced by the stratigraphic record throughout the Cenozoic era. The only major exception seems to have been an important wet period in Oligocene times, and possibly another still earlier in the Palaeogene period. Throughout this long history of dominantly arid conditions, active arc-related volcanism remained a constant additional influence on geomorphology, climate and sedimentation.

From a geological viewpoint the region corresponds to an active collisional continental border with high seismic activity, a magmatic–stratigraphic record ranging from Ordovician up to Holocene in age, with a magmatic belt of Miocene–Holocene age to the east. This magmatic belt includes hundreds of volcanic systems, many of which show fumarolic activity (Fig. 7.6).

Products of this volcanism, which has continued since at least Miocene times, are mostly found within intra-arc basins and the forearc area.

An older magmatic belt (Late Cretaceous–Eocene), parallel to but to the west of the Miocene–Holocene belt (Scheuber et al. 1994), was originally constructed along the area occupied today by the Central Depression and easternmost part of the Coastal Cordillera. This older volcanic belt once defined the east–west watershed, which then lay further west than today, when there was more land existing to the west of present-day Chile. In this setting for tens of millions of years the intra-arc, forearc and backarc basins have accumulated saline and siliciclastic material produced by both magmatic arcs (e.g. Scheuber et al. 1994). Isotopic information has shown that many of the salts in these evaporitic basins have a volcanic genesis (Spiros & Chong 1996). One consequence of this geological setting is the resulting high salinity in the present-day Atacama Desert and nearby regions, with huge volumes of salts leached from volcanic soils and rocks, as well as those deposited directly in the basins. The high mobility of these salts, the presence of other evaporites in the region (such as Mesozoic gypsum/anhydrite deposits), the climatic conditions, and the topographically enclosed geomorphology together have produced a classic example of what has been described as a ‘saline domain’ (Chong 1988). This whole regional saline system, further enhanced by the action of marine spray and dripping fogs, creates exceptional conditions in which salts are found everywhere: in the rocks, in the hydrological system, and even as abundant wind-blown particles in the atmosphere. In the rocks these salts are seen in soils and palaeosols, as saline crusts and efflorescences, in the porosity and structural open spaces in rocks, as saline horizons in alluvial and lacustrine sedimentary successions and in the filling of evaporitic basins. Within the water system, many of the superficial and underground waters of the Atacama Desert are brackish or brines. Finally, salty, marine-derived airborne particles are abundantly present in the dripping fogs (camanchacas), and they contribute to saline soil formation (Rech et al. 2003b).

The most obvious expression of this highly saline domain is the presence of salt flats or salars that correspond to evaporites formed in closed (endorheic) basins, in geomorphological traps due to volcanic rock deposition, in tectonic depressions, or in the distal part of coalescent alluvial fan systems. During the early sedimentary history of these basins perennial to ephemeral lakes are commonly formed, but the high evaporation rates later exceed inflow so that the water bodies evolve to become

Fig. 7.5. Geomorphological sketch of the Andean orogen in northern Chile, looking to the north. 1, Pacific Ocean; 2, Coastal Cordillera; 3, Central Depression; 4, Preandean Range; 5, Preandean Basins; 6, Puna (Altiplano = High Andes = Cordillera Principal).

Fig. 7.6. Panoramic view showing close relationship between saline basins and active volcanism in the High Andes. In the foreground an active fumarole (Volcán Lastarria, 25°09’S, 68°31’W) and in the lowest part of the basin a salt flat at the foot of the volcano. The basin receives all the volcanic material from the volcano and can be considered as part of the volcanic system.
playa lakes or salt flats. Past repetition of similar conditions, in particular during late Tertiary times, has produced multiple evaporitic episodes, even within the same basins (e.g. Salar de Llamara; Fig. 7.7)

A salt flat, or salar, is a body of salt and siliciclastic sediment in the lowest part of an endorheic basin. It is common to find a range of different evaporite bodies in the same basin, as well as interconnecting lakes with brackish water or brines. These rather complex hydrological systems can involve the exchange of waters so that sudden salinity changes can occur (Chong 1984). The surface of the salt flats can be dry, wet or saturated saline crusts and/or efflorescences that change when the basin is flooded (Fig. 7.8). Inflow is mostly from underground water, minor drainage systems and springs which, close to volcanoes, can be of thermal origin. In saline crusts a number of saline structures can be seen, including common desiccation polygons with the cracks infilled with salts by evaporation of the shallow water table, and sink holes that become lakes (Fig. 7.9). A long-established salt flat like the Salar de Atacama can show concentric zonation with sedimentary outer rings displaying playa lakes, then inner zones with salts interfingering with other sediments and, in the central and lowest part, rings of less soluble salts (sulphates) surrounding a core containing the most soluble salts (chlorides) (Fig. 7.10). In contrast, younger salt flats, or those associated with coalescent alluvial fan systems, are subject to more frequent flooding which dissolves and reprecipitates salts and crusts to produce a more unstable and complex mixture of detrital and saline units. Despite their extreme salinity, salars are oases for an abundant macrofauna and specialized flora (xerophytes, halophytes), as well as including an amazing amount and diversity of microorganisms. These very fragile environments are seriously threatened by abusive water exploitation (Figs 7.11, 7.12 & 7.13).

Salars can show a progressive fossilization with time, a good example of this being provided by the Salar Grande, which is the only evaporitic basin of the Coastal Cordillera and today receives practically no inflow (Fig. 7.10). Salt flats in the process of fossilization are seen in the Central Depression where they still receive inflows from the High Andes (Fig. 7.7). Fully functional salt flats, receiving abundant meteoric (and locally juvenile) water inflows, are seen today only in the High Andes. Bearing in mind that all recharging inflows are derived from the east, and that aridity is increasing westwards, a gradual fossilization of salt flats from west to east can be expected. Given the complex diachronity of salt deposition and diagenesis in these systems, it is perhaps not surprising that salt flat classification has tended to be based on purely geographic location (Chong 1984, 1988) without considering age, composition of brines, economic importance, type of crusts, biodiversity, degree of
fossilization, or any other factors. Thus the Coastal Cordillera has the Salar Grande (Fig. 7.10), the Central Depression has a range of salt flats of different ages, composition and characteristics (Fig. 7.7), the Preandean salt flats comprise two main evaporitic bodies hosted in tectonic basins between the Altiplano and the Preandean Range of Domeyko (Figs. 7.8, 7.14), and, finally, the easternmost belt of evaporitic basins include fully functional Andean saline lakes and flats (Figs. 7.15, 7.16).

The main anions in the brines and crusts are sulphates, chlorides, minor borates and rarely carbonates in combination with sodium, calcium, magnesium and potassium. Common minerals are gypsum, halite, bloedite, thenardite, mirabilite and ulexite (Table 7.1) and brines are mainly enriched in sulphates.
and chlorides. The nearby presence of active volcanism, with associated high geothermal gradients, allows the concentration of unusual salts, notably borates and lithium/potassium chlorides. In other cases, inflows with high silica content are favourable for the development of diatomites.

From an industrial minerals viewpoint, the economically valuable Group 1 deposits (Table 7.1) are Neogene nitrates and iodine (see next section), as well as the borates and chlorides described below. Other resources, in the saline domain of northern Chile include gypsum/anhydrite, diatomites and sodium sulphate.

**Borates**

Borates in Chile are related exclusively to salt flats, in particular the Andean salars where they appear in layers (mantos) called barras or as nodules called papas (potatoes). The barras are interbedded with fine-grained sediments close to the surface and overlie a very shallow water table (Fig. 7.16). Their typical thickness is several decimetres, although locally they can exceed 2 m. Other sources of borates are the brines associated with the Salar de Atacama. The only economic borate mineral is ulexite, a sodium/calcium borate used in boric acid manufacture and also in fertilizer mixtures. Chile is the second most important borate producer in South America.

**Lithium chloride**

This appears in high concentrations (up to 0.14%) in brines in the Salar de Atacama. The richness of this IM ore can be appreciated by comparing these concentrations with those of other world deposits such as Salar de Uyuni in Bolivia (0.025%), Salar de Hombre Muerto in Argentina (0.07%), and Silver Peak in the USA (0.025%). Chile exports lithium carbonate and chloride and is the world leader in these unusual mining operations, producing the equivalent of 8000 tonnes of metallic lithium per year.

**Potassium chloride**

This is also present in the Salar de Atacama brines and the main commercial product is potassium sulphate, currently sold only in the domestic market.

**Sodium chloride (salt)**

Salar Grande is a fossil salar some 35 km long, 3–4 km wide, and with a depth of between 80 and 120 m. The fill is exclusively halite with grades over 90% NaCl, being virtually free from other salts (sulphates, for example, appear only as traces) or interfingering sedimentary horizons. The quality and volume of reserves of this Salar, plus its relative proximity to the sea and the presence of a specialized harbour, make this one of the best salt ore deposits in the world. Its yearly production is around 5 million tonnes of industrial salt, chemical salt and salt for roads (Fig. 7.17).

**Gypsum/anhydrite**

Reserves of these resources are so huge they are almost inexhaustible, but they are currently exploited only on a small scale to serve domestic demand. The main deposits correspond to marine evaporites of Jurassic age.

**Diatomites**

These are exploited from lacustrine–volcanic successions of Neogene age, mostly in the High Andes, and are exclusively continental.

**Sodium sulphate**

This salt is a by-product of the nitrate industry (see below). It is also exploited on a small scale from horizons of thenardite associated with saline soils (Fig. 7.18). Similar soils are associated with low grade ulexite horizons, but are not currently exploited (Fig. 7.19).

**Nitrate and iodine ore deposits (G.C.D.)**

Chilean nitrate and iodine ore deposits have been exploited for more than 150 years, with the main commercial product, sodium nitrate, being called salitre. These deposits are world
class in terms of economic importance, and geologically exceptional because they are a water-soluble salt assemblage that has remained stable over a geological timescale. Furthermore, some of these salts are rarely found in nature, and certainly not in the volume and economic grade seen in Chile. Key publications on these ores begin with Gandarillas & Ghigliotto (1908), followed by Ericksen (1963, 1979, 1981, 1983), Chong (1991, 1994) and Chong & Pueyo (1991, 1992). In addition there are several important publications, mostly very recent, dealing with specific aspects of the deposits, notably their genesis, mineralogy, and relationship with saline soil formation (Searl & Rankin 1993; Collao et al. 2000, 2002; Bohlke et al. 1977, 2003; Urbansky et al. 2001; Spurlock et al. 2003).

Over the last 50 years the Chilean nitrate industry has attracted not only a huge number of publications but also its own mining–technological terminology. Terms still in use include the following: caliche is the nitrate ore with associated salts (Fig. 7.20); pampa is used to describe the places where salts immediately underlie the soils but do not crop out (Fig. 7.21); calicheras means mine workings (Fig. 7.22); oficina salitrera is the setting around which is grouped the administration, lodge camps, beneficiation plants and spoil dumps (the latter being

Fig. 7.19. Saline soil including a ulexite horizon (Cerro de la Joya, 21°51'S, 69°23'W). These horizons were once exploited but the ulexite grade was so low that the operation became uneconomic.

Fig. 7.20. Blasted and rotated block including caliche (hammer zone) and overlying crust (saline conglomerates). Enlarged part shows detail of black caliche.

Fig. 7.21. ‘Pampa’, where nitrates immediately underlie the detrital desert soil but do not crop out. In this case relief is flat but in the nitrate industry lexicon it is not relevant.

Fig. 7.22. ‘Calichera’ (mine workings). The exploitation zone is to the right whereas to the left are the different levels where the caliche was stored (‘canchas’) and in the extreme upper left a stock of caliche is piled.
referred to specifically as *torta* (Fig. 7.23); *repasos* is the word used to include all material that was left behind in old workings and that can be exploited now through heap leaching, recovering nitrates and iodine from ore formerly considered to be too low grade.

There are few trustworthy records documenting the number of *oficinas salitreras* historically active in the nitrate business, with figures ranging from 'more than one hundred' up to 350–400. Today only two companies are actively mining new ore, with another two using *repasos* as ore, and the only active *oficina* is María Elena (Fig. 7.24 and Chapter 13). Commercial products traded today are sodium and potassium nitrates, elemental iodine and some iodine compounds, salt mixtures used as fertilizers and sodium sulphate as a by-product. Yearly production is about 1 million t of nitrate, some 10–11 t of iodine, and sodium sulphate production in recent years has been around 60 000 t. These figures place the country at the head of natural nitrate and iodine production in the world: in fact, given the relative value of the ores, the mining activity could more accurately be referred to as the Chilean iodine industry, with nitrates relegated to a coproduct.

**Ore deposit location and geological setting**

Nitrate and iodine ores are distributed along a north–south belt stretching for some 700 km (latitude 19°30′–22°24′S) through the Chilean desertlands, with a relatively narrow but variable width reaching some tens of kilometres between longitude 69°30′ and 70°30′ (Fig. 7.25). The main ores lie in the areas of the former *oficinas salitreras*, although the deposit as a whole may be considered as continuous because between these former

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**Fig. 7.23.** Aerial view of the former oficina salitrera Chacabuco, Segunda Región de Antofagasta (23°08′S, 69°37′W). In the foreground are the ‘calicheras’ and to the left the spoil dumps.

**Fig. 7.24.** Oficina María Elena in the Segunda Región de Antofagasta (22°20′S, 69°40′W) is the active oficina salitrera *sensu stricto*. The large stocks in the foreground are salitre.

**Fig. 7.25.** Nitrate belt location in northern Chile.
mines there are unexploited low grade and, probably, undiscovered deposits. The belt is interrupted by alluvial, aeolian, lacustrine and fluvial sequences deposited under mostly arid conditions over Pleocene–Pleistocene time. A lack of ore continuity to the south is associated with wetter palaeo-climatic conditions unfavourable to the preservation of these deposits, and with the development of fluvial systems that have destroyed former deposits. The northern limits of the nitrate belt are also coincident with late Tertiary fluvial systems, although here there is the additional influence of the destruction of land areas by subduction erosion.

From a geomorphological viewpoint the belt partly occupies two of the main morphotectonic units of northern Chile: the westernmost part of the Central Depression, and the easternmost part of the Coastal Cordillera. However, isolated deposits can be found reaching westwards almost to the present-day coastal cliff, and eastwards to the western border of the Precordillera. More specifically, the ore deposits are located between the main branches of the Atacama Fault System to the west and the Domeyko Fault System (West Fissure) to the east. The ore belt is broadly coincident with the palaeo-geographical location of the Late Cretaceous–Eocene Volcanic Arc sensu Scheuber et al. (1994). Even when host rocks of nitrate deposits are of different ages and lithology, they are always related in some way to a key unit referred to as the Chastic Saline Series (CSS: late Oligocene to Neogene). The CSS comprises mostly siliciclastic and volcaniclastic sandstones and conglomerates produced by the erosion and resedimentation of rocks from the pre-existing Late Cretaceous–Eocene volcanic arc. This key stratigraphic unit includes rocks deposited under a range of sedimentary environments that include fluvial, aeolian, lacustrine and alluvial, but all of which were developed under mainly arid conditions. The upper parts of the CSS includes lacustrine rocks and evaporites mainly composed of sulphates and chlorides. The CSS outcrop is always located to the west of the former Late Cretaceous–Eocene volcanic arc, blanketing the present-day topography.

The ore deposits

The ore, or caliche, comprises an insoluble siliciclastic component, the grain size of which varies from clay to gravel, and various salts. The latter include both geologically commonplace minerals (sulphates, chlorides, borates, and, more rarely, carbonates) as well as rare and interesting varieties of nitrates, iodates, chromates, dichromates, chlorates and perchlorates, with the main cations being Na, Ca, K and Mg. Reserves of these deposits are estimated to reach hundreds to thousands of millions of tons. The saline paragenesis, with the minerals precipitating from former brines, fills any free space available in the main volcaniclastic rocks. The geometry of mineralized bodies mostly forms either stratiform layers (manto) or stockworks of veins and veinlets, and their maximum depth is about 15–20 m in exceptional cases. Salts precipitate according to their relative solubility in the following sequence: silicates (zeolites), calcite, calcium, sodium, potassium and magnesium sulphates, sodium, potassium and magnesium nitrates and sulphates, chromates, borates and perchlorates (Pueyo et al. 1998). The resulting mineralogy is complex and several completely new minerals have been described over the last thirty years (Ericksen et al. 1974, 1986, 1989; Mrose et al. 1970). Table 7.2 lists the assemblage of salts from a typical Chilean nitrate deposit, their identification having been achieved using a combination of optical, XRD, and SEM–EDS analytical methods.

<table>
<thead>
<tr>
<th>Chlorides</th>
<th>Halite</th>
<th>NaCl</th>
</tr>
</thead>
<tbody>
<tr>
<td>Anhidrite</td>
<td>CaSO₄</td>
<td></td>
</tr>
<tr>
<td>Bloedite</td>
<td>Na₂Mg(SO₄)₂*4H₂O</td>
<td></td>
</tr>
<tr>
<td>Glauberite</td>
<td>Na₂Ca(SO₄)₂</td>
<td></td>
</tr>
<tr>
<td>Gypsum</td>
<td>SO₄*2H₂O</td>
<td></td>
</tr>
<tr>
<td>Lowelite</td>
<td>Na₂Mg(SO₄)₂*15H₂O</td>
<td></td>
</tr>
<tr>
<td>Polihalite</td>
<td>Ca₂Mg₆(SO₄)₂*7H₂O</td>
<td></td>
</tr>
<tr>
<td>Starkeyite</td>
<td>MgSO₄*4H₂O</td>
<td></td>
</tr>
<tr>
<td>Thennardite</td>
<td>Na₂SO₄</td>
<td></td>
</tr>
<tr>
<td>Carbonates</td>
<td>Calcite</td>
<td>CaCO₃</td>
</tr>
<tr>
<td>Nitrates</td>
<td>Nitratine</td>
<td>NaNO₃</td>
</tr>
<tr>
<td>Nitre</td>
<td>KNO₂</td>
<td></td>
</tr>
<tr>
<td>Darpskite</td>
<td>Na₂[(NO₃)₂(SO₄)₂]²*H₂O</td>
<td></td>
</tr>
<tr>
<td>Humberstonite</td>
<td>K₂Na₂Mg₂[(NO₃)₂(SO₄)₂]²*6H₂O</td>
<td></td>
</tr>
<tr>
<td>Borates</td>
<td>Kaliborite</td>
<td>KMg₂H₂[BO₃(OH)₆]²*4H₂O</td>
</tr>
<tr>
<td></td>
<td>Probertite</td>
<td>NaCa[B₂O₃(OH)₆]²*3H₂O</td>
</tr>
<tr>
<td>Iodates</td>
<td>Lautarite</td>
<td>CaIO₃</td>
</tr>
<tr>
<td></td>
<td>Hectorfloresite</td>
<td>Na₆[(IO₃)₂(SO₄)₂]²</td>
</tr>
<tr>
<td></td>
<td>Fuenzalidade</td>
<td>K₄[Na₂,K₆][Mg₈,[SO₄]₁₂[IO₃]₁₂]*12H₂O</td>
</tr>
<tr>
<td>Chromates</td>
<td>Dietzite</td>
<td>Ca₆[(IO₃)₂CrO₄]</td>
</tr>
<tr>
<td>Oxides</td>
<td>Hematite</td>
<td>Fe₂O₃</td>
</tr>
<tr>
<td>Silicates</td>
<td>Quartz</td>
<td>Na₂Mg₆[(SO₄)₂]₂[IO₃]₁₂</td>
</tr>
</tbody>
</table>

Data from Pueyo et al. (1998)

Sedimentary deposits

These ores form the most important of the nitrate deposits and are consequently exploited more than the others, not least because of their substantial volumes (Fig. 7.26). They are hosted in the CSS, the ore of which is known as Caliche negro, with lithologies ranging from breccias, conglomerates and alluvial sandstones to various kinds of volcaniclastic rocks (Fig. 7.20). All these rocks have a light to dark tan colour, and miners have separated sedimentary units which, from top to bottom, are called chusca (and/or losa), costra, caliche, and a basement unit called conjelo when cemented and coba when friable. Caliche is very well cemented by salts that cannot be detected with the naked eye: primary porosity and all type of cavities and microfractures are filled with salts. Besides Caliche negro, a second ore type named Caliche blanco is recognized and corresponds to a high grade ore composed almost exclusively of nitratine and halite that may have either a massive, translucent aspect or can form fibres when it fills fractures. Other, more exceptional caliches have been described in the literature, such as blue or violet caliches (due to radioactive decomposition of the salt in halite), and yellow-orange caliches (due to the presence of glauberite or iodates and chromates). Caliche layers (mantos) can be up to 2 m thick, the average being 1.0–1.5 m, although there are many exceptions to this rule. The distribution of nitrate grade is irregular, as are vertical variations in caliche type.

Deposits in rocks

These ores occur as layers (mantos), veins and veinlets of white caliche hosted in zones of structural weakness: faults, joints and bedding planes (Figs 7.27, 7.28 & 7.29). Their thickness can vary from millimetres up to more than 1 m, and their horizontal distribution can be thousands of square metres, in several cases reaching tens of thousands of square kilometres. The host rocks

| Table 7.2. Recognized minerals in caliche from Pedro de Valdivia ore deposit |
|---------------------------|-----------------|-----------------|
| Salt                      | Mineral         | Chemical formula |
| Chlorides                 | Halite          | NaCl            |
| Anhidrite                 | CaSO₄           |
| Bloedite                  | Na₂Mg(SO₄)₂*4H₂O |
| Glauberite                | Na₂Ca(SO₄)₂     |
| Gypsum                    | SO₄*2H₂O        |
| Lowelite                  | Na₂Mg(SO₄)₂*15H₂O |
| Polihalite                | Ca₂Mg₆(SO₄)₂*7H₂O |
| Starkeyite                | MgSO₄*4H₂O      |
| Thennardite               | Na₂SO₄          |
| Carbonates                | Calcite         | CaCO₃           |
| Nitrates                  | Nitratine       | NaNO₃           |
| Nitre                     | KNO₂            |
| Darpskite                 | Na₂[(NO₃)₂(SO₄)₂]²*H₂O |
| Humberstonite             | K₂Na₂Mg₂[(NO₃)₂(SO₄)₂]²*6H₂O |
| Borates                   | Kaliborite      | KMg₂H₂[BO₃(OH)₆]²*4H₂O |
|                          | Probertite      | NaCa[B₂O₃(OH)₆]²*3H₂O |
| Iodates                   | Lautarite       | CaIO₃           |
|                          | Hectorfloresite | Na₆[(IO₃)₂(SO₄)₂]² |
|                          | Fuenzalidade    | K₄[Na₂,K₆][Mg₈,[SO₄]₁₂[IO₃]₁₂]*12H₂O |
| Chromates                 | Dietzite        | Ca₆[(IO₃)₂CrO₄] |
| Oxides                    | Hematite        | Fe₂O₃           |
| Silicates                 | Quartz          | Na₂Mg₆[(SO₄)₂]₂[IO₃]₁₂ |

in pre-existing rocks, although there are some exceptions to this, as described below under ‘Miscellaneous’ (Erickson 1983; Chong 1994).
can be virtually any lithology or age (Palaeozoic to Tertiary). The filling is white, translucent to opaque nitratine plus halite that frequently appears as fibres up to 1 m long with crystals up to 0.5 cm in size. Ericksen & Mrose (1972) have described associated veins of halite and bloedite. In the case of the manto layers, these are sometimes clearly linked via fractures (Fig. 7.9), sometimes with the appearance of having been injected under hydraulic pressure, an effect that can also result from salt recrystallization (Fig. 7.30). The nitrate grade can be very high and in exceptional cases it reaches almost 100%.

Fig. 7.26. Profile of nitrate exploitation in a calichera. In this case, caliche is located close to the feet of the person serving as scale. The rest of the sequence does not represent any of the units described in the literature and can be designated simply as overburden.

Fig. 7.27. Vein of massive white caliche. Composition is almost pure nitratine with less halite.

Fig. 7.28. Thick caliche blanco manto exploited in underground mining. Oficina California, Región de Tarapacá (19°40’S, 70°03’W). In this case the caliche blanco displays long fibres demonstrating fracture infilling.

Fig. 7.29. Thin manto of caliche blanco hosted in limestone of Jurassic age. The caliche blanco was injected into open spaces between sedimentary planes. Mantos are linked by fractures.

Fig. 7.30. Fragmentary and fractured rock included in caliche blanco. This can be interpreted as due to hydraulic pressure at the time of injection. Another interpretation invokes rupture due to crystallization processes.
Minerals are mainly nitratine and halite with minor sulphates (less than 3%) and iodates, chlorates, potassium, calcium and magnesium borates. Recently veins of pure humberstonite has been found hosted in ryholitic ignimbrites in the Oficina Domeyko (23°47'S, 69°21'W).

Miscellaneous deposits

Eriksen (1983) has described anomalous concentrations of nitrates in evaporitic basins that were exploited in the past, but although once exhausted, could be ‘harvested’ again some years later because salts had been redepósited within them. The mechanism for this phenomenon involves leaching of nitrate deposits by underground waters, consequent enrichment of brines, transportation to closed basins under an arid climate and, finally, the enrichment of crusts due to capillary action of brines. The only example of this known from the nitrate belt occurs in the salt flat known as Salar del Carmen, (23°38'S, 70°17'W), but it is not currently of any economic importance.

Another type of ore deposit, known as ‘mud dykes’, occurs in fractures that have been filled with detrital sediments (and sometimes volcanic ash) from the top downwards, and cemented with nitrate-rich salts. A distinctive type of mud dyke is infilled with fine-grained, brown-red sediments, a high grade of nitrate ores have been forming in the Chilean Andes. Within the upper part of the sequence places the rocks on the Cretaceous–Tertiary boundary, emphasizing the length of time over which nitrate preservation.

The filling process can be repeated several times, as demonstrated by sedimentary lamination parallel to the fracture margins. These dykes can be seen cutting both the CSS as well as underlying basement rocks.

In the Cerro Birrete area (23°01'S, 69°40'W) Chong (1994) has described a nitrate-rich lacustrine–volcanic sequence cut by a geothermal system that has produced severe hydrothermal alteration, with manganese and travertine veins and a dense sulphate stockwork. A radiometric age of 64 Ma from a lava in the andesite stockwork. A radiometric age of 64 Ma from a lava in the volcanic sequence produces the rocks on the Cretaceous–Tertiary boundary, emphasizing the length of time over which nitrate ores have been forming in the Chilean Andes. Within this system lacustrine sediments, clays from palaeo-geyser conduits, salt relicts and saline crusts all contain extremely high grades of nitrate (higher than 25%) and iodine (higher than 2000 ppm). The interpretation for this previously unrecognized type of deposit involves former brines pervading the system and becoming enriched in nitrates and iodine.

Among other ‘oddities’, Eriksen (1981) mentioned a little-known nitrate deposit associated with the Miocene volcano Maricunga (27°00'S, 69°13'W). Also, in the Oficina Domeyko (23°47'S, 69°21'W), another type of deposit is currently under investigation: hosted in Lower Tertiary volcanic rocks it comprises mainly veins, mantos and irregular bodies of white caliche containing unusual veins of pure humberstonite and nitratine-halite.

Ore genesis

The genesis of the nitrate deposits has been the subject of scientific discussion since Darwin’s comments about them in the nineteenth century. In 1920 Whitehead stated ‘Many quite divergent hypotheses on the origin of the deposits have been suggested with the universal acceptance of none’, a comment that remains valid today. A brief overview on ideas on ore genesis is provided below.

More than 30 theories have tried to explain the genesis of these strange nitrate deposits which, according to Eriksen (1983), ‘...are so extraordinary that, were it not for their existence, geologists could easily conclude that such deposits could not form in nature’. Theories include inorganic as well as organic possibilities, but none satisfactorily provides all the answers. Chong (1994) has thus argued that ‘Nitrate genesis cannot be addressed as a single question. Several aspects must be considered’, this presumably being the case because there are several processes at work to produce the ores. In recent years some publications have emphasized the importance of isotopic ratios to explanations of nitrate genesis (e.g. Bohlke et al. 1997, 2003), arguing for a probable atmospheric origin. However, such an approach has so far failed to explain the source of the full suite of associated elements, such as iodine. Even when an atmospheric genesis for the nitrates can be convincingly demonstrated, this still does not identify the elemental source (marine, terrestrial, volcanic), only that the elements have oxidized in the atmosphere via ozone (Admunson, pers. comm.). Any working hypothesis that fully explains the origins of the Chilean nitrate ores needs to elucidate the formation of the different types of caliche, the way the elements have become concentrated, the post-depositional geological evolution, as well as how and why they have been preserved.

Three key points are relevant to the discussion.

1. There was a common genesis for most of the elements via volcanic activity in the latest stages of the Late Cretaceous–Eocene volcanic arc and/or in the initial phases of the Miocene–Holocene arc. The presence of such huge amounts of normally extremely rare elements must be linked to the subduction-dominated geotectonic setting of this Andean region in which there are planetary-scale anomalous enrichments in copper, molybdenum, rhenum, gold, tin, lithium and borates (e.g. Mather et al. 2003; Snyder & Fehn 2001)

2. The concentration and deposition of these unusual elements and compounds, together with more common ones such as sulphates, chlorides and borates, have occurred in volcanic rocks and soils over geological timespans (tens of millions of years). The Tertiary climate of this region has been persistently arid, punctuated by episodes of extreme aridity and a few periods of higher than normal rainfall (Alpers & Brimhall 1988; Hartley & Chong 2002; Hartley 2003; Houston & Hartley 2003). During the latter, the action of enhanced amounts of runoff promoted oxidative weathering, transportation, deposition and then concentration through evaporation. Brines transported down the regional slope via shallow drainage systems infiltrated the ground surface, blanketing the topography and depositing salts from the surface downwards. Meanwhile, concentrated brines moving between the CSS and older rocks beneath formed a network of veins and mantos of white caliche.

3. After deposition in an arid to hyperarid environment, the nitrate ore units were capped by further volcanic and sedimentary input into the basins, with increasing aridity generating more evaporitic salt, lowering the water table, and thus aiding nitrate preservation.

Fig. 7.31. Blasted rock block in nitrate workings (‘calicheras’). The darker tabular body to the right with the hammer as scale is a mud dyke enriched in nitrate.
Summary

This chapter has provided an overview of the origin, exploration and exploitation of Chilean industrial minerals. A wide range of different types of industrial minerals are present within Chile that have formed from a diverse suite of processes related to either sedimentary, metamorphic and/or igneous activity. The most important industrial minerals in terms of their contribution to the Chilean economy are the evaporite deposits of the salt flats of northern Chile, particularly the nitrate and iodine ores. The process of formation of these ores is still debated but must be strongly linked to the subduction-zone setting and the arid climatic regime that has prevailed in northern Chile for millions of years.

G. Chong thanks FONDECYT Project 1030441 for funding his research on northern Chile Cenozoic basins related to industrial minerals including nitrate deposits.
Chilean water resources

JOSÉ F. MUÑOZ (coordinator), BONIFACIO FERNÁNDEZ, EDUARDO VARAS, PABLO PASTÉN, DIEGO GÓMEZ, PABLO RENGIFO, JAIME MUÑOZ, MESENSA ATENAS & JUAN C. JOFRÉ

This chapter examines the main characteristics of surface waters and groundwater deposits in Chile. The extreme variations in Chilean climate are reflected directly by huge differences in hydrological conditions, from the deserts of the north to the temperate rainforests of the south. The mountainous geomorphology, the presence of major basins, and the influence of the Pacific Ocean and the South Pole on oceanic currents and air masses, all also affect water distribution across the country. Anthropogenic demand for water resources, mostly for municipal wastewater, industry and agriculture, has created pollution problems that are currently being dealt with by application of new environmental legislation. Such problems are particularly acute in the north, where scarce water deposits, already commonly contaminated by naturally occurring metalliferous deposits, have been affected by extensive mining operations. In the centre of the country, where most of the population lives, the main challenges to a high quality water supply have been more associated with treating municipal wastewater. Further south, threats to clean water resources are often associated with effluents from cellulose plants and aquaculture.

Geographic background

The striking variations in Chilean climate from north to south, and therefore water supply, stem mainly from geographic location, topography and atmospheric circulation. The importance of geographic location derives from the position of the country along the southwestern side of the South American continent, with a corresponding strong influence from the Pacific Ocean and the South Pole. Movements of Antarctic and Subantarctic water currents and masses of polar air affect the whole country. Similarly, the spectacular topography of Chile has a dominant influence on climate and precipitation patterns. The mountains of the Andes and Coastal Cordillera form geographic barriers that block the maritime influence of the Pacific Ocean on their eastern slopes and in the Central Valley. Superimposed on these two main mountain ranges are transverse belts running from the Andes to the sea, and these help segment the country into both broad climatic zones and more local microclimatic areas.

The third major influence on Chilean climate, namely atmospheric circulation, mainly involves the impact of the South Pacific anticyclone, located in front of the Chilean coast between 20°S and 40°S latitude, in the proximity of the 100°W longitude meridian. The anticyclone shows an annual shift of 10°S latitude from north to south, reaching its southernmost location during the summer (January–March). This high-pressure weather system induces a descending flow of warm dry air from the upper atmosphere, generating the clear skies characteristic of the northern part of the country. When the anticyclone shifts towards the south it blocks the path of storm fronts moving towards central Chile. The occurrence of major climatic events, such as long droughts or large floods, is strongly controlled by the location and persistence of this South Pacific anticyclone, and is also influenced by El Niño/Southern Oscillation (see Chapter 11).

Climatic zones and hydrographical regions

Chile can be divided roughly into five distinct climate zones from north to south (Toledo & Zapater 1989), each of which has characteristic precipitation and temperature patterns and geographic distribution of principal basins (Fig. 8.1). In addition to the background climatic conditions, geomorphology imposes restrictions on the geographic homogeneity of the land, and conditions both the behaviour of, and the demand for, the surface water resources. The coastal geomorphology, the Andean mountains and the presence of many valleys, have historically controlled the location and size of cities, other populated areas and agricultural zones.

The northern zone (Norte Grande), from 17°S to 27°S latitude, is mostly desert, and permanently subjected to warm anticyclonic influences. The Atacama Desert extends from the coast to the central areas, where almost no rainfall occurs and very low air humidity is common. Along the eastern part of this region, which comprises the high Altiplano plateau, there is some rainfall from December to March (200–300 mm/year) due to the influence of the Amazon River basin. In this zone, rivers are ephemeral and the hydrology is conditioned by the desert landscape and the low rainfall they receive. Although there are some exorheic drainage systems, with sporadic or permanent surface flow, most water resources occur within endorheic (closed) systems which are widespread and abundant. The internal drainage and evaporative nature of such closed systems has produced many closed playa lakes or salares, some of which have evolved to arheic (or inactive) basins like those located in the Atacama Desert.

Moving south from the northern desert, the next climatic zone (Norte Chico) is semi-arid (between 27°S and 32°S latitude). This is a zone of transition, with winter rainfall but with substantial annual variations in precipitation quantities. During a dry year, the conditions are similar to those in the northern regions, while in a relatively rainy year, the climate resembles that further south.

South to the Aconcagua River, as far as the Imperial River (between 32°S and 38°S latitude), there is a zone of Mediterranean climate, known as the Central Zone. Rainfall here is concentrated during the cold winter season, while the summer season is of a dry character, and rainfall annual variation decreases from north to south. The main geomorphologic characteristic of the Central Zone is the Andean Cordillera, which has a greater height and width than further north, resulting in less available land for agriculture, but is correspondingly able to store important water reserves in the form of snow in the high peaks. In this zone appears the Central Valley, separating the Andean Cordillera from the Coastal Cordillera, where there
is space for agriculture on fertile soils located on the floodplain banks of rivers.

South of the Imperial River, as far as the Gulf of Reloncavi (between 38°S and 42°S latitude), the climate is humid and temperate (South Zone). Progressively increasing rainfall is caused by constant frontal perturbations moving across the area as the influence of the Pacific anticyclone diminishes southwards.

Finally, from the Chacao channel down to Cape Horn (between 42°S and 56°S latitude) the climate is cold and humid, with relatively consistent constant rainfall during the whole year (Austral Zone). This area is subdivided into the Pacific side, with very cold weather and strong rainfall (annual average of over 3000 mm), and an eastern side, with considerably less rainfall (average of 300 mm/year). The region is characterized by archipelagos and a narrow, continental Subandenn side, with very cold weather and strong rainfall (annual average of 50-1000 mm/year). There are also changes in rainfall patterns from east to west across the country, from the shoreline to the Andean mountains. The northern and central areas show increasing precipitation from the coast towards the Andes, whereas in the southernmost areas this tendency is reversed, with extensive rainfall on the coast and dry areas in the Andean mountains (Niemeyer & Cereceda 1984). In central Chile the volume of surface water varies as greatly from north to south as from east to west, as a result of the presence of the Andes, which source and store water resources for the whole region (Fernández 1997). Figure 8.2 shows annual average values of rainfall from north to south as a function of latitude, based upon the records of several pluviometric stations in the area between Copiapó and Temuco. It is clear from this figure that the amount of precipitation increases towards the south, being very scanty in all the stations in the north. Also towards the south, greater differences between coast and mountain range rainfall appear but the resource is, in general, much more abundant, although with a considerably greater interannual variability. Annual rainfall north of the latitude 33°S is less than 350 mm/year. A large east-west variability is observed south of latitude 36°S, due to the presence of a central valley and the existence of coastal mountain ranges and the Andes.

Surface water resources

Rainfall

The amount of rainfall and evaporation in each area of the country has a strong dependence on the local climate. Incrementally increasing rainfall in more southerly latitudes is clear throughout the country, with almost no precipitation in Arica, 50-1000 mm/year in the Central Zone, and over 3000 mm/year in the far south. There are also changes in rainfall patterns from east to west across the country, from the shoreline towards the Andean mountains. The northern and central areas show increasing precipitation from the coast towards the Andes, whereas in the southernmost areas this tendency is reversed, with extensive rainfall on the coast and dry areas in the Andean mountains (Niemeyer & Cereceda 1984).

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Figure 8.3 indicates this variability between different years, expressed as the ratio between the standard deviation of the annual rainfall in relation to the mean, and it constitutes a measurement of the magnitude of the changes observed from year to year. In the north these changes are significant and abrupt in relation to the average values, whereas in the south they are small, with a smooth transition throughout the territory. This is illustrated by situations such as in La Serena where between 1919 and 1985 the driest year registered only 4 mm of rain, while the wettest recorded was 307 mm. With an annual average of 98 mm, these results represent variations between 4% and 300%. In contrast, in Osorno, for a similar period, the average was 1433 mm, with the driest being 1054 mm and the wettest being 1879 mm, representing respectively 74% and 131% of the average. These figures emphasize how in the north rains are scarce, with great interannual differences, and as they increase towards the south, the annual variation in amount diminishes. This figure shows that north of latitude 33°S the coefficient of variation is greater than 0.5 while the regions in southern Chile have a coefficient of variation in the order of 0.35.
Monthly precipitation is distributed differently depending on the geographic region and climate (MOP, Manual de Carreteras 2001). In central and southern Chile it is concentrated in winter months whereas in the south it is distributed uniformly during the year (Fig. 8.4). In the north (San Pedro de Atacama) rainfall occurs in the summer months in the Andes. In central Chile (represented by La Serena, Santiago, Talca, Concepción) rainfall occurs mainly in the winter months (May–August). Further south (Valdivia, Puerto Aysen and Punta Arenas) rainfall is more uniformly distributed during the year. Precipitation levels directly affect surface water volumes in the different river basins.

**Runoff**

Figure 8.5 shows how total rainfall is distributed between runoff and abstractions, mainly evaporation–transpiration, as a function of latitude, in 15 river basins of the Central Zone. It illustrates how precipitation transforms into evaporation in the
Fig. 8.4. Graphical comparison of mean annual rainfall on selected cities.

River basins in the north, with significant runoff being generated only south of the Aconcagua River. Thus, as shown in Table 8.1, rivers in the north have limited streamflow and few events with substantial volumes of runoff, since they receive little precipitation which mostly evaporates or transpires when falling on dry ground. As an example, the Elqui River catchment area has an average annual precipitation of 180 mm, but less than 1% of this flows to the sea. In river basins in the south, however, precipitation is abundant and generally falls on humid ground, converting easily into runoff. An example of this is the Imperial
River, whose average annual precipitation in the basin is 1640 mm per year, of which 65% flows to the Pacific coast. The previous example shows that although in the Imperial River the annual precipitation is ten times that of the Elqui River, runoff volumes are 300 times greater. In this way, the percentage of runoff tends to increase from north to south, due to the precipitation increase and cooler climates.

River basins exhibit spatial differences, because of their large storage capacity, and they decrease the effect of temporal rainfall variations, improving water availability as compared to the availability of rainfall. Nevertheless, in a zone of strong differences such as central Chile, runoff also shows important fluctuations over the years and months. Thus northern rivers contain small water volumes with sporadic runoff, in which the important resources come from a small number of floods, with relatively prolonged intervals of water shortage. Rivers in the south have a permanent regime with abundant winters and periods of annual flow. Table 8.2 shows average, standard deviations and skew coefficients for annual flows (expressed in mm) for different basins in Chile. Figures 8.6 and 8.7 show time series of monthly average volumes in a northern and a southern river.

Flow distribution within the year is also quite different in the north and in the south depending on the origin of the water (Fig. 8.8). In the north, rivers either have larger flows in the summer period due to rainfall occurring in the upper Andes, or almost constant flow throughout the year when the principal flow component is groundwater (e.g. Loa River). In central Chile monthly flow variation due to snowmelt in the summer months is important and there is also an increase during the winter months (e.g. Colorado River). Further south, winter precipitation is more important than summer snowmelt, since the Andes mountains are lower, so that only one period of maximum flow is observed (e.g. Bio-Bio and Allipen rivers).

**Use of surface water resources**

Current total water use in Chile is approximately 72 533 hm³/year, including consumptive and non-consumptive uses (Salazar 2003). Consumptive use is approximately 20 498 hm³/year, 84.5% of which is used in irrigation, 6.5% in industry, 4.5% in mining activities, and 4.4% by the urban population. However, as might be expected from the human geography of Chile, behind this overall distribution of water demand, there are major regional variations. The main usage, 12 614 hm³/year, is concentrated in the Central Zone (33–38°S), between Santiago and Valdivia, with the two large cities of Valparaiso and Santiago accounting for most water usage (the Metropolitan Region of Santiago alone uses 7726 hm³/year). From 33°S to 38°S, irrigation is the most important water use, although hydroelectric power generation is concentrated in Regions VI and VII (35–37°S). Further south still, consumptive uses are generally small. With regard to northern Chile (38–32°S), water demand is distributed more or less equally between irrigation, mining, urban and industrial uses, and water availability is very limited, being less than 1000 m³/capita/year and dropping to

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**Table 8.1. Impact of evaporation on runoff in different zones of Chile**

<table>
<thead>
<tr>
<th>Zone</th>
<th>Basin</th>
<th>Surface (km²)</th>
<th>Rainfall (mm/year)</th>
<th>Runoff (mm/year)</th>
<th>Potential evaporation (mm/year)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Norte</td>
<td>Laulau</td>
<td>2350</td>
<td>N/A</td>
<td>20</td>
<td>1941</td>
</tr>
<tr>
<td>Grande</td>
<td>Tarapacá</td>
<td>17253</td>
<td>54</td>
<td>1</td>
<td>N/A</td>
</tr>
<tr>
<td></td>
<td>Loa</td>
<td>33570</td>
<td>75</td>
<td>8.5</td>
<td>2000</td>
</tr>
<tr>
<td></td>
<td>Salar de</td>
<td>15620</td>
<td>2786</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td></td>
<td>Atacama</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Norte</td>
<td>Copiapó</td>
<td>18800</td>
<td>94</td>
<td>0.2</td>
<td>93</td>
</tr>
<tr>
<td>Chico</td>
<td>Elqui</td>
<td>9645</td>
<td>179</td>
<td>0.8</td>
<td>171</td>
</tr>
<tr>
<td></td>
<td>Limari</td>
<td>11760</td>
<td>274</td>
<td>20.1</td>
<td>249</td>
</tr>
<tr>
<td></td>
<td>Petorca</td>
<td>1964</td>
<td>254</td>
<td>N/A</td>
<td>264</td>
</tr>
<tr>
<td>Central</td>
<td>Aconcagua</td>
<td>7573</td>
<td>529</td>
<td>12.8</td>
<td>395</td>
</tr>
<tr>
<td></td>
<td>Maipo</td>
<td>15157</td>
<td>663</td>
<td>207</td>
<td>423</td>
</tr>
<tr>
<td></td>
<td>Rapel</td>
<td>13710</td>
<td>960</td>
<td>399</td>
<td>544</td>
</tr>
<tr>
<td></td>
<td>Maule</td>
<td>20865</td>
<td>960</td>
<td>39</td>
<td>544</td>
</tr>
<tr>
<td></td>
<td>Biobio</td>
<td>24782</td>
<td>1891</td>
<td>1277</td>
<td>599</td>
</tr>
<tr>
<td>South</td>
<td>Imperial</td>
<td>10654</td>
<td>1217</td>
<td>2041</td>
<td>N/A</td>
</tr>
<tr>
<td></td>
<td>Valdivia</td>
<td>9902</td>
<td>2307</td>
<td>2956</td>
<td>N/A</td>
</tr>
<tr>
<td></td>
<td>Maullín</td>
<td>4298</td>
<td>1789.9</td>
<td>1441</td>
<td>N/A</td>
</tr>
<tr>
<td>Austral</td>
<td>Aysén</td>
<td>11456</td>
<td>1800</td>
<td>3000</td>
<td>1500</td>
</tr>
<tr>
<td></td>
<td>Serrano</td>
<td>7347</td>
<td>&lt;60</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td></td>
<td>Cisne</td>
<td>5193</td>
<td>184.3</td>
<td>N/A</td>
<td>N/A</td>
</tr>
</tbody>
</table>
Table 8.2. .4~ernee.sm,aior-d deieiiniio,~n,idskcli, roei/icio,ir q f o i r i i i ~ o l f l ~ ~ d((ferri~t
~ ~ . ~ f o rb<t.siii.~in Cliile
Lat. (S)

Basin

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values

Elqui R.

Limari R.

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Aconcagua R.

Maipo R.

Rapcl R.

Mataquito R.

Maule R.

r t a t j 111

Bio-Bio R.

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Bvc-.o R.

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R~

T i ~ i b i oat Vaiillar
Clara at Rivadavia
Elqui at Algarrobal
Elqui at Alrnendral
Hurtado at Angostura de Pangue
Grande at Agua Chica
Rio Grande at Puntilla San Juan
Grande at paloma 1
Illapel at Huinfil
Choapa at Puente Negro
Choapa at Salamancd
Putaenda at Resguarda los Patos
Blanco at Blanco
Aconcagua at Blanco
Aconcagua at Chacabuquita
Maipo at Cabimbao
Estero Arrayan at la Montosa
Mapocho at 10s Almendros
Maipo at el Manzano
Volcai~at Queltehues
Cachapovl at Puente Termas
Tinguiririca baja Los Rriooes
Rapel at Corneche
Teno at los Quefies
Claro at 10s Quefies
Colorado a t lunta can Palos
Palos a t Junta con Colorado
Claro at Camarico
Melado at la Lancha D G A
Maule at Armerillo D G A
Potagan a t Yerbas Buenas
Ancoa at el Morro
Longavi a t la Quiriquina
Cauquenes at el Arrayan
Nuble at San Fabian
Chillan at Esperanza
Itata a t Cholguan
Renegado a t Invernada
Diguillio at San Lorenzo
ltata at balsa Nueva Aldea
Estero Pichi~olcuraat lo Gatica
Pol&lra.at Valie Cuatro Juntas
Polcura at Balseadero
Laja at Tucapel
Duqueco at Villucura
Queuco at Puente Queuco
Bio-Bio at Rucalhue
Bureo at Mulchen
Mulchen at Mulchen
Maileco at Collipulli
Puren at Tranaman
Traiguen a t Victoria
Cholchol at Cholchol
Cautin at Rari-Ruca
Cautin zit Caj6n
Allipen a t los Lauieles
Puyehue at Quitratue
Tolten at Villarrica
Huanehue a t Desague Lago Calafquen
Hua-Hum at Desembocadura Lago
Piiihueic
Fui at Desague Lago Pirihueico
Pilmaiquent at San Pablo
Pilmaiquen bajo el Salto
\laullin at Llanquihue

29"57'
29"59'
30"OO'
2999'
30"26
30°42'
30"42'
30"41'
31-34,
31°41'
31°47'
32"31'
32"55'
32"54'
32"50'
33-47
33"21'
33"22'
3395'
33'48'
34"15'
34"41'
33"59'
35'00'
35"OO'
3596'
35"16'
3i010'
35"il'
35"42'
35"47'
35"54'
36"14'
'6"OI'
6"f4'
6"47'
7"09'
36-02,
36"5@
36"39'
37-19,


Mean Monthly Flow at: Tránsito en junta el Carmen 28.6° South Latitude

Fig. 8.6. Chronological series of mean monthly runoff of a northern river.

Mean Monthly Flow at: Ñuble en San Fabián 36.5° South Latitude

Fig. 8.7. Chronological series of mean monthly runoff of a southern river.

less than half this amount in some places. Thus water in northern Chile is very much a scarce and precious resource, the availability of which constrains human activities. Drinking water is mostly provided from surface sources (73%), although a number of small towns and rural communities use groundwater when available.

Groundwater resources

Hydrogeological setting

The groundwater resources of Chile depend directly on the hydrogeological properties of each region and on the underlying soils and geology (Salazar 2003). Aquifer characteristics from representative river basins from the country are presented in Table 8.3, based on the previously defined hydrological and geological regions. From a hydrological viewpoint, geological influences on groundwater resources in northern Chile (Norte Grande zone) are dominated by the presence of abundant Tertiary and Quaternary volcanic rocks, especially in the extremely high Altiplano. Water draining from the Altiplano flows either eastward towards Bolivia, or westward into arid basins lying east of the Coastal Cordillera. This westward flowing water moves through rocks of very low permeability, and so it is mostly confined to near river channels. Much of the
water accumulates in aquifers in the basins between the High Andes and the Coastal Cordillera, in zones of high or moderate permeability such as the Tamarugal Pampa. Given the extreme evaporation and climatic conditions, these aquifers only rarely have an exit, usually evaporating instead to give rise to salt plains like the Salar de Atacama. The aquifers located in the Norte Grande zone, which are almost totally unconfined, have recharge values estimated to reach 9698 hm³/year.

Similar hydrological patterns are displayed by the Norte Chico zone further south of the north zone, with the main difference being that waters usually have an exit connecting to the ocean. Thus 'salar' salt plains, so characteristic of the Atacama Desert, are eliminated in favour of fluvial clay deposits and the formation of more confined aquifers, the recharge values of which are estimated to reach about 4906 hm³/year.

In the Central Zone, the waters again flow from the Andes through rocks of generally low permeability, and accumulate in the Central Valley. Aquifers in this depression are formed in zones of moderate to high permeability, where the water is accumulated easily due to the presence of a wall of very low permeability provided by the batholithic intrusions of the Coastal Cordillera. The recharge of these Central Valley aquifers reaches 187 614 hm³/year, emphasizing the abundance of water supplied compared to the north, with very high permeability due to the abundance of fluvioglacial sediments and with groundwater levels lying very close to the surface. The aquifers are highly interconnected and the majority have exits to the sea, following the river channels. The recharge of the aquifers reaches 244 617 hm³/year.

The hydrology of the Austral Zone in the far south has been very little studied, but is clearly strongly influenced by the Andean chain, the height of which subsides slowly towards the south. Aquifers are generally scant or very shallow due to the predominantly low permeability sediments, and in the few sectors where a central valley can be distinguished, the sediments are highly consolidated. There is no Coastal Cordillera here to act as a hydrogeological barrier, and the Central Valley gives way southwards to a largely submerged area with hundreds of islands, gulls and estuaries. Under these conditions, Quaternary geological formations tend to dominate aquifer distribution. The recharge of these aquifers is estimated to reach 273 208 hm³/year.

**Use of groundwater**

The use of groundwater resources in Chile depends on the concentration of population and economic activities that require this resource, and on its availability, which as we have seen is directly related to the climate of the different regions. The use of this resource in the far north (Norte Grande zone) involves 984 wells, rising to 1426 wells in the Norte Chico zone, and to 8350 wells in the central zone. There is no detailed registry of the use of water resources in the south of Chile, mainly due to the abundance of superficial water resources. The greatest demand for groundwater resources, in terms of volume, is predictably in the Central Zone, at 157 140 hm³/year, whereas the greatest percentage of groundwater use is in the north (Norte Chico zone and the far north combined), reaching 14 338 hm³/year. There are again no measurements available in the southern regions due to the abundance of the resource. It is clear from this demand distribution that Norte Chico zone and more northern areas are the most hydrologically sensitive and stressed due to the lack of superficial water resources, and thus require the greatest attention in measurement and protection.
Table 8.3. Soils, lithology and hydrogeology units of principal aquifers in Chile

<table>
<thead>
<tr>
<th>Zone</th>
<th>Basin</th>
<th>Soils and Lithology</th>
<th>Hydrogeology units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Norte</td>
<td>Lauca</td>
<td>Silstones mudstones with calcareous intercalations, conglomerates and/or pyroclastic rocks. Histosols with high contents of organic material and salt content. Mollisols rich in calcium.</td>
<td>1) Fractured volcanic rocks occurring as lava flows, tuffs and breccias intercalated with Tertiary–Quaternary siliciclastic continental sediments of medium permeability. 2) Unconsolidated, highly permeable Quaternary sedimentary deposits of fluvial, glacial, alluvial and lacustrine origin.</td>
</tr>
<tr>
<td>Grande</td>
<td></td>
<td></td>
<td>Part of the broader unit known as the Pampa del Tamarugal. Originating in the Andean mountains, the waters flow through fractured volcanic rocks of Tertiary–Quaternary age (andesitic flows, breccias and tuffs of medium permeability), accumulating in the pampa depression within unconsolidated and highly permeable deposits. Rocks of the Coastal Cordillera block drainage of the basin to the west, forcing the underground waters to flow into the River Loa basin.</td>
</tr>
<tr>
<td>Norte</td>
<td>Tarapacá</td>
<td>Pleistocene–Holocene sedimentary rocks: alluvial, colluvial and lacustrine deposits, gravel deposits, sands and silstones. Histosols, mollisols, entisols. Entisols derived from coarse sediments with high salt content and elevated pH.</td>
<td>The upper part comprises Tertiary–Quaternary fractured volcanic rocks (andesitic lavas, breccias and tuffs) of medium permeability. At Calama the waters are blocked underground due to plutonic and hypabyssal rocks of low permeability. The aquifer flows from east to west parallel to the River Loa, through carbonate rocks (limestones and calcareous shales) of medium permeability. In the Intermediate Depression the waters pond and join with those of the Pampa del Tamarugal in unconsolidated Quaternary fluvial, alluvial and aeolian deposits.</td>
</tr>
<tr>
<td>Grande</td>
<td></td>
<td></td>
<td>1) Fractured volcanic rocks as lava flows, breccias and tuffs, intercalated with continental elastic sediments of Tertiary–Quaternary age and medium permeability. 2) Unconsolidated fluvial, glacial, alluvial and lacustrine deposits of Quaternary age and high permeability. The salar aquifer is a closed system, recharged by waters from the NNE and south, which drain into the Salar and evaporate.</td>
</tr>
<tr>
<td>Norte</td>
<td>Loa</td>
<td>Upper Miocene–Pliocene sediments: classic examples of alluvial fan, colluvial and fluvial deposits, with conglomerates, sandstones and silstones. In the north there are peaty soils and the presence of salts in vertical profile. In the ENE some soils show an impermeable tosa. The soils to the west are characterized by sandy loam to sandy textures, with a loamy silt layer at depth.</td>
<td>Upper sector has very low permeabilities due to the presence of volcanosedimentary, volcanic, hypabyssal and plutonic rocks. Between the Paipote quebradas and the River Copiapó are Tertiary volcanlastic rocks and Cretaceous sediments and volcanic rocks also with very low permeabilities. In the area around the Lautaro Reservoir are Palaeozoic plutons and Jurassic volcanosedimentary rocks. From Copiapó as far as Caldera, waters run through Tertiary volcanosedimentary rocks, and highly permeable unconsolidated sediments.</td>
</tr>
<tr>
<td>Grande</td>
<td></td>
<td></td>
<td>The entire basin comprises unconsolidated fluvial gravels, sands and silts that follow the courses of the major rivers, and their recent terraces and floodplains. Soils are classified as brown soils. In the coastal zone, alluvial soils on marine terraces and in the valley bases have evolved from both marine and continental sediments. These are referred to as coastal zone soils and mollisols, with a brown colour and fine texture, formed from sands and silts. Clays dominate in the upper part of the terraces. In the River Elqui basin, the soils are dominantly red lithosols that comprise clay, with accumulations of silt in rock fractures.</td>
</tr>
<tr>
<td>Norte</td>
<td>Salar de</td>
<td>Miocene to Quaternary evaporitic sediments (salphates, chlorides, carbonates) with fine-grained detrital layers locally containing borax and/or lithium. Very rare soils, virtually all of which are part of the saline Salar de Atacama. The zone bordering the Salar shows soils of entisol type. Soils poorly stratified due to the aridity, extreme salinity, and elevated pH.</td>
<td>1) Fractured volcanic rocks as lava flows, breccias and tuffs, intercalated with continental elastic sediments of Tertiary–Quaternary age and medium permeability. 2) Unconsolidated fluvial, glacial, alluvial and lacustrine deposits of Quaternary age and high permeability. The salar aquifer is a closed system, recharged by waters from the NNE and south, which drain into the Salar and evaporate.</td>
</tr>
<tr>
<td>Grande</td>
<td>Atacama</td>
<td></td>
<td>The upper part comprises Tertiary–Quaternary fractured volcanic rocks (andesitic lavas, breccias and tuffs) of medium permeability. At Calama the waters are blocked underground due to plutonic and hypabyssal rocks of low permeability. The aquifer flows from east to west parallel to the River Loa, through carbonate rocks (limestones and calcareous shales) of medium permeability. In the Intermediate Depression the waters pond and join with those of the Pampa del Tamarugal in unconsolidated Quaternary fluvial, alluvial and aeolian deposits.</td>
</tr>
<tr>
<td>Norte</td>
<td>Copiapó</td>
<td>North: Lower to Upper Cretaceous continental sediments and volcanic rocks with rare intercalations of marine sedimentary and volcanic breccias, andesitic lavas, sandstone conglomerates, lacustrine calcareous silstones with fossil flora. SW: Upper Cretaceous continental lavas and volcanosedimentary succession (andesites, trachytes and rhyolitic volcaniclastic rocks). NE: Lower to (early) Upper Cretaceous intrusive rocks (diorites, pyroxene- and hornblende–monzodiorites, granodiorites, hornblende–biotite monzodiorites) associated with Fe, Cu, Au mineralization. Fluvial sediments mixed with detritus derived from tributary canyons (quebradas) feeding into the main valley: sandstones, silts and shales predominate in the lower part of the valley whereas coarser sediments occur higher up with finer sediments confined to small lakes or dammed areas.</td>
<td>Upper sector has very low permeabilities due to the presence of volcanosedimentary, volcanic, hypabyssal and plutonic rocks. Between the Paipote quebradas and the River Copiapó are Tertiary volcanlastic rocks and Cretaceous sediments and volcanic rocks also with very low permeabilities. In the area around the Lautaro Reservoir are Palaeozoic plutons and Jurassic volcanosedimentary rocks. From Copiapó as far as Caldera, waters run through Tertiary volcanosedimentary rocks, and highly permeable unconsolidated sediments.</td>
</tr>
<tr>
<td>Chico</td>
<td></td>
<td></td>
<td>The entire basin comprises unconsolidated fluvial gravels, sands and silts that follow the courses of the major rivers, and their recent terraces and floodplains. Soils are classified as brown soils. In the coastal zone, alluvial soils on marine terraces and in the valley bases have evolved from both marine and continental sediments. These are referred to as coastal zone soils and mollisols, with a brown colour and fine texture, formed from sands and silts. Clays dominate in the upper part of the terraces. In the River Elqui basin, the soils are dominantly red lithosols that comprise clay, with accumulations of silt in rock fractures.</td>
</tr>
<tr>
<td>Norte</td>
<td>Elqui</td>
<td></td>
<td>The entire basin comprises unconsolidated fluvial gravels, sands and silts that follow the courses of the major rivers, and their recent terraces and floodplains. Soils are classified as brown soils. In the coastal zone, alluvial soils on marine terraces and in the valley bases have evolved from both marine and continental sediments. These are referred to as coastal zone soils and mollisols, with a brown colour and fine texture, formed from sands and silts. Clays dominate in the upper part of the terraces. In the River Elqui basin, the soils are dominantly red lithosols that comprise clay, with accumulations of silt in rock fractures.</td>
</tr>
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</table>
Table 8.3. (Continued)

<table>
<thead>
<tr>
<th>Zone</th>
<th>Basin</th>
<th>Soils and Lithology</th>
<th>Hydrogeology units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Norte</td>
<td>Limari</td>
<td>All the water resources are found within unconsolidated fluvial deposits (gravels, sands, silts) found along the modern river courses, and their recent terraces and floodplains. In the central part of the basin are Upper Jurassic-Lower Cretaceous volcanic rocks comprising basaltic lavas, andesitic to dacitic domes, breccias and agglomerates, with intercalations of continental and marine dark limestones with neutral or slightly alkaline pH. Silts and recent sediments predominate in the valley bottoms, in contrast to more clay-rich higher terraces. The soils of the lowest terraces are thin and poorly evolved, clayey, and nitrogen-poor.</td>
<td></td>
</tr>
<tr>
<td>Norte</td>
<td>Petoica</td>
<td>The zone is characterized by the presence of a huge amount of eruptive igneous rocks (cropping out both as lavas and volcanic rocks) which are interstratified with Upper Cretaceous to upper Miocene marine and continental sediments. There are also abundant ancient fluvial sands and gravels, with indications of weathering and rounding. The soils are andisols, inceptisols and mollisols characteristic of Region V in Chile. Andisols are well graded soils derived from a granitic protolith, with clay at depth. The inceptisols, are dark red and derived from marine terraces. The mollisols are alluvial soils with a high content of organic and other nutrients useful for agriculture.</td>
<td></td>
</tr>
<tr>
<td>Central</td>
<td>Aconcagua</td>
<td>Pleistocene-Holocene fluvial deposits: gravels, sands, and silts. Lower-middle Miocene eruptive rocks forming partially eroded volcanic complexes comprising lavas, breccias, domes and pyroclastic rocks. Lower-Upper Cretaceous volcanic and sedimentary sequences including breccias, andesitic lavas, porphyritic andesites (coconets), sandstones and calcareous silstones. Gravelly alluvial soils with rounded pebbles, with organic-rich soils and fine deposits in areas of lower energy deposition. Andisols, inceptisols and mollisols. Andisols are well graded soils derived from a granitic protolith, with clay at 3 depth. The inceptisols, are dark red and derived from marine terraces. The mollisols are alluvial soils with a high content of organic and other nutrients useful for agriculture.</td>
<td></td>
</tr>
<tr>
<td>Central</td>
<td>Maipo</td>
<td>The Maipo basin is infilled with fluvial and fluvioglacial sediments and volcanic ashes, as well as having a basement of Palaeozoic and Mesozoic granite rocks, and Cretaceous sediments and volcanic rocks. In the High Cordillera there are poor soils on a deeply eroded rocky substrate. In the foothills are stratified, well drained and highly fertile soils. They are medium (loamy silts and loamy sandy loams) to fine grained (clayey sand to very clayey), with some coarser areas (loamy sands and silts). The Mapocho cones and the coastal zone display coarse soils (gravely sands and ripios).</td>
<td></td>
</tr>
<tr>
<td>Central</td>
<td>Rapel</td>
<td>Sulphur-bearing rocks common due to the influence of the Tinguiririca volcano. Volcanic and sedimentary rocks: basaltic to andesitic breccias and andesites, pyroclastic rocks, andesites and rhyolites, basaltic to dacitic lavas, epiclastic to pyroclastic rocks, sandstones, calcareous siltstones. At the coast are dark red soils derived from marine terraces, and soils derived from a granitoid substrate with a high clay content. In the interior valley are alluvial alfisols, mollisols and entisols. In the Cordillera there are sandy soils derived from glasy volcanic materials.</td>
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</table>

Due to the presence of a granitoid and volcanic basement, the rocks in the high source regions are practically impermeable. The tectonically controlled Intermediate Depression displays medium to high permeability, in contrast to the Coastal Cordillera where Palaeozoic granitoid rocks and Cretaceous volcanic and sedimentary deposits display only medium to low permeabilities.

Due to the presence of a granitoid and volcanic basement, the rocks in the high source regions are practically impermeable. The tectonically controlled Intermediate Depression displays medium to high permeability, in contrast to the Coastal Cordillera where Palaeozoic granitoid rocks and Cretaceous volcanic and sedimentary deposits display only medium to low permeabilities.

This sector is divided into five aquifers which flow in different directions. The first runs parallel to the River Cachapoal; the second drains north-south through Angostura de Paine; the third follows the Tinguiririca basin; the fourth begins at Tinguiririca and drains through San Fernando; and the fifth comes from the Coastal Cordillera and ponds in the Rapel reservoir. In general the aquifers are confined within plutonic rocks with very low permeability. At the coast the rocks can be highly permeable whereas in the Central Valley they are moderately impermeable.
### CHILEAN WATER RESOURCES

<table>
<thead>
<tr>
<th>Zone</th>
<th>Basin</th>
<th>Soils and Lithology</th>
<th>Hydrogeology units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Central</td>
<td>Maule</td>
<td>Siluroian- Carboniferous muscovite schists and slates, and Carboniferous-Pelmo granites. Lower Jurassic intrusive rocks: diorites, gabbros, pyroxene monzodiorites, quartz-diorites, granodiorites, and hornblende-biotite tonalites. Soil textures mostly clayey in the main valley and Precordillera, with some areas rich in volcanic ash. Some zones show sediments of alluvial origin with variable amounts of lithic fragments.</td>
<td>Quaternary lavas and pyroclastic rocks of dacitic, andesitic and basaltic composition and low permeability. Waters accumulate in the Intermediate Depression due to being blocked by impermeable Palaeozoic plutonic and hypabyssal rocks of the Coastal Cordillera.</td>
</tr>
<tr>
<td>Central</td>
<td>Biobío</td>
<td>Late Lower Cretaceous continental basaltic and andesitic lavas and breccias. Upper Cretaceous-Palaeogene volcanosedimentary sandstones, conglomerates, andesitic and dacitic lavas, silstones and limestones. Peleistocene basic to intermediate lavas and volcanic centres. In the higher areas are remnant soils on a deeply eroded rocky substrate, whereas lower down are highly fertile trumosas and silt soils, as well as sandy areas of low fertility. The soil protolith is mostly volcanic ash.</td>
<td>Low lying estuarine-deltaic areas show high permeabilities, in contrast to the Coastal Cordillera where the granitoid and metamorphic rocks show medium to low permeabilities. The Central Depression contains basaltic-andesitic lavas, tuffs and alluvial and kibaristic deposits with high permeabilities. In the Andean Cordillera are poorly permeable granites and granodiorites.</td>
</tr>
<tr>
<td>South</td>
<td>Imperial</td>
<td>Palaeozoic-Triassic metapelites, metachert, metasiltites and gneisses. Oligocene-Miocene marine sandstones and coquinas, volcanic breccias, lavas and pyroclastics. Plio-Pleistocene basaltic lavas, tuffs and conglomerates. Highly fertile and organic-rich andisols (trumosas), gray clayey soils derived from volcanic material, medium grained soils with clay at depth, and clayey alluvial soils with a franca-arcollosa texture.</td>
<td>Higher part of basin shows low permeabilities within volcanic formations. Underground waters follow the river channels as far as the Intermediate Depression where they accumulate in unconsolidated alluvial deposits, with a notably low water table (below 6 m). From here the aquifer follows the river courses to the coast.</td>
</tr>
<tr>
<td>South</td>
<td>Valdivia</td>
<td>Palaeozoic-Triassic metapelites, metachert, metasiltites and gneisses. Carboniferous-Pelmo-Middle to Upper Jurassic intrusive rocks: granites, granodiorites, tonalites, diorites, monzodiorites, pyroxene and hornblende diorites. Peleistocene-Holocene sandstones, fluvial and glacioclustrine muds, sands, gravels and silts. Mostly ‘Nahuelbuta’ soils of high altitudes derived from altered mica schists and phyllites. On the slopes are clayey-sandy soils with a loamy clay surface and more clay-rich at depth. In the northern zone the soils are dark red or yellow-red, loamy silt or sandy silt.</td>
<td>Upper part: poorly permeable Tertiary-Quaternary volcanic lavas, breccias and tuffs, plus mudstones, limestones and sandstones. In the Central Valley the aquifer drains through highly permeable glacial and alluvial sediments. The Coastal Cordillera comprises metamorphic rocks of low permeability that splits the aquifer into two.</td>
</tr>
<tr>
<td>South</td>
<td>Maullín</td>
<td>Pleistocene-Holocene maraines, fluvio-glacial and glacioclustrine sediments, silty sands, gravels and silts. Quaternary volcanic rocks: stratovolcanoes and other volcanic complexes, basaltic to rhyolitic lavas, basaltic andesitic to dacitic pyroclastics. Ferruginous-encrusted trumosas (radii) soils predominate. Soils derived from volcanic ashes, with very fine to fine loamy sandy textures, with loamy silt zones. High amounts of organic material and roots, and at depth they are sandy-clay to loamy clays. All have acidic pH.</td>
<td>Upper areas: Jurassic-Tertiary impermeable granitoid intrusions and hypabyssal rocks. In the Central Valley the aquifer runs through unconsolidated glacial infill comprising highly permeable maraines and alluvial materials. Coastal Cordillera forms an impermeable barrier of Palaeozoic sedimentary and metamorphic rocks.</td>
</tr>
<tr>
<td>Austral</td>
<td>Aysén</td>
<td>Upper Jurassic-Lower Cretaceous marine sediments: littoral and platformational limestones, mudstones, calcareous sands, sandstones and coquinas. Quaternary volcanic rocks: stratovolcanoes and other volcanic complexes, basaltic to rhyolitic lavas, domes, and basaltic andesitic and rhyolitic pyroclastic deposits (mostly calcalkaline). Soils derived from glassy volcanic rocks show coarse texture (loamy sands to very sandy), low fertility and poor moisture retention. Trumosa-type soils are derived from volcanic ashes and contain high organic content.</td>
<td>Poorly studied. In the Andean Cordillera there is generally low permeability except the highly permeable ground associated with the River Coyhaique. Underground aquifers only found around the River Aysén. Elsewhere in the basin there is a lack of aquifer due to low (or virtually zero) permeability.</td>
</tr>
<tr>
<td>Austral</td>
<td>Serrano</td>
<td>Pliocene-Pleistocene basaltic lavas, tuffs and conglomerates. Eocene-Miocene marine sandstones and mudstones. Upper Cretaceous sandstones, conglomerates, mudstones, limestones. Miocene granodiorite, diorite and tonalite intrusions. Soils poorly studied but include abundant histosols.</td>
<td>Tertiary volcanosedimentary rocks are prominent within the basin, although little information is available.</td>
</tr>
<tr>
<td>Austral</td>
<td>Cisnes</td>
<td>Oligocene-Miocene basaltic to rhyolitic lavas, pyroclastic and epiclastic rocks. Miocene granodioritic and tonalitic intrusions. Palaeozoic-Triassic metapelites and metabasites. Soils little studied but can be divided into those in the forests (the more common) and those of the steppe.</td>
<td>Higher part comprises unconsolidated Quaternary sediments. Within the valley the waters drain through intrusive and hypabyssal rocks.</td>
</tr>
</tbody>
</table>
Water quality and environmental issues

The following description is based on three distinct geographic zones (North, Central and South) which correspond to the following administrative regions: I to IV (17°30'-32°16'S), V to VIII (including the Metropolitan Region, 32°16'-38°30'S), and IX to XII (38°30'-56°30'S), respectively. Greater emphasis is given to the description of the North Zone, considering that major mining operations, and the highest stress on water quality and quantity, are found there. We firstly describe the most relevant catchments and the key natural factors that control water quality throughout the country. Secondly, the anthropogenic factors that control water quality are outlined with respect to two components: a description of the economic activities with environmental relevance, and an overview of Chilean environmental policy and regulations which will shape the future of the quality of water resources in Chile.

Characterization of water quality and its controlling factors

Water quality varies widely throughout the country (Cadenas and Idepe 2004) and is controlled by a complex interaction of hydraulic, hydrological, geological and anthropogenic factors (Table 8.4). In the north, water chemistry commonly contains a strong contribution from naturally occurring ore deposits to the levels of metals and metalloids in waters. Elevated levels of arsenic, boron, lithium, sodium and potassium are present (Table IX). Water quality varies widely throughout the country (Table VIII), reflecting both natural and anthropogenic factors that control water quality. Selection of the most relevant catchments and the key natural factors that control water quality throughout the country are outlined in Table IX. Present-day water quality in Chile is low due to low flow rates. Sporadic rain events during the 'Bolivian winter' season (December-March) cause transient surface runoff that brings high sediment loads to rivers, whereas normally strong solar radiation generates high levels of water evaporation, increasing the concentration of chemical species. A lack of comprehensive measurements and quantitative geochemical models makes it difficult to differentiate the natural background level of metals and metalloids from the contribution of the same species from anthropogenic sources. Notably, high concentrations of boron and arsenic (up to 50 mg/L, for example, have been linked to thermal waters of volcanic origin (Klohn 1972). A recent study of arsenic concentrations in sediments from the Elqui River drainage system has shown strong enrichment due to a combination of naturally exposed mineralized hydrothermal alteration zones and mining activity in the river source region (Oyarzun et al. 2004).

Andean metallogenic deposits also contribute to the levels of metals in Central Zone waters, although there is substantial dilution compared to the North Zone, and consequently a better overall water quality. Higher surface runoffs coupled with increasing, but still limited, vegetation coverage initially produce more turbid waters in the northern section of the Central Zone. However, as vegetation cover increases southwards, so turbidity decreases. Lower solar radiation and higher stream flows in the Central Zone make concentration by evaporation a negligible factor compared to the North Zone. Towards the south of Central Zone more abundant clays and limestone make waters more alkaline.

The South Zone is characterized by excellent water quality. Flow rates are substantially higher and the vegetation coverage is dense, reducing erosion and suspended solids, but increasing the dissolved organic carbon from the degradation of vegetation. Lakes from ice and snowmelt towards the southern end show higher turbidities; however, this turbidity decreases down to negligible levels as the water moves downstream from the lake drainage. Although normally of good quality, some waters in the southernmost part of the South Zone show high levels of aluminum and other metals due to local metallic ore deposits (e.g. aluminum and zinc in the Aysén river).

Anthropogenic factors

Present-day water quality in Chile is controlled by past and current human activities, while water quality in the future will be controlled by current environmental policies and regulations. A general view of the main economic activities for selected catchments is shown in Table 8.5. The activities that traditionally have contributed to water pollution throughout the country are untreated municipal wastewater, industrial effluents, agriculture and mining. Mining is an important source of metals in the far north, whereas forestry-related pollution associated with cellulose plants is the most important in the central and southern areas, while aquaculture is becoming an important contamination source specifically in the waters of Southern Chile.

Untreated municipal wastewater has been historically the single most important contributor to water pollution, mainly affecting biochemical oxygen demand (BOD) and introducing pathogens. However, the construction of new wastewater plants has reduced this problem substantially in the last few years, improving from 8% of total urban population with wastewater treatment in 1989 to 66% in 2003. Table 8.6 shows wastewater treatment coverage by the companies that provide sanitary services to 99.6% of the urban population. It is projected that the wastewater treatment coverage will increase from 66% in 2003 to 99% in 2014, according to national wastewater treatment plans.

Industrial effluents have been recognized from a total of 2500 industries, with an estimated 1780 of these being potential wastewater generators. The main receiving water bodies are rivers (67%), lakes (3%), coastal waters (6%), irrigation open channels (2.9%), and soil (14.4%) and surface waters (15%). Since the year 2000 a new regulation for industrial discharges to marine and continental waters has been in place (DS00/2000), in an attempt to control physical and chemical parameters, organics, metals, microbiological parameters and nutrients. Pollution from agriculture operations has been linked to fertilizer and pesticide application. The diffuse nature of this kind of pollution makes it difficult to estimate its contribution to environmental levels, at least in terms of nutrients, although lakes in the Central Zone in particular have been affected by eutrophication potentially attributed to agriculture. The scale of pollution associated with Chile's important mining industry is only now becoming recognized and documented. For example, in the northern administrative region IV, an aging copper smelter, the government agency for geology and mining, has detected a total of 343 mineral tailings as of December 2004 (Table 8.7). About 25% of these waste deposits were abandoned active plants, 42% were related to dismantled plants, and 33% were related to inactive operations. Overall, about one half of the tailings did not have a permit, and just 2% had an environmental permit. Efforts are just starting to assess the extent of the problem of contaminated sites of all types throughout the country, and it can be expected that over the next few years this issue will become important as the public starts to evaluate the risks derived from their existence.

Another environmental factor linked in part to water resources concerns the establishment and management of protected areas in each catchment (Fig. 8.4). Chile has a natural parks system established by law (SNASPE: the acronym means national system of state-protected land). Currently, it covers 14 million hectares, equivalent to 19% of the country. The system establishes three management categories: national parks (31), national reserves (48) and natural monuments (15). A national park is an extensive area where unique, distinctive or representative biodiverse environments of Chile are found. National
### Table 8.4. General description of the geological factors and related water quality parameters for selected catchments

<table>
<thead>
<tr>
<th>Catchment</th>
<th>Geological factors</th>
<th>Water quality parameters</th>
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<tbody>
<tr>
<td><strong>North Zone</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lauca</td>
<td>Chemical composition is determined by high salinity overflows from Cotacotani lake and llixiviates from ignimbrites and andesites. High carbonates, calcium and magnesium suggest connection with dolomitic rocks from Osaya formation. High solar radiation coupled and small flows produce increased salt concentrations.</td>
<td>Good to fair Salinity, boron, arsenic, cyanide. Lauca river in Japa: 2 m^3/s; boron: 2 ppm; conductivity: 1000 umho</td>
</tr>
<tr>
<td>Tarapacá</td>
<td>The upper part of the catchment presents high concentrations of metals. Metallogenetic streaks are believed to be the biggest contributors. Soil quality and high solar radiation produces high salinity. Water quantity and quality are variable and depend on the season.</td>
<td>Good to poor Salinity, boron Tarapacá river in Pachica: 0.05–0.5 m^3/s; sodium: 300 ppm; chloride: 800 ppm; boron: 13 ppm; conductivity: 2000 umho</td>
</tr>
<tr>
<td>Loa</td>
<td>High arsenic and salts content occurring naturally due to Salado river (salado means salty in Spanish). Intense rain events have been coupled to water quality episodes involving extremely high arsenic and metal concentrations in suspended solids. Although some studies indicate anthropogenic causes, the evidence is not conclusive.</td>
<td>Good to poor Sodium, chloride, boron, arsenic Loa river in Quillagua: boron: 40 ppm; conductivity 8000 umho</td>
</tr>
<tr>
<td>Salar de</td>
<td>High concentrations of salts with major contribution from local soil lithology (evaporites). Potassium, boron and lithium are mined in evaporation ponds. Detailed geochemical studies are being carried out by mining companies, involving groundwater flow models coupled to geochemical balances.</td>
<td>Good to fair Metals and ions</td>
</tr>
<tr>
<td>Atacama</td>
<td>Local lithology coupled to low stream flows introduces ions along the watershed. Moderate salinity, boron, arsenic, low sodium, high sulphate.</td>
<td>Good Metal and ions</td>
</tr>
<tr>
<td>Copiapó</td>
<td>Two metallogenetic streaks, F10 and F4, contribute strongly to the background metal concentrations. Quality determined by groundwater from the Pucarí reservoir.</td>
<td>Good</td>
</tr>
<tr>
<td>Elqui</td>
<td>Metallogenetic streaks contribute to background metal concentrations. Quality determined by groundwater from the Pucarí reservoir.</td>
<td>Good</td>
</tr>
<tr>
<td>Limari</td>
<td>Local lithology contributes to background metal concentrations. Shallow groundwaters impact surface water quality with high chlorine, sulphate and magnesium.</td>
<td>Good to poor Boron, arsenic, sulphate. Conductivity: 2000 umhos</td>
</tr>
<tr>
<td>Ptoceo</td>
<td>Metallogenetic streaks contribute to background metal concentrations, particularly manganese and copper. Abundant sedimentary material from colluvial and alluvial deposits. High sediment content during storm events due to low vegetation.</td>
<td>Good to fair Metals and ions</td>
</tr>
<tr>
<td><strong>Central Zone</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Aconcagua</td>
<td>Metallogenetic streaks contribute to metal concentrations, particularly manganese and iron.</td>
<td>Good to poor Manganese, iron, copper</td>
</tr>
<tr>
<td>Maipo</td>
<td>Local lithology produces background concentrations of copper, aluminium, chromium, lead and molybdenum, and alkaline conditions. Groundwater adds sulphate and manganese downstream.</td>
<td>Good to poor Copper, aluminium, chromium, lead and molybdenum High calcium, magnesium and sulphates.</td>
</tr>
<tr>
<td>Rapel</td>
<td>Metallogenetic streaks contribute to metal concentrations, particularly copper, manganese and molybdenum.</td>
<td>Good to poor Copper, manganese, molybdenum</td>
</tr>
<tr>
<td>Maule</td>
<td>Rich clay and limestones from volcanic origin (verificar)</td>
<td>Good to fair</td>
</tr>
<tr>
<td>Biobío</td>
<td>High rainfall and vegetation coverage produces outstanding water quality.</td>
<td>Excellent to fair</td>
</tr>
<tr>
<td><strong>South Zone</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Imperial</td>
<td>Volcanic activity strongly controls water quality, especially the active Llaima volcano. 'Trumao', a characteristic soil in this catchment, introduces acidic conditions. However, lower in the catchment, limestone introduces alkalinity, counteracting the effects of the trumao.</td>
<td>Good to poor</td>
</tr>
<tr>
<td>Valdivia</td>
<td>Some metal contribution from groundwaters from soil leaching.</td>
<td>Excellent to good</td>
</tr>
<tr>
<td>Maulin</td>
<td>Organic contribution from 'ñadi' soil.</td>
<td>Excellent to good</td>
</tr>
<tr>
<td>Aysen</td>
<td>High aluminium contents in the waters are due to clay particles. Chromium, copper, boron and manganese are also elevated.</td>
<td>Good Alumium, copper, boron, chromium and manganese</td>
</tr>
<tr>
<td>Serrano</td>
<td>High water quality from ice and snowmelt. Moraine material is carried down the stream.</td>
<td>Excellent to good</td>
</tr>
<tr>
<td>Cianes</td>
<td>Some copper and aluminium from local ore deposits.</td>
<td>Excellent</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Catchment</th>
<th>Human activities</th>
<th>Investment* (billions of US$)</th>
<th>Protection areas†</th>
</tr>
</thead>
<tbody>
<tr>
<td>North Zone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lauca</td>
<td>Tourism, agriculture</td>
<td>1. Mining: 20.6</td>
<td>NP: Lauca</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2. Energy generation: 2.9</td>
<td>NR: Las Vicuñas</td>
</tr>
<tr>
<td>Tarapacá</td>
<td>Mining, agriculture</td>
<td>3. Manufacture: 1.9</td>
<td>None</td>
</tr>
<tr>
<td>Loa</td>
<td>Mining, agriculture</td>
<td>4. Land development: 1.6</td>
<td>CS: Loa estuary and Quillagua Oasis</td>
</tr>
<tr>
<td>Salar de Atacama</td>
<td>Mining, agriculture</td>
<td></td>
<td>NR: Los Flamencos</td>
</tr>
<tr>
<td>Copiapó</td>
<td>Mining, agriculture, commerce</td>
<td>5. Sanitation: 0.7</td>
<td>CS: Copiapó estuary and the blooming desert</td>
</tr>
<tr>
<td>Elqui</td>
<td>Agriculture, mining</td>
<td></td>
<td>CS: Condroiasco, Punta de Teatinos, Estero Guanta, Arrayán, Quebrada Honda, and Estero Derecho.</td>
</tr>
<tr>
<td>Limari</td>
<td>Agriculture, mining</td>
<td></td>
<td>NP: Fray Jorge</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>NM: Fichasca</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>CS: El Durazno and Cuesta El Espino</td>
</tr>
<tr>
<td>Petorca</td>
<td>Mining, agriculture</td>
<td></td>
<td>CS: Altos de Petorca, Alicahue, and Santa Inés</td>
</tr>
<tr>
<td>Central Zone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Aconcagua</td>
<td>Agriculture, mining, industry</td>
<td>1. Land development: 7.6</td>
<td>NP: La Campana</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2. Energy generation: 5.6</td>
<td>NR: Río Blanco</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3. Manufacture: 2.7</td>
<td>CS: Cordillera Melón</td>
</tr>
<tr>
<td></td>
<td></td>
<td>4. Forestry: 2.4</td>
<td></td>
</tr>
<tr>
<td>Maipo</td>
<td>Services and manufacture</td>
<td></td>
<td>NM: El Morado</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2. Energy generation: 5.6</td>
<td>NR: Río Claro, national reserve</td>
</tr>
<tr>
<td></td>
<td></td>
<td>4. Forestry: 2.4</td>
<td></td>
</tr>
<tr>
<td>Rapel</td>
<td>Agriculture and mining</td>
<td>5. Sanitation: 2.1</td>
<td>NP: Las Palmas de Cocalán</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>NR: Río Cipreses</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>CS: Las Cardillas and Alto Huemal</td>
</tr>
<tr>
<td>Biobio</td>
<td>Manufacture, agriculture, service</td>
<td></td>
<td>NP: Laguna del Laja</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>NR: Nuble, Maluco, Malalcahuello, and Ralco.</td>
</tr>
<tr>
<td>South Zone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Imperial</td>
<td>Agriculture, services</td>
<td>1. Manufacture: 3.1</td>
<td>NP: Conguillio</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2. Forestry: 2.6</td>
<td>NR: Malalcahuello</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>NM: Contulmo</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>CS: Cerro Andeul, Monteul, Ñielol-Rucamanque, Purén, and Villa Las Araucarias.</td>
</tr>
<tr>
<td>Valdivia</td>
<td>Agriculture, cattle</td>
<td>3. Aquaculture: 1.6</td>
<td>NR: Mocho-Choshuenco</td>
</tr>
<tr>
<td></td>
<td></td>
<td>4. Land development: 0.6</td>
<td>MN: Valdivia</td>
</tr>
<tr>
<td></td>
<td></td>
<td>5. Energy generation: 0.5</td>
<td>CS: Curinanco, Cordiller a de la Costa.</td>
</tr>
<tr>
<td>Maullín</td>
<td>Tourism, fishing</td>
<td></td>
<td>NP: Vicente Pérez Rosales</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>MN: Llanquihue</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>CS: Cascadas-Volcán Osorno, Río Maullín</td>
</tr>
<tr>
<td>Aysén</td>
<td>Fishing, mining</td>
<td></td>
<td>MN: Dos Lagunas</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>NR: Trapamada, Río Simpson, Coyhaique, and Cerro Castillo</td>
</tr>
<tr>
<td>Serrano</td>
<td>Tourism, cattle</td>
<td></td>
<td>NP: Torres del Paine, Bernardo O’Higgins</td>
</tr>
<tr>
<td>Cisnes</td>
<td>Tourism, cattle</td>
<td></td>
<td>NP: Queulat</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>NR: Lago Las Torres, Lago Carlota.</td>
</tr>
</tbody>
</table>

*Investment in projects that entered the environmental impact assessment system between 1992 and 2004
†NP, national parks; NR, national reserve; NM, natural monument; CS, conservation site
Data for investment in projects that entered the environmental impact assessment system were obtained from the Conama database. Data for economic activities and protected areas in each catchment were adapted from Cade-Idepe (2004).
Table 8.6. Projections of wastewater treatment coverage until 2014

<table>
<thead>
<tr>
<th>Administrative region</th>
<th>Company</th>
<th>Urban population in 2003</th>
<th>Percentage of national urban population</th>
<th>With wastewater treatment (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>2003</td>
</tr>
<tr>
<td>North Zone</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>I</td>
<td>Essat S.A.</td>
<td>415 247</td>
<td>2.9</td>
<td>96.1</td>
</tr>
<tr>
<td>II</td>
<td>Essan S.A.</td>
<td>465 104</td>
<td>3.3</td>
<td>98.9</td>
</tr>
<tr>
<td>III</td>
<td>Emessat S.A.</td>
<td>245 514</td>
<td>1.7</td>
<td>77.5</td>
</tr>
<tr>
<td>IV</td>
<td>Escco S.A.</td>
<td>520 116</td>
<td>3.7</td>
<td>95.2</td>
</tr>
<tr>
<td>Central Zone</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>V</td>
<td>Essal S.A.</td>
<td>1 430 500</td>
<td>10.1</td>
<td>77.6</td>
</tr>
<tr>
<td>V</td>
<td>Coopagua Ltda.</td>
<td>4400</td>
<td>0</td>
<td>37.7</td>
</tr>
<tr>
<td>Metropolitan Region</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Aguas Andinas S.A.</td>
<td>5 561 081</td>
<td>39.3</td>
<td>64.2</td>
<td>89.4</td>
</tr>
<tr>
<td>Smapa</td>
<td>618 446</td>
<td>4.4</td>
<td>98.4</td>
<td>100</td>
</tr>
<tr>
<td>Aguas Cordillera S.A.</td>
<td>433 095</td>
<td>3.1</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Servicomunal S.A.</td>
<td>71 882</td>
<td>0.5</td>
<td>88.5</td>
<td>88.9</td>
</tr>
<tr>
<td>Aguas Manquehue S.A.</td>
<td>17 801</td>
<td>0.1</td>
<td>46.4</td>
<td>46.8</td>
</tr>
<tr>
<td>Aguas Los Dominicos S.A.</td>
<td>15 653</td>
<td>0.1</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>VII</td>
<td>Aguas Nuevo Sur Maule S.A.</td>
<td>633 809</td>
<td>4.5</td>
<td>34.5</td>
</tr>
<tr>
<td>VI and VIII</td>
<td>Essbio S.A.</td>
<td>2 145 702</td>
<td>15.2</td>
<td>75.4</td>
</tr>
<tr>
<td>South Zone</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>IX</td>
<td>Essar S.A.</td>
<td>594 516</td>
<td>4.2</td>
<td>12.9</td>
</tr>
<tr>
<td>X</td>
<td>Essal S.A.</td>
<td>538 585</td>
<td>3.8</td>
<td>50.7</td>
</tr>
<tr>
<td>XI</td>
<td>Aguas Décima S.A.</td>
<td>131 758</td>
<td>0.9</td>
<td>92.2</td>
</tr>
<tr>
<td>XII</td>
<td>Aguas Patagonia de Aysen S.A.</td>
<td>69 343</td>
<td>0.5</td>
<td>71.4</td>
</tr>
<tr>
<td>XII</td>
<td>Esmag S.A.</td>
<td>148 986</td>
<td>1.1</td>
<td>10.6</td>
</tr>
<tr>
<td>National</td>
<td></td>
<td>14 061 538</td>
<td>99.4</td>
<td>66</td>
</tr>
</tbody>
</table>

Source: Superintendencia de Servicios Sanitarios (SISS).

Table 8.7. Classification of mineral tailings in administrative region IV according to plant status and permits, as of December 2002

<table>
<thead>
<tr>
<th>Plant status</th>
<th>Number of Plants</th>
<th>Tailings with Total mining agency permit</th>
<th>Tailings without mining agency permit</th>
<th>Total permit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Operating</td>
<td>32</td>
<td>31</td>
<td>51</td>
<td>83</td>
</tr>
<tr>
<td>Dismantled</td>
<td>82</td>
<td>78</td>
<td>65</td>
<td>143</td>
</tr>
<tr>
<td>Stopped</td>
<td>59</td>
<td>59</td>
<td>56</td>
<td>117</td>
</tr>
<tr>
<td>Total</td>
<td>173</td>
<td>168</td>
<td>172</td>
<td>343</td>
</tr>
</tbody>
</table>

Source: Sernageomin.

Environmental policy and regulations

The environmental agenda proposed by the government for 2004-2006 views pollution control and improvement of environmental quality as a high priority. The programme for the control of aquatic contamination involves the following activities: (i) identification of water bodies with environmental priority due to existing or threatened pollution; (ii) development of formal control instruments; (iii) development of sanitation programmes and decontamination bonds; (iv) enforcement of environmental rules and standards; (v) development of institutional capabilities; (vi) development of better practices for environmental data management, such as a discharge and contaminant transfer register; and (vii) development of integrated water management plans in priority catchments. All these lines of work complement the international agenda, particularly bilateral agreements (e.g. Chile-Canada, free trade agreement with the USA) and collaboration with international organizations (e.g. IADB, PNUMA, OEIA, UNESCO and the World Bank). The standards with highest priorities that will be developed in the next few years are: (i) secondary water quality standards for the priority catchments, Loa, Elqui, Aconcagua, Maipo-Mapocho, Cachapoal, Biobío, Llanquihue, Aysén and Serrano; (ii) secondary water quality standards for the priority coastal waters, namely VII region sounds, XI region fjords, and XII region sediments.

The natural resources in national parks should be only minimally altered by human action and have a self-perpetuating condition. Biotic species or geological formations in national parks have a special educational, recreational and scientific interest. A national reserve is an area where natural resources must be used and preserved with special concern, since they may be prone to degradation or they may be important for the wellbeing of the community. A natural monument is a generally limited area with singular interest from the scenic, cultural or scientific point of view, where native species live or where relevant geological sites are found. In terms of management, the most important characteristic of the system is that all natural resources in national parks, including vegetation, wildlife and water resources, cannot be used for economic purposes. Furthermore, natural resources in national parks must be protected, whereas natural resources in national reserves may be used with sustainability. New projects that are located in protected areas and need to enter the national environmental impact assessment system (SEIA) are subject to careful scrutiny prior to approval, and extensive monitoring during their operation, in order to ensure that their operation is compatible with the objectives of the protected area.
A framework environmental law was passed in 1992, which started an ambitious programme that defines a set of standards that will control wastewater emission and environmental water quality for Chilean water resources. All new wastewater discharges to surface waters, groundwater and sewers must meet the new standards, whereas the old discharges must submit a programme designed to meet the same standards. Figure 8.9 shows how any type of wastewater discharge is regulated. Discharge to sewer systems is regulated by DS609/98 (NE1), industrial and treated water discharges to rivers or lakes are regulated by DS90/2000 (NE2), and discharges from wastewater treatment plants (including treated waters) to groundwater are regulated by DS46/2002 (NE4). The regulations may also include specific discharge standards like the molybdenum and sulphate standard that is being developed for Caren Creek. Overall, considerable progress is being made in assuring a high standard of water quality in Chile, although given the widespread natural and anthropogenic contamination of hydrological systems in the north, in places there is still much work to be done.
Neotectonics

JOSE CEMBRANO, ALAIN LAVENU, GONZALO YAÑEZ (coordinators), RODRIGO RIQUELME, MARCELO GARCÍA, GABRIEL GONZÁLEZ & GERARD HÉRAIL

Nazca–South American plate interaction provides a classic example of a subduction-type convergent margin (e.g. Dewey & Bird 1970), in which the continental lithosphere overrides the oceanic plate. Evidence of subduction processes having taken place along the western margin of South America since at least Triassic times has been thoroughly described in the literature (e.g. Mpodozis & Ramos 1989) and elsewhere in this book. Although subduction has been an essentially continuous process along the Andes, its impact on the geological evolution of the continent varies with time and along-strike (e.g. Jordan et al. 1983b, 2001). Plate kinematics, the subduction of passive and/or active ridges, fracture zones, plate age at the trench, and climate change have all been invoked as controlling factors for continental plate geological evolution and segmentation (e.g. Jarrard 1986; Gutscher et al. 2000b; Yañez et al. 2001; Lamb & Davis 2003; Yañez & Cembrano 2004; Sobolev & Babeyko 2005). Figure 9.1 shows the tectonic and morphologic elements that shape the geological evolution of the Andean margin. The tectonic segmentation shown on this figure provides a useful reference for the geological evolution of the Andes.

The Andean mountain chain trends NNE–SSW between latitudes 18°S and 48°S with slight local variations in strike but with significant latitudinal changes in morphotectonic configuration (Fig. 9.1). Between 18°S and 28°S, the Andes consist, from west to east, of the Coastal Cordillera, Central Depression, Precordillera and Western (Main) Cordillera with the Altiplano and Puna to the east. The region between 28°S and 33°S, where the subducting slab is subhorizontal, has no Central Depression or active volcanism, and as a counterpart there are Neogene uplifted blocks of the broken foreland cropping out widely in NW Argentina (Jordan et al. 1983b). South of 33°S, the Chilean Andes consist of a Coastal Cordillera, a Central Valley and a Main Cordillera that is called the Patagonian Cordillera south of 39°S (Fig. 9.1). On the outer forearc, the coastline is characterized by Quaternary marine terraces that have been uplifted to over 780 m a.s.l. in Peru (Ortlieb & Macharé 1990).

In northern Chile, the Altiplano and Main Cordillera show the overall highest elevations of the Andes, with mean elevations of 4000 m and some mountains reaching over 6000 m. In contrast, the southern Chilean Andes have an average elevation of 1500 m and only locally reach 4000 m. These gross differences in topography are consistent with a crustal thickness of almost 70 km under the Altiplano and Western Cordillera but only 30 km in the south Andes (Tassara & Yañez 2003).

The main crustal-scale margin-parallel fault zones of the Chilean Andes are the Mesozoic–Cenozoic Atacama Fault System (AFS) in the Coastal Cordillera of northern Chile, the Cenozoic Domeykó Fault System in the Precordillera of northern Chile and the Mesozoic–Cenozoic Liquiñe–Ofqui Fault Zone (LOFZ) in the Main Cordillera of southern Chile. The AFS is the most important tectonic feature of the Chilean forearc, whereas the LOFZ is an intra-arc fault that extends along the margin of the Miocene–Pliocene and present-day volcanic arcs (Fig. 9.1).

In the first part of this chapter the relative importance of plate kinematics and related controlling factors are reviewed. The second part deals with the interplay between Neogene tectonic activity and landscape evolution of the central Andes forearc. Finally, the third and last section examines the stress field within the central and southern Chilean Andes since Pliocene times, using kinematic analysis of neotectonic fault systems.

Late Tertiary plate kinematics

Nazca (Farallon)–South American plate interaction has been documented in the literature since the mid-1980s, based on analysis of seafloor spreading magnetic anomalies (e.g. Pilger 1983; Pardo-Casas & Molnar 1987; Cande et al. 1987; Tebbens & Cande 1997). Recently, Somoza (1998) carried out a new compilation for late Tertiary kinematics of Nazca–South America plate interaction, taking into account all the previous reconstructions and recent information in the Antarctic–Africa–South American area, as well as applying the Nuvel1-A solution for present constraints. Figure 9.2 shows the kinematics of Nazca (Farallon)–South American Plate convergence using the relative poles of rotation of Somoza (1998). Results are computed for the five tectonic segments of the Andean convergence. Some first-order conclusions are clearly displayed in this figure, the most obvious of which is the dramatic change in convergence 25 Ma ago after the break-up of the Farallon Plate. Prior to that time, convergence was relatively slow (c. 8 cm/year) and with a large obliquity (c. 50°E), favouring the development of dextral strike-slip faulting in the continental plate. Later on, the convergence increased its rate (to more than 14 cm/year), reducing the convergence angle to values in the order of 10–20° (measured counterclockwise from the trench orthogonal). This high rate of convergence between 26 and 10 Ma roughly coincides with an extensional phase recognized in the forearc region of the southern Andes (Jordan et al. 2001). High rates of convergence and the development of an extensional regime contradict the common perception that shortening is linked to high convergence rate regimes; this is an important issue which is linked to plate coupling and is discussed later. During the last 20 Ma, convergence rate has decreased steadily from 12 cm/year to values in the order of 7 cm/year after the last 5 Ma (e.g. Somoza 1998, 2005), although the snapshot provided by GPS results (Angermann et al. 1999) indicates a present-day value of only 6.3 cm/year (Angermann et al. 1999; Kendrick et al. 2003). It is important to note that this decrease in the convergence rate is coincident with inversion of the Miocene basin developed during the previous extensional regime (Jordan et al. 2001), as well as being approximately coincident with the building of the main Andean mountain chain.
and widespread Pliocene compression. Since 20 Ma, obliquity has remained at low values (c. 10–15º), but with some increases between 20 and 15 Ma and within the last 5 Ma (c. 20ºE). Along-strike variations are second-order features compared to the general pattern described in the previous paragraphs. One of the most relevant second-order features of the convergence regime is the duration and timing of the high convergence rate period. Whereas in the northern segment it lasted for more than...
10 million years with a peak at 22 Ma, in the southern region the convergence peak occurred 3 million years earlier, and lasted less than 10 million years. Such a difference can be attributed to along-strike variations of plate coupling and continental lithosphere response.

The role of the Juan Fernandez Ridge in the flattening of the subducting slab has been recently revisited (see for instance Pilger (1981, 1983), Gutscher et al. (2000b), Yañez et al. (2001, 2002), and elsewhere in this chapter). Flat-slab segmentation explains both the absence of volcanism in the active margin and the transition to a more compressive regime where intrusive bodies are emplaced (e.g. Hollings et al. 2003), and deformation migrates to the foreland region (e.g. Jordan et al. 1983b). The kinematics of the Juan Fernandez Ridge during late Tertiary times has been presented in Yañez et al. (2001). Figure 9.3 shows a summary of the ridge path during this time period, considering the hot-spot reference frame proposed by Gordon & Jurdy (1986). Although a new plate reconstruction presented by Somoza (1998) incorporates some variations in the plate kinematics (especially within the last 10 million years), the main features of the plate interaction remain the same as represented in the Gordon & Jurdy (1986) model. The collision of the Juan Fernandez Ridge with the South American margin migrated southward at a rate of more than 200 km/Ma during the time

![Image](image_url)
seismic expressions of unusual mechanical behaviour including phase transitions and water release from a probably saturated crust. Excessive water release may also be relevant to the anomalous magmatism associated with the source of the Miocene giant copper deposits of the High Andes (e.g. Garrido et al. 2002).

**Active and passive ridge collision and migration: the connection with the flat-slab segmentation**

Several models have been postulated to explain flat-slab segmentation along active margins (for a review see Gutscher 2002); however, a growing amount of empirical and theoretical evidence (e.g. Yañez et al. 2001; Gutscher 2002) suggests a direct link with the larger buoyancy of subducted passive ridges. The subduction of the oceanic plate underneath the continental plate results simply from buoyancy differences, with the density of the subducting slab being mostly controlled by the advection of low temperature isotherms associated with the cooling path of the oceanic plate (e.g. McKenzie 1967). This cooling path is perturbed when a hot-spot generates a volcanic chain that becomes a passive ridge younger and more buoyant than the surrounding plate. However the dip angle of subduction is the result of a delicate equilibrium between the thermally/compositionally controlled negative buoyancy of the subducting slab and the pressure forces associated with the asthenospheric corner flow (e.g. Turcotte & Schubert 2002).

The greater buoyancy attributed to the subduction of passive ridges produces both thermal resetting and phase changes with depth. Thermal resetting is a transient phenomenon that depends on the size of the heat source and the time span between the thermal event and the subduction of the ridge. Assuming a linear heat source of 10 km radius, the thermal effect decays to 20% in 6 million years, whereas for a 25 km radius the thermal decay takes more than 40 million years.

On the other hand the subducting plate is not uniform mantle pyrolite but, because of melting at the ridge axis, it has segregated into a basaltic ocean crust (c. 5 km thick), and residual harzburgite underlain by ordinary pyrolite (e.g. Ringwood 1991). According to Ringwood (1991), thermally equilibrated density contrasts, with respect to the reference pyrolite mantle, are positive in the basaltic layer down to the 650 km discontinuity (200 kg/m$^3$ in the first 400 km and 50 kg/m$^3$ in the deeper part), below which it becomes negative (−200 kg/m$^3$). On the other hand the depleted harzburgite is less dense than the reference mantle down to 650 km depth (−80 kg/m$^3$ in the first 400 km and −30 kg/m$^3$ downwards), and below the 650 km transition zone it is more or less neutral. In consequence, the net buoyancy difference between normal and thicker oceanic lithosphere is determined by the thickness of the basaltic layer and the depleted harzburgite mantle.

The thickness of the basaltic layer underneath the Juan Fernandez Ridge is about 8–10 km (Contreras & Vera 2003), whereas the thickness of the depleted harzburgite mantle below is not well constrained. If we assume a linear relationship with the thickness of the enriched basaltic layer, we can estimate a thickness of at least 50 km for this depleted harzburgite layer (which is normally about 25 km: Ringwood 1991). The net effect on the buoyancy of the subducting plate is a density contrast of −16 kg/m$^3$ for a 650-km subducting plate of 100 km thickness. In addition, the shape of the Juan Fernandez Ridge (Contreras & Vera 2003) provides a first-order estimate of the heat source size (c. 10 km radius). Its effects at the trench have a time lag of 11 Ma (Yañez et al. 2001), so that the temperature of the thermally perturbed slab would rise about 10%, producing a density contrast of −5 kg/m$^3$. Figure 9.4 shows the theoretical torque associated with the pressure and buoyancy forces as a function of the dip angle for realistic values (see figure captions for details). Both competing effects decay as the slab becomes thinner with time.

**Fig. 9.3.** Upper panel: predicted track of the Juan Fernandez Ridge in a hot-spot reference frame (rotation poles from Gordon & Jurdy 1986). Lower panel: the trace of the Juan Fernandez Ridge (segmented blue line) over the seismicity of central Chile (modified from Pardo et al. 2003b). In the upper panel the time evolution of the Juan Fernandez Ridge and the South American margin are shaded from dark (present time) to light (22 Ma before present), with a time interval of 1 Ma.

Notice that the South American margin is slowly moving to the west, whereas the track of the Juan Fernandez Ridge is moving towards the east at a much faster rate (at least three times faster). Seismicity in the lower panel is indicated by light dots, and the contours of the Benioff plane by dashed white lines. Notice the good correspondence between the track of the Juan Fernandez Ridge and the cluster of seismicity aligned in the ENE direction. Focal mechanism along this seismic cluster is variable as indicated by some selected events.
vertical, although when the density contrast is larger the negative buoyancy torque is consistently much greater. An equilibrium condition is achieved for progressively smaller dip angles when the density contrast is being reduced. Focusing on the Juan Fernandez Ridge, the semi-quantitative result of Figure 9.4 indicates that a reduction of the order of $-21 \text{ kg/m}^3$ reduces the net buoyancy by more than 40%, this in turn inducing a reduction in the dip angle of $15–20^\circ$.

The path of the Juan Fernandez Ridge during the last 20 million years, shown in Figure 9.3, predicts a southward migration of the flat-slab during this period. Present evidence for the presence of the Juan Fernandez Ridge is recorded in the anomalous seismicity pattern along the track of the ridge underneath the continent (Fig. 9.3). Focal mechanisms from Pardo et al. (2003b) suggest a complex interaction with the upper plate and metamorphic effects linked with water release as the subducted plate heats up. Geological evidence (Kay & Mpodozis 2002) consistently shows that magmatism related to flat-slab subduction migrates southward with the same timing as that of the Juan Fernandez Ridge path. It may be concluded, therefore, that flat-slab segmentation is a transient phenomenon that depends on the anomalously enhanced thickness of subducted oceanic crust, and on the associated release of heat.

**The role of plate coupling in the build-up of the Andes**

Two different hypotheses have recently been presented to explain the role of plate coupling on Andean orogenesis. Yañez & Cembrano (2004) postulate that plate coupling is mostly controlled by the age of the oceanic plate and only secondarily by convergence velocity. Stronger coupling is achieved for older (cooler) plates and slower convergence regimes. On the other hand, Lamb & Davis (2003) predict that sediment fill reduces the shear stress at the trench, weakening plate coupling, but increasing the down-dip extent of the coupling zone. In this latter model, late Cenozoic changes in sediment fill linked to climate change triggered the uplift of the Andes. Both effects successfully explain the uplift of the Altiplano region in the last 20 Ma, and both are probably reinforcing the other where older oceanic crust and the sediment-starved trench are spatially coincident (Fig. 9.5). However, it is argued here that, as a general mechanism, the climate hypothesis has some limitations, notably that it is not relevant to pre-late Cenozoic time. Prior to the opening of the Drake Passage (late middle Eocene–early Miocene) climate conditions in the western margin of South America ranged from tropical rainforest and paratropical rainforest to woodland (Frakes et al. 1992). Also, given the fact that present relief is a first-order expression of plate coupling, the rather gradual southward reduction of relief contrasts with the sharp reduction of trench fill to the north of $33^\circ$S (see in Fig. 9.1 the transition between a deep trench to the north of this latitude). In addition, tectonic erosion processes carry a large amount of sediment downwards, in almost the same fashion as filled trench environments (C. R. Ranero, pers. comm., 2005). Plate coupling, mainly controlled by the age of the oceanic plate and the convergence velocity, also provides a reasonable explanation for the apparently contradictory nature of the plate kinematics previously mentioned: higher compressive regime when the convergence rate is lower and vice versa. Higher convergence velocity decreases the age of subducted crust at the trench and also reduces the effective viscosity at the slip zone (given a power law rheology at the plate contact zone), reinforcing a reduction in plate coupling. The opposite is consistently true for slowing the convergence velocity, which enhances higher coupling. A sharp transition in the plate coupling is predicted across offset fracture zones of great age, such as the Challenger Fracture Zone in Central Chile (Fig. 9.1). In this
case the model predicts the occurrence of c. east–west trending deformation zones in the overriding plate that accommodate the deformation, expressed for example as the Melipilla Fault Zone at the latitude of Valparaiso (33.5ºS) (Yañez & Cembrano 2002), and the origin of at least some of these can be linked to this differential coupling across fracture zones offsetting oceanic crust with markedly differing age.

**Discussion**

Nazca–South America plate interaction during late Tertiary times has had a fundamental impact on the geological evolution of the active margin of South America, with transient events modulating the long-lived continuous activity of the subduction factory. Long-term geological effects on the overriding plate are probably controlled by pre-existing heterogeneities on a lithospheric scale (e.g. Tassara & Yañez 2003); these regulate the way in which the continental plate accommodates the deformation forced by the tectonic boundary conditions. Key elements in the stress regime at the continental lithospheric margin are absolute plate velocities, plate coupling and subduction of buoyant slabs (e.g. Yañez & Cembrano 2004; Sobolev & Babeyko 2005). All these parameters are inherently variable and episodic with time and space. A good example of this episodicity is shown in the evolution of the Nazca (Farallon) Plate during late Tertiary times. Prior to the break-up of the Farallon Plate (25 Ma) the oceanic plate was moving in a NE direction, producing highly oblique convergence at the plate margin. With the plate moving NE, the presence of the Juan Fernandez hot-spot in the middle of the Pacific generated a volcanic chain aligned in the same direction. After the break-up of the Farallon Plate, however, a new regime was established with a much faster and highly oblique convergence. This abrupt change resulted in the subduction of the Juan Fernandez Ridge underneath the continent and the arrival of progressively younger (and hotter) oceanic crust at the South American trench. Subduction of more buoyant oceanic lithosphere allowed the generation of a flat-slab segment that migrates southward, following the path of the thermally and compositionally perturbed subducted slab. Plate segmentation by the flat-slab has had a pronounced effect on the geological evolution of the continental margin. In the flat-slab segment the reduction of the asthenospheric wedge induces an eastward migration of the volcanism and the deformation front. Such tectonomagmatic events probably favoured the development of the giant late Miocene ore deposits of the main Cordillera (Garrido et al. 2002). The much faster convergence rate and the arrival of younger oceanic crust at the trench also reduces the coupling between plates (Yañez & Cembrano 2004), enhancing the development of an extensional regime in the forearc region and therefore the accumulation of a thick pile of sediments (i.e. Abanico Formation). Reduction in the plate convergence in the last 10 million years induced the opposite effect in plate coupling, with progressively older oceanic crust at the trench and lower strain rate at the slip layer. This new tectonic boundary condition generated the inversion of the previously developed basin and regional uplift of the Andes.

The generalized description of late Tertiary Nazca (Farallon)–South American plate interaction outlined in the previous paragraph emphasizes the relevance of episodic events in the geological evolution of the continental margin. Thus, from 40 to 25 Ma there was a highly oblique convergence at a rate of c. 8 cm/year, then a more orthogonal and higher convergence rate of 14 cm/year (25–15 Ma), and finally a slowdown in convergence velocity to 7 cm/year (< 12 Ma). Such transient events are mostly controlled by heterogeneities and kinematic changes in the oceanic plate. The geological record in the Andean evolution is in fact plagued by episodic events restricted in time and space. For instance, the episodic slowdown in plate convergence rates since 20 Ma is clearly coeval with discrete shortening events of the Andes build-up, as will be documented in the following sections.

**Neogene tectonic activity and landscape evolution of the central Andes forearc, northern Chile (R.R., M.G., G.G. & G.H.)**

This section examines Neogene tectonic activity in northern Chile, between 18ºS and 28ºS, and its relation to the present-day landforms. This region is emphasized specifically because it has remarkable and well exposed examples of interactions between tectonic activity and morphogenesis. Exceptional preservation and exposure of the landscape has been made possible due to the arid to hyperarid climatic conditions in the region which have prevailed at least since middle Miocene times (Sillitoe & McKee 1996; Alpers & Brimhall 1988, 1989; Hartley & Chong 2002). The Neogene tectonic activity of the region has been studied through analysis of landscape evolution, although the exact nature of the deformation mechanisms operating in the forearc and arc during the Neogene is still far from being fully understood. However, there is no doubt that relating tectonics to geomorphological evolution has the potential to better constrain the deformation history and give new insights into recent regional deformation mechanisms of northern Chile.

**Geomorphological and geological setting**

In the forearc region of northern Chile, between the latitudes of Arica (18ºS) and Copiapó (28ºS), five north–south elongated physiographic units can be differentiated. These units are, from west to east, the Coastal Cordillera, the Central Depression, the Precordillera, the Preandean Depression and the Western Cordillera (Fig. 9.6). The geological record of these physiographic units shows eastward migration of a subduction-related magmatic arc from the Jurassic period to the present. The present-day position of the magmatic arc in the Western Cordillera was reached about 25 Ma (e.g. Coira et al. 1982). The Mesozoic to Cenozoic migration of the magmatic arc has
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Fig. 9.6. Location and physiographic units of northern Chile: CC, Coastal Cordillera; CD, Central Depression; DC, Domeyko Cordillera; P, Precordillera; PD, Preandean Depression; WC, Western Cordillera. The main fault systems of the forearc region are shown: the Atacama Fault System (AFS) in the Coastal Cordillera and Domeyko Fault System (DFS) in the Precordillera and Domeyko Cordillera. Locations of Figures 9.8 (A), 9.9 (B) and 9.11 (C) are indicated.

been accompanied by the development of two north–south orientated megafault systems (e.g. Scheuber & Reutter 1992): the Atacama Fault System, located in the Coastal Cordillera, and the Domeyko Fault System, located in the Precordillera (Maksaev 1990; Reutter et al. 1996) (Fig. 9.6). In addition to the movement accommodated by these two megafault systems, clockwise rotation around a vertical axis linked to the formation of the Bolivian Orocline has been recorded in the forearc region by palaeo-magnetometry (e.g. Arriagada et al. 2000). The two major longitudinal segments of the central Andes east of the Western Cordillera are the Altiplano and Puna (Fig. 9.6). These regions represent two different Cenozoic geological domains that can be distinguished on the basis of their deformation styles and lithospheric composition (Whitman et al. 1992, 1996; Tassara & Yañez, 2003; Allmendinger et al. 1997). Similar and corresponding first-order variations also seem to be reflected in the arc and forearc regions.

Coastal Cordillera
This can be traced for more than 1000 km along-strike between 18°S and 28°S. The average elevation is about 1500 m a.s.l., reaching a maximal elevation of 2700 m a.s.l. The western flank of the Coastal Cordillera is formed by a prominent and continuous coastal cliff of 1000 m average height and a maximum relief up to 2000 m (Mortimer et al. 1974). The main structural system of the Coastal Cordillera is the Atacama Fault System (AFS) which is located in the inner part of the range (Fig. 9.6). In contrast to its western flank, the topographic transition between the Coastal Cordillera and the Central Depression is smooth and much less abrupt.

Central Depression
Between 18°S and 23°S, this is a relatively smooth and flat surface with average altitude of 1500 m a.s.l. This flat surface represents the top of a Neogene infill, composed of rhyolitic tuff and continental sedimentary deposits, that together reach a stratigraphic thickness of up to 1000 m. South of Antofagasta, at 23°S, the flat surface of the Central Depression is interrupted by topography characterized by isolated hills and ranges (monadnocks) that rise over an undulating surface formed by coalescent and overlapping alluvial fans. These alluvial fans represent the final stages of the Neogene infill which in this region does not surpass 500 m in thickness (Riquelme 2003). The elevation of this portion of the Central Depression varies from 1200 to 2000 m a.s.l.

Precordillera
This is a prominent positive morphostructure that can be followed between 18°S and 22°30′S reaching altitudes between 3500 and 4500 m. Contrasting morphological, geological and structural features allow the subdivision of this morphostructural unit into two segments: a northern segment (18°S to 22°30′S) referred in the geological literature only as Precordillera (e.g. Muñoz & Charrier 1996), and a southern segment (22°30′S to 27°30′S) indistinctly called the Precordillera or Cordillera de Domeyko (e.g. Mpodozis & Ramos 1989). In this text, for the sake of clarity, we will refer to this geomorphological unit as the Precordillera for the northern segment and the Cordillera de Domeyko for the southern one. The Precordillera forms the piedmont of the Western Cordillera. The northern and highest part of the Precordillera is covered by Miocene ignimbrites which are affected by west-vergent monoclinal folding. Along its southern extension, in the highest part of the Precordillera, deep structural levels represented by a basement of Palaeozoic sedimentary and igneous rocks are exposed and thrust over Jurassic and early Cretaceous sedimentary units. Further south, the general structural framework of the Cordillera de Domeyko is dominated by an anastomosing north–south trending fault system linked by strike-slip and reverse faults, the Domeyko Fault System (DFS). The highest parts of the Cordillera de Domeyko are formed by late Palaeozoic igneous and volcanic rocks (e.g. Mpodozis & Ramos 1989). The Cordillera Domeyko borders an intermontane basin, the Preandean Depression, to the east (Fig. 9.6). This basin has an elevation of 2500 to 3000 m and contains the major Atacama, Punta Negra and Pedernales salars.
Western Cordillera

In northern Chile the Western Cordillera is dominated by a continuous Neogene volcanic chain which reaches elevations of up to 6893 m a.s.l. (Ojos del Salado stratovolcano). It forms the western border of the Bolivian Altiplano basin which lies at 3800–4200 m a.s.l. Between 18°S and 19°30’S, the Neogene volcanoes overlie deformed Oligocene–early Miocene volcanic and sedimentary rocks (Salas et al. 1966; Lahsen 1982; García 2001), which themselves unconformably cover Proterozoic–Palaeozoic metamorphic basement in the Belén area.

Neogene structural framework

The recent fault scarps of the northern Chilean forearc and arc are best exposed along the Coastal Cordillera, in contrast to most of the other parts of the region which are characterized instead by poor geomorphological expressions of underlying faults. An exception is provided by the topographic boundary region between the Precordillera and Central Depression which is controlled by broad, open folds affecting the Oligocene–Miocene strata.

Neogene structures of the Coastal Cordillera

The geomorphology of the Coastal Cordillera is characterized by prominent, fault-related topographic features (Fig. 9.7) which offset a 20–14 million-year-old mature planar surface at the top of the range (Dunai et al. 2005). The most significant structures of the Coastal Cordillera are north–south trending faults and lineaments, grouped within the AFS (Arabasz 1971; Brown et al. 1993). Other locally important structures include a group of nearly east–west striking reverse faults exposed along the Coastal Cordillera between Quebrada Camarones and Rio Loa (Fig. 9.8) valley (Allmendinger et al. 2005).

The AFS extends over 1000 km along the continental margin from south of Iquique (21°S) to La Serena in the south (29°45’S; Fig. 9.6). It has been divided into the Salar del Carmen, Paposo and Salado segments (Naranjo 1987; Thiele & Pincheira 1987) (Fig. 9.6). The earliest movement along the AFS has been constrained to Jurassic to Early Cretaceous times, contemporaneous with arc magmatism in the Coastal Cordillera (Naranjo & Puig 1984; Hervé 1987b; Scheuber & Andriessen 1990; Marinovic et al. 1995; Scheuber & González 1999). The most recent activity of the AFS is expressed by large fault scarps that deform the surface of the Coastal Cordillera in the Salar del Carmen and Paposo segments (González & Carrizo 2003; Fig. 9.7B,C). However, farther to the south, in the Salado Segment, the youngest tectonic activity does not form fault scarps, but is expressed by a sharp geomorphological contrast between both sides of the AFS (Riquelme et al. 2003).

Fig. 9.7. Fault scarp related to normal faults of the outer forearc. (A) Flat iron formed in Plio-Pleistocene marine sediment of the Mejillones Peninsula. Country rock formed by Palaeozoic metamorphic rock is in fault contact with Miocene sedimentary rocks. Small scarp along the foot of the flat iron is related to a reverse fault connected with normal faulting and backward tilting of the sedimentary cover. View looking to the west. (B) Fault scarp formed by the reactivation of the Atacama Fault System near Antofagasta. View looking to the NW. (C) Fault in line scarp of the Atacama Fault System close to Salar Grande Area. View looking to the NE. (D) Reverse, east–west orientated fault scarp, 300 m high, showing the spectacular deformation of the Miocene topography of the Coastal Cordillera east of the Salar Grande.
In the Paposo and Salar del Carmen segments, Neogene tectonic activity has disrupted the morphology and produced spectacular fault scarps (Arabasz 1971; Okada 1971; Mortimer 1980; Naranjo 1987; Hervé 1987a; Armijo & Thiele 1990; Delouis et al. 1998; González et al. 2003) (Fig. 9.7). Most of these authors suggest a predominantly normal movement with a dominant vertical component that has been responsible for the relative uplift of the western side of the AFS. According to the kinematics of some faults along the Salar del Carmen segment, the near-surface strain field is characterized by east–west extension (Delouis et al. 1998; González et al. 2003) (Fig. 9.8). This extension shows a spectacular expression in the Peninsula de Mejillones area, where north–south striking normal faults offset Miocene to Pleistocene marine deposits (Armijo & Thiele 1990). In this area vertical displacement varies between 300 and 600 m. Other normal faults are exposed in the northern extreme of the Coastal Cordillera near the town of Pisagua where several normal faults offset Neogene ignimbrites (Allmendinger et al. 2005).

The uplift of the western block of the AFS in the Paposo Segment resulted in the tectonic capture of east-to-west directed valleys (Naranjo 1987; Hervé 1987a). In the Salado Segment of the AFS, Neogene vertical slip has cut off the westward flowing rivers, denying them access to the Pacific. The generation of such closed basins east of the AFS has allowed the deposition of >300 m of middle–upper Miocene sedimentary deposits known as the ‘Atacama Gravels’ (Riquelme et al. 2003). Both large-scale and local studies of drainage system morphometry have revealed a more incised landscape to the west of the AFS. This situation has been interpreted as evidence for uplift of the western block along the AFS, at least from the mid-Miocene onwards (Riquelme et al. 2003).

Neogene structures of the Central Depression and Precordillera (Altiplano segment)

Neogene tectonic activity near Arica is represented mostly by west-verging gentle folds and high-angle reverse faults. These structures affect Oligocene–Miocene ignimbrites and clastic deposits (Fig. 9.9). The most significant structure in the Precordillera in this region consists of a north–south to N40W-striking fold (Oxaya Anticline; Salas et al. 1966; Muñoz & Charrier 1996; García 2002), with an axis extending over 50 km strike length, and wavelength reaching up to 30 km (Fig. 9.9B). This structure has been interpreted as related to the Neogene gravitational collapse of the Western Cordillera (Wörner et al. 2000b). However, field observations indicate that it results from the upward propagation of the Ausipar Fault during a relatively short-lived reactivation interval in late Miocene times (12–10 Ma; Garcia 2001; García & Hérail 2005). Across this fold the topography shows a difference in height of about 1500 m between western lowland and eastern highland. The amount of shortening, as determined by balanced cross-section methods, is c. 100 m which requires that the Ausipar Reverse Fault must be interpreted as a high angle structure at depth (Fig. 9.9B). Below the 26–19 Ma Oxaya Formation, the Ausipar fault has displaced a Mesozoic folded sequence over 1500 m between western lowland and eastern highland. The amount of shortening, as determined by balanced cross-section methods, is c. 100 m which requires that the Ausipar Reverse Fault must be interpreted as a high angle structure at depth (Fig. 9.9B). Below the 26–19 Ma Oxaya Formation, the Ausipar fault has displaced a Mesozoic folded sequence over 1500 m between western lowland and eastern highland. The amount of shortening, as determined by balanced cross-section methods, is c. 100 m which requires that the Ausipar Reverse Fault must be interpreted as a high angle structure at depth (Fig. 9.9B). Below the 26–19 Ma Oxaya Formation, the Ausipar fault has displaced a Mesozoic folded sequence over 1500 m between western lowland and eastern highland.
East of the Iquique and Pica area (20°30'S), between the Precordillera and Central Depression, the topography shows a major step of 2500 m due to another major monoclinal flexure (Galli & Dingman 1962; Victor et al. 2004). South of Pica, the boundary between the Precordillera and Central Depression is represented by a system of east-dipping reverse faults that uplift basement blocks of the Sierra de Moreno. Movement on this fault system has largely been constrained to the Mesozoic to early Cenozoic (Skarmeta & Marinovic 1981). However, movements accommodated by some of these faults cut or gently fold Miocene strata and indicate minor Neogene reactivation. Similar minor Neogene faulting has also been documented on the eastern border of the Sierra de Moreno (Reutter et al. 1996).

**Neogene structures of the Cordillera de Domeyko (Puna segment)**

Much of the tectonic configuration of the Domeyko Cordillera is controlled by the north–south trending Domeyko Fault System (Fig. 9.6). This structure results from a thick-skinned uplift of a late Palaeozoic plutonic and volcanic basement that is bounded by a system of reverse faults and folds, whose vergence alternates along-strike (Coira et al. 1982; Reutter & Scheuber 1988; Mpodozis & Ramos 1989). Reverse faults and folds affect a Late Cretaceous–Miocene cover and have resulted in the inversion of a Mesozoic backarc marginal basin since middle Cretaceous times (Muñoz et al. 2000, 2002). The Cenozoic timing and kinematics of the structures related to the DFS are poorly constrained; however, it was active during the lifespan of the Eocene to early Oligocene magmatic arc, showing both important strike-slip movement as well as shortening (e.g. Reutter et al. 1991, 1996; Tomlinson & Blanco 1997; Niemeyer 1999). Both eastern and western foothills of the Domeyko Cordillera are mainly composed of extensive alluvial fan sediments and pedimentation-related deposits accumulated from middle–late Miocene times to the present day (Sáez et al. 1999; Kober et al. 2002; Riquelme 2003). Active faulting and
folding attributed to the Neogene reactivation of DFS structures has been reported in the eastern and western flank of the Domeyko Cordillera. Neogene faulting has been documented to the east, in the Salar de Atacama area (e.g. Jolley et al. 1990; Flint et al. 1993; Mpodozis & Clavero 2000; Muñoz et al. 2000, 2002; Jordan et al. 2002) (Fig. 9.10). On this flank, intermittent local uplift due to thrust tip propagation controls the Neogene topographical and sedimentological evolution. The appearance of a distinctive series of discrete sub-basins (e.g. Llano la Paciencia and Pampa Viscachita), as well as eastward progradation of alluvial fans, both result from the activity of an east-vergent thrust system. The Cordillera de la Sal is an intrabasinal physiographic unit running down the eastern border of the Atacama Salar. It has been interpreted as a thin-skinned contractional feature folding and thrusting a Mio-Pliocene sedimentary sequence some 2000 m thick (Flint et al. 1993) (Fig. 9.10). West-dipping blind thrusting in the Atacama Salar represents the youngest tectonic activity in this area (Jordan et al. 2002). Active faulting and folding occur without topographic expression, and net reverse offset across the faults, which can reach up to 700 m during Pliocene to Recent times, decreases to the east (Jordan et al. 2002). Late Miocene to Recent reactivation of DFS-related structures, evident from fault scarps, has recently been reported in the Salar de Punta Negra area (Soto et al. 2005).

Around the latitude of Antofagasta (23°S), Neogene strike-slip activity on the eastern side of the Domeyko Cordillera has been reported (Reutter et al. 1996; May et al. 1999). In contrast, further south Plio-Quaternary tectonic activity without significant strike-slip movements has been recorded on the western flanks of the Domeyko Cordillera and in the westernmost part of the Central Depression (25°30'S and 26°30'S: Audin et al. 2003) (Fig. 9.11). These latter authors observed that the relief of these more southerly areas results from north–south trending active vertical faults in the topographically higher pampas (western flank of the Domeyko Cordillera), and active folding around a north–south trending axis in the lower pampas. In the higher pampas the fault scarps cut through alluvial deposits and are distributed parallel to older tectonic structures (Fig. 9.11). Major vertical, and in some cases composite escarpments, up to 5–10 m high, can be recognized on the northern prolongation of the Sierra Castillo Fault, a main fault of the DFS at this latitude (Fig. 9.11). The dip direction of these escarpments does not change along-strike. Audin et al. (2003) identified no strike-slip movement and proposed that the vertical throw could be associated either with normal faults or with overturned thrusts reflecting along bedding planes and showing a normal sense of displacement.

**Neogene structures of the Western Cordillera**

Structures affecting the Neogene sequences in the Western Cordillera are best exposed to the east of Arica where north–south to NNW–SSE structures form a complex fold and thrust belt. This belt affects the late Oligocene–middle Miocene rocks of the Lupica, Chucal, Macusa, Joracane and Putani formations, all of which are partially, progressively and unconformably covered by late Miocene–Holocene volcanic and sedimentary rocks (Muñoz & Charrier 1996; García 2001). In the Visviri, Colpitas, Churiguaya and Chucal areas, the deformation is east-vergent, whereas in the Putre–Bélén area the vergence is to the west (Fig. 9.9B). In the region of Churiguaya a gentle growth-anticline is exposed, with the anticlinal flanks dipping less than 35°, and with a wavelength of around 20 km. In the Putre–Bélén area, Neogene folds and thrusts produce the greatest change of altitude in the region, from an average of c. 3200 m to the west up to c. 5300 m in the east, across a horizontal distance of only 15 km (Fig. 9.9B). In these areas, the Lupica Fm is folded into anticlines and synclines that are thrust over the subhorizontal sequences of the Oxaya and Huaylas formations and middle Miocene volcanic rocks.
Fig. 9.11. (A) Morphotectonic interpretation, on Aster Satellite image, of the western flanks of Domeyko Cordillera between 25°30′S and 26°S approximately. In this area, alluvial fans are affected by vertical faults. (B) Aerial photograph and (C) interpretation of the vertical faults, in Pampa Exploradora around 3500 m. S1 to S4 are successive nested alluvial surfaces. The scarps are linear and continuous forming vertical and composite scarps about 5–10 m high, the dip direction of which (toward the east) never changes along-strike. No lateral offset can be identified along these fault traces. The scarps are coupled with neighbouring secondary synthetic faults that create small half-graben where Quaternary sediments are trapped. (D) Inset shows type example of a major normal fault scarp, with 30 m throw, cropping out in the Precordillera of the El Salvador region (view looking north to the Pampa Exploradora’s flank). Main photo shows north-looking view of river-cut cross-section of a fault scarp at the Pampa Doña Ines Chica.
The three most important reverse faults extend over 50 km and dip between 20° and 60° to the east. Around Belén, the contractional system is characterized by a large asymmetric anticline of 35 km wavelength, which involves thrust sheets of Precambrian–Palaeozoic basement and, to the west, sheets of syntectonic coarse-grained conglomerates and volcanic rocks. In the Chucal–Macusa area, west-vergent and east-vergent folds and faults are exposed (Riquelme 1998; García 2001; Charrier et al. 2005b). In Chucal, an east-vergent growth anticline of 5 km wavelength has flanks dipping from 10° to 80°. Along its western border, the anticline is cut by the 40°–50° east-dipping Jaropilla reverse fault, which displaces the lower Luripica Fm over the Chucal Fm. To the west, the Macusa folds are short, tight (0.5 to 2 km in wavelength) and west-vergent, forming a narrow zone of deformation. The Precambrian–Palaeozoic basement is not exposed in the Chucal–Macusa area. Three balanced cross-sections in the Western Cordillera, east of Arica, indicate a minimum post-18 Ma shortening ranging from 6 to 7 km which translates to 20% to 40%. Recent (late Holocene) tectonic activity on the border of the Western Cordillera is evidenced by shallow (<20 km) microseismic activity (Comte et al. 1999).

To the south, at the latitude of Pisagua, folding and faulting related to uplift have been reported to affect Miocene lavas and ignimbrites (Puchuldiza, Condoriri and Utayane formations; Lahsen 1982). East of Iquique and Antofagasta, the Western Cordillera of northern Chile largely consists of Pliocene–Holocene stratovolcanoes, and Neogene tectonic features are either not exposed or have not been described. In Pica, shortening across the Precordillera and Western Cordillera has been calculated to be around 5 km (Vicet al. 2004). East of Copiapó, in the Claudio Gay Cordillera, Neogene east- and west-vergent deformation is represented by gentle folding and basement-involving high-angle reverse faulting (Mpodozis & Clavero 2002).

**Large-scale landscape configuration**

Northern Chile can be viewed as comprising two north–south elongated neotectonic units: an inner forearc and arc domain, and an outer forearc domain. The inner forearc and arc domain comprises parts of the Central Depression, the Precordillera, the Cordillera de Domeyko and the Western Cordillera, whereas the outer forearc domain is dominated by the Coastal Cordillera.

The outer forearc domain is characterized by north–south striking normal faulting (and locally by east–west faulting), a phenomenon characteristic of, and restricted to, the Coastal Cordillera. In addition to this faulting, long-term uplift has affected the Coastal Cordillera in northern Chile, and has a major topographic expression in the prominent (up to 1000 m high) coastal cliff (Fig. 9.12). Late Pleistocene to Holocene marine terraces exposed at the foot of the Coastal Cliff register the most recent uplift pulses in the coastal area. The terraces are distributed in a staircase pattern extending from 3 to 170 m a.s.l. (Leonard & Wehmiller 1991; Niemeyer et al. 1996; Ortlieb et al. 1996). Exceptionally high elevations of late Pleistocene to Holocene marine terraces are reached on the Mejillones Peninsula where some terraces are exposed at elevations above 300–400 m a.s.l. (González et al. 2003). Uplift rates calculated on the basis of the age of the marine terraces and their position relative to the present-day sea level indicate values between 0.05 and 0.5 mm/year (Ortlieb et al. 1996; Delouis et al. 1998).

The deformational regimes described above indicate that the outer forearc is undergoing east–west extension orthogonal to the trench, north–south shortening and forearc uplift. These major deformation processes have been active over a time span of Miocene to late Pleistocene. Because the extensional and the contractional structures (north–south shortening) are restricted to the Coastal Cordillera, above the zone where the Nazca Plate and South American Plate are seismically coupled, it is reasonable to assume that these first-order processes are related to plate interaction. East–west directed extension has been interpreted to be a result of tectonic erosion (Niemeyer et al. 1996), upper plate bending (Delouis et al. 1998; González et al. 2003) and coseismic extension (Delouis et al. 1998; González et al. 2003). North–south shortening can be explained as an effect of the subduction occurring along a segment of the central Andes concave to the ocean geometry. The limited geographic distribution of reverse faults which mark the north–south shortening, which is restricted to the inner part of the central Andean oroclinal bend, strengthens this interpretation. Forearc uplift seems to be the most widely distributed tectonic process, acting over the entire length of the forearc. A number of hypotheses have been postulated to explain the long-term uplift of the forearc. The most commonly applied explanation is the underplating of crustal material beneath the Coastal Cordillera, along the seismically coupled zone (Armijo & Thiele 1990; Delouis et al. 1998). In this context, uplift would be the result of a deep-seated process of crustal thickening. An alternative
process is the progressive indentation of the Nazca Plate beneath the South American plate. Because east–west extension and forearc uplift are regionally the most widely distributed type of tectonic activity, it is reasonable to infer that they represent the direct morphological expression of plate convergence in the outer forearc.

The inner forearc and arc domain of northern Chile is characterized by double-vergent folding and thrusting that can be related to overall east–west shortening and uplift affecting the high central Andes. In the Altiplano segment, the Precordillera was mainly uplifted by west-vergent folding in connection with reactivation of high-angle reverse faults with large vertical throws (García 2001; Víctor et al. 2004; Pinto et al. 2004b; Farias et al. 2005b). The recorded vertical throws along these reverse faults can alone explain the differences in elevation between the Precordillera and Central Depression. The Western Cordillera is configured by west-vergent moderate-angle reverse faults involving larger shortening than in the Precordillera, and it forms part of the fold and thrust belt that affect the backarc. However, two questions concerning the role that the folding and faulting play on the Neogene uplift of the Altiplano can be raised: (1) What was the elevation of the Altiplano at the beginning of the Neogene? (2) Does the mentioned shortening explain the Neogene uplift of the Western Cordillera?

In relation to the first question, sedimentological evidence from the Bolivian Altiplano and Central Depression indicates that the Western Cordillera and Precordillera were uplifted areas of high relief subject to erosion during the Oligocene (Lamb et al. 1997; Horton et al. 2001; García 2001). Towards the south, between the latitudes of Antofagasta and El Salado, fission track data indicate that an important episode of exhumation occurred in the Domeyko Cordillera in Eocene–early Oligocene time (Maksaev & Zentilli 1999). Riquelme et al. (2003) documented this exhumation process from field data and argued that the Precordillera was already at least 2500–3000 m above sea level at that time. Thus, most of the present-day altitude of the inner forearc and arc region was probably reached prior to Neogene times.

In relation to the second question, various authors argue that the main uplift episode of the Altiplano–Puna occurred during the Neogene (e.g. Isacks 1988; Baby et al. 1997; Gregory-Wodzicki 2000). However, assuming an important pre-Neogene altitude in the Western Cordillera, Neogene shortening by itself cannot explain the present-day elevation of the Altiplano. In order to reach its present-day altitude, geophysical and petrological studies consider additional processes have contributed to crustal thickening (e.g. Giese et al. 1999). Additional crustal thickening from magmatic addition (e.g. Gill 1981; Hoke et al. 1994; Kono et al. 1989; Francis & Hawkesworth 1994), lithospheric thinning and removal of the subcrustal lithosphere (Isacks 1988; Whitman et al. 1996; Lamb et al. 1997; Beck & Zandt 2002), underplating of material removed from the forearc by subduction erosion (e.g. Schmitz 1994; Baby et al. 1997; Hartley et al. 2000), or ductile mass transfers within the lower crust (Kley et al. 1999; Beck & Zandt 2002; Gerbault et al. 2002; Husson & Sempere 2003) have been proposed.

The effects of crustal thickening and uplift in the inner forearc domain have been to produce a westward regional tilting and/or crustal flexuring (e.g. Isacks 1988; Lamb et al. 1997). At the latitude of Copiapó–El Salvador, deep vertical incision in the forearc region can only be explained by post-10 Ma tilting of the inner forearc and arc domain (Riquelme et al. 2003). Neogene shortening without significant strike-slip motions is recorded in the Domeyko Cordillera. Field evidence shows that this latest deformation episode corresponds to minor reactivation of pre-existing structures since late Miocene times (e.g. Jordan et al. 2002; Audin et al. 2003; Soto et al. 2005). Fault reactivation can occur under stress levels lower than those needed for creating new faults (Handin 1969). Moreover, if a previous fault pattern is present, fault reactivation rather than formation of new faults is frequently observed, even if the consecutive deformation episodes are not coaxial, in particular if the latest deformation episodes are smaller (e.g. Brun & Nalpas 1996; Dubois et al. 2002). Thus, it is probable that the latest tectonic activity observed in the Precordillera and Cordillera de Domeyko corresponds to low-intensity episodes of deformation that reactivated Eocene–Oligocene structures originally related to uplift of the western Andean margin within a compressive regime.

Convergence controlling the shortening of the forearc is expressed by thrusting of western border Precordillera over the Central Depression. In contrast, forearc uplift localized on the Coastal Cordillera is a deep-seated process and not directly connected with near-surface faulting. Given this background, the popular idea that the Atacama Fault System controls the uplift of the Coastal Cordillera has to be reconsidered. East–west extension and north–south shortening control only the second-order morphostructural features of the forearc by faulting and differential uplift.

To summarize, two longitudinal structural domains can be proposed for the Neogene in the forearc region: an outer forearc domain comprising the Coastal Cordillera, and an inner forearc and magmatic arc domain comprising the western part of the Central Depression, Precordillera, Domeyko Cordillera and Western Cordillera. The outer forearc exhibits north–south striking normal faults consistent with east–west directed extension, as well as forearc uplift evidenced by the landscape features (coastal cliff >1000 m high, Late Pleistocene–Holocene marine terraces). Both east–west extension and forearc uplift result from the interaction of the seismically coupled Nazca and South America plates. East–west extension may be attributed to a number of processes: tectonic erosion, upper plate bending or coseismic extension. Forearc uplift results from underplating of crustal material beneath the Coastal Cordillera or from progressive indentation of the Nazca Plate beneath South America. Trench-parallel shortening is restricted and localized to where the plate boundary has a concave-wards-sea geometry (i.e. the oroclinic bend of the central Andes). The inner forearc and magmatic arc domain is characterized by high-angle reverse faulting and folding (Altiplano segment) and minor reactivation of older structures (Puna segment). This most recent tectonic activity in the inner forearc region is related to the uplift of the western Andean margin within a compressive regime. Uplift localized in the outer forearc in the Coastal Cordillera is a deep-seated process, while shortening controlled by convergence is expressed by thrusting and folding in the inner forearc region.

Pliocene–Pleistocene state of stress for the central and southern Chilenan Andes (A.L. & J.C.)

This section summarizes the nature and significance of Plio-Pleistocene deformation of the central and southern Chilenean Andes between 33°S and 46°S (Fig. 9.13). In contrast with northern Chile, where extremely arid conditions allow the superb preservation of fault scarps and other structural features associated with Neogene to Recent deformation, the more humid climate and concomitant higher erosion rates of central and southern Chile make direct observation of many morphotectonic structures difficult. One way to overcome this limitation was to map a number of well-exposed structural sites located close to or on large-scale lineaments believed to represent major fault zones located at the boundaries between the main morphological units of the Andes. The outcrop-scale mapping of these structural sites consisted in identifying main rock units and the nature, kinematics and relative age of
mesoscopic faults and folds. Fault-slip data from each site were then used to calculate one or more stress tensors, believed to represent different tectonic events. Different authors have treated fault-slip data for the Andes as yielding local stress (e.g. Lavenu & Cembrano 1999b) or strain tensors (e.g. González et al. 2003; Allmendinger et al. 2005). Whether it is more convenient or accurate to obtain stress or strain from field data is the subject of a long-standing discussion that is beyond the scope of this review (e.g. Bott 1959; Carey & Brunier 1974; Marrett & Allmendinger 1990). However, for internal consistency we will assume that the principal directions of stress are more or less coaxial with those of incremental strain for regions that have not undergone significant block rotations, an assumption that can be considered fair for large-scale deformation studies such as the one addressed here.

In this part of the chapter we provide an integrative account of the most geologically significant structural sites of the central and southern Chilean Andes for which a Pliocene and/or Pleistocene stress or strain tensor has been calculated from fault-slip data. We have selected those sites for which good structural data can be placed into a regional context based on a good constraint of both kinematics and age. For the sake of clarity on data presentation and discussion, we have divided the central (33–39°S) and southern (39–46°S) Chilean Andes into margin-parallel, morphostructural domains that exhibit internally consistent geology and kinematics.

Although some authors have restricted the term ‘neotectonics’ to the last recorded deformation event of an area, we deliberately include Pliocene to Recent deformations to better understand the nature of the link between temporal and spatial changes in plate boundary conditions (rates, subduction angle, degree of coupling) and tectonics of the upper plate.

**Geological and morphotectonic framework**

The Andes of central and southern Chile consist of three tectonically distinct parallel domains: (i) a forearc zone located between the Peru–Chile Trench and the Main Cordillera, in which the Coastal Cordillera and Central Depression are
located; (ii) a magmatic arc in the Main Cordillera, which is the active volcanic zone; and (iii) a foreland zone in Argentina where the most recent deformations are documented (e.g. Costa et al. 2000b; c; Diraison et al. 1998; Folguera et al. 2004) (Fig. 9.13).

**Coastal Cordillera**

The Coastal Cordillera between 33°S and 46°S is a discontinuous morphological feature made up of Mesozoic volcanic-sedimentary rocks intruded by coeval plutonic belts. Late Palaeozoic to Jurassic granitoids and metamorphic rocks occur along the coast between 36°S and 46°S, west of the Coastal Cordillera, and represent the basement upon which the Andes have been built (Hervé 1987a, b). A few patches of Cenozoic sedimentary units occur west of the Coastal Cordillera proper along the coast at Navidad (33°S) and Peninsula de Arauco (37°S).

The Coastal Cordillera varies in altitude from up to 2 km at 33°S to c. 0.5 km at 46°S. Available geological and fission track data suggest that the Coastal Cordillera of central Chile underwent episodic southward exhumation of 100 Ma, and that of a major compressional/transpressional event (e.g. Cembrano et al. 2003; Arancibia 2004; Parada et al. 2005b). Evidence for neotectonic deformation along the Coastal Cordillera of central and southern Chile is scarce and of poorly known age (e.g. Lavenu & Cembrano 1999b). However, a few sites show good evidence of Pliocene and Pleistocene deformation; these will be described below.

**Central Valley**

The Central Valley (sometimes also referred to as the Central Depression) is a north–south trending morphostructural domain running between the Coastal Cordillera and the Main Cordillera between 33°S (close to Santiago) and 46°S (Ofqui Isthmus). The valley is filled with 1500 to 5000 m of Oligocene to Recent continental and sometimes marine (?) deposits (Elgueta et al. 2000a), and whilst never more than 75 km wide, it concentrates most of Chile's population. The western and eastern boundaries of the Central Valley have long been thought to be normal faults defining a graben structure (Aubouin et al. 1973a; Laugenie 1982; Cisternas & Frutos 1994). However, more recent work (e.g. Lavenu & Cembrano 1999b; Farias et al. 2005b; Fock et al. 2005) has shown that these boundaries are locally represented by eroded fault scarps with complex kinematics. The eastern limit of the Central Valley from 33°S to 46°S is currently regarded as an eroded fault system made up of Miocene to Pliocene strike-slip and west-verging reverse faults (e.g. Fock et al. 2005; Cembrano et al. 2002). The boundary between the Central Valley and the Main Cordillera from 39°S to 46°S, corresponds well with the western branch of the Liquiñe–Ofqui Fault Zone (LOFZ), marked by a straight NNE-trending lineament. Limited field evidence along this lineament shows a complex system of Pliocene (c. 4 Ma) dextral strike-slip and reverse dextral to brittle fault zones developed on the west flank of the Patagonian Batholith Miocene belt (e.g. Lavenu & Cembrano 1999b; Cembrano et al. 2000, 2002). The LOFZ, which controls the morphology of channels and fiords, places deep structural levels of the Main Cordillera represented by the Patagonian Batholith, on top of upper structural levels of the Central Valley documented here by mid-Tertiary volcanic and sedimentary rocks (Lavenu & Cembrano 1999b; Pankhurst et al. 1999; Cembrano et al. 2002; Thomson 2002).

**Main Cordillera**

The Main Cordillera between 33°S and 39°S is characterized by folded and faulted Mesozoic–Cenozoic volcanosedimentary units intruded by Late Miocene to Pliocene plutonic rocks and the present-day volcanic arc. Altitudes vary between c. 6000 m in the north down to 3000 m in the south, with an average width of 80 to 100 km. The most important regional-scale structures at these latitudes are a series of east-verging Neogene fold and thrust belts that developed within and to the east of the Main Cordillera in the frontal Cordillera and foreland of Argentina (Ramos 1988a; Charrier et al. 2002; Giambiagi et al. 2003). Neogene deformation on the Chilean side of the Cordillera is more limited. However, recent work by Farias et al. (2005b) suggests that the western part of the Main Cordillera east of Santiago has been differentially uplifted during the Neogene by a west-verging fault system that may still be active (San Ramon fault, see Fig. 9.13).

Neotectonic deformation along and east of the Main Range between 37°S and 39°S is well-represented by the Antiñir–Copahue Fault System in Argentina and by the northern termination of the Liquiñe–Ofqui Fault System in Chile (e.g. Folguera et al. 2004). The Antiñir–Copahue system (Fig. 9.13) is formed by an east-verging fan of high-angle transpressional and transtensional faults affecting Pliocene to Pleistocene volcanic rocks. Some recent deformation of unconsolidated deposits has also been reported (Folguera et al. 2004), which emphasizes the importance of these faults as active Andean structures. The Antiñir–Copahue system of Argentina gives way southward to the Liquiñe–Ofqui Fault Zone. This fault system is at these latitudes represented by a dextral transtensional array of master NNE-striking right-lateral faults and a subsidiary set of NE- and ENE-striking oblique-slip and normal faults. Both strike-slip and normal faults are spatially and temporally associated with stratovolcanoes and caldera structures such as the El Agrio Caldera (Melnick et al. 2002; Folguera et al. 2004, 2005).

The most important Neogene regional structure of the Main Cordillera between 39°S and 46°S latitude is the c. 1000-km-long Liquiñe–Ofqui Fault Zone (Hervé 1976; Hervé et al. 1979b), which consists of two overlapping NNE-striking master faults joined by a series of en echelon NE-striking subsidiary faults forming a strike-slip duplex structure (Fig. 9.13). Field observations combined with Ar/Ar and fission track thermochronology along the LOFZ document a main event of dextral transtensional dextral deformation at c. 4 Ma, followed by brittle compressional to strike-slip deformation after 3.8 Ma and 1.6 Ma, respectively (Hervé 1994; Cembrano et al. 1996, 2000, 2002; Lavenu & Cembrano 1999b; Thomson 2002).

**Pliocene state of stress between 33°S and 46°S**

Previous work shows that the state of stress in the Andes during latest Pliocene time (1.8–3 Ma) had two main characteristics: it was compressive, and it had an approximate east–west maximum horizontal stress direction throughout the whole Andean chain. This deformation is well documented from Ecuador to southern Chile and from the Pacific coast to the Subandean zone in Peru, Bolivia, and foreland regions in Argentina (Lavenu et al. 1989, 1995; Allmendinger et al. 1989; Lavenu & Mercier 1991; Mercier et al. 1992; Urreiztieta et al. 1996; Niemeyer et al. 1996; Diraison et al. 1998; Lavenu & Cembrano 1999b; Gonzalez et al. 2003). A summary overview of Pliocene deformation for the central and southern Andes is provided by Farías et al. (2005) which show the complications resulting from fault-slip data inversion for the sites described in this study plus others selected from the literature (e.g. Martinez 1980; Jordan et al. 1983b; Lavenu & Mercier 1991).

**Northern segment between 33°S and 39°S: Coastal Cordillera**

**Navidad (34°S).** The Miocene Navidad basin, west of the Coastal Cordillera of central Chile exhibits evidence of several deformation events of Miocene to Pliocene age (Lavenu & Encinas 2005). After a widespread extensional Miocene to Pliocene episode, the basin underwent an important uplift event followed by NW–SE directed extension. During the late Pliocene, a compressive regime (∇σ1 = N250°E) deformed the basin deposits as indicated by mostly north–south striking reverse faults. The last deformation episode is represented by north–south striking normal faults compatible with east–west extension of the outer forearc (Fig. 9.15).
Fig. 9.14. Summary figure showing the results of the microtectonic analysis of Pliocene faults from central and southern Chile and other parts of the Andes, for comparison. The compressional east–west tectonic regime documented in Chile has also been recorded in the Altiplano, Eastern Cordillera, and well into the foreland (e.g. Martinez 1980; Jordan et al. 1983b; Lavenu & Mercier 1991).

Península de Arauco (38°S). Dense forest and limited access to the area prevent systematic structural studies of the Arauco Peninsula, although two broad morphostructural domains can be identified. The first of these is an eastern domain consisting of a Palaeozoic metamorphic basement and a Triassic–Cretaceous sedimentary cover. The second is a western domain (Arauco Peninsula) which forms a low-altitude zone marked by marine terraces close to the coast: an upper terrace between 180 and 200 m, two intermediate ones at 75–100 m and 50 m, and a lowermost one at 25 m. This coastal block consists of Eocene deposits and a Mio-Pliocene basin, with the latter forming a graben flanked to the west and east by NNE-striking normal faults (Martinez-Pardo & Osorio 1968; Ferraris & Bonilla 1981; Martinez-Pardo 1990).

Northern segment between 33ºS and 39ºS: Central Valley
At 36ºS, the boundary between the Central Valley and the Main Range is marked by regional faults striking N30°E. These faults cut Oligocene–Miocene volcanic rocks, which document a vertical displacement (west-side-down) of c. 1000 m. Mesoscopic faults exposed along the eroded fault scarp on these volcanic rocks are mainly strike-slip, compatible with either east–west or north–south compression. The western flank of the Central Valley is characterized by a system of north–south to N10°E master faults and N30–40°E subsidiary faults. Mesozoic and Neogene rocks exhibit mesoscopic strike-slip and reverse faults with minor evidence for normal faults. The latter have been interpreted as gravitational structures associated with topography.

Northern segment between 33ºS and 39ºS: Main Cordillera
Los Andes, Farellones, Mina El Teniente (33–34°S). A systematic study of the Neotectonic deformation along the Main Range was undertaken in the San Felipe–Los Andes region, at the town of Farellones east of Santiago and at El Teniente copper deposit, covering a region of central Chile between 33ºS and 35ºS. A 4.5 Ma breccia pipe (Cuadra 1986), the Braden Breccia, cuts through the core of El Teniente copper deposit and provides an excellent opportunity to map Pliocene–Pleistocene (?) deformations. Kinematic analysis of fault-slip data from the Braden Breccia document two independent and superposed compressional events. An older event, characterized by a N60°E-trending σ1, and a younger event with a north–south trending σ1. Additional fault-slip data coming from a nearby younger 2.8 Ma lamprophyric dyke (Cuadra 1986), record only the later north–south compressional event, confirming a likely
Fig. 9.15. Field evidence for ESE–WNW compressional deformation in the Neogene Navidad basin of central Chile. A reverse fault places Cretaceous to Neogene strata on top of a Palaeozoic granodiorite. This fault is interpreted as an early normal Neogene fault reactivated as a reverse fault during compression. Reverse reactivation is supported by overprinting kinematic indicators on the fault surface (Lavenu & Encinas 2005).

Southern segment between 39°S and 46°S: Coastal Cordillera Taitao Peninsula (46°S). Only a few studies of brittle deformation have addressed the Neogene tectonics of the southermost part of the Coastal Cordillera; available data only describe the geometry of faulting at the regional scale close to the Nazca–South America–Antarctica triple-junction (e.g. Forsythe & Nelson 1985; Forsythe et al. 1986). According to Veloso (2001), a Pliocene compression–transpression event, with σ1 close to N260°E, is documented in the Taitao Ophiolite (dated 11–6 Ma) and the Seno Hoppner pluton (dated 6.2–5.5 Ma).

Southern segment between 39°S and 46°S: Main Cordillera Reloncavi (41°S). On both sides of Estuario Reloncavi, centimetre- to decametre-scale striated fault planes cut all the plutonic units ranging in age from Cretaceous to Miocene, whereas Quaternary sedimentary and volcanic rocks do not show any evidence of faulting. Most faults are steeply dipping
 (>60°), with striae ranging in pitch from subhorizontal to subvertical. Two distinct homogeneous populations of faults were identified in the Reloncaví region. The first is represented in Figure 9.17, which includes faults with a predominant strike-slip component, and a few with predominant dip-slip component. Faults with strike ranging between N337°E and N065°E (average N020°E) show right-lateral slip sense; faults striking between N084°E and N115°E (average around N100°E) are left-lateral; and faults striking WNW are reverse-slip. The kinematic analysis of these mesoscopic faults is consistent with a compressional regime suggested by the first fault population:

\[
\begin{align*}
\sigma_1 &= 261°, 05°; \\
\sigma_2 &= 171°, 07°; \\
\sigma_3 &= 28°, 82°.
\end{align*}
\]

A second diagram (Fig. 9.17) shows the second homogeneous population of faults identified; most have a significant down-dip component, although some strike-slip faults are also included. These younger faults will be further described in the Pleistocene deformation section.

**Fig. 9.16.** Regional map showing the location of structural sites in the Main Cordillera of Central Chile for which fault-slip data were obtained.

Puyuhuapi, Queulat, Puerto Cisnes (44°S). Four sites were measured in intrusive rocks with late Miocene to Pliocene Ar/Ar dates (5.5 and 10 Ma (Hervé et al. 1993) and 5.3 and 13.3 Ma (Cembrano 1998)) and two were measured in Patagonian batholithic rocks of uncertain age. The main direction of compression in this case is also close to east–west, the directions of \( \sigma_1 \) being between ENE–WSW (N237°, 8°) and WNW–ESE (N101°, 8°). In most cases, the tectonic regime is found to be compressional; in some cases it is a strike-slip compressional regime (Fig. 9.18). Another, younger fault population has been identified at some of these sites and will be described in the next section on Pleistocene deformation.

**Pleistocene state of stress between 33°S and 46°S**

In contrast with the generalized Pliocene compression throughout most of the Chilean Andes between 33°S and 46°S, fault-slip analysis on several sites from the coast to the main range...
suggests that during Pleistocene time, rather than one generalized tectonic episode, there were several main deformational events of different nature and spatial distribution.

**Northern segment between 33°S and 39°S Coastal Cordillera**

Two types of Pleistocene deformations are recognized along the Chilean forearc. The first is east–west extensional deformation that affects large portions of the outer forearc of northern Chile including the Mejillones Peninsula and the northern Atacama Fault Zone (23°S), the Caldera coast area (27–28°S), the Altos de Talinay Peninsula (30–31°S) near La Serena, and the Arauco Peninsula (37–38°S) (e.g. Marquardt 1999; Benado 2000; González et al. 2003; Marquardt et al. 2004). All these sites lie along the coastline at a distance of 70–90 km from the offshore trench. West- or more frequently east-dipping normal faults cut uplifted Quaternary marine terraces and their deposits. On the Caldera coast and Mejillones Peninsula, the extensional deformation is younger than 400 ka. The second type of deformation is a north–south shortening in the Coastal Cordillera, the Central Valley and the Main Cordillera, west of the volcanic arc.

**San Antonio (34°S).** Close to the coast, in the Coastal Cordillera along the road between Melipilla and San Antonio, Neogene to Pleistocene marine sediments of the Navidad Formation and Potrero Alto strata (Valenzuela 1992; Wall et al. 1996) are cut by a subhorizontal erosion surface. This surface forms a pediment of late Pliocene to early Pleistocene age. A Palaeozoic granitic gneiss has been emplaced over Pleistocene sandy sediments of the Potrero Alto strata along c. east–west striking reverse faults. Mesoscopic (decimetric to metric) faults with centimetre- to metre-scale displacements affecting both the Palaeozoic gneisses and the Pleistocene strata show well-defined striae. Fault-slip data from this site are compatible with a N185°E-trending, horizontal σ1, and a vertical σ3 (Lavenu & Cembrano 1999b) (Fig. 9.19).

**Península de Arauco.** East of the peninsula, between Curanilahue and Ramadillas, there is an eroded fault scarp. This scarp locally exhibits north–south striking normal faults that place Neogene–Pleistocene (?) marine sediments against the Palaeozoic–Mesozoic basement. Fault-slip data from these localities were insufficient to obtain a local stress tensor; however, all normal faults are compatible with roughly east–west trending extension of uncertain age.

**Northern segment between 33°S and 39°S: Central Valley Esperanza (38°S).** Pleistocene terrace sediments filling the Central Valley at Esperanza are cut by reverse faults showing metre-scale displacements. Striae on these faults are compatible with a subhorizontal σ1 = N359°E and σ2 = N90°E (Lavenu & Cembrano 1999b). A similar geometry and kinematics can be found farther south in the Central Valley at Victoria, where Pleistocene fluvial sediments of the Quilleco polygenetic cone (Thiele et al. 1998) are cut and displaced by a reverse fault striking N150°E. This fault lies along a lineament cutting across
the Palaeozoic basement. Fault-slip analysis documents a
north–south to NNE compressional axis (Fig. 9.20).

**Victoria (38°S).** Pleistocene sediments of the Quilleco polygenetic cone (Thiele et al. 1998) exposed on the bank of the Río Traiguén are affected by a N150°E-striking reverse fault with a metre-scale vertical throw. This fault also lies along a lineament cutting across the Palaeozoic basement suggesting local Quaternary reactivation. A limited number of faults and striae are compatible with a north–south to NNE compressional axis similar to that calculated from Esperanza (Fig. 9.20).

**Northern segment between 33°S and 39°S: Main Cordillera**

*Cajón del Maipo (34°S).* Five levels of fluvial terraces occur along the left margin of the Maipo river SE of Santiago. These terraces lie on top of a late Oligocene–early Miocene basement belonging to the Abanico and Farellones formations of central Chile (Elgueta et al. 2000a). Close to the Colorado river confluence with the Maipo river, these terraces, located between 880 and 1050 m above sea level, are well exposed in a gravel quarry. The highest (i.e. oldest) terrace is formed by conglomerates and sandstones with intercalated, probably reworked, ash layers. These can be considered as part of a 0.45 Ma ignimbrite deposit which is widely distributed throughout the Maipo Valley and Central Valley close to Santiago (Stern et al. 1984a). The fluvial deposits of the highest terrace are folded and faulted, showing cumulative displacements of c. 11 m. Striae on c. east–west striking reverse fault planes and on imbricate cobbles are compatible with a subhorizontal $\sigma_1$ trending N338°E. Shortening directions, as obtained from drag folds, are consistent with a NNW compression (Fig. 9.21).

**Mina El Teniente (34°S).** A systematic study of recent deformation was undertaken at the El Teniente mine, located in the main range. The main source of fault-slip data was a 2.8 to

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**Fig. 9.19.** Field evidence for Pleistocene, north–south trending compression in the forearc of Central Chile, near San Antonio. Mesoscopic faults with centimetre- to metre-scale reverse sense of displacement affect both the Palaeozoic gneisses and the Pleistocene strata showing well-defined striae. $\sigma_1$, as obtained from fault slip data, trends N185°E.
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Fig. 9.20. Pleistocene sediments in Esperanza and Victoria (37°S) are affected by east–west striking reverse faults, compatible with a north–south to NNE–SSW compressional stress regime.

3.9 Ma lamprophyric dyke, which occurs inside and outside the mine (Cuadra 1986). Twenty-eight mesoscopic striae-bearing faults measured in the lamprophyric dyke are compatible with subhorizontal principal stress directions \( \sigma_1 = N23^\circ E \) and \( \sigma_3 = N293^\circ E \), respectively. This strike-slip stress tensor is believed to be Pleistocene. Thirteen mesoscopic striae-bearing faults measured in the Braden Breccia are compatible with \( \sigma_1 = N180^\circ E \) (Fig. 9.22).

Laguna del Laja–Lonquimay–Copahue (37°S). Folguera et al. (2005) recognized a set of post-5 Ma NE and ENE faults making up a transtensional imbricate fan structure which is spatially associated with Recent volcanism. The LOFZ merges north into the Antiñir transpressional fault system in Argentina, which represents the present-day active tectonic front of the Andes between 38°S and 39°S latitude (Fig. 9.13).

Southern segment between 39°S and 46°S: Coastal Cordillera There is a complete lack of published field data on the geometry and kinematics of Pleistocene deformation along the Coastal Cordillera at these latitudes. Our limited regional observations suggest that this part of the forearc has accommodated little if any Pleistocene deformation. This is somewhat surprising considering that this region corresponds well to that affected by the great 1960 Valdivia earthquake.

Southern segment between 39°S and 46°S: Central Valley

Fresia-Nueva Braunau (40°S). Although at 40°S no important lineaments are observed as marking the boundary between the Coastal Cordillera and the Central Valley, sediments in the latter contain evidence for a mid- to late Pleistocene NNE-trending compression event. Mid-Pleistocene fluvioglacial sediments, coeval with the Quilleco Polygenetic cone deposits (Thiele et al. 1998), are affected by a weak but significant deformation. Fault striae present on pebbles and cobbles are compatible with a subhorizontal compression ranging from N204°E to N26°E at Fresia and N48°E at Nueva Branau yielding a NNE-trending average compression direction (Fig. 9.23).

Ancud–Lajas (41°S). Pleistocene fluvioglacial sediments occurring at Chiloe Island, similar to those found at Fresia and Nueva Branau, also are deformed and faulted. Thirty mesoscopic faults were measured at two sites (Ancud 1 and 2) belonging to a Pleistocene moraine deposit dated as younger than 0.76 \( \pm 0.21 \) Ma (Vergara & Munizaga 1974). The analysis of ten faults at Ancud 1 yielded a principal stress direction \( \sigma_1 = N17^\circ E \), with a horizontal \( \sigma_3 = N280^\circ E \); Ancud 2 site, with 20 faults, gave a principal stress direction \( \sigma_1 = N29^\circ E \) (Fig. 9.24A,B). Eight faults affecting a 25.6 \( \pm 0.7 \) Ma ignimbrite (Stern & Vergara 1992) at Ancud 3 are compatible with a horizontal compression \( \sigma_1 = N156^\circ E \) (Lavenu & Cembrano
At Lajas site (Fig. 9.24C), the same Pleistocene sediments show faults and striae compatible with $\sigma_1 = \text{N}341^\circ\text{E}$ (Lavenu & Cembrano 1999b).

**Southern segment between 39°S and 46°S: Main Cordillera Liquiñe (39°S).** According to Rosenau (2004), results from drainage network anomalies around Liquiñe are consistent with lineament analysis for the same area highlighting the importance of north–south to NE–SW subvertical fracture zones in the tectonic development of the region. The latter structures have been interpreted by López-Escobar *et al.* (1995a) and Rosenau (2004) as tension cracks spatially associated with Pleistocene volcanism. Furthermore, the N55–60°E orientation of volcanic features would reflect the direction of the main principal stress $\sigma_1$ in the southern Andes intra-arc zone.

**Cuesta Los Añiques and Lago Caburgua (39°S).** A 8.1 Ma granite exposed in a quarry along one of the main lineaments of the LOFZ exhibits numerous NNE-striking right-lateral faults and ESE-striking left-lateral faults compatible with subhorizontal principal stresses $\sigma_1 = \text{N}238^\circ\text{E}$ and $\sigma_3 = \text{N}147^\circ\text{E}$ (Fig. 9.25A). This stress tensor may represent a post-Pliocene dextral transpressional deformation along the LOFZ that is also well documented farther south. Another fault population, made up of NNE-striking normal–left-lateral faults are in turn compatible with a different, extensional stress tensor with NE-trending $\sigma_3$. The age and significance of this deformation is unknown, but may be related to the development of the Central Valley at this latitude. Major NNE-trending and ENEn-trending lineaments of the LOFZ cut an 8.5 Ma granodiorite along the eastern margin of Lago Caburga. There, the great majority of mesoscopic faults show strike-slip kinematics, yielding a N228°E-trending $\sigma_1$. Although the resulting stress tensor is compressional in nature, it is fairly consistent with Pliocene dextral transpression along the LOFZ (Fig. 9.25B).

**Reloncavi and Volcan Apagado (41–42°S).** Farther south, the NNE-trending faulted margins of the Reloncaví fiord define the LOFZ at c. 41ºS. Plurimetric to hectometric vertical faults developed on both Cretaceous and late Miocene granites show superposed sets of striae: an early down-dip set and a late, subhorizontal set. Faults striking 040–110° (mainly 60–80°) are left-lateral reverse-slip. Faults striking 325–25° (mostly 10–20°) show right-lateral reverse to normal slip. This fault population is compatible with a N221°E-trending $\sigma_1$, interpreted as a Pleistocene event related to dextral displacement along the LOFZ (Fig. 9.26). Reverse faults with decimetric displacement affect semi-consolidated pyroclastic deposits spatially associated with Apagado Volcano (Fig. 9.26). These deposits are attributed to the Plio-Pleistocene ‘Estratos de Contao’ unit as defined by Alarcón (1995). Fault-slip analysis for the stress tensor yields a subhorizontal principal compression direction $\sigma_1$ trending N212°E. This is consistent with Pleistocene, intra-arc, dextral transpression.
Hornopiren (42°S). The Cholgo channel, east of Hornopirén, follows the regional trend of the LOFZ at these latitudes (Fig. 9.26). Here, over 50 mesoscopic faults cutting a Pliocene granitoid (3.59 Ma; Schermer et al. 1995) yielded a well-constrained subhorizontal principal stress direction $\sigma_1 = N236^\circ E$ (NE–SW) and a subhorizontal $\sigma_3$ (Fig. 9.26). This stress tensor is compatible with dextral strike-slip displacement on NNE-striking faults of the LOFZ. The fact that Hornopirén Volcano is elongated NE is consistent with Pleistocene to Recent dextral strike-slip deformation of the volcanic arc domain (e.g. Dewey & Lamb 1992).

Puyuhuapi, Queulat and Puerto Cisnes (44°S). Miocene diorites and granodiorites cut by Pliocene (3.8 Ma) mylonitic shear zones are exposed along the LOFZ, south of Puyuhuapi (44°S, Fig. 9.26) (Arancibia et al. 1999; Cembrano et al. 2002). Mesoscopic faults overprint the earlier ductile fabrics and document a dextral transpressional regime with $\sigma_1 = N52^\circ E$ (NE–SW) and $\sigma_3 = N321^\circ E$ (NW–SE) and with $\sigma_1 = N211^\circ E$ (NNE–SSW) and $\sigma_3 = N116^\circ E$ (WNW–ESE). This deformation post-dates a 1.6 Ma cooling age for biotite from a granodioritic unit (Cembrano 1998; Lavenu & Cembrano 1999b) and so can be considered as Pleistocene. The direction of the maximum horizontal stress is subparallel to an elongated cluster of monogenic alkaline volcanoes, suggesting that dextral transpression is still active (López-Escobar et al. 1995a).

In the southern part of the Andes, and along the Liquiñe–Ofqui Fault Zone (Fig. 9.17), published fault-slip data for the intra-arc region between 39°S and 47°S indicate that the state of stress was transpressional, with $\sigma_1$ oriented in a NE–SW direction. The Pleistocene transpressional deformation has been dated as younger than 1.6 Ma (Lavenu & Cembrano 1999b). Partitioning of deformation is recorded along and across the western continental margin between 33°S and 47°S (Fig. 9.27). Horizontal compression trending north–south to NNE–SSW has affected the Quaternary deposits at the coast, in the Central Valley and in the Main Cordillera, and McKinnon & Garrido de la Barra (2003) have also observed the north–south trending shortening as Quaternary. The intra-arc zone, where the LOFZ is located, has been affected by NE–SW transpressive deformation with dextral movement along the main fault zone. This north–south to NE–SW deformation has been dated as after 2.4 Ma in the Santiago region (33°S), after $0.76 \pm 0.21$ Ma in Chiloé Island (43°S) (Vergara & Munizaga 1974) and after 1.6 Ma in the Puyuhuapi area (45°S). To the east of the LOFZ, the remainder of the Cordillera and the Argentinian Subandean zone have little or no evidence of Quaternary deformation (Diraison et al. 1998). Recent contributions (Melnick et al. 2002; Folguera et al. 2004) suggest the continuation of the northernmost trace of the LOFZ into the eastern Andean hillslope between 37°S and 38°S, through the Antiñir–Copahue Fault, where Quaternary activity and gravity-related phenomena have been reported. Thus, the LOFZ seems to accommodate much of the upper plate deformation, with no significant deformation after the middle Miocene recognized in the Argentinian Subandean zone between 33°S and 52°S (Costa et al. 2000a; Cembrano et al. 2002).

**Relationship between neotectonics and seismicity for central and southern Chile**

Previous studies have shown that the subducted oceanic Nazca Plate beneath South America exhibits strong along-strike variations in the dip angle that suggest segmentation of the downgoing slab (Barazangi & Isacks 1976; Cahill & Isacks 1982; Barazangi & Isacks 1983; Dilek & Jackson 1998).

In the southern part of the Andes, and along the Liquiñe–Ofqui Fault Zone (Fig. 9.17), published fault-slip data for the intra-arc region between 39°S and 47°S indicate that the state of stress was transpressional, with $\sigma_1$ oriented in a NE–SW direction. The Pleistocene transpressional deformation has been dated as younger than 1.6 Ma (Lavenu & Cembrano 1999b). Partitioning of deformation is recorded along and across the western continental margin between 33°S and 47°S (Fig. 9.27). Horizontal compression trending north–south to NNE–SSW has affected the Quaternary deposits at the coast, in the Central Valley and in the Main Cordillera, and McKinnon & Garrido de la Barra (2003) have also observed the north–south trending shortening as Quaternary. The intra-arc zone, where the LOFZ is located, has been affected by NE–SW transpressive deformation with dextral movement along the main fault zone. This north–south to NE–SW deformation has been dated as after 2.4 Ma in the Santiago region (33°S), after $0.76 \pm 0.21$ Ma in Chiloé Island (43°S) (Vergara & Munizaga 1974) and after 1.6 Ma in the Puyuhuapi area (45°S). To the east of the LOFZ, the remainder of the Cordillera and the Argentinian Subandean zone have little or no evidence of Quaternary deformation (Diraison et al. 1998). Recent contributions (Melnick et al. 2002; Folguera et al. 2004) suggest the continuation of the northernmost trace of the LOFZ into the eastern Andean hillslope between 37°S and 38°S, through the Antiñir–Copahue Fault, where Quaternary activity and gravity-related phenomena have been reported. Thus, the LOFZ seems to accommodate much of the upper plate deformation, with no significant deformation after the middle Miocene recognized in the Argentinian Subandean zone between 33°S and 52°S (Costa et al. 2000a; Cembrano et al. 2002).
The shape of the Nazca Plate beneath central Chile, which is inferred from the top of the local and teleseismically recorded seismicity in the zone, shows a sharp transition around 32–33°S from subhorizontal slab to the north to a 27°-dipping slab at intermediate depths to the south. The central Chile subduction zone is very seismically active, with magnitude 8 interplate thrust earthquakes having occurred historically along its entire length (Comte et al. 1986; Beck et al. 1998). The zone between 31°S and 33°S has been the nucleation site for several large thrust events (e.g. 1944, Ms = 7.4; 1971, Ms = 7.5; 1977, Ms = 7.4; 1985, Mw = 8.0), intraplate events below the subduction interface, and events within the subducted slab (1965, Ms = 7.5). The Punitaqui earthquake of 15 October 1997 (Mw = 7.1) is an example of an intraplate earthquake in the flat-slab zone (Pardo et al. 2002b).

South of latitude 32–33°S, the slab steepens in dip from nearly 10° to 25–30° (Pardo et al. 2002a). Based on teleseismic and local data, Barazangi & Isacks (1976) proposed that this change in dip is accommodated by a tear in the slab or by a continuous southward flexure or ramp. More recently Cahill & Isacks (1992) have proposed a continuous flexure as the transition between the flat-slab to the north and the steep subduction zone to the south of 33°S.

The great San Antonio–Valparaiso earthquake (Ms = 7.8) in central Chile occurred on 3 March 1985. Fault slip was concentrated at a depth between 30 and 40 km within the interplate zone and induced coseismic coastal uplift (Comte et al. 1986; Barrientos 1988). There is no field evidence for reactivation of surface faults as a result of this relatively deep crustal earthquake.

In south-central Chile, Campos et al. (2002) document shallow earthquakes beneath the Andean volcanic chain. In the Central Valley, two events that were located at depths of 28 and 42 km on N20–30°E faults are puzzling. Their focal mechanisms are compatible with a northward displacement of the Coastal Cordillera block relative to the Central Depression.

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**Fig. 9.23.** Pleistocene compression recorded in gravel deposits of the ‘Rodados Multicolores’ in the Central Depression close to Fresia (40°S).
These displacements are consistent with well documented compressional to transpressional deformation associated with north–south to NNE-trending compression in the Central Valley and volcanic arc, respectively. This in turn suggests that the roughly north–south coseismic deformation need not be compatible with the interseismic east–west deformation indicated by GPS data.

After the Lonquimay/Navidad crater erupted on 25 December 1988 in southern Chile (38°S), seismicity increased over the next few days (Barrientos & Acevedo 1992). According to these authors, the largest shallow crustal event ($M_w=5.3$) of the series took place on 24 February 1989, at between 6 and 10 km depth. The focal mechanism corresponded to a right-lateral compressive (transpressional) strike-slip fault that coincides with the strike of the Liquiñe–Ofqui Fault Zone. This focal mechanism shows a NE–SW to NNE–SSW trending compressional stress axis that corresponds to the Quaternary surface transpressional stress documented by hectometric to kilometric faults in the field.

The largest earthquake recorded in the world was the great Valdivia earthquake of 22 May 1960 (Plafker & Savage 1970), which was estimated at $M_s=8.5$ (minimum) and $M_w=9.4$. The earthquake occurred along the subduction interface thrust zone (25 km depth, 1000 km length rupture, 200 km down-dip width, and displacement of 20–40 m) and produced coseismic vertical displacement (warping uplift and subsidence) in the coastal zone. No precise focal mechanism was available at the time of the earthquake. There is no convincing evidence for surface fault activity associated with this or other subduction interface earthquakes in historic times. According to Plafker & Savage (1970), the southern part of the Central Valley, between 38°S and 46°S, experienced post-seismic deformation, mainly vertical displacements.

In southern Chile, Chinn & Isacks (1983) reported a teleseismically recorded shallow crustal earthquake near the Hudson Volcano (46°S, 73°W) on 28 November 1965. With a depth of 33 km and $M_s=6$, this earthquake could have occurred on a NNE–SSW trending right-lateral fault that would belong to the right-lateral Liquiñe–Ofqui Fault Zone, which is located about 25 km to the west. This earthquake is consistent with our neotectonic observations along the LOFZ. Further south, in Laguna San Rafael (46.75°S, 73.90°W), sag ponds and dextral strike-slip indicators along reactivated fault planes are compatible with the NE–SW transpression documented elsewhere.

**Discussion**

Published analyses of fault-slip data show that the stress field in many areas throughout the central and southern Andean forearc switched from widespread Pliocene east–west compression to Pleistocene north–south compression or north–south and east–west extension near the Pacific coast. In contrast, the state of stress in the backarc above the flat-slab zone (Sierras Pampeanas in Argentina) has remained compressional and east–west during Pliocene and Pleistocene times. These differences in regional stresses cannot be easily accounted for by plate kinematics alone because, along the central Andes, the convergence vector has not changed significantly during the past 6 Ma. Furthermore, when considering the whole Andean chain from forearc to backarc regions, different regions record different stress fields during both the Pliocene and Pleistocene as a function of the distance from the trench, slab geometry and topography. Which then are the processes responsible for the nature and spatial distribution of these different tectonic regimes?

On the Pacific Coast, the east–west trending extension is the most recent deformation observed, and it corresponds to the elastic rebound of the continental plate edge after large subduction interface earthquakes. Displacement results from two
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deformation processes: (1) interseismic accumulation of elastic strain in relation to the subduction coupling; and (2) coseismic strain release during the earthquake.

How can we explain that in the central and southern Chile forearc the compressional direction became north–south? The following processes can account for the north–south compression: (1) strike-slip partitioning arising from oblique convergence favours a potential northward displacement gradient of a forearc sliver against a buttress producing north–south shortening (e.g. Beck et al. 1993); (2) forearc northward motion is enhanced by the collision of successive portions of the Chile Ridge; (3) the kinematics of the forearc deformation depends on the geometry of the plate boundary and the convergent margins concave towards the subducting plate, thus causing margin-parallel shortening as a ‘buttress effect’ (Beck 1991; McCaffrey 1992, 1996; Allmendinger et al. 2005); and (4) another ‘buttress effect’ can also be the abrupt shallowing in the slab dip from 30° south of 33°S to subhorizontal north of 33°S. Allmendinger et al. (2005) also show good evidence of Neogene trench-parallel shortening in the forearc of northern Chile. In their interpretation, margin-parallel shortening arises from a north–south elastic velocity gradient at a locked, curved plate boundary, as modelled by Bassis et al. (2001). As a comparison, in North America, Johnson et al. (2004) have similarly interpreted the north–south margin-parallel shortening of the Cascadia forearc to be a consequence of oblique (NE–SW) subduction of continental and oceanic plates at a curved margin.

Subduction of oceanic ridges does not appear to induce long-term regional compressional deformation, but rather enhances continental margin uplift. There is no conclusive evidence of more significant or widespread deformation under compression above subducted oceanic ridges: during the Pliocene, for example, compression along all the active margin of the Andes was oriented east–west.

Deformation caused by ridge subduction is observed close to the trench and is associated with an increase in the seismic activity at the subduction zone. Different types of deformation in the trench sedimentary fill and/or in the accretionary wedge are observed with or without seismic activity such as in the Andes, Marianas and Costa Rica (Cloos & Shreve 1996; von Huene & Ranero 2003).

The coastline above a subducting ridge shows a significant increase in surface uplift. This is the case for the Nazca Ridge, above which Quaternary and Pliocene marine terraces and their deposits are uplifted to high altitudes not seen in any other part of the Pacific coast of South America. The 125 ka marine terraces (MIS 5) are as high as 105 m a.s.l. and the Pliocene terraces reach altitudes of 780 m a.s.l. (Ortlieb & Macharé 1990) (Fig. 9.28).

Fig. 9.25. Fault-slip data and stress tensor for Pleistocene deformation along the northern portion of the LOFZ at Los Añiques (A) and Caburga (B). A subhorizontal σ1 trending NE–SW is compatible with right-lateral displacement along the LOFZ.
Near latitude 33°S, the Juan Fernandez Ridge deforms the thinned crust at the continental margin and inland through the forearc, volcanic arc and backarc. According to von Huene et al. (1997) it is difficult to correlate existing tectonic structures on the overriding plate with the arrival of the ridge; instead they relate it to previous tectonic events, although other authors have proposed otherwise (Yañez et al. 2002). There is no evidence of uplifted Quaternary marine terraces on the continental margin affected directly by the predicted subduction of the Juan Fernandez Ridge. On the contrary, a few degrees to the north, in La Serena–Altos de Talinay and Caldera, Quaternary marine terraces have been uplifted as much as 200 m a.s.l. without an oceanic ridge being subducted. Similarly, over other oceanic ridges or seamounts worldwide, no significant uplift is noticed at the forearc region. Generally the coastal zone uplift has marine terrace heights which could be classified as ‘normal’, with a MIS 5 terrace at 25 to 30 m a.s.l. and the upper Quaternary terraces at 200 m a.s.l.

After 22–25 Ma, DeMets et al. (1990) and D. E. Engebretson (pers. comm. 1995) indicate slight increases within an overall slowdown in the convergence rate at 4.9 Ma in the early Pliocene, and near 2 Ma in the latest Pliocene. On the contrary, during the same time, Somoza (1998) indicates a slowdown in the convergence rate which is coeval with a global event during early Pliocene times and with an east–west trending compression that affected the whole mountain range. After 2.4 Ma another, less intense global change affected the chain in a differential way, yet corresponds to Engebretson and DeMets
models. In effect, there are changes in the recorded stress field in Chile as follows: after 5.5 Ma (Puerto Cisnes), 4.0 Ma (Mina El Teniente), 2.4 Ma (Mina El Teniente) and, during the Pleistocene, after 1.6 Ma (Puyuhuapi) and middle Quaternary (Chiloé).

**Short-term upper plate deformation: GPS data**

One key question in neotectonics is how long-term and short-term deformation data can be reconciled. For instance, it seems reasonable to compare the recent stress field obtained from fault-slip data with that implied by modern GPS measurements. Published GPS data concerning deformation above the Nazca–South America plate boundary in the central Andes consistently show west–east velocity vectors that are nearly parallel to the convergence vector (Norabuena et al. 1998; Bevis et al. 1999, 2001; Kendrick et al. 1999, 2001; Klotz et al. 2001; Brooks et al. 2003). These authors interpret these displacements as the result of the interseismic crustal velocity field. During large subduction earthquakes, slip along the plate interface induces coseismic east–west extensional deformation in the upper plate, roughly parallel to the slip vector convergence (Fig. 9.18). This was the case for the 1960 southern Chile earthquake (Plafker & Savage 1970; Barrientos & Ward 1990) and the 1995 Antofagasta earthquake (Delouis et al. 1998), both of which were located in the outer forearc in the western Coastal Cordillera area. Another paper (Ruegg et al. 2002) interprets the N89–90°E GPS-derived velocity vectors relative to stable South America as reflecting interseismic strain accumulation above the Nazca–South America subduction zone.

Following Delouis et al. (1998) we assume that the interseismic compression in the overriding plate is due to the plate motion, while the plate interface is locked at shallow depth, as illustrated in Bevis & Martel (2001). Therefore, within the southern forearc of Chile, we can interpret the north–south neotectonic compression deduced from microfault analysis to be the result of coseismic deformation, oriented perpendicular to the slip vector and due to the partition of deformation inside the forearc and along its edge with the backarc region (Fig. 9.19). The focal mechanisms reported by Campos et al. (2002) appear to support this interpretation.

![Fig. 9.27. Summary figure showing the results of the microtectonic analysis of Pleistocene faults from central and southern Chile and other parts of the Andes, for comparison. East–west extension is recorded along the Pacific Coast, at the Arauco Peninsula; north–south to NE–SW compressive and strike-slip regimes are documented in the forearc and the intra-arc zones in Chile; east–west compressional tectonic regime in the Argentinian central foreland (SE of Sierras Pampeanas) (e.g. Bettini 1980; Costa et al. 2000b).](image-url)
In the southern part of the central Andes (central and southern Chile) we suspect that the slip partitioning into the forearc corresponds to the development of a forearc sliver (Fig. 9.29) as proposed in the model of Fitch (1972), with the LOFZ being the limit between the forearc and the backarc region of the South American Plate.

Bevis et al. (2001) and Bevis & Martel (2001) regard the Andean Cordillera as a rigid microplate located between Nazca and South American plates. In fact, in the middle and southern parts of the central Andes, one cannot consider this really a rigid portion because it is cut into several thin slivers, parallel to the trench and the main direction of the Cordillera. The partitioning of Cenozoic deformation was achieved along several principal right-lateral reverse faults that trend N10°E, and form a transpressional flower structure (Fig. 9.29) (Cembrano et al. 2002). This slip partitioning could correspond in part to a ‘smoothly distributed plastic’ deformation within ‘a mechanically coherent forearc’ as in the Bevis & Martel (2001) model, but the lack of structural or offset stratigraphic markers does not allow us to determine the amount of displacement along the LOFZ and within the forearc slivers.

In a recent article, Allmendinger et al. (2005) have demonstrated that interseismic elastic deformation and permanent deformation show surprisingly similar spatial and temporal patterns. In particular, the sense and amount of long-term block rotations in the central Andes Bolivian Orocline, as revealed by palaeo-magnetic data are consistent with the instantaneous picture given by GPS data when extrapolated in time. Consistency between elastic and permanent deformation would reflect that both types of deformation are driven by the same stress field.

Conclusions

Neotectonic studies along and across the central and southern Chilean Andes document a number of different tectonic regimes during Pliocene and Quaternary times. Extensional, compressional, transpressional and transtensional deformations are each recorded in various morphostructural zones and several mechanisms are proposed to explain these tectonic changes in time and space. Because we have access only to surface faults, the local analysis of fault-slip data does not necessarily reflect the general state of stress of the Cordillera which has been under a relatively constant convergence regime for the past 10 Ma, both in magnitude and direction. However, there is a clear regional pattern to stress fields that are well correlated with major tectonic domains such as the forearc, volcanic arc and foreland.

1. Segments of the coast, close to the trench, are mostly affected by east–west trending extension, a consequence of the elastic rebound of the upper plate which is subjected to uplift and westward displacement during and after compressional earthquakes of the subduction interface zone. Alternatively, outer forearc extension can be attributed to interseismic folding and associated stretching of the surface (e.g. González et al. 2003; Allmendinger et al. 2005)

2. Inside the forearc zone, farther away from the trench, a weak but well documented north–south compression (i.e. central and southern Chile) results from coseismic deformation partitioning of oblique convergence and a concomitant northward lateral displacement gradient of the forearc block against a transversal buttress (i.e. the heterogeneity on the subducting Nazca Plate near latitude 33°S). There is still little evidence of Plio-Quaternary surface deformation in the inner forearc of northern Chile.

3. The Andean intra-arc zone between 39°S and 46°S has undergone NE-SW transpression (Liquiñe–Ofqui Fault Zone) as a consequence of coseismic and/or interseismic deformation partitioning.

4. The Andean orogenic front above the Pampean flat-slab and the Sierras Pampeanas undergoes east–west compression with a variable strike-slip component, although local north–south compression during the Quaternary cannot be ruled out.

5. Current GPS data showing periods of interseismic shortening parallel to the convergence vector and coseismic extension are only partly compatible with active structures observed in the field. Only the outer forearc shows evidence of coseismic crustal extension; the inner forearc does not record significant deformation amenable to interseismic
east–west shortening. The north–south compression documented in the inner forearc of central-southern Chile can be attributed to coseismic deformation partitioning of oblique convergence.

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Fig. 9.29. Kinematic partitioning of oblique plate convergence. 1. Geocartoons showing three states of the deformation of the active margin (a) during interseismic (b) and coseismic (c) deformation before normal state (d). 2. Schematic block diagram showing how heterogeneous compressional and transpressional deformation accommodates oblique convergence along and across the plate boundary (state (c) of the geocartoons). Convergence is resolved into east–west extensional deformation in the outer forearc zone, north–south contraction in the inner forearc zone, and transpressional deformation concentrated in the volcanic arc zone. LOFZ, Liquiñe–Ofqui Fault Zone.
Earthquakes in Chile

SERGIO E. BARRIENTOS

No recorded human generation in Chile has escaped the damaging consequences of large earthquakes. More than ten events with magnitudes equal to or greater than magnitude 8 have taken place during the twentieth century alone. Among these earthquakes is the 1960 event, the largest earthquake ever recorded since the beginning of instrumental seismology. Such extreme seismic activity is a result of the interaction of the Nazca, Antarctic, Scotia and South American plates in southwestern South America where Chile is located (Fig. 10.1).

In general, seismogenic zones in Chile are basically well established: large shallow (0–50 km) thrust earthquakes along the coast; large deeper (70–100 km) tectonic as well as compressional events within the subducting Nazca Plate; and very shallow seismicity (0–20 km) in a few places, such as the cordillera region of central Chile and the southern extremity of the continent by the Magellan Strait. Deeper seismicity (150 to 650 km) occurs further to the east, beneath Bolivia and southwestern Argentina (Figs 10.2 & 10.3).

The large thrust earthquakes, responsible for most of the damage recorded in history, are located along the coast from Arica (18°S, the northernmost extreme of coastal Chile) to the triple-junction at Taitao Peninsula (46°S). With magnitudes that can reach values well over eight, these events are usually accompanied by noticeable coastal elevation changes and, depending on the amount of seafloor vertical displacement, by catastrophic tsunamis. Their rupture zones are limited to the coupled region between the Nazca and South American plates which extends down to 45–53 km depth (Tichelaar & Ruff 1991) and their lengths could reach well over 1000 km. Their spatial and time characteristics have been studied (Kelleher 1972; Barrientos 1981; Martin 1991; Nishenko 1985; Ramirez 1988; Beck et al. 1998), so that the hazard due to these large events is well recognized and understood. Return periods for magnitude 8 events are of the order of 80 to 130 years for any given region in Chile, but about 12 years when the country is considered as a whole. Megathrust earthquakes seem to have much longer return periods, of the order of a few centuries for any given region (Cifuentes 1989; Barrientos & Ward 1990; Cisternas et al. 2005). The 1939 Chillán event does not belong to this suite of thrust earthquakes; however, it has been the most damaging earthquake in terms of loss of human life in Chilean history.

North of the Chile triple-junction, the Nazca Plate is being subducted beneath South America (SA) at a rate of about 65 mm/year (Angermann et al. 1999). South of the junction as far as 57°S, the Antarctic Plate (AN) is being subducted at a considerably slower rate, between 10 and 20 mm/year (Pelayo & Wiens 1989). The slow convergence rate and the type of subducted structures are probably the main reasons for the relative lack of seismicity on the western margin of South America south of the Chile triple-junction. The fundamental tectonic difference between the region north of the triple-junction and that to the south is that the Chile Ridge has been recently subducted in the latter.

South of the western end of the Magellan Strait (c. 52°S), the AN is no longer being subducted under the SA, but under the Scotia Plate (SC) (Winslow 1982; Pelayo & Wiens 1989; Klepeis 1994), so that the southernmost tip of South America is not on the SA but on the SC. The SA–SC boundary is a diffuse boundary displaying mostly left-lateral strike-slip relative motion, which runs approximately along the Magellan Fault (Katz 1964), from the easternmost tip of Tierra del Fuego (Strait of Magellan) to Seno Fagnano, Seno Almirantazgo and into the NW Magellan Strait. North of c. 52°S, the SA and AN converge normally at a rate of 19 mm/year. South of this point, part (c. 7 mm/year) of the convergence rate is accommodated by the SA–SC boundary along the Magellan Fault, and the AN and SC converge at a lower rate of 13 mm/year. Historical earthquakes in 1879 and 1949 (in which two events took place on 17 December, separated by about 8 hours), all of them with estimated magnitudes above 7, are the largest earthquakes recorded in the region.

More recently, the hazard produced by large, intermediate-depth tensional earthquakes within the subducting Nazca Plate has been recognized. These events are caused by faulting within the subducting plate at depths between 80 and 110 km with apparently high stress drops (Kausel 1991). Extreme examples of these type of events are the M = 8 Calama earthquake in 1950 (Kausel & Campos 1992), the M > 8 Chillán event in 1939 (Beck et al. 1998) and the most recent large event (M = 8.0) of 30 June 2005 which took place within the Nazca Plate about 200 km east of Iquique, the largest city in the region. Malgrange & Madariaga (1983) have reported a suite of tensional events of this type within the Nazca Plate in the >7 magnitude range, of which the 1965 La Ligua event stands out for the destruction it produced.

Additionally, complex stress interaction gives rise to down-dip compressional events at about 60–70 km depth. These events can reach magnitudes over 7, as recently reported by Lemoine et al. (2001) and Pardo et al. (2002b), in particular for the very damaging earthquake of October 1997 near Illapel.

Another seismogenic region, which has become the subject of recent studies, is that located at shallow depths in the Andean cordillera in the central part of Chile. Godoy et al. (1999) and Barrientos et al. (2004b) have carried out structural and seismicity studies to understand this shallow active region, in which the largest known earthquake (less than 10 km depth) took place on 4 September 1958 (M = 6.9; Lomnitz 1961; Pinder 1961). This activity is probably the surface consequence of the change in dip angle with which the subducting Nazca Plate underthrusts (flat subduction) beneath the South American Plate (Gutscher 2003).

Large events and characteristics of the seismogenic region

The large Chilean earthquakes have inevitably commanded the most attention since seismicity began to be recorded in the country (Silgado 1985). Only recently, in the last couple of decades, has a significant effort towards understanding the fine structure of the Chilean seismogenic zone materialized. The advancement in quantity and quality of instruments of global networks and the deployment of permanent and temporary local and regional networks has been the key to elucidating the
detailed knowledge of the geometry of subduction zones (Wadati–Benioff) and their characteristics such as earthquake focal mechanisms, state of stress, coupling extent, and presence of double seismic regions.

The first systematic seismological effort in Chile was initiated in the first decade of the twentieth century after the occurrence of the significant 1906 Valparaíso earthquake. The government of President Pedro Montt decided to invite Ferdinand Montessus de Ballore (1889–1923) to initiate seismological work in the country. Parts of his plans are reflected in the first page of the first issue of the Bulletin of the Seismological Society of America (1911) which includes a map of the existing seismic network in Chile (Fig. 10.4). This network includes a central station in Santiago comprising a two-component horizontal Wiechert pendulum weighing 183 kg and a vertical one with a mass of 163 kg, two 100 kg Bosch-Omori pendulums and an 850 kg Stiattesi two-component pendulum. Four stations of second order in Tacna (now part of Peru), Copiapo, Osorno and Punta Arenas completed the National Seismic Network, each of them equipped with a one-component horizontal Wiechert pendulum of 200 kg. To detect local earthquakes, 29 Agamennone seismoscopes (pendulums recording on curved smoked glass) were distributed across an equal number of locations along the country.

Another major contribution of Montessus de Ballore was his summary work on the seismic history of the region published in several volumes between 1911 and 1916, depicting in detail the major earthquakes and tsunamis which took place before the time of his writings. Perhaps the first regional-scale effort was that initiated by the Carnegie Institution of Washington in northern Chile in the 1960s with the installation of five instruments inland from Antofagasta. One of the first local systematic efforts in central Chile was the study of the aftershocks of the MW = 8.1 event in Central Chile which took place in March 1985 (Comte et al. 1986). After this study, not only the characteristics of the aftershocks of the 1985 event were recorded and analyzed, but some of the first tomography studies in Chile were carried out after a wealth of data were acquired by means of a permanent seismic network complemented by a temporary network of seismographs (Comte et al. 1986; Pardo et al. 2003a). The current seismographic network at the beginning of the twenty-first century is depicted in Figure 10.4. These instruments are mainly composed of short-period sensors and 16-bit digitizers. A few stations around Santiago include broadband sensors with 24-bit digitizers. The stations located in Iquique (IQQ), Calama (LVC), Easter Island (RPN), Las Campanas Observatory (LCO) at the latitude of the city of La Serena, Peldehue (PEL) and Coihaique (COI) are part of international networks such as IRIS, GEOSCOPE and GEOFON, and the National Seismological Service, presently part of the University of Chile in Santiago, regularly publishes seismic activity occurring in central Chile.

Permanent and temporary seismic networks have been deployed from coast to cordillera, in collaborative efforts among different institutions, from north to south covering almost the entire territory. These studies have permitted not only the characterization of the three-dimensional velocity structure of the region, but have also allowed attenuation studies and a clear definition of the seismogenic zone. A comprehensive report of these efforts is summarized below, depicting the underthrusting characteristics of the earthquake mechanisms of very large earthquakes (Fig. 10.6).
Fig. 10.2. Relocated seismicity (black open circles) on the western flank of South America (Engdahl & Villaseñor 2003). Eight profiles show how the seismicity changes with depth as a function of distance from the trench (upward-pointing arrows, right panels) revealing the different behaviour of the Nazca Plate after subduction and the response of the overriding South American Plate in terms of shallow seismicity. The triangles represent Quaternary volcanic edifices.
(which is now a suburb of Arica). In his evaluation of earthquake potential along the circum-Pacific belt, Nishenko (1985) categorized this region as a seismic gap, together with the region south of Antofagasta, earlier identified by Kelleher (1972). The southern part of this region—Mejillones Peninsula to Paposo—was reactivated in 1995 with a moment magnitude $M_w = 8.0$ event, described in the next section. On the northern extreme of this region, an $M_w = 8.4$ earthquake took place in southern Peru in June 2001, which is considered the gap-filling event of most of the area subjected to the effects of the 1868 $M_T = 9.0$ (tsunami magnitude; Abe 1979) earthquake (Fig. 10.6). The rupture region of the 2001 event extended for 320 km up to the coastal location of Ilo in southern Peru, about 150 km north of the Arica bend (Giovanni et al. 2002). As is the case in most of the studied large events, the rupture propagated mainly from north to south (Beck et al. 1998).

Fig. 10.3. Dashed and solid lines represent contours of equal seismic depth according to Gudmundsson & Sambridge (1998) from 0 to 650 km, every 50 km, and Engdahl & Villaseñor (2003) at depths of 0 (trench), 75, 125, 200, 275, 550 and 600 km. The small squares represent the locations of seismic stations of the Seismological Network of Chile as of 2005. Network efforts around 18°S and 23°S are academic collaborations between the University of Chile and French institutions.

Fig. 10.4. Seismic Network of Chile as presented by F. Montessuss de Ballore after the occurrence of the 1906 Valparaíso earthquake (Bulletin of the Seismological Society of America, 1911).

Arica–Tocopilla (17–22°S)

Very large earthquakes have taken place historically in this region, among them the most recent underthrusting event corresponds to the 1877 $M_T = 9.0$ earthquake, where $M_T$ is the tsunami magnitude. A rupture zone of nearly 450 km (Kausel & Campos 1992) produced a large tsunami that flooded most of the coastal cities in the area. Significant subsidence was observed along the coast near Iquique and Alacran Island (which is now a suburb of Arica). In his evaluation of earthquake potential along the circum-Pacific belt, Nishenko (1985) categorized this region as a seismic gap, together with the region south of Antofagasta, earlier identified by Kelleher (1972). The southern part of this region—Mejillones Peninsula to Paposo—was reactivated in 1995 with a moment magnitude $M_w = 8.0$ event, described in the next section. On the northern extreme of this region, an $M_w = 8.4$ earthquake took place in southern Peru in June 2001, which is considered the gap-filling event of most of the area subjected to the effects of the 1868 $M_T = 9.0$ (tsunami magnitude; Abe 1979) earthquake (Fig. 10.6). The rupture region of the 2001 event extended for 320 km up to the coastal location of Ilo in southern Peru, about 150 km north of the Arica bend (Giovanni et al. 2002). As is the case in most of the studied large events, the rupture propagated mainly from north to south (Beck et al. 1998).

These two very large shocks that took place in the nineteenth century, in particular the 1868 event, which was felt from Guayaquil (Ecuador) to Valparaíso, did not go unnoticed. The
estimate maximum rupture lengths of 680 and 510 km with corresponding displacements of nearly 14 and 10 m, thus the estimated magnitudes $M_w$ would be 9.0 and 8.9, respectively, both significantly larger than those presented by Kausel & Campos (1992). The southern extreme of the 1868 rupture reached beyond the Arica bend, farther to the south than the dislocation associated with the recent 2001 event. Thus, a stretch of nearly 500 km along the coupled region of southern Peru–northern Chile (Ilo–Arica–Mejillones Peninsula) has not been subjected to significant earthquakes since 1868–1877. Comparison of tsunami amplitudes recorded in Japan, at Hakodate, indicate maximum values reaching 2 m (Soloviev & Go 1984) and 3 m (Lockridge 1985) for the 1868 event compared to maximum values of 1 m for the 2001 event.

Comte & Suárez (1995) and Comte et al. (1999) characterized the initial dip angle of the Wadati–Benioff region at 20°E and confirmed that this region also represents a double seismic region as has been observed in other subduction environments. Relatively large extensional events (normal faulting) within the subducting Nazca Plate, with hypocentres beneath the coast, have taken place recently, the event of 8 August 1987 (18.96°S, 70.02°W) with $M_w = 7.2$ at 80 km depth being the largest and most recent example (Fig. 10.7).

David et al. (2005) reported on crustal seismicity in the forearc of Arica as registered by several temporary seismographic deployments in addition to the southernmost stations of the Peruvian National Network and the permanent network of Arica, installed in 1994. A possible association of shallow seismicity with the Incapuquio Fault System (southern Peru) is reported, and the Copaqilla–Tignamar Thrust and Fold Belt (in northern Chile) might also be involved. These two systems form the limit between the Pacific Piedmont and the Western Cordillera. Farther to the south, a relatively large ($M_w = 6.3$) shallow event took place with its epicentre below 19.44°S, 69.18°W at a depth of 15 km (see Fig. 10.7). The Harvard moment tensor solution can be interpreted as a right-lateral displacement on a north–south plane dipping 46°E. This is an unusually large magnitude for a shallow event in the region.

Another important source of seismic activity is the region where the top part of the subducting plate reaches about 100 km depth and undergoes normal faulting, as was the case for the 13 June 2005, $M_s = 7.9$ event (20.02°S, 69.17°W) at 95 km depth. This event caused severe destruction in the towns and villages in the epicentral area, near Chiapa (Fig. 10.7).

Antofagasta–Taltal (22–25°S)

The 1995 Antofagasta earthquake, $M_w = 8.1$, which extended from Mejillones Peninsula to Paposo, is probably the best studied earthquake in Chile to date. Ruegg et al. (1996) presented the first characterization of the earthquake based on repeated Global Positioning System (GPS) observations and inversion of seismograms at teleseismic distances. GPS measurements indicated that the southern part of Mejillones Peninsula was uplifted about 15 cm, the city of Antofagasta itself was in a nodal axis (no vertical change) and maximum subsidence (34 cm) took place just inland behind Cerro Paranal, the highest point in the region, at 2635 m above mean sea level. Maximum horizontal displacement reached about 0.8 m to the WSW with respect to the easternmost observation point of the profile. Klotz et al. (1999) reported on 70 repeated GPS measurements observed in the area affected by the 1995 event. According to these observations, the city of Antofagasta was shifted 80 cm westward and horizontal displacements reached 10 cm as far as 300 km away from the trench. Another interesting result of their study is that the slip angle (rake) was 66°, implying that the oblique convergence between the Nazca and South American plates is accommodated by oblique convergence slip, not by slip partitioning. Additionally, they observed that the present-day crustal shortening of 3–4 mm/year is much slower than the average value of 8–14 mm/year for
the last 27 million years. A continuous GPS instrument placed in the southern part of the city of Antofagasta showed no precursory displacement or immediate post-seismic adjustments (Klotz et al. 1999). Additional indicators of elevation changes along the coast were provided by Ortlieb et al. (1996) on the basis of death of coralline algae (lithotamnium) in the intertidal environment. As this coralline alga dies its colour turns from pink to white, leaving an infralittoral fringe, which depends on the amount of elevation; this had a maximum thickness of 80 cm at the southwestern tip of the Mejillones Peninsula, while the coast at Antofagasta suffered practically no vertical change, in agreement with repeated GPS observations. Coastal uplift is noticeable, about 10 cm, again about 100 km south of Antofagasta (near Punta Tragagente). The preferred solution of Ruegg et al. (1996) involves a 180-km-long rupture on a plane dipping between 19° and 20° to the east. The source time function comprised three main pulses propagating southward; the main pulse of moment release (second) coincides with the location of the Harvard centroid calculated by Dziewonski et al. (1996).
Fig. 10.7. The estimated rupture zone (it is not known whether the Mejillones Peninsula participated, hence the dashed line) of the 1877 great earthquake lies between the sites of the Mw = 8.0 July 1995 Antofagasta and the Mw = 8.4 June 2001 southern Peru earthquakes. Three large normal-fault-type (tensional) earthquakes took place in 1950 (M = 8.0), 1987 (M = 7.2) and 2005 (M = 7.9). Earthquakes, such as the one of July 2001 (M = 6.3) at 15 km depth, at the western flank of the Andes, are not reported very often but it is clearly consistent with what has been observed further to the south, in the Precordillera at the latitude of Antofagasta. Volcanoes are open triangles and GPS observations (1996–1999; Chlieh et al. 2004) indicate shortening rates of the order of 30 mm/year at the coast.
Delouis et al. (1997), using data collected from a permanent network around Antofagasta complemented with additional temporary instruments, reported on a study of aftershocks within the context of this area experiencing a possible precursor to the impending earthquake in northern Chile (which now corresponds to the southern extension of the 2001 Arequipa earthquake and the northern extreme of the 1995 Antofagasta event, 23°S; see Fig. 10.7). These authors also reported on the peak ground acceleration of 0.28 g in the east–west component of a strong ground-motion sensor recorded in the city of Antofagasta. Subsequently, Graeber & Asch (1999) and Husen et al. (1999, 2000) studied the velocity structure of the region around the 1995 event, using both aftershocks of this event as well as regularly occurring events by means of offshore and inland networks, giving rise to one of the most detailed views of three-dimensional velocity variations and location of aftershocks for any event in Chile (Fig. 10.8). Geometrical definition and characterization of the seismogenic zone, from 20 to 46 km depth along the Wadati–Benioff zone, were among the main results of these studies. Figure 10.9 shows the two profiles, north and south of 23°S, both defining the Wadati–Benioff zone as well as an important shallow seismicity (Graeber & Asch 1999), part of which is displayed in Figure 10.10 (Belmonte 2002).

Interferometric Synthetic Aperture Radar (InSAR) was recently developed as a tool to determine the distance along the line of sight from radar satellites. Repeated observations give an estimation of the changes in distance. The first applications of InSAR to earthquake and fault monitoring in Chile have been the works of Pritchard et al. (2002) and Chlieh et al. (2004) on the 1995 earthquake. While the first study emphasizes the analysis of the coseismic displacement field, the second study concentrates on the evaluation of the post-seismic displacements. In their study, Pritchard et al. (2002) recover the whole ‘coseismic’ displacement field (combination of images taken in 1992 and 1997). The results of their work are consistent with other estimations of the order of 5 m for coseismic displacement (Ruegg et al. 1996; Ortlieb et al. 1996, Ihmlé & Ruegg 1997). One of the major outcomes of their study is that, at a resolution of few centimetres, no displacement was observed along the Atacama Fault segment just above the active portion during 1995. The Atacama Fault is a major feature that extends from south of Iquique at 22°S in its northern extreme to around 25°S, close to Taltal, and bounds the Coastal Cordillera on its eastern flank. Arabaz (1968, 1971) unsuccessfully tried to find displacement along the Atacama Fault associated with the December 1966, Ms = 7.8 Taltal earthquake.

In addition to the InSAR information, Chlieh et al. (2004) used the difference of GPS positioning along 40 benchmarks measured between 1996 and 1999 to refer the InSAR fringes to a known displacement field. They interpreted the observed surface deformation as mostly the result of aseismic slip along the down-dip extension (transition zone) from 35 km to 55 km depth as well as post-seismic displacement in the northern...

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**Fig. 10.8.** Hypocentral distribution of 789 well-located events (after Husen et al. 2000). Lower left shows an east–west cross-section and upper right shows a north–south cross-section. Stars mark the hypocentre determination of the Antofagasta earthquake by NEIC and Harvard CMT, indicating a southward rupture. The rupture extended roughly from 23.4°S to 25.0°S. At the lower right corner a histogram depth distribution is shown. Triangles and circles denote on- and offshore stations.
lateral extension of 25–45 km, under the Mejillones Peninsula, a prominent geomorphological feature just NW of Antofagasta. There is general agreement that this is the southern limit of the 1877 rupture (Delouis et al. 1997; Chlieh et al. 2004).

As was the case in the region to the north, extension occurs further along the subduction zone. A significant extensional event (Mw = 8.0) took place in December 1950 at 100 km depth near the city of Calama (Kausel & Campos 1992). The mode of rupture of this event has been interpreted as a normal fault rupturing the whole Nazca Plate, initiating at 100 km depth and extending downward.

As has been the case in the Arica region, Kono et al. (1985) examined the unusual east–west orientation of the compressive field for the 17 January 1977 earthquake (24.9°S, 68.7°W). They proposed, because of its depth at 120 km, that it might be indicative of a double Benioff zone. Many tomographic as well as velocity–structure studies have been carried out in this region utilizing mainly the aftershock of the 1995 event as well as regular seismicity (Haberland & Rietbrock 1998; Ancoop Working Group 1999; Schmitz et al. 1999; Graeber & Asch 1999; Bock et al. 2000)

**Copiapó (26–29°S)**

Farther to the south is the Copiapó region, a site of major earthquakes in 1819 and 1922. Lomnitz (2004) reported that the 1819 event had two large foreshocks, on 3 and 4 April. The main event, of magnitude 8½, on 11 April, was felt over a radius of 800 km and caused a major tsunami which destroyed the port of Caldera. According to Lockridge (1985), the 1922 earthquake generated a local tsunami with maximum wave heights of 9 m and a farfield amplitude in Japan of 39–70 cm. Kelleher (1972), from the analysis of the aftershocks recorded in La Paz, Bolivia, deduced a rupture extension of about 470 km in a north–south orientation. Willis (1929), in the spectacular report of his field expedition to study the 1922 event, reported that the towns of Copiapó, Vallenar and La Serena were among the most severely damaged. Three main shocks within the first few minutes were reported by local residents of Copiapó, Vallenar and La Serena, hence Willis concluded that the 1922 earthquake was the result of three events with different epicentres; this is further verified by Beck et al. (1998) when analysing the first 75 s of the P-wave train in which three main pulses

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**Fig. 10.9.** Distribution of crustal seismicity (black points) north and south of c. 23°S. Grey points refer to PISCO data (Graeber 1997) and Wadati–Benioff zone seismicity. Figure from Belmonte et al. (2006).
are observed in the inversion for the source time function (Fig. 10.11). The northern part of the 1922 rupture zone was partially reactivated in 1983 with an event of $M_w = 7.8$ (Dziewonski et al. 1983), with hypocentre at 38 km producing a minor tsunami. Beck et al. (1998) proposed that the 1922 segment failed in multiple-asperity-type rupture. A heterogeneous style of faulting in the 1922 rupture zone is further suggested by the occurrence of two previous earthquakes in 1796 and three in 1819. The evidence thus implies that this region does not always fail in a single rupture, but at least sometimes the energy is released in clusters of medium to large events.

Comte et al. (2002) analysed a two-month period of data recorded during a deployment of on- and offshore instruments carried out in 2001. Consistent with other studies, the Wadati–Benioff zone in the coupled region (thrust interplate contact that generates large events) has an average dip angle of 20°, the same as that derived for other regions of Chile.

**La Serena–La Ligua (30°–33°S)**

P-waves recorded at teleseismic distances are consistent with an underthrusting mechanism for the 1943 Illapel $M_w = 7.9$ event (Beck et al. 1998). This earthquake is much smaller and simpler than that of 1922 (see Fig. 10.12). A single pulse of duration 24–28 s for a time function on a source with a shallow dip to the east satisfies well the constraints of the teleseismic waveforms (Beck et al. 1998). As is characteristic of the largest underthrust earthquakes in Chile, a local tsunami, which damaged some fishing boats at Los Vilos, was observed along the coast. The farfield tsunami in Japan reached maximum amplitudes of 30 cm (Lockridge 1985). Kelleher (1972), based on aftershock S–P times recorded in Bolivia, estimated the extent of the rupture zone to be about 360 km.

In October 1997, a $M_w = 7.1$ intraplate earthquake took place at roughly 31°S, 71.2°W, very close to the town of Punitaqui. Apart from the significant damage produced in the closest towns, the most important characteristic of this event is that it was the result of a nearly vertical fault with rupture initiation at 68 km depth, probably a result of the unbending of the Nazca Plate when reaching subhorizontality to the east of 70.5°W (Pardo et al. 2002b). Lemoine et al. (2001) interpreted this downward rupture as the result of down-dip compression. Choy & Kirby (2004) calculated an apparent stress drop $\tau_a$ ($\tau_a = \mu E / M_s$, where $\mu$ is rigidity; $E$ is the radiated energy; and $M_s$ is the moment of the earthquake.) of 4.4 MPa (44 bar) for this event, the largest in the suite of all normal faults worldwide between 1987 and 2001.

Prompted by the study of two significant earthquakes in 1965 ($M_s = 7.1$, $mb = 6.4$) and 1971 ($M_s = 7.5$, $mb = 6.6$, with $M_s$ being the surface-wave magnitude and $mb$ the body-wave magnitude) near the town of La Ligua, on the northernmost extension of the 1906 event, Malgrange et al. (1981) concluded that shallow underthrust events and deeper tensional-type events coexist along the coast from southern Peru to central Chile. Both events produced roughly the same peak ground acceleration in Santiago (0.19g and 0.17g). Additionally, in their comparison of these two events, Malgrange et al. (1981) point to the lower stress drop of the 1971 underthrusting earthquake (3.8 MPa or 38 bar) in comparison with the 9.1

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**Fig. 10.10.** Distribution of crustal seismicity after Belmonte (2002). Dark circles are shallow events (0–20 km depth) and grey squares represent locations of stations used during the PISCO experiment. The majority of the shallow seismic activity is located in the western boundary of the volcanic chain as well as in the Precordillera Fault System and the Salar de Atacama. Open triangles represent Quaternary volcanoes.

**Fig. 10.11.** Three main pulses (75 s) of the source time function of the November 1922 derived from the P-waveform recorded at De Bilt, the Netherlands (DBN). Solid and dashed lines represent the observed and modelled signals, respectively. The focal mechanism was derived from P-wave first motions (Beck et al. 1998).

**Fig. 10.12.** P-waveforms at station DBN (De Bilt, the Netherlands) of the 1922 (Dist = 99°), 1943 (Dist = 101°), 1928 (Dist = 104°) and 1939 (Dist = 105°) earthquakes. Observe the difference in amplitude between the 1922 Copiapo event and the rest.
conquerors, who kept written records after 1541, when the city of Santiago was founded because it was part of the first settlements of the Spanish in Chile. This is the region with the longest historic earthquake record in South America.

Valparaíso–Pichilemu (33–35°S)

This is the region with the longest historic earthquake record in Chile because it was part of the first settlements of the Spanish conquerors, who kept written records after 1541, when the city of Santiago was founded, with Valparaíso being its main sea-port.

The earthquake in central Chile on 3 March 1985 corresponds to the last example of the sequence that has been taking place regularly in the region every 82 ± 6 years. The events that constitute the sequence 1647, 1730, 1822, 1906 and 1985 present several common features: offshore epicentre location, rupture lengths of over 150 km, systematic coastal uplift and small tsunami compared to the size of earthquakes, with the possible exception of the tsunami associated with the 1730 event. Although the recurrence intervals have been very regular, the rupture lengths differ substantially; the largest rupture, close to 100 km, was that associated with the 1730 event (Comte et al. 1986; Barrientos & Kaufel 1988). Lomnitz (2004) has argued that the epicentre of the first recorded earthquake in this sequence, in 1647, might have been inland, either in the High Andes in a similar manner to the 1958 shallow event, or in what has been recently recognized as the San Ramon Fault (Rauld 2002). The San Ramon Fault trends north–south along the western flank of the Andean foothills, along the easternmost limit of the city of Santiago. However, detailed seismic monitoring of the Santiago basin during the last 30 years has not provided any evidence of seismic activity along this fault.

One of the first detailed accounts of coastal elevation changes associated with large earthquakes in this region is provided in the journal kept by Mrs Maria Graham (1824) during her short stay on the coast of central Chile. She experienced and witnessed the consequences of the 1822 earthquake, and reported elevation changes of the coast for nearly 160 km, in particular of nearly 1 m at Valparaíso and about 1.2 m at Quintero (30 km north of Valparaíso).

Okal (2005) re-evaluated the two earthquakes that took place within 30 minutes of each other on 17 August 1906 in the Alue-tian Islands and central Chile, through the analysis of mantle waves; he concludes that the latter was a regular subduction zone event with moment of $2.8 \times 10^{20}$ dyne·cm and argues that its size has been overestimated by previous authors, with the exception of Kanamori (1977), and that its rupture length did not exceed 200 km. Steffen (1907) reports coastal uplift from Llico (34.8°S) to Los Vilos (31.9°S), along more than 300 km. These seismically determined smaller coseismic moments can be reconciled with Steffen’s coastal observations if we consider an important post-seismic displacement, as was the case of the 1985 central Chile event (Barrientos 1997).

An mb = 5.5 foreshock took place about 13 s before the initiation of the 1985 Mw = 8.0 earthquake; Comte et al. (1986) place both epicentres, at 33.24°S, 71.85°W, h = 17 km, within the region in which more than 200 precursors took place up to eight days earlier. Christensen & Ruff (1986), based on the deconvolution of P and PP waves, concluded that the rupture took place at depths between 10 and 40 km and the energy was released in two pulses, separated by 16 s, with the second one propagating to the south approximately 75 km away from the first. They also noted that the aforeshock region was much larger than the proposed rupture. The mean focal mechanisms coincide with low-angle thrusting events (dip angles around 10° to 20°). Through the study of Rayleigh waves (R1, R2, R3 and R4) recorded at GEOSCOPE and IDA stations, Monfret & Romanowicz (1986) inferred that the rupture of 100–150 km propagated southward or to the SW depending on the type of surface waves used for the analysis. They estimated source time functions of 60–80 s with a moment of $1.2 \times 10^{21}$ N m ($1.2 \times 10^{21}$ dyne·cm) or Mw = 8.0. Zhang & Kanamori (1986) used the spectra of long period waves (150–300 s) to estimate the source longitudinal extent, and they found a 160-km-long rupture orientated on a 10°E azimuth (Mo = $1.2 \times 10^{21}$ N m).

Precise relocation of the 1985 aftershocks allowed Pardo et al. (1986) to conclude that the maximum area involved in the rupture included a region of 200 km by 90 km on a plane dipping approximately 10°E, with most activity within 10 and 45 km depth. Average focal mechanisms coincide with low-angle fault planes. Choy & Dewey (1988) studied the details of rupture initiation and concluded that the earthquake consisted of three stages within 30 s. They also noted that deeper aftershocks showed larger stress drops and their fault dip angles changed to 25–30° when compared to shallow aftershocks (10°).

Another interesting feature of this region is the presence of shallow seismicity (<20 km depth) in the Andes (Barrientos et al. 2004a). This seismicity extends from the Sierras Pampeanas in Argentina (Alvarado et al. 2005) and continues along the Andes to the south (Fig. 10.13). The limit of the southern extension is not well defined because this region lacks the appropriate distribution of sensors to characterize low-magnitude seismicity, but a relatively large (Mw = 6.4) event took place in August 2004 on the border between Chile and Argentina (35.17°S, 70.53°W, at 16 km depth) on a N21°E azimuth fault dipping 61° to the east with almost pure right-lateral displacement responding to a mainly WSW–ESW compressive axis. In September 1958 a shallow M = 6.9 earthquake took place in Las Melosas, in the Andes along the Maipo Valley, less than 60 km away from central Santiago (Fig. 10.14). This event reached intensity X in the Modified Mercalli Intensity Scale causing landslides which closed the roads and inflicted significant damage to a hydroelectric power plant at Los Quelchues (Lomnitz 1961); the focal mechanisms calculated at that time are presented in Figure 10.14. Six years of shallow Andean seismicity (Fig. 10.14) analysed by Barrientos et al. (2004a) show concentrated activity associated with volcanism (panels D and G) together with NW–SE trending features. No surface rupture evidence has been found to be associated with this activity.

Araneda et al. (1989) also used a 25–30° dip angle to model the static deformation revealed by repeated measures of over
a 100-km-long levelling line between Santiago and Algarrobo, located about 100 km away on the coast. Barrientos (1988) estimates the slip distribution from the information provided by the levelling lines, tide gauges at Valparaíso, and two limnigraphs located at the extremes of Rapel Lake, towards the southern part of the rupture region (Fig. 10.15). The latter two observations imply that a significant amount of immediate post-seismic displacement was associated with this event.

As has been the case in other regions, the Santiago area is also exposed to the effects of relatively large intermediate-depth (80–110 km) tensional-type earthquakes. In September 1945 a $M = 7.1$ event took place at 90 km depth which produced a peak ground acceleration of 0.13g (Lepe & Torres 1950) in Santiago compared to the 0.05g recorded with the same instrument for the 1958 event at a closer hypocentral distance. Barrientos et al. (1997) estimated that an earthquake magnitude 7.5 in this intermediate-depth zone could present recurrence periods of approximately 110 years in the region between 32°S and 37°S, generating intensities of the order VIII (Modified Mercalli Scale).

**Pichilemu–Concepción (35–37°S)**

This part of Chile marks the transition between well-defined seismic zones to the north and south, and has been recognized as a region with moment release deficit (Barrientos 1990;
Campos et al. (2002). On a few occasions past ruptures from north or south have penetrated into this region. This was the case in 1906 (Ms = 8.3) and most likely also during the 1647 and 1730 events, extended from the north and, in 1751, rupturing from the south. The last large magnitude thrust event in this region took place in 1928.

On 20 February 1835 a large earthquake took place in this region. It mainly affected the Concepción area where the damage it produced and the ensuing tsunami were described in detail by Darwin in his book ‘Voyage of the Beagle’. At the time of the earthquake, Darwin and Robert Fitz-Roy, the Captain of HMS Beagle, were in the area of Valdivia, 350 km south of Concepción. Darwin, who visited the Concepción area 12 days after the earthquake occurrence, included in his reports how the earthquake was felt, the destruction of the city and evidence of uplift of the Concepción Bay and Santa María Island, by nearly 1 m and over 3 m respectively.

Beck et al. (1998) analysed in great detail the 1928 Talca and the 1939 events which took place in this region. They concluded that the 1928 Mw = 7.9 event is a shallow underthrusting event with rupture propagating to the south. It generated a 1.5 m local tsunami, therefore much of the deformation must have taken place under the sea bed between the coast and the trench. A 28-s duration of the P-pulse with a rupture velocity of 3 km/s indicates that the 1928 event probably extended for approximately 90 km. From the S–P times for aftershocks recorded in La Paz, Bolivia, it appears that the aftershocks occurred south of the main shock across a north–south area of about 150 km (Kelleher 1972).

The 1939 Chillán event with Ms = 7.8, has been the most disastrous earthquake in terms of human life in Chile in historical times: 28 000 deaths and significant damage in the Central Valley between Linares (35.8°S) and Los Angeles (37.5°S), in particular to the city of Chillán (Lomnitz 2004). The consequences of this event were so dramatic that the President of Chile, Pedro Aguirre Cerda, formed a Commission to deal with the catastrophe, to study the cause of such massive destruction. The report of the Commission included a special

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**Fig. 10.14.** Epicentres of shallow (<20 km depth) earthquakes in the central Chilean Andes during the period 1986–2001 as determined by Central Chile Seismic Network (solid squares). The dashed lines in regions A and B correspond to the lineaments along the Pocuro Fault System (Box A) and Olivares River (Box B) respectively (from Barrientos et al. 2004b).
Arauco Peninsula–Taitao Peninsula (38–45°S)

The great Chilean earthquake of 22 May 1960, with a surface wave magnitude of 8.5 and a moment magnitude of 9.5, was the largest event recorded in the last century (Kanamori 1977). The quake and subsequent tsunami affected a region inhabited by two and a half million people, and caused over two thousand fatalities (Sievers et al. 1963; Plafker & Savage 1970). This event has a special place in seismological history, because it provided experimental confirmation of the idea that earthquakes can cause free oscillations of the Earth (Benioff et al. 1961). Nearly all the important cities in south-central Chile from Concepción to Puerto Montt suffered severe damage from shaking which exceeded intensity VIII according to the Modified Mercalli Intensity scale (MMI). Intensities above VIII were registered in areas of poor or unconsolidated soil conditions and sites exposed to topographic amplification. In Chile, for large under-thrusting events beneath the coast, the isoseismal corresponding to intensity VIII roughly mimics the size of the rupture length.

Cifuentes & Silver (1989) determined a magnitude of $M_w = 8.1 \times 10^{23}$ N m, for the foreshock with a rupture length of about 150 km, that took place 33 hours earlier in the northernmost extension of the rupture region (37.03–38.74°S; Plafker & Savage 1970) of the 1960 $M_w = 9.5$ giant earthquake. Cifuentes (1989) suggested that the seismicity shows a southward migrating pattern during these 33 hours. Krawczyk et al. (2003) suggested that the hypocentre should be located about 80 km to the west from that reported by Cifuentes (1989), at the interface between the Nazca and South American plates. The total rupture length of the sequence is of the order of 1000 km.

Plafker & Savage (1970) reported coastal elevation changes produced by the 1960 event from Peninsula de Arauco to Peninsula de Taitao, a 1000-km-long segment along the coast of southern Chile. These measurements, carried out in 1968 and based on intertidal algal environment modification and evidence from dead barnacles, balanus and trees, revealed extreme values of 6 m of uplift in Isla Guamblin and 2 m of subsidence in the city of Valdivia. Repeated measurements of triangulation as well as levelling were used in their estimation of the earthquake size: a nearly 1000-km-long dislocation with 20 to 40 m of fault displacement. Later, Plafker (1972) reanalysed the static deformation and deduced a causative fault 120 km wide by 1000 km long, dipping 20°E with 20 m of slip. Assuming a rigidity modulus of $5 \times 10^{10}$ Pa, the total seismic moment reaches $1.2 \times 10^{23}$ N m or $M_w = 9.3$. Kanamori & Cipar (1974) estimate a moment magnitude of 9.5.

Barrientos & Ward (1990) used the same information collected by Plafker & Savage (1970) to infer a variable slip model of rupture to identify patches of nearly 40 m of displacement in the northern half of the rupture and only 15 m in the southern portion of the rupture (Fig. 10.19) producing a total moment of $1.8 \times 10^{23}$ N m ($M_w = 9.4$). They proposed that these types of event recur nearly every 400 years, a hypothesis later strengthened by Cisternas et al. (2005) from the records of tsunami sands deposited in Maullín, to the west, on the open coast at the latitude of Puerto Montt. Eight occurrences of this type of deposit allowed them to estimate a mean recurrence of the order of 280 years. Curiously, the major event that took place in November 1837 (see Fig. 10.6), which caused a large tsunami across the Pacific Ocean (6 m runup at Hilo, Hawaii; 2 m in Manceba Island, near Valdivia), did not produce a sand deposit in Maullín, so the rupture must have been restricted from Chiloé Island to the south. This is consistent with the lower values of displacement in the southern portion of the 1960 event (Fig. 10.19).

Post-seismic displacements associated with this large event have been reported by Barrientos et al. (1992), with observations made in 1991–1992 at several sites previously reported by Plafker & Savage (1970). These observations, together with

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Fig. 10.15. (a) Daily differences of water level at two limnigraphs located at the extremes of Rapel Lake (dots) and its 30-day running average (solid line). (b) Daily values of mean sea-level as a function of time recorded at the Valparaiso tide gauge (dots). The solid line represents a 30-day average. The time of the 3 March 1985 earthquake is denoted by a vertical dashed line in both figures (Barrientos 1997). For references of the locations of Valparaiso and Rapel, see Figure 10.13.
measurements at the Puerto Montt tide gauge, indicated continuous uplift of the region. Nelson & Manley (1992) reported on the uplift of Mocha Island before, during and after the 1960 event, with nearly the same amount of elevation change, 1.5 m coseismic and another 1.5 m of post-seismic uplift. Khazaradze et al. (2002) interpreted the post-seismic horizontal deformation observed between 1994 and 1996 (Klotz et al. 2001) as viscous relaxation of the mantle.

Taitao Peninsula–Puerto Natales (46–52°S)

This is the region in Chile with the lowest rate of seismicity. No field seismic studies have been carried out in the zone except for that of Murdie et al. (1993) centred around the Taitao Peninsula. Nine three-component digital seismic stations installed for a period of two months recorded about three local events per day (Fig. 10.20). Nearly 50 well-recorded events allowed Murdie et al. (1993) to establish two main observations: (a) a lineament of epicentres interpreted as an active segment of a subducted transform fault, and (b) extensional mechanisms for earthquakes located along the predicted subduction ridge. One of their main conclusions is that extension in the slab influences the overlying crust, giving rise to dextral transtension, and that the pattern of extension is reflected in the development of a basin south of the northward relative subducting ridge.

Fig. 10.16. P-wave first motion focal mechanism and waveforms for the 1939 earthquake (Beck et al. 1998).

Fig. 10.17. Observed P-waveforms (solid lines) at stations in Ottawa (Canada) and Berkeley (US) and the modelled signals (dashed line) using the focal mechanism shown in the previous figure and source time function on the left. The earthquake depth is 100 km (Beck et al. 1998).
The Magallanes region is generally perceived as fairly aseismic when compared to the rest of the country. Although there might be some truth in this perception, because few shocks are felt during any given year, the seismic hazard is still fairly high. Historical reports indicate that major earthquakes have been felt in Punta Arenas (in 1879 and 1949), probably occurring along the Magellan Fault (Lomnitz 1970; Kleipis 1994). Because Magallanes was a generally uninhabited place in those days, very little is known about the 1879 event, except that it had a significant magnitude, probably greater than 7 (Lomnitz 1970).

The SISRA (catalogue of earthquakes for South America, Centro Regional Para América del Sur, 1985) catalogue lists a magnitude of 7½ for the 1879 event. Two large events that took place on 17 December 1949, the first occurring at 06:53 hours and the second at 15:07 hours, have been listed with magnitudes of 7¾ in the SISRA (catalogue). Recent estimations from waveform modelling of teleseismic records of the 1949 events also show that the magnitude estimates are in the range of > 7 (P. Alvarado, pers. comm.). Figure 10.21 shows the different epicentre locations for the 1949 events. The seismograms (vertical component) at Huancayo (Peru) show that the second event produced somewhat larger surface wave amplitudes. Also, the second event produced higher intensities in Punta Arenas. Smalley et al. (2003) report a cut on a road at the eastern end of Lago Fagnano, liquefaction in the floodplain of Rio Grande and 150 km SW of Punta Arenas (Lomnitz 1970) as well as up to 5 m of horizontal offsets of fences in a farm located at the intersection of the Magallanes–Lago Fagnano Fault. Further examination of GPS data in this region is presented in the next section.
Adaros et al. (1999) reports on a high concentration of earthquakes with focal depths in the range 30–50 km on the western coast of Chile between 52°S and 54°S. They emphasized that few earthquakes took place in Tierra del Fuego itself, but without associating them with any morphological surface structure. This study was based on a five-seismometer deployment for two years in 1997 and 1998.

Summary of the state of stress in the convergent margin (Nazca–South America)

One approach to establish the state of stress in the subduction region is through the observations of deformations, which can be derived from observations using conventional geodetic methods, GPS observations or a combination of both. These
observations are clearly restricted to locations where instruments can be installed, therefore these are land-based methods. A detailed application of the GPS observations in Chile is presented in the next section.

Another approach to examine the state of stress is through indirect observations, such as the type of faulting associated with earthquakes. When an earthquake takes place in a particular region, the waves it generates contain information about the magnitude, energy release, surrounding velocity structure, depth and other information, such as type and orientation of the activated fault. Thus, the information on type of faulting and its conjugate, or earthquake moment tensor, can be extracted from the waves recorded at several different stations worldwide.

The seismology group in Harvard University (http://www.harvard.edu/projects/CMT) has carried out routine computations of centroid-moment inversion for earthquakes with magnitude larger than 5.5 worldwide since the mid-1970s. Figures 10.22 to 10.25 show the centroid-moment tensor solutions for the earthquakes in the region of interest (from north to south) during a 26-year period (1980–2005, inclusive). In these four figures, larger symbols represent earthquakes with moment magnitudes (Mw) larger than 6.0 and the dashed line represents the axis of the Peru-Chile trench. Figures 10.22 to 10.24 show moment tensors compatible with low-angle thrust earthquakes distributed along the coast. Further inland, these turn mainly into mechanisms of tensional type, perhaps at depths of the order of about 70–80 km.

Fig. 10.22. Locations of earthquakes and focal mechanisms listed in the Harvard Moment Tensor site for the period 1980–2005 (inclusive), 17–25°S. The larger circles represent earthquakes with moment magnitudes greater that 6.0. The dashed line corresponds to the location of the Peru-Chile trench. In general, focal mechanisms consistent with low-angle thrust are observed offshore and along the coast; in contrast, focal mechanisms consistent with tensional faulting are observed inland.
Figure 10.23 shows thrust-type activity in the region bounded by the coast and the trench. A concentration of activity is evident around 23°S and 24°S, mostly aftershocks of the July 1995 earthquake. The $M > 5.5$ seismicity decreases significantly to the east when compared to the northern region, but the pattern is maintained, mostly tensional faulting.

Major exceptions to the low-angle reverse faulting close to the coast are earthquakes labelled A–D in Figure 10.24. In this figure, B corresponds to the 1997 event described for La Serena–La Ligua above. Events labelled C and D correspond to normal faults activated as a result of the extension of the upper part of the Nazca Plate due to its initial downward bending. Fromm et al. (2006) studied in detail the main event ($M_w = 6.7$) of 9 April 2001 and its 142 localized aftershocks. The seismic clustering in conjugated planes correlates with ridge-parallel fractures observed by bathymetric surveys. Additionally, there are four shallow events (hypocentral depths < 20 km) taking place beneath the Andes south of 34°S, labelled S (Fig. 10.20).
Two of these events show a strike-slip mechanism (right-lateral slip, assuming a north-south fault plane). The northernmost event evidences a NE–SW compression, similar to the strike-slip events, resulting in a roughly 45° angle fault plane dipping to the NE or the SW. The mechanism of the fourth event represents a vertical fault.

Figure 10.25 shows an anomalous pattern when compared to the previous three figures. All moment tensors located offshore south of 39°S indicate an extensional regime. This means that the Nazca Plate is still being pulled, most likely due to post-seismic adjustment after the giant 1960 earthquake. The rectangle and star at 38°S represent the 1960 epicentral locations proposed by Cifuentes (1989) and Krawczyk et al. (2003), where rupture began its propagation southward, eventually reaching the Taitao Peninsula, close to 46°S. Two other interesting features can be observed in this figure: (a) the mechanism of the event south of the city of Chillán, and (b) the shallow earthquakes with strike-slip mechanisms labelled S. As mentioned earlier, the 1939 Chillán earthquake was particularly catastrophic. The fault mechanism proposed by Beck et al. (1998) for the 1939 event is consistent with the type of faulting derived for the earthquake located just south of Chillán, which took place in September 1986 at around 80 km depth. Two shallow ($h < 20$ km) events of interest are those labelled S: if the north–south plane is the rupture plane, it means that these events represent right-lateral displacement which would be consistent with the expected displacement along the Liquiñe–Ofqui Fault (described in other chapters and in Cembrano et al. 1996).

Comte & Suárez (1994) used complementary information of the state of stress in the convergent zones at depth of about 100 km, from the analysis of data recorded in two microseismic experiments in northern Chile, at the latitudes of the cities of Iquique and Antofagasta (see Fig. 10.5) and at depths of about 100 km. They found evidence of a double seismic zone in which down-dip tensional events were consistently shallower than a family of compressional earthquakes. They interpreted the extensional regime as a result of basalt to eclogite transformation in the subducted oceanic crust, which also produces compression in the underlying mantle. Local studies such as this (Comte & Suárez 1994) can only be carried out with local, denser seismic networks.

Global Positioning System efforts

The rapid convergence rate between the Nazca and South American plates makes this region an excellent laboratory to test ideas about the seismic cycle. In fact, the 1995 Antofagasta earthquake took place in an area where previous GPS observation had been made. Both Ruegg et al. (1996) and Klotz et al. (1999) reported a maximum horizontal displacement of the order of 80 cm to the west at the coast near Antofagasta. Current estimations of the displacement field at the coast are of the order of 20–30 mm/year in the opposite direction, revealing the deformation to which the region is subjected during the interseismic period of stress accumulation.

Three large continuing efforts can be identified (Fig. 10.26): (a) South America Geodynamics Activities (SAGA) (Klotz et al. 2001; Khazaradze & Klotz 2002); (b) Central Andes GPS Project (CAP) (Bevis et al. 1999, 2001, 2004; Kendrick et al. 1999; Smalley et al. 2003); and (c) local efforts (Ruegg et al. 1996, 2002; Chlieh et al. 2004). All these works present results of GPS observations in Chile, with SAGA and CAP at a global level and the more localized efforts of Ruegg in northern Chile and around the area of Constitución–Concepción. Results of repeated observations in different epochs have been shown already in Figures 10.7 and 10.18. Figures 10.22 and 10.23 show the observation points and results of the SAGA effort.

The velocity field of the coastal region between 26°S and 37°S is characterized by convergence roughly parallel to the relative convergence direction of the Nazca and South America plates at a rate of 3.5 cm/year (Klotz et al. 2001) (Fig. 10.27). This rate decreases with distance from the coast. The authors interpret these observations as velocities produced by the interseismic strain accumulation due to 100% locking of the interface between the Nazca and South America plates. To the north of 26°S and to the south of 37°S the regions are dominated by post-seismic effects of the 1995 and 1960 events.

Bevis et al. (2001) invoked both elastic loading (as was done in the previous case, by locking of the boundary between the Nazca and South America plates) and backarc convergence

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**Fig. 10.24.** Locations of earthquakes and focal mechanisms listed in the Harvard Moment Tensor site for the period 1980–2005 (inclusive), 29–39°S. Focal mechanisms of earthquakes labelled A and B (the latter corresponds to the October 1997 event) represent faulting on a vertical plane within the subducting Nazca Plate with the trench-ward block down-dropping with respect to the eastern block. Events C and D represent tensional faulting within the Nazca Plate. Four events taking place in the Andes south of latitude 34°S marked with S correspond to shallow events (hypocentral depth < 20 km).
(5–6 mm/year, or 8.5% of the total Nazca–South America convergence rate) to explain the deformation (velocity field) of the southern Peru–northern Chile region. They concluded that in this region there is no evidence of slip partition.

Additionally, Figure 10.21 shows the estimated displacement in the area close to the Magallanes–Lago Fagnano Fault zone (Pelayo & Wiens 1989) and the measurements after repeated GPS observations in 1998, 1999 and 2000 (Smalley et al. 2003). Repeated GPS observations indicate that the deformation zone is not restricted to a linear feature but, as expected in strike-slip regimes, the stress accumulation is dependent on the locking depth of the brittle region. Smalley et al. (2003) modelled this velocity field as a vertical fault with 15 km locking depth and a velocity of 6.6 mm/year. From reports of 5 m offsets during the 1949 event, Smalley et al. (2003) suggest a return period of the order of 750 years for this segment of the fault. In addition, GPS reveals that the whole southern Patagonia region is moving to the east in relation to stable South America at a rate of approximately 1.5 mm/year.

Brooks et al. (2003) proposed an interesting model to explain the deformation field across the Andes between 26ºS and 36ºS with respect to the craton. By selecting this region, they avoided possible contamination of the data with the post-seismic effects of the 1995 earthquake in northern Chile and the post-seismic effects of the 1960 southern Chile event. Measurements carried out between 1993 and 2001 reveal that Andean locations show eastward velocities ranging from approximately 35 mm/year close to the Nazca boundary to basically zero (within the error ellipses) in the craton. No oblique deformation is obvious from these measurements. Their results suggest that the oceanic Nazca Plate is fully locked while the continental backarc boundary creeps continuously at around 4.5 mm/year.

Fig. 10.25. Locations of earthquakes and focal mechanisms listed in the Harvard Moment Tensor site for the period 1980–2005 (inclusive), 37–44ºS. Most of the seismic activity is concentrated offshore the trench south of 38.5ºS. Focal mechanisms reflect an extensional regime up-dip of the region involved in the 1960 rupture. The large white square represents the epicentral location of the main events of the 1960 sequence (Cifuentes 1989), the small square represents the epicentre of its main precursor, one day before, and the large star is the proposed epicentral location of the main event according to Krawczyk et al. (2003).
Seismic hazard assessment in Chile

Seismic hazard is, by definition, one of the two major components of seismic risk assessment, the other being vulnerability. The product of seismic hazard and vulnerability of structures provides the basis on which key planning and evaluation studies can assess the different risk levels for existing and future buildings and works of infrastructure at any given site. The main goal of seismic hazard assessment is to estimate, within a period of interest, the characteristics of the expected ground motion, which can be expressed in terms of intensity, ground displacement, velocity or acceleration. The possible consequences of this movement on the ground itself (e.g. liquefaction) and the soil response are also part of the earthquake hazard estimation.

In principle, a long history of records at any site should provide enough information to estimate future ground motion. Unfortunately, this is not possible everywhere because not all sites have been adequately covered by appropriate instrumentation, and the long history of records is not long enough compared with the return periods of large earthquakes, which are the most dominant in the estimation of hazard. In the case of Chile, these large events, with magnitudes of the order of 8, involve the whole width of the coupling region within the subduction zone and have return periods of about 100 to 150 years (Lomnitz 2004). Furthermore, it has been postulated that magnitude 9+ earthquakes might have recurrence periods of the order of 300 years or more (Cisternas 2005; Barrientos & Ward 1989). Lomnitz (2004) states that even at these long scales, the moment release is not constant. In the case of central Chile he proposes that there has been an acceleration of moment release after the large 1822 earthquake.

First estimations of seismic hazard were carried out by Lomnitz (1969) who, using a probabilistic method, only considered events larger than 7½ because smaller ones do not significantly contribute to the seismic hazard and therefore can be disregarded. According to Lomnitz (1969) these medium to large-magnitude events produced a Modified Mercalli Intensity (MMI) over VI. To enable probabilistic estimations, he assumed that the seismic sources follow a Poisson distribution and the isoseismal VI represents a level of 0.1g in acceleration.

This idea is expressed as $R$, in the equation

$$R = 1 - e^{-\frac{n_i}{T}}$$

in which $R$ is the seismic hazard (Lomnitz used the term ‘risk’), $T$ is the length of the whole earthquake record, $n_i$ is the number of events that produce an MMI of VI or greater during $T$, and $t$ is the design period.

Explicit attenuation relationships were not used in this approach, although they were included in the determination of the isoseismal VI (MMI), with the maximum extent of the hazard depending on the attenuation of the intensity with distance. Lomnitz (1969) published estimated probabilities with an exceedence of 0.1g over a period of 30 years, emphasizing the fact that what is important is the number of times this level (0.1g) is exceeded, not by how much.

More recently Barrientos (1980) and Martin (1991) have developed probabilistic estimations of the seismic hazard for the country as a whole. Based on the method proposed by Algermissen & Perkins (1976), both studies include a characterization of the seismicity based on: (a) determination of the seismogenic sources, (b) earthquake ‘productivity’, defined as the number and relative size of the events in terms of the distribution (log $N = a - b M$), and (c) the expected maximum magnitude that earthquakes can reach within each region. According to Algermissen & Perkins (1976), in a no-memory Poisson model, estimation of the return periods of an event exceeding a given level of acceleration (or ground motion) is the reciprocal of its annual probability of exceedance. In calculations of seismic hazard, as Lomnitz had proposed in 1969, the desire is to estimate the probability of exceeding the level of ground motion within a certain period $T$ (e.g. the lifetime of a...
structure), then the return period is related to that probability according to: \( P(t) = 1 - t/T \).

Barrientos (1980) divided the seismogenetic zone in eight different regions (four coastal and four cordilleran) with different values of \( a \) and \( b \), the coefficients of the frequency–magnitude distribution. In his study, the subdivision of coastal and cordilleran regions was based on those events located shallower or deeper than 50 km, respectively, recognizing the down-dip limit of the rupture zones of large subduction events.

An attenuation relationship on Modified Mercalli Intensity of the type

\[ I(r) = 1.38M_s - 3.74 \log(r) - 0.0006r + 3.91 \]

where \( M_s \) is the event magnitude based on surface waves and \( r \) is the epicentral distance or the distance from the site of interest to the closest point of the fault (Fig. 10.28), allowed Barrientos (1980) to estimate intensities with a 10% probability of being exceeded in 50 years.

Martin (1991) estimated accelerations not to be exceeded within a period of interest (50 and 100 years). For this, he collected all maximum vertical and horizontal accelerations recorded on rock due to earthquakes in Chile and summarized them in an attenuation relationship of acceleration as a function of magnitude and distance of the type:

\[ a = \frac{71.3}{(R + 60)^{0.35}} \]

The expected peak accelerations as a function of distance (\( R \)) for different magnitude events taking place at a depth of 30 km are shown in Figure 10.29. Peak accelerations nearly half that of gravity are expected in the neighbourhood of the surface projection of the fault for a large-magnitude event, while a magnitude 5 tremor would not generate more that 5% of g. Figure 10.30 shows the peak accelerations with a 10% chance of being exceeded during a 50-year term. As expected, higher peak
accelerations are located closer to the coast due to the location and depth of the large thrust events that dominate the seismicity in the region. Two zones are of additional interest in this map, both dominated by shallow (less than 20 km depth) seismicity. One of these is in response to plate shortening (in the Andes, between 30°S and 35°S latitude, and continuing into Argentina; Alvarado et al. 2005), whereas the other is attributed to transcurrent displacement (Magallanes region) which give rise to earthquakes such as those occurring in 1949 (Klepeis 1994).

In these approaches, there has been no attempt to include aspects related to local effects, which depend on the geometric and physical conditions of the site. Localized studies have been attempted in various cities in Chile but no uniform and standardized approach has yet been proposed. Additional elements of seismic hazard which will eventually be included in estimations are, among others, rupture directivity and variations of stress-drop associated with inter- and intraplate events, examples of the effects of which are beginning to emerge.

Discussion

The dominant aspect of seismicity in Chile is controlled by the subduction of the Nazca Plate under the South America

Fig. 10.28. Intensity attenuation as a function of distance for different magnitudes according to Barrientos (1980).

Fig. 10.29. Horizontal acceleration attenuation as a function of distance for different magnitudes according to Martin (1991).

Fig. 10.30. Peak accelerations with a 10% chance of being exceeded during a 50-year term (Martin 1991). Higher peak accelerations are expected closer to the coast due to the location and depth of the large thrust events that dominate the seismicity in the region. Two zones are of additional interest in this map, both dominated by shallow (less than 20 km depth) seismicity: the Cordilleran region between 30°S and 33°S and the southern extreme, in Tierra del Fuego (53–55°S). The next generation of this type of map should include the effects of higher stress-drop events (7–100 km depth) tensional faulting and the recognized southward extension (at least to 35°S) of the shallow cordilleran region.
Plate. This subduction, however, is not regular throughout the convergence margin. The age of the subducting Nazca Plate varies from 90 Ma at 17°S to 0 at the triple-junction near 46°S. Furthermore the region between 26°S and 33°S is characterized by the gradual southward decrease of the dip angle of subduction, the absence of a central valley between the Coastal Cordillera and the Andes, and lack of recent volcanism due to the buoyancy of the subducting Juan Fernández Ridge (Gutscher et al. 2000b).

It is clear that, with the possible exceptions of Mejillones and Arauco peninsulas (and certainly the Taitao Peninsula), there is no particular segmentation of the rest of the coupled region along the coast of Chile. The 1000-km-long 1960 rupture had been previously partially ruptured by the 1737 and 1837 events, with only the 1575 earthquake being of a similar size. Analysis of tsunami deposits indicates that 1960-type events recur in the region about every 300 years. The central Chile region is another example of variability of consecutive ruptures; the largest event in the sequence of historic earthquakes, which begins in the mid-1500s with the Spanish conquest, is the 1730 event which had a source length of more than 500 km. One interesting observation is that all the underthrusting earthquakes studied in any detail have shown that their rupture begins in their northern part, and their ruptures propagate to the south.

Not only do the large magnitude earthquakes differ in size from cycle to cycle but also in the manner in which the convergence between the two plates is accommodated. The July 1995 Antofagasta earthquake showed only a small post-seismic readjustment towards the north of the coseismic rupture, under the Mejillones Peninsula. In contrast, continuous records of tilt and tide gauges at the coast above the rupture region in Vaparaiso revealed that an eight- to ten-month process was dominant in accommodating the deformation. Another is the case of the 1000-km-long rupture produced by the 1960 megathrust event, with post-seismic readjustments still being measured by GPS techniques 45 years after its occurrence.

Another interesting aspect of seismicity in Chile is the increasing evidence for shallow seismicity in the Andes and/or its foothills throughout the whole country. Even though no surface evidence of rupture was found for the September 1958 Las Melosas (Upper Maipo Valley, SE of Santiago) earthquake (M = 6.9), it was clearly documented as a shallow event, Lomnitz (1961) assigned a local intensity of X (MMI). A recent example of a shallow event (M = 6.4) in August 2004 reminds us of the southern extension of these seismic sources and the need for a reconsideration of their importance to seismic hazard evaluations. Another relatively large shallow event (M = 6.3) took place in July 2001 on the western flank of the Andes in northern Chile. Temporary networks have demonstrated the existence of shallow crustal seismicity in varying degrees along the country.

In summary (Fig. 10.31), the potential areas most subject to large-magnitude earthquakes lie above the coupling region between the Nazca and South American plates, which corresponds to the contact region between the trench and approximately 45–53 km depth along the Wadati–Benioff region. Earthquakes in intraplate regions within the subducting Nazca Plate, which can reach magnitudes of the order of 8, are characterized by higher stress drops than events that take place in the contact between these two plates, thus producing higher accelerations at the surface. Large magnitude earthquakes are also generated at the transcurrent system between the South America and the Scotia plates. We have come a long way from the early reports of Maria Graham and Charles Darwin, and the overall controls on Chilean seismicity are clearly much better understood. Whilst future studies will continue to improve our knowledge of the seismic pattern, much work also needs to be done on prediction and damage limitation in this spectacularly earthquake-prone country.

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The chapter examines ancient and modern geological and oceanographic processes relevant to that part of the Pacific Ocean lying offshore from the Chilean mainland. Initial overviews are presented of submarine geomorphology, plate tectonic background, Chilean physiography, oceanic circulation and present-day climate. These are followed by an examination of modern sedimentation, which looks at controls on marine productivity, the roles of plankton, organic matter and biogenic input fluxes, calcareous and siliceous organisms, and the input of terrigenous sediment into the marine environment. These modern processes are then compared with past sedimentation patterns and palaeo-productivity, in a further attempt to better understand the driving forces behind ancient and modern environmental variability.

The team of authors assembled for this chapter reflects a long-term cooperative effort between Chilean and German oceanographers. In 1991, the universities of Concepción, Austral de Chile, Católica de Valparaíso and del Mar instigated a scientific cooperative programme with the Department of Geosciences of the University of Bremen, Germany. Several projects funded by the German Ministry for Science and Education in the fields of biogeochemistry (JGOFS: Joint Global Ocean Flux Study—Chile) and palaeoceanography (CHI PAL: Späthiszäitige Variationen der Paläoproduktivität im östlichen Sudpazifik, SO-102 (Hebbeln et al. 1995); PUCK: Wechselwirkungen zwischen Produktivität und Umweltbedingungen am chilenischen Kontinentalhang, SO-156 (Hebbeln et al. 2001)), as well as Fondecyt-funded projects for studying biogeochemistry and palaeoceanographic processes, were conducted along the Chilean coast (Fondap I and II, Fondecyt no. 10000912 and no. 1040968). Key scientific goals of these projects include a better understanding of the El Niño—Southern Oscillation (ENSO) cycle and other large-scale phenomena, and how the Oxygen Minimum Zone and the Antarctic Intermediate Waters may influence, or be influenced by, global climate change.

Tectonic and geomorphological setting

Plate tectonic controls

As has been detailed elsewhere in this book, the Chilean continental margin is characterized by the subduction of two oceanic plates (Nazca and Antarctic) beneath a continental plate (South American) at the Chile Trench (Fig. 11.1). The oceanic plates are separated by the Chile Ridge, where seafloor spreading between the Nazca and Antarctic plates is presently occurring at a rate of approximately 64 km/Ma. The Chile Ridge strikes nearly parallel to the trench and therefore collides with the Chile Trench in a highly oblique orientation (Fig. 11.1). At approximately 46°S the Chile Ridge is subducted beneath the South American Plate. Here the Nazca and Antarctic plates meet the continental crust of the South American Plate to form the Chile triple-junction. North of the triple-junction, the Nazca Plate is subducted beneath the South American Plate in an ENE direction and at a rate of 8.4 cm/year, while to the south the Antarctic Plate is subducted in an easterly direction at only 2.4 cm/year (Bangs & Cande 1997). Subduction beneath South America is believed to have been continuous since Jurassic times. The subduction of the Chile Ridge began at approximately 14 Ma when it collided with the Chile Trench near Tierra del Fuego. Therefore, the triple-junction has migrated north to its present position. The most apparent effects of the subduction of young crust from the Chile Ridge are a shallow trench and the rapid narrowing of the forearc region in the vicinity of the triple-junction (Cande et al. 1987).

Accretionary and non-accretionary or erosional episodes are probably linked to temporal variations in trench sediment thickness. Bangs & Cande (1997) described thick trench sediments that correlate with accretion along the southern Chile margin, and thin trench sediments that correspond with non-accretion or tectonic erosion near the Chile Ridge and northwards from the Juan Fernández Ridge.

Earthquake locations and the determination of focal mechanisms have allowed analysis of the geometry of the Wadati-Benioff zone beneath western South America (Cahill & Isaacs 1992; Frutos 1981; Jordan et al. 1983). In northern Chile (20°S to 27.5°S), the Wadati–Benioff zone is characterized by a steep angle and dips 30°E. Further south, a small horizontal ‘bench’ begins to form in the slab at between 100 and 125 km depth, and between 29°S and 32°S the slab dips nearly horizontally in this depth range. Above this flat-slab zone, the descending oceanic plate initially has a concave-up geometry (at 50–100 km depth), followed by the long horizontal segment extending 300 km to the east, before it finally descends to below 125 km further eastwards. This segment of the slab coincides spatially with the intersection of the Juan Fernández Ridge. South of 33°S the Wadati–Benioff zone returns to its original dip of around 25–30°E and south of 39°S restricted available seismic information indicates that the slab is slightly flatter with a dip of 15–20°E.

Subaerial physiography

On land, the underlying subduction geometry described in the previous section influences topography and volcanism in the
Fig. 11.1. Tectonics of Chilean continental margin and segmentation of the Andean margin: Central Andes (15–33.5°S) and southern Andes (33.5–47°S); Altiplano (15–23°S), Puna (23–28°S) and Frontal Cordillera (28–33.5°S) within the central Andes; Principal Cordillera (33.5–39°S) and Patagonian Cordillera (39–47°S) in the southern Andes. Contour lines represent the depth of the Nazca Plate. Triangles mark the location of active volcanoes. Thick line shows the position of the Chilean trench and adjacent small numbers indicate the thickness of the trench sediment wedge (in kilometres). Large numbers point out the age of the Nazca Plate. Grey bands correspond to the ridges (Nazca, Juan Fernández, Chile) and thin lines represent principal fracture zone (FZ) of the Nazca Plate: Challenger (CHFZ), Mocha (MFZ), Valdivia (VFZ), Chiloé (ChFZ) and Guaya (GFZ). Modified from Tassara & Yañez (2003).
and volcanic rocks, and outcrops of the metamorphic basement. The elevation of this part of the cordillera is high, reaching maximum values of up to 7000 m a.s.l. around 33°S.

South of 33°S, both geology and geomorphology change abruptly (Lowrie & Hey 1981) as the dip of the subduction zone once again increases, and Plio-Quaternary volcanism and alluvial basins reappear (Fig. 11.2). The geology of the Coastal Cordillera south of 33°S is marked by abundant, primarily low-grade metamorphic rocks (Zeil 1986), with Palaeozoic plutons also being common southwards to about 38°S (Ruiz & Corvalan 1968). The Chilean Central Valley is filled with up to 4000-m-thick sequences of alluvial sediments (Zeil 1986), but south of 41°S has been invaded by the sea. Further east, the basement of the Andes consists mainly of Mesozoic plutons south of 41°S, in contrast to the pre-Pliocene andesitic to rhyolitic volcanics and sediments that crop out further north between 33°S and 41°S (Zeil 1986). The Plio-Quaternary volcanics occurring throughout this Andean segment are more basic in chemistry than in northern Chile (Thornburg & Kulm 1987b).

As has been detailed in Chapter 8, the hydrology of Chile displays prominent changes across latitude, reflecting the changing climate. The region north of 27.5°S is characterized by prevailing arid conditions and dry valleys with only sporadic water runoff. Sediments derived from the Andes are mostly trapped in the alluvial basins east of the Coastal Cordillera because transverse valleys in the Coastal Cordillera which reach the Pacific coast are almost absent. Today, therefore, eroded material reaches the ocean mainly by aeolian transport and occasional flash-floods, predominantly from the Coastal Cordillera (Lamy et al. 1998a). From 27.5°S to 33°S, rivers originating in the Andes generally cut through the Coastal Cordillera but river discharges are very low (<1 km³/a) (Milliman et al. 1995; Chapter 8). Further south the density of river systems increases rapidly and fluvial runoff is significantly higher (up to 21 km³/a) (Milliman et al. 1995). Thus, in central and southern Chile seasonal to year-round precipitation allows perennial rivers to transport huge sediment loads from the Andes onto the continental slope, while the component of material derived from the Coastal Cordillera is relatively low (Lamy et al. 1999).

Taken together, the climatic and physiographic latitudinal segmentation of Chile results in substantial north–south variations in the supply of terrigenous material to the continental margin offshore. This broad pattern, as described in more detail later in this chapter, is clearly reflected in the grain size and composition of modern surface sediments and late Quaternary sedimentation rates along the Chilean continental margin.

**Submarine geomorphology**

Offshore, the Chilean continental margin is characterized by a narrow continental shelf of variable width, a steep continental slope, and a deep-sea trench (Chile Trench) that can be subdivided into three morphological segments which are separated by tectonic discontinuities (Thornburg & Kulm 1987a, b). North of 33°S latitude, the continental shelf is very narrow, rarely exceeding a width of more than 5 km. South of 33°S the shelf is wider but variable: 10 km to the north of the Mataquito river (33°S), and usually between 40 and 60 km wide further south, although it reduces to 12 km at the latitude of the Arauco Peninsula (37–38°S), and widens to 100 km in the far south (46°S). The shelf edge lies at a depth of 120–150 m.

The continental slope is generally steep and reaches maximum inclinations of 10° to 15° in its lower part. The ascent from the trench is often broken by structural troughs which form.
sediment traps (Scholl et al. 1970). Both the shelf and the continental slope are intercepted by 11 submarine canyons, situated between La Serena and Chiloé, which discharge directly into the trench. Canyons are sparse north of 33°S but become more frequent further south where large submarine fans are developed on the lower continental slope and in the trench (Thornburg & Kulm 1987a, b).

The best studied canyons off Chile are the San Antonio and Biobío canyons (Fig. 11.3). The Biobío submarine canyon (36°49’S) is 102.4 km long with a low sinuosity (1.3). Its head is located at 300 m from the shoreline and at a depth of 15–20 m, from where it extends down to the trench, reaching a depth of 4570 m and developing an extensive submarine fan. Bathymetric data reveal a canyon with a ‘V’-shaped cross-section in the upper part, and a ‘U’-shaped cross-section along the rest of its path, with a flat bottom due to the accumulation of turbidite deposits. The path of the canyon is structurally controlled by three principal directions: north–south to N10°W, N50–60°E and N50–60°E (Pineda 1983c). The San Antonio submarine canyon (33°30’S) is 170 km long with a sinuosity of 1.25, and extends from near San Antonio harbour across the continental slope, where there are extensive turbidite deposits, down to the trench (Hagen et al. 1996).

The Chile Trench south of 18°S maintains a uniform strike and lies about 100 to 150 km off the coast. The average depth of the trench is 6000 m between the Nazca and Juan Fernández ridges. At 20°S the trench reaches its maximum depth of 8000 m, which decreases gradually towards the south. The volume of sediment in the trench is closely correlated with the annual rainfall on the adjacent continent, and thus increases substantially from northern to southern Chile. The Chile Trench can be subdivided into three geomorphological segments according to the sediment distribution: (1) from 18°S to 27.5°S the trench is only locally filled with sediments which are less than 0.1 km thickness; (2) from 27.5°S to the Juan Fernandez Ridge (33°S) the trench sediment wedge is nearly continuous with a moderate volume of sediment (less than 10 km wide and 0.5 km thick); (3) from 33°S to the Chile Ridge (45°S), the trench fill increases in volume and eventually buries the trench completely (20 km or more wide and 1.5 to 2.3 km thick). South of the Chile Ridge the trench sediment wedge maintains its thickness at around 2.3 km (Thornburg & Kulm 1987a; Bangs & Cande 1997). Between the Chile Ridge and the Juan Fernandez Ridge, the trench is slightly inclined towards the north, causing northward sediment transport through a winding axial channel over 650 km in length that serves as a flow-channel for turbidity currents.

Oceanic circulation

The surface ocean circulation in the East Pacific off Chile is dominated by the northward flowing Peru–Chile (or Humboldt) Current (PCC) and the southward flowing Cape Horn Current (CHC), which both originate at 40–45°S where the Antarctic Circumpolar Current (ACC) approaches the South American continent (Boltovskoy 1976) (Fig. 11.4a). The northward deflection of the ACC is primarily responsible for the initiation of the Peru–Chile Current, which stretches all along the South American west coast before turning westwards close to the equator to form the South Equatorial Current. Those waters of the ACC that become advected southwards by the CHC are finally transported to the Atlantic Ocean via the Drake Passage.

Off the Chilean coast occasionally the PCC can be divided into an oceanic branch (PCCocean) and a coastal branch (PCCcoast) separated by the poleward flowing Peru–Chile Countercurrent (PCCC) (Fig. 11.4b), which is located 100–300 km offshore and transports subtropical surface water to the south. The PCCocean (also termed Chiloé Coastal Current CCC) Fig. 11.4c) extends to c. 100 km off the coast and is characterized by a significant admixture of low salinity surface waters derived from the Chilean fjord region. Close to the coast these surface water masses are underlain by the poleward flowing Gunther Undercurrent (GUC; Equatorial Subsurface Water) (Fig. 11.4b.e), which is mainly located at water depths between 100 m and 400 m over the shelf and the continental slope.

Deeper currents include the Antarctic Intermediate Water (AAIW), which flows towards the equator at water depths between 400 m and 1200 m (Fig. 11.4c), and which sometimes approaches the surface during periods of strong upwelling (Strub et al. 1998), although normally it is the Gunther Undercurrent that supplies the upwelling waters (Morales et al. 1996). Below the Antarctic Intermediate Water, slowly southward flowing Pacific Deep Water prevails (PDW; Fig. 11.4c). The hydrography of surface and subsurface waters in this part of the South Pacific has been examined and described in detail by Avanzini et al. (1989), Ingle et al. (1980), Martens (1981), Shaffer et al. (1993, 1997, 2000) and Strub et al. (1998).

The PCC forms one of the most prominent eastern boundary currents (EBC) in the world’s oceans. In such EBC regions, dominant equatorward alongshore wind stress induces an offshore surface Ekman transport resulting in the upwelling of relatively cold, nutrient-rich subsurface waters nearshore. Due to this continuous and intense coastal upwelling, the Peru–Chile Current is one of the most important high-productivity regions in the world (Berger et al. 1987). The upwelling process

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Fig. 11.3. The distribution of canyons along the Chilean margin (modified from Thornburg & Kulm 1987a).
and its main impact on the marine environment is essentially confined to the waters over the shelf and the upper slope region, but the influence of coastal upwelling can be seen at distances as far as 400 km offshore (Strub et al. 1998; Thomas 1999). The upwelling-induced nutrient injection stimulates biological productivity and consequently phytoplankton pigment concentrations in near-coastal waters can exceed 6 mg/m³ (Thomas et al. 1994) and annual production rates of >200 g C/m² occur (Berger et al. 1987).

The latitudinal pattern of southward-increasing pigment concentrations has some similarity with the pattern of alongshore wind-forcing. While north of c. 33°S winds are upwelling-favourable throughout the year, and become progressively weaker with decreasing latitude, further south wind-forcing shifts seasonally from mean upwelling- to mean downwelling-favourable conditions, while south of 39°S prevailing onshore-blowing westerlies prevent upwelling (Strub et al. 1998). Based on this observation it is interesting to note that, also at c. 45°S,
an area south of the main upwelling region and already in theealm of the Southern Westerlies (i.e. persistent downwelling-
favourable winds), pigment concentrations are rather high.
This is probably due to the fact that in this region the high-
nutrient, low-chlorophyll waters of the ACC mix with iron-rich
coastal waters.

This general setting of the PCC is disturbed at irregular inter-
vals of a few years by El Niño-Southern Oscillation (ENSO)
events. The Chilean coast lies within the southern reaches of the
oceanographic and atmospheric perturbations caused by such
ENSO events. For instance, off northern Chile sea surface
temperatures were increased by 3°C during the strong 1982/83
El Niño event, the thermocline was 300 m below the normal
level (Fonseca 1985), and the phytoplankton biomass was
considerably decreased (Aviria & Muñoz 1987).

Present climate

The climate of Chile is controlled by its unique geographic
location on the western flank of the Andes and bordering the
Pacific coast of South America, with the country occupying a
4200 km long but very narrow strip mostly less than 400 km
wide. However, this extreme latitudinal extent (17°S to 56°S)
is not reflected in contrasts in mean annual air temperature,
which is remarkably uniform throughout the country, ranging
from about 19°C at sea level in the extreme north to 5°C in
southernmost Chile (Miller 1976). This relatively small
temperature gradient is principally caused by the compensating
effect of the Perú–Chile Current, bringing cool Subantarctic
water masses up to northern Chile, and by the effective
isolation of the country from climatic influences of the South
American interior by the Andes.

In contrast to these relatively uniform temperatures, precipi-
tation patterns in Chile show what is probably the most pro-
nounced latitudinal gradients on Earth, ranging from hyperarid
conditions in the northern third of the country (Atacama Desert)
to extremely high rainfall in the mountains of southern
Chile, which are amongst the wettest extratropical regions in
the world (Miller 1976). Except for a small highland area in the
extreme NE, which receives tropical summer rain, precipitation
in Chile is entirely related to the rain-bearing Southern Wester-
lies which therefore principally define the climatic zonation
of the country (Fig. 11.2). Due to the orographic rise of moist
Pacific air masses, rainfall in the Andes is significantly higher
than in low elevation areas of the same latitude, leading to a
northward displacement of the Chilean climatic zones in the
mountains.

The track and intensity of cyclonic storms of the Southern
Westerlies are controlled by the strength and latitudinal posi-
tion of the subtropical anticyclone in the SE Pacific and the
circum-Antarctic low pressure belt (Cerveny 1998). The sea-
sonal shift of the Southern Westerlies is between 5° and 10°
of latitude. In the southern winter the storm tracks are centred
around 40°S and the related rainfall reaches a mean northern
limit of c. 31°S (Fig. 11.2). Further north, up to c. 27°S, only
occasional winter rain events occur, and are related to an
extreme northward penetration of atmospheric perturbations
from the westeringies (Miller 1976). As the major influence of the
SE Pacific anticyclone is basically restricted to lower levels of
the atmosphere, the passage of atmospheric troughs originating
in the westerlies once or twice a year can bring precipitation to
high altitudes even as far north as the Bolivian Altiplano (Miller
1976). These disturbances are often linked to a cut-off and
northward displacement of cold polar air masses and the
collision with warm, humid tropical air masses from the NE
(Rutland & Fuenzalida 1991; Messerli et al. 1998). Annual
rainfall increases significantly south of 31°S, at low elevations,
from 200 mm/a to 2000 mm/a at 41°S and even higher
values in the Andes. Central Chile between 31°S and 37°S is
characterized by a typical Mediterranean-type climate with only
up to 5% of rainfall during the southern summer. Between 37°S
and 42°S summer rainfall increases sharply towards the average
northern summer limit of the Southern Westerlies (Fig. 11.2).
South of 42°S Chile is traversed regularly by cyclones of the
westerlies throughout the year, leading to frequent rain and
extraordinary annual precipitation values of up to 7500 mm/a
(Miller 1976). The amount of rainfall varies locally and depends
largely on exposure to the moist Pacific air masses.

Interannual rainfall variability, especially in central Chile, is
strongly related to ENSO (El Niño years) (Pittock 1980; Karoly
1989; Rutland & Fuenzalida 1991). During the warm phase of
ENSO, a weakening of the SE Pacific anticyclone results in a
northward shift of the Southern Westerlies leading to
wet anomalies in central Chile. Conversely, a strengthening
and southward expansion of the anticyclone results in more
poleward located westerly storm tracks reducing winter rain in
the Mediterranean climate zone of Chile.

An additional important regional climatic phenomenon is
the presence of persistent coastal fogs and marine stratus clouds
along the coast of Chile, especially north of c. 30°S (Fig. 11.2).
This coastal cloud cover is caused by subsiding warm air along
the eastern periphery of the SE Pacific anticyclone above the
cold Humboldt Current system, resulting in a prominent
temperature inversion in the lower atmosphere. Though only
supplying negligible amounts of measurable rainfall, the coastal
fogs, known as Camanchacas, greatly reduce insolation and
evaporation and allow a comparatively dense vegetation cover
(Miller 1976; Aravena et al. 1989).

Modern sedimentation

Marine productivity

The biological pump, a mechanism that links surface primary
production with deeper waters, transferring particulate organic
carbon (POC) to the deep ocean, has a first-order impact on the
CO₂ concentration of the atmosphere, and therefore on Earth
climate. This POC pump is driven mainly by the vertical flux of
faecal pellets and aggregates (reviewed by Turner 2002).
However, zooplankton contribute to the detritus pool not only
with faeces, but also with pseudo-faeces, exuviae and endadiers.
Larger animals tend to carry out more extensive vertical migra-
tion and are likely to deposit particles, gases and dissolved
compounds as excretion products at greater depth than smaller
animals. Thus, the species composition of the zooplankton,
in terms of size and behaviour (including vertical distribution
patterns), is important in determining the magnitude of the ver-
tical organic flux (Longhurst & Harrison 1989). In the waters
off Chile, faecal material contributes a significant proportion of
the total POC. A study of samples collected at 2300 m depth
in the oceanic region off Coquimbo showed faecal material
averaging 46.6% of total POC for the period July 1993–1998.

Vertical fluxes of bioelements are tightly coupled with the
dominance of different phytoplankton functional groups on
both ecological and geological time scales. In the PCC off
Chile, seasonal (central PCC) and permanent (northern PCC)
upwelling activity brings nutrients to the surface and partially
explains the high productivity of these coastal areas. This sce-
nario of high nutrient concentrations and turbulence seems to
favour the development of diatoms that predominate in the
water column and in the surface sediments between 34°S and
38°S, and between 41°S and 43°S (Romero et al. 2001). By
contrast, at 30°S, high abundances of coccolithophorids and
coccoliths found in a sediment trap located at 2300 m depth
suggest that this functional group is important at this latitude of
the PCC (González et al. 2004). Whether silica-precipitating
atoms dominate over carbonate-precipitating coccolithophorids, or vice versa, seems to depend on the switch from highly turbulent, nutrient-enriched regimes (diatom-dominated) to stable, nutrient-depleted regimes (coccolithophorid dominated). These two functional groups, one based on silicate and the other on carbonate, greatly influence biogeochemical cycles in the modern ocean (Iglesias-Rodriguez et al. 2002). Tozzi et al. (2004) postulated that nutrient pulses, with different frequencies and intensity, regulate the abundance and dominance of coccolithophorids and diatoms. Thus, in highly turbulent, nutrient-rich coastal upwelling ecosystems of the central-southern PCC, the dominance of diatoms in surface sediments (Romero et al. 2001) seems to support Tozzi et al.'s (2004) hypothesis. In contrast, nutrient pulses at low frequencies appear to favour coccolithophorids, regardless of the concentration of nutrients. At 30°S, the high number of coccolithophorids collected in deep sediment traps (González et al. 2004) might be associated with less upwelling-favourable winds (Rutlant et al. 2004).

Among the most widely used sedimentary indicators for the productivity of an overlying water column are organic carbon and biogenic opal. Although organic carbon in marine sediments can also have a terrigenous origin, in the highly productive Chilean upwelling region it appears that this biogenic constituent is almost exclusively derived from marine sources (Hebbeln et al. 2000a).

The contents of organic carbon and biogenic opal in the sediments from the Chilean slope display a consistent pattern. Both have their lowest contents in the sediments from around 27°S, from where they increase slightly towards 33°S. Distinct maxima have been found between 35°S and 37°S, with slightly lower values farther to the south between 41°S and 43°S (Fig. 11.5). A comparison of the organic carbon and biogenic opal contents, as sedimentary proxies for productivity, with the satellite-derived average pigment concentration in the surface waters (Thomas et al. 1994), used as a surface water proxy for productivity, reveals some similar features (Fig. 11.5). However, highest organic carbon and biogenic opal contents have been found at 36°S, while highest pigment concentrations occur farther south at 42°S. Combining information on Holocene or interglacial sediment accumulation rates (Lamy et al. 1998a, 2001; Marchant et al. 1999; D. Hebbeln, unpubl. data) with average organic carbon and biogenic opal contents for these areas, allows a rough estimate for accumulation rates along the Chilean continental slope (Hebbeln et al. 2000a). It appears from such calculations that the estimated accumulation rates of organic carbon, biogenic opal and also biogenic barium (see Klump et al. 2000) closely resemble the pigment concentration (Fig. 11.5), indicating the suitability of these proxies as (paleo-) productivity indicators in this region.

Another useful proxy is the stable nitrogen isotope (δ¹⁵N) value of the organic matter. Between 27°S and 43°S the δ¹⁵N values show a broad range from 9.1‰ to 15.2‰, with a significant decrease of δ¹⁵N towards the south (Hebbeln et al. 2000a). Such a pattern points to a much higher nutrient availability in the south compared to the north. Thus, this data set is also in accordance with the observed higher pigment concentrations and the higher productivity in the southern part of the study area.

These observations are puzzling. Due to the oceanographic and atmospheric setting, coastal upwelling is a common feature north of 38°S. However, south of this border, prevailing onshore winds prevent upwelling. Nevertheless, satellite-derived pigment concentrations as well as all the productivity proxies described above, indicate highest productivities in the study area south of 38°S, i.e. south of the upwelling area. As discussed by Hebbeln et al. (2000a), the δ¹⁵N data might be a key to understanding the nutrient dynamics in the waters off Chile. δ¹⁵N data tend to increase with increasing distance from the nutrient source as available nutrients are gradually consumed (Farrell et al. 1995). The distinct south-to-north δ¹⁵N gradient along the Chilean continental slope, in addition to the parallel latitudinal trend in total productivity, might indicate that the ultimate source of all the nutrients consumed in this area lies south of 38°S.

In the south there are two possible nutrient sources: the Antarctic Circumpolar Current and the Chilean hinterland. The Antarctic Circumpolar Current is a typical high-nutrient, low-chlorophyll region, in which the consumption of the nutrients (i.e. nitrate) is hampered by the lack of micronutrients (i.e. iron). When the nutrient-rich waters of the Antarctic Circumpolar Current reach the Chilean coast, the availability of iron increases dramatically, depending on the input of iron by fluvial and wind transport from the hinterland and by benthic exchange processes. Thus, the high productivity found south of 38°S along the Chilean coast most likely results from the interaction of the high-nutrient, low-chlorophyll waters of the Antarctic Circumpolar Current with high iron availability in the coastal waters of Chile. From there this mix of macro- and micronutrients is transported northward by the Peru–Chile Current, where these nutrients are recycled several times within the coastal upwelling system and gradually consumed on their way north, as indicated by the δ¹⁵N data.

Fig. 11.5. Percentages and accumulation rates of organic carbon, biogenic opal, and carbonate in surface sediments along the Chilean continental slope (Hebbeln et al. 2000a; Romero & Hebbeln 2003) in relation to surface ocean pigment concentration derived from satellite data (Thomas et al. 1994).
Calcareous organisms

Field studies on calcareous organisms off Chile are limited to time-series sediment traps located at 2300 m depth and 100 nm (nautical miles) off Coquimbo (30°S, Marchant et al. 1998, 2004; González et al. 2004). Studies of planktic foraminiferal flux patterns at this site between 1991 and 1998 reveal slightly higher fluxes during austral late winter–early summer than in autumn–early winter, contributing on average c. 40% to the total carbonate flux. Besides planktic foraminifera, coccolithophorids are the most important contributors to the total carbonate flux with highest fluxes in August and September. A distinct sequence in the fluxes of different organism groups in response to the onset of upwelling conditions, starting with a bloom of diatoms (Romero et al. 2001), followed by coccolithophorids, and finally planktic foraminifera, displays the typical pattern of secondary producers trailing the primary producers (Marchant et al. 2004). The planktic foraminiferal assemblage is dominated by five species accounting for ca. 90% of the total fauna (Table 11.1).

Planktic foraminifera and coccolithophorids also dominate the total carbonate preserved in surface sediments off Chile between 23°S and 44°S (Hebbeln et al. 2000a; Mohtadi et al. 2005). Pteropods seem to play a minor role in calcium carbonate vertical fluxes and surface sediment accumulation despite their high species abundance in Chilean waters (39 species; M. Marchant, unpubl.). The planktic foraminiferal abundances in surface sediments vary from 1 to 1200 shells per cm³ (Fig. 11.6A; Mohtadi et al. 2005). This range is clearly linked to the increasing fluvial input of terrigenous sediments towards the south causing an enhanced southward dilution of the biogenic signal (Scholl et al. 1970; Milliman et al. 1995). Generally, low abundances of planktic foraminifera (< 78 individuals/cm³) occur south of 33°S. Highest values are found between 30°S and 33°S (Fig. 11.6A). The planktic foraminiferal assemblage includes a maximum of 27 species, with six species contributing approximately 95% of the total fauna (Table 11.1).

The planktic foraminiferal assemblage in surface sediments is in good accordance with the sediment trap assemblage at 30°S (Table 11.1), suggesting that this proxy is a reliable tool to monitor surface water conditions off Chile. The distribution pattern of planktic foraminifera along the PCC shows two different spatial maxima (Fig. 11.6B): (1) in the upwelling region north of 24°S and between 30°S and 33°S, where high contributions of Globigerina bulloides, Neogloboquadrina pachyderma sin. and Globoquadrina glutinata characterize a typical upwelling fauna in these areas; (2) south of 39°S in a predominantly high-productivity environment under non-upwelling conditions, with enhanced relative contributions of N. pachyderma dext. and N. dumerilii and lower contributions of G. bulloides, G. glutinata and N. pachyderma sin.

Relative abundances of N. pachyderma dext. are highest in the offshore samples, where sea surface temperatures (SSTs) are higher and nutrient contents lower (Fig. 11.6C), and in surface samples south of 39°S, where SSTs are lowest and nutrient contents highest (Fig. 11.6B). Hence, the distribution pattern of N. pachyderma dext. seems to be controlled more by stratification of the water column than by temperature or nutrient availability. Likewise, G. bulloides and N. pachyderma sin. are the most important contributors to the total carbonate flux. Besides planktic foraminifera, coccolithophorids are the most important contributors to the total carbonate flux with highest fluxes in August and September. A distinct sequence in the fluxes of different organism groups in response to the onset of upwelling conditions, starting with a bloom of diatoms (Romero et al. 2001), followed by coccolithophorids, and finally planktic foraminifera, displays the typical pattern of secondary producers trailing the primary producers (Marchant et al. 2004). The planktic foraminiferal assemblage is dominated by five species accounting for ca. 90% of the total fauna (Table 11.1).

Table 11.1. Planktic foraminiferal assemblage from the sediment trap at 30°S and from surface sediments between 24°S and 44°S

<table>
<thead>
<tr>
<th>Species</th>
<th>Sediment trap (30°S)* (average %)</th>
<th>Surface sediments (24°-44°S)* (average %)</th>
</tr>
</thead>
<tbody>
<tr>
<td>N. pachyderma dext.</td>
<td>60</td>
<td>57</td>
</tr>
<tr>
<td>G. bulloides</td>
<td>5-20</td>
<td>15</td>
</tr>
<tr>
<td>N. dumerilii</td>
<td>2-8</td>
<td>10</td>
</tr>
<tr>
<td>N. pachyderma sin.</td>
<td>3-5</td>
<td>8</td>
</tr>
<tr>
<td>Globigerina glutinata</td>
<td>&lt;1</td>
<td>3</td>
</tr>
<tr>
<td>Globorotalia inflata</td>
<td>&lt;1</td>
<td>2</td>
</tr>
<tr>
<td>Globigerinella calida</td>
<td>4-9</td>
<td>&lt;1</td>
</tr>
<tr>
<td>Other species</td>
<td>10</td>
<td>5</td>
</tr>
</tbody>
</table>

* Data from Marchant et al. (2004)
† Data from Mohtadi et al. (2005)

Siliceous phytoplankton

The occurrence of preserved siliceous phytoplankton in modern, late-Holocene sediments along the Chilean coast reflects present-day productivity conditions of surface waters in the SE Pacific. Diatoms overwhelmingly dominate the siliceous microfossil community preserved in surface sediments between 27°S and 43°S off Chile (Fig. 11.7; Romero et al. 2001; Romero & Hebbeln 2003). The pattern of diatom concentration closely relates to the amount of opal (Fig. 11.7) and coincides with the pattern offered by remote sensing estimations of phytoplankton pigment concentrations in surface waters of the PCC (Thomas 1999). Between 18°S and 33°S, where lower pigment concentrations are measured, only secondary peaks of diatom and opal values are observed (Romero et al. 2001; Romero & Hebbeln 2003). In contrast, higher diatom concentration and opal content in surface sediments at 34°-38°S and 41°-43°S correspond well with estimated higher pigment concentrations (Thomas 1999). Off northern Chile, the diatom concentration is generally one order of magnitude lower than further north off Peru (Schwab & Schöniger 1981). This divergence also reflects differences in surface water productivity along the PCC: almost throughout the year higher pigment values and a further seaward extension of chlorophyll filaments occur north of 18°S than south of it (Thomas 1999).

The composition of the diatom thamnocytoenosis strongly reflects late Quaternary conditions and clearly distinguishes the occurrence of coastal upwelling off northern and central Chile from the influence of water masses originating off southern Chile (Fig. 11.8) (Romero et al. 2001; Romero & Hebbeln 2003). Though favourable conditions for intense phytoplankton proliferation characterize surface waters along the whole Chilean coast (Strub et al. 1998), Coastal Zone Color Scanner derived pigment concentrations reveal strong north–south variations (Thomas 1999). Nutrient enrichment of surface waters off the Chilean coast due to upwelling results in intense production of diatoms, reflected by the predominance of Chaetoceros spores as far as around 38°S. Though the occurrence of upwelling is limited to a narrow coastal band (less than 40 km wide) (Morales et al. 2001), the upwelling diatom signal is recorded in deep-sea sediments beyond the continental slope (Romero & Hebbeln 2003). Offshore streaming chlorophyll filaments trace the seaward transport of the coastal upwelling diatom signal (Romero et al. 2001; Mohtadi et al. 2004). The predominance of the warm-water silicoflagellate Ditychaspis
**Fig. 11.6.** Distribution of planktic foraminifera (> 150 μm) in surface sediments along the Chilean continental slope between 24°S and 44°S.

(A) Number of specimens per cubic centimetre sediment; (B) Relative abundances of dominant species of planktic foraminifera. Note the different scales for each panel. (C) Relative abundances of three selected species versus the depth of the samples, indicative of their distance to the coast (modified from Mohr et al., 2005).

*mesaensis* (Takahashi & Blackwelder 1992) and tropical/subtropical diatoms (Fig. 11.8B) reflects the occurrence of subtropical surface waters in the northern and central areas off Chile.

The highly diversified diatom community off 34–38°S points to changing hydrographic conditions. Coastal upwelling-associated diatoms abruptly decrease south of c. 38°S, being progressively replaced by non-upwelling representatives,
The original source rock signal of the different geological terranes of Chile between 25°S and 43°S is best preserved in bulk mineralogical characteristics and the elemental composition of the slope sediments. Bulk mineral assemblages and major element compositions are not significantly altered by continental chemical weathering because of high morphologic gradients and short transport distances throughout the study area. This is reflected, for example, in generally low quartz contents, the dominance of plagioclase, and high amounts of pyroxenes and amphiboles that emphasize the low maturity of the sediments along the Chilean continental margin. Climate influences sediment composition less by weathering but more by its impact on continental hydrology which controls the relative source rock contribution from the Coastal Cordillera versus the Andes. Similarly, these source rock contributions also primarily control the relative abundance of clay minerals. Climate influences on the clay mineralogy include the continental neoformation of smectite and, especially, changes in illite crystallinity. The latter reflects latitudinal variations in weathering activity in Chile corresponding to the climatic zonation. Clay mineral assemblages have not been altered significantly by marine authigenic processes. Finally, the grain-size distribution of the terrigenous sediments is primarily controlled by climate through its influence on the mode of sediment input (aeolian versus fluvial) and the grain size of the source material (physical versus chemical weathering).

Off northern Chile at 27°S the arid climate results in strong aeolian input of terrigenous material, as indicated by high silt/clay ratios and coarse median silt grain sizes (Fig. 11.9a), and low chemical weathering intensity, as reflected, for example, in high illite crystallinities (Fig. 11.9b). The absence of larger rivers transporting material from the Andes to the coast results in mineralogical and elemental composition characteristics typical for a Coastal Cordillera provenance. These include
comparatively high amounts of quartz, amphibole, K-feldspar and mica contents, less plagioclase and pyroxenes, and markedly lower Fe/A as well as smectite/illite ratios in surface sediments of this region. Such data are consistent with predominantly felsic plutonic source rocks with subordinate amounts of metamorphic rocks and basaltic to andesitic volcanic lithologies.

Further to the south (i.e., c. 33°S), increased precipitation allows rivers originating from the Andes to cut through the Coastal Cordillera and to supply Andean material, mostly basaltic-andesite source rocks, to the continental slope. This Andean source is clearly reflected in higher plagioclase and pyroxene, less quartz, K-feldspar and mica contents, higher Fe/A, and lower illite/smectite ratios (Fig. 11.9c). These data from the slope are consistent with results based on analyses of trench sands in this region whose origin has been attributed to Andean source rocks, especially to Quaternary volcanics at the northern limit of active volcanism of the Southern Volcanic Zone (Thornburg & Kulm 1987b; Chapter 5). The continental hinterland around 33°S transect is still characterized by relatively coarse-grained source material due to prevailing mechanical weathering under a semi-arid climate and high morphologic gradients. But as winter rain increases, sediment is mainly supplied by rivers. The observed strong offshore fining trend of the bulk and silt grain size at short distances indicates a strong energy gradient occurring off river mouths. Aeolian sediment input is less important. The fluvially supplied material is probably mainly deposited by hemipelagic processes, as turbidity currents are channelized off central Chile (33–38°S) and are restricted to submarine canyons and fan systems (Thornburg et al. 1990), which have not been sampled.

The most immature bulk mineralogical assemblage recognized so far occurs in the 35°S to 36°S area (i.e., very high plagioclase and pyroxene contents, the lowest amphibole amounts and comparatively low quartz contents). This is in agreement with the findings of Baba (1986) for river samples and Thornburg & Kulm (1987a) for trench sands at similar latitudes. The very immature character of the sediments is in contrast to maximum chemical weathering intensities expected by the climatic conditions in this area, i.e., a combination of high rainfall and temperature. These high chemical weathering rates are reflected, for example, in poor illite crystallinities in the 33°S and 35°S area (Fig. 11.9b). However, immature bulk mineral assemblages seem to contradict this finding. Apparently chemical weathering is sufficient for a strong hydrolysis of illite (resulting in poor crystallinities) but not for significant

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**Fig. 11.8.** Relative abundance of four diatom groups, expressed as percentage of the whole diatom assemblage, in surface sediments collected in the SE Pacific, off Chile: (A) coastal upwelling; (B) tropical/subtropical; (C) southern, cold waters; (D) benthic. The solid line represents the average abundance for each degree of latitude between c. 22°S and 44°S (modified from Romero & Hebbeln 2003). The pigment concentration in surface waters (dashed line) is derived from Coastal Zone Color Scanner measurements (http://daac.gsfc.nasa.gov).
alteration and dissolution of minerals of the source rocks, due to high erosion rates and short transport distances. Baba (1986) suggested that the high rainfall induces rapid rock erosion which offsets the increased weathering intensity. Source rocks in the 33–38°S segment include abundant metamorphic rocks in the Coastal Cordillera and dominantly volcanic rocks (basalts to basaltic andesites) in the Andes. Bulk and clay mineralological as well as elemental composition data indicate a dominance of debris from Andean source rocks transported by the high discharging river system. The mineral assemblage of the 35°S transect reveals the lowest variability within the continental slope transsects of the study area reflecting extended drainage areas and the dominance of the Andean source rocks. Thornburg & Kulm (1987a) also noted the distinctive nature of trench sands in this area due to an overwhelming contribution from highly erodible Andean volcanic rocks. Significantly finer-grained bulk and silt grain sizes at 35°S (Fig. 11.9a) can be explained by finer-grained source material due to abundant volcanic source rocks and to increasing precipitation which promotes chemical weathering. Moreover, topographic gradients in the hinterland are less pronounced than further north and some of the coarser-grained terrigenous material becomes trapped in both the Chilean Longitudinal Valley and on the shelf, which becomes wider in this area.

In the southernmost part of the study area (41°S to 43°S), the heterogeneity of the bulk mineralogy and grain-size data within the transect increases. While feldspar and quartz contents are comparatively low, amphibole and mica amounts reach their maximum values. The sediments are probably derived from various source rocks including metamorphic rocks of the Coastal Cordillera, and volcanic and plutonic lithologies from the Andes. Due to the extreme width of the shelf, including the sea-covered forearc basin and the dissection and partial submergence of the Coastal Cordillera into an archipelago, the provenance of surface samples cannot be related directly to single drainage basins. The variability of the data within the slope environments results from varying source rock contributions and likely shelf dynamics, as well as resedimentation processes on the slope as indicated by surface samples with more positively skewed, coarser grain-size distributions that may represent distal turbidites. Thornburg & Kulm (1989b) also observed a mixed igneous and metamorphic provenance for trench sediments around 42°S and explained the diversity of the mineralogical assemblage by mixing processes in fluvial and shallow-marine environments.

Surface sediments from areas west of the Peru-Chile Trench contain significant amounts of terrigenous fine silt. This material may reach the pelagic areas by aeolian transport in the northern part of the study area as has also been proposed for northern Chile and Peru (Krisske et al. 1980). In the southern part, the trench does not form a barrier to the seaward transport of terrigenous material, owing to its thick sediment infill. Therefore, distal turbidity currents can reach out into the ocean west of the trench axis and deposit silt material. In this area prevailing onshore westerly winds prevent aeolian sediment input. The bulk and clay mineral assemblages do not differ significantly from the adjacent slope samples and the regional patterns of the slope samples are particularly well recorded by the bulk mineralogy and illite crystallinity (Fig. 11.9).

Past sedimentation patterns and palaeo-environmental implications

Palaeo-productivity

A number of recent publications concerned with palaeoceanography and palaeo-productivity reflect increasing interest in Quaternary climate dynamics along the PCC (e.g. Marchant et al. 1999; Klump et al. 2001; Hebbeln et al. 2002; Dezelic et al. 2004; Mohtadi & Hebbeln 2004; Mohtadi et al. 2004; Romero et al. 2006). These studies suggest that the strength and the position of the Southern Westerlies belt and the ACC, which are presently responsible for nutrient supply and upwelling intensities off Chile (Thomas et al. 1994; Thomas 1999; Marchant et al. 1998; Hebbeln et al. 2006a, b; Romero

Fig. 11.9. Sedimentological data of surface sediments from the Chilean continental slope: (a) Silt/clay ratio and the median of the silt fraction; (b) illite crystallinity; (c) Fe/Al and smectite/illite ratios. All data are average values for the respective continental slope transects. For more details see Lamy et al. (1998b).
The following palaeo-productivity reconstructions are mainly based upon results from Mohtadi et al. (2004) & Mohtadi and Hebbeln (2004), who analysed five sediment cores between 24°S and 33°S, and results from two marine records at 35°S (Romero et al. 2006) and 41°S (D. Hebbeln, unpubl. data; Lamy et al. 2002) (Figs 11.10 & 11.11). Generally, higher marine productivity off Chile occurred during the last glacial compared to the early and mid-Holocene. During the late Holocene, palaeo-productivity increased between 24°S and 41°S (Fig. 11.10). Past records of planktic foraminiferal faunal composition off central and northern Chile reveal distinct periods with relatively decreased upwelling intensity congruent with low productivities before ca. 32 000 cal yr BP, the most intense upwelling and highest productivities between 32 000 and 16 000 cal yr BP, and lowest productivities after 16 000 cal yr BP towards the present (Fig. 11.11). Interestingly, the onset and ending of these periods do not show any latitudinal preferences. Relative variations in the palaeo-productivity calculated from accumulation rates (AR) of biogenic compounds generally support these findings, although some local differences occurred, mainly during the last deglaciation (Fig. 11.10).

The predominately southward increase in marine productivity under present-day conditions can also be observed in marine records between 24°S and 41°S for the last 40 000 years. This pattern implies that the same circulation systems have been responsible for productivity variations off Chile over this time span. The comparatively consistent pattern of planktic foraminiferal fauna across ca. 10° of latitude throughout the last 40 000 years implies large-scale changes in the oceanic circulation synchronously affecting the whole area, i.e. changes in the position and/or advection of the PCC/ACC. During the last glacial, a more northerly position of the ACC probably led to higher productivities off Chile compared to the Holocene (Hebbeln et al. 2002; Mohtadi & Hebbeln 2004; Mohtadi et al. 2004). According to these authors, a northward shift of the climatic zones brought the ACC (the main macronutrient source, in particular phosphate and nitrate; Levitus et al. 1994), and the Southern Westerlies (the main onshore precipitation source contributing to micronutrient supply, e.g. iron; Hebbeln et al. 2000a; Dezileau et al. 2004) closer to the study area between 24°S and 33°S, boosting marine productivity during the last glacial. In addition, a northerly position of these zonal systems would successively result in a steeper hemispheric temperature gradient, stronger zonal and meridional winds, and intense coastal upwelling. Stronger advection of the PCC/ACC during glacial periods has also been proposed by investigations north of the study area, off Peru (Feldberg & Mix 2002, 2003), and the Galapagos Islands (Le et al. 1995), and in the Eastern Equatorial Pacific (Pisias & Mix 1997; Mix et al. 1999).

Different palaeo-productivity patterns deduced from biogenic compounds can be related to local differences in this region during the last deglaciation. At 33°S and further south, the northward displacement of the Southern Westerlies belt led to the highest palaeo-productivities around the Last Glacial Maximum (LGM), when the climate zones were at their northernmost position (e.g. Lamy et al. 1998a) (Fig. 11.10). However, highest palaeo-productivities north of 33°S occurred before and after the LGM, suggesting additional or different controls on palaeo-productivity. Continental records from northern Chile do not show any influence of the Southern Westerlies since late Pleistocene times (e.g. Fox & Streecker 1991; Veit 1992, 1996; Ammann et al. 2001; Grosjean et al. 2001). These results might explain the different patterns of biogenic components in the marine sediments north of 33°S. Within this area, the period of highest productivity during the deglaciation coincides well with the observed late glacial humid phases in north Chile.

Fig. 11.10. Comparison of the normalized accumulation rates of biogenic compounds combined with the palaeo-productivity index (PP, thick black lines), which is the mean of the normalized accumulation rates of organic carbon (dashed lines), opal (thin black lines) and CaCO3 (grey lines). The data have been normalized to a 0–1 scale (1) by subtracting the minimum value from all values within one data set and (2) by dividing them through the maximum minus minimum value (data from Mohtadi & Hebbeln 2004).
and Bolivia caused by a more southerly position of the Inter-Tropical Convergence Zone (ITCZ) (Messerli et al. 1992; Seltzer 1992, 1994; Servat et al. 1995; Clapperton et al. 1997; Clavtyon & Clapperton 1997). Enhanced onshore precipitation and flooding of the shelves probably launched additional nutrient sources, leading to highest productivities in this part of the PCC during deglaciation. The exact effect of the ITCZ on the marine palaeo-productivity off north Chile remains to be determined in future studies. However, it seems that palaeo-productivity off northern Chile is additionally controlled by variations in the tropical climate.

During the early and middle Holocene, the zonal systems were at their southernmost position (Markgraf 1995) corresponding with lowest marine productivity within the past 40,000 years (Mohdahi & Hebbeln 2004) (Fig. 11.10 & 11.11). During late Holocene times, productivity increased most likely due to a northward shift in the position of the zonal systems. This shift is indicated in, for example, glacier advances in the Andes (Grosjean et al. 1998), palaeosols in northern Chile (Velei 1996), as well as in pollen records (Heusser 1998b; Villagrán & Varela 1990), which show a general shift to more humid conditions in northern and central Chile over the late Holocene. In addition, an intensification of ENSO events occurred in the last 7000 years when rapid fluctuations in the planktic foraminiferal relative contributions started (Fig. 11.11). The timing of the onset (ca. 7000 cal yr BP) and the intensification (ca. 5000 cal yr BP) of ENSO in the marine records off Chile are consistent with records off Peru (Keefer et al. 1998), and lake sediments from Ecuador (Reddoll et al. 1999). High rainfall in central and northern Chile, Peru and Ecuador seems to have been provoked by intensified El Niño activities (McGlone et al. 1992; Markgraf 1998), contributing to enhanced marine productivity in this part of the world ocean during the late Holocene.

### Sea surface temperatures

Presently available sea surface temperature (SST) records along the Chilean continental margin are restricted to the central (ca. 33–35°S; Kim et al. 2002; Romero et al. 2006) and the southern parts (ca. 41°S, Kaiser et al. 2005; Lamy et al. 2002, 2004) of the Chilean continental margin. The southern records from core GeoB 3313 and ODP Site 1233 provide an exceptional high time resolution and clearly resolve millennial to multicentennial-scale variations. The alkene-based SST reconstruction at ODP Site 1233 reaches back 70 thousand years (ka) (Kaiser et al. 2005; Lamy et al. 2004). The main features of this unique record are the occurrence of coolest temperatures around 65 ka BP (marine isotope stage MIS 4), comparatively warm temperatures in early MIS 3, cold temperatures during the peak glacial from 45 to 19 ka BP without a clearly defined last glacial maximum (LGM), a 6°C SST warming over Termination I, and an early Holocene optimum (Fig. 11.12D). Important to note are pronounced millennial-scale variations in the order of 2°C that clearly correlate to temperature fluctuations as known from Antarctic ice-cores (Blunier & Brook 2001) and from marine records elsewhere in the southern hemisphere mid-latitudes (Fig. 11.12).

The 6°C warming over Termination I appears to be characteristic for the region as it is similarly observed in the alkene SST records at 33°S and 35°S (Kim et al. 2002; Romero et al. 2006) as well as in terrestrial records (e.g. Denton et al. 1999). It further matches results from a modelling study that fits the modelled ice extent to the empirical evidence on the extension of the Patagonian ice-sheet during the LGM, suggesting that a 6°C temperature decrease relative to present day is required (Huybrechts et al. 2002). A further important feature of the Termination I record at Site 1233 is the occurrence of a major cooling event (ca. 0.8°C) that matches the Antarctic Cold Reversal (Fig. 11.12D) and clearly preceeds the northern hemisphere Younger Dryas (YD) during which a pronounced warming of c. 2°C to early Holocene values is observed. Thus the SST
Fig. 11.12. Comparison of the ODP Site 1233 SST record in the SE Pacific with other SST proxy records from the southern hemisphere, and the Antarctic air temperature record of Byrd ice-core, during the past 70 ka. (A) Oxygen isotope record on planktic foraminifera in the South Atlantic (Ninnemann et al. 1999). (B) Mg/Ca SST reconstruction in the SW Pacific (Pulskke et al. 2003). (C) Mg/Ca SST reconstruction in the South Pacific (Mashlofeta et al. 1999). (D) Alkenone SST reconstruction at ODP Site 1233 (this study). (E) Oxygen isotope record of the Byrd ice-core (Blumier & Brook 2001). ACR, Antarctic Cold Reversal; A1 to A5, Antarctic warm events after Blumier & Brook (2001). After Kaiser et al. (2005).

record opposes results from terrestrial records within the Chilean Lake District region and Isla Grande de Chiloé that have been interpreted in terms of a YD cooling (e.g. Moreno et al. 2001), though it has been recently suggested that the deglacial cold reversal in NW Patagonia started earlier (at c. 14.7 to 13.4 ka BP), and that the YD interval is rather characterized by fire disturbances (Hajdas et al. 2003; Moreno 2004b) that may not necessarily imply cooling.

The SSTs reach a maximum in the early Holocene (c. 11 to 8 ka BP) and generally decrease thereafter, reaching modern levels in the late Holocene (Fig. 11.12D). A warmer and probably also drier (than today) climate during the early Holocene is consistent with regional terrestrial records (e.g. Abarzua et al. 2004) and the marine records of terrestrial palaeo-climates in northern and central Chile (Lamy et al. 1998a, 1999). Furthermore, most Antarctic ice-core records show a widespread early Holocene optimum between 11.5 and 9 ka BP (Masson et al. 2000). This early Holocene optimum was not documented in the earlier published SST record based on the short core (GeoB 3313-1) drilled at the same location as Site 1233 that covers only the last c. 8 ka (Lamy et al. 2001, 2002). However, this short record provided some interesting details on millennial to centennial scale SST variations during the middle and late Holocene and their relation to continental climate changes (see above). The main observation is that SST changes on these timescales generally correlate to changes in Antarctica. The same applies for continental rainfall changes in the middle Holocene but these changes appear to become out of phase with SST changes in the late Holocene, possibly related to the onset of the modern state of ENSO (Lamy et al. 2002).
Continental palaeo-climates derived from terrigenous sediment input changes

Due to the extreme latitudinal rainfall gradient, reaching from hyperarid conditions in the Atacama Desert through semi-arid Mediterranean-type conditions in central Chile to extremely high rainfall values in the southern mountains, the terrigenous sediment input to the ocean, and thus offshore sedimentation rates, increases significantly to the south. This pattern of increasing sedimentation rates, which has been broadly known since the early studies by Scholl et al. (1970), is clearly revealed by dated sediment cores recovered during two cruises with the German R/V Sonne (Hebbeln et al. 1995, 2001) and is further consistent with data based on the recently drilled ODP Sites 1233, 1234 and 1235 (Mix et al. 2003). Holocene sedimentation rates increase from ca. 5 cm/ka off arid northern Chile (Lamy et al. 1998a, 2000), ca. 10 cm/ka off semi-arid north-central Chile (Lamy et al. 1999; Marchant et al. 1999), ca. 50 cm/ka off humid central Chile (D. Hebbeln. unpubl. data), ca. 100 cm/ka at 41°S off year-round humid southern Chile (Lamy et al. 2001, 2004), to ca. 200 cm/ka at c. 44°S (D. Hebbeln, unpubl. data) where rainfall reaches extremely high values. Sedimentation rates were even higher during the last glacial, reaching ca. 10 cm/ka at 27°S, ca. 50 cm/ka at 33°S, and ca. 200 cm/ka at 41°S. These pronounced north-south gradients at the Chilean continental margin provide a unique opportunity to reconstruct late Quaternary palaeo-environmental changes both on the continent and in the coastal ocean, with time resolutions ranging from Milankovitch scale (i.e. ca. 20 000 to ca. 100 000 year variability) off arid northern Chile to centennial scale off year-round humid southern Chile.

On Milankovitch time scales, there is strong evidence for precession-driven palaeo-environmental variations in the Norte Chico (27°S) marked by increased onshore precipitation.
related to relatively northward positions of the Southern Westerlies during precession maxima (including the LGM) and vice versa (Lamy et al. 1998a, 2000; Stuart & Lamy 2004). These precessional cycles are evident in different sedimentological parameters that suggest substantial changes of sediment provenance, weathering regimes in the source areas, and modes of sediment input to the ocean. During times of precession minima the coarse median grain sizes of the terrigenous silt fraction reflect a predominantly aeolian sediment input (Fig. 11.13). Low smectite/fiilitie and Fe/Al ratios (Fig. 11.14a) point to a high contribution of plutonic source rocks from the Coastal Cordillera and, thus, to the absence of perennial rivers.

![Fe/Al ratio graphs](image)

**Fig. 11.14.** Terrigenous sediment input changes (exemplified by Fe/Al ratios) on different time scales as recorded in three sediment cores along the Chilean continental slope from north to south. (a) GeoB 3375-1 covering the last c. 120 000 years (after Lamy et al., 2000). (b) GIK 17748-1/GeoB 3302-1 covering the last c. 30 000 years (J. Klump, unpubl. data). (c) GeoB 3313-1 covering the last c. 8000 years (after Lamy et al., 2001).
delivering material from the Andes. In combination with high illite crystallinities (Fig. 11.13), indicating minimal chemical weathering, these data clearly indicate severely arid conditions during precession minima. Conversely, during time spans coincident with precession maxima, finer grain sizes (Fig. 11.13) can be interpreted in terms of increased fluvial sediment supply, while higher smectite/illite and Fe/Al ratios (Fig. 11.14a) record a stronger Andean volcanic source rock signal. In addition, lower illite crystallinities point to an increase in the intensity of chemical weathering. Thus, for precession maxima our data suggest a semi-arid climate with seasonal precipitation similar to the modern Mediterranean climate further south, where perennial rivers led to increased fluvial sediment input from the Andes. Besides showing dominantly precession-driven rainfall changes over the Andes, the record also reveals millennial-scale changes in weathering intensity over the Chilean Coastal Cordillera, these most likely induced by changes in coastal fog occurrences (Lamy et al. 2000). As the frequency and intensity of coastal fogs along the Chilean coast is today also related to the position of the Southern Westerlies, we can reasonably assume that the past record indicates rapid latitudinal shifts of this wind belt throughout at least the last 80 000 years.

Further south, off central Chile (c. 33°S), higher sedimentation rates allow a more detailed reconstruction of climatic evolution over the last c. 30 000 years (Lamy et al. 1999). As with the Norte Chico record, clay mineralogical and element composition data clearly record changes in the relative contribution of Andean versus Coastal Cordillera source rocks. However, as increased precipitation has enabled rivers originating from the Andes to cut through to the coast, sedimentological data instead record the relative dilution of the Andean signal by additional input of Coastal Cordillera material during comparatively humid intervals. Thus, we find a lower contribution of clay mineralogical assemblages and Fe/Al ratios characteristic of Andean provenance during more humid phases. As shown, for example, by the Fe/Al ratio record (Fig. 11.14b), the sedimentological records clearly suggest more humid conditions during the LGM, which is consistent with the Norte Chico record. The deglaciation is characterized by a trend towards a more arid climate that reached its maximum during mid-Holocene times (8000–4000 cal yr BP). The late Holocene was marked by more variable palaeo-climates with a return to generally slightly more humid conditions.

Off humid southern Chile (41°S), where extremely high sedimentation rates allow a reconstruction of terrigenous sediment input changes over the last 80 000 years on centennial time scales, Holocene ‘long-term’ trends are generally consistent with the results off central Chile (Fig. 11.14c) (Lamy et al. 2001). Clay mineral and element composition data suggest a higher input of Andean material during the mid-Holocene, indicating less dilution by Coastal Cordillera source rocks and thus less rainfall. However, in this presently year-round humid region, hydroclimatic conditions were probably always humid and never reached semi-arid (Lamy et al. 2001). On millennial to multicentennial scales, terrigenous input changes reveal dominating bands of variability centred on c. 900 and 1500 years, suggesting significant rainfall changes and thus latitudinal shifts of the Southern Westerlies on these time scales (Lamy et al. 2001). Recent ODP drilling off southern Chile at the same location now offers the chance to extend the very high resolution records into the last glacial (Lamy et al. 2004). Under these glacial conditions, ODP Site 1233 was located in an ideal position to monitor Patagonian ice-sheet (PIS) extent variations by recording compositional changes in the regional terrigenous sediment input to the continental margin. For this purpose, the Fe content of the bulk sediment is taken as a proxy for changes.

Fig. 11.15. Iron content changes (ten-point moving average; plotted inverse) as a proxy for Patagonian ice-sheet changes and alkenone sea surface temperatures (SST) from 8 to 50 ka yr at ODP Site 1233 (modified after Lamy et al. 2004). Graphical correlation lines illustrate the variable ice-sheet response time to SST changes offshore. Ice-sheet maxima based on terrestrial data are from Denton et al. (1999).
in the PIS extent, assuming that glacial erosion processes strongly enhanced the glaciocluvial sediment flux from Fe-rich basaltic volcanics in the Andes during the last glacial, and ice-sheet advances provided more Fe-rich material that was subsequently transported to the continental margin by rivers (Lamy et al. 2004). A modelling study of PIS changes during the late glacial points to a close dependence of ice extent on offshore SSTs (Hulton et al. 2002). However, by comparing the PIS extent record with SST changes (see below) reconstructed on the same well-dated archive, the data from Site 1233 suggest a variable ice-sheet response time of up to c. 1000 years (Fig. 11.15). Terrestrial data suggest that the PIS retreated rapidly from the Chilean Lake District at c. 17.5 ka BP (Denton et al. 1999), which is clearly documented in a significant decrease of sea surface salinities at Site 1233, suggesting large inputs of melt-water (Lamy et al. 2004). Thereafter, the record of Site 1233 becomes more and more dominated by rainfall changes as discussed above.

**Driving forces for palaeo-environmental variability**

Based on the marine records discussed in the previous sections, it appears that both the terrestrial palaeoclimate and the palaeoceanography of the Peru–Chile Current system (PCC) (at least off Chile) are primarily controlled by latitudinal shifts of the dominating atmospheric (the Southern Westerlies), and oceanographic (the Antarctic Circumpolar Current: ACC) circulation members. The position of the Westerlies and the ACC is controlled by the location of the sub-polar low pressure belt and the strength and position of the SE Pacific anticyclone (Cerveny 1998), allowing the potential for both high (southern) latitude and tropical Pacific forcing mechanisms.

A tropical impact is most obviously visible on orbital time scales, i.e. in the precessional band as precession-driven changes in insolation in the tropics are much more pronounced than at high latitudes. Such changes are thus primarily observed in proxy records off northern Chile (Lamy et al. 1998; Stuut & Lamy 2004). On shorter time scales tropical forcing
mechanisms, possibly involving long-term ENSO changes, have been suggested for palaeo-productivity changes in the north (Mothadi & Hebbeln 2004) and partly for rainfall variability in southern Chile (Lamy et al. 2001, 2002).

In particular, the records off southern Chile suggest that the major large-scale control of palaeoceanographic changes off Chile is related to latitudinal shifts of the northern margin of the ACC originating from Antarctica and the surrounding Southern Ocean. At least on millennial time scales, these latitudinal shifts appear to be part of a hemisphere-wide pattern as observed for example in SST records around the southern hemisphere mid-latitudes that correlate to temperature fluctuations over Antarctica (Fig. 11.12). This clear ‘Antarctic timing’ pattern is consistent with the often-discussed seesaw mechanism of interhemispheric climate change (e.g. Broecker 1998), but the very similar temperature pattern around Antarctica in different ocean basins could also imply a larger involvement or even a source of millennial-scale climate variability in the southern hemisphere most likely involving changes in the extent of sea-ice (Lamy et al. 2004; Kaiser et al. 2005).

The latitudinal shifts of the ACC appear to affect not only the mid-latitude SSTs but also the large-scale SST pattern along the PCC system further to the north. Kaiser et al. (2005) recently reconstructed SST gradients covering the complete latitudinal range of the system up to the equator and showed that the subtropical gyre circulation was displaced by several degrees of latitude equatorwards during cold MIS 2 and 4, primarily originating from a northward shift of the ACC resulting in particularly enhanced SST gradients in the southern PCC (Fig. 11.16). This configuration enhances the equatorward flow of cold water within the PCC. Conversely, during relatively warm periods such as early MIS 3 and the early Holocene optimum, the oceanic circulation in the PCC system was weakened and the ACC, as well as the associated westerly wind belt, moved southward.

Following the model in which high productivity in the southern Peru–Chile Current is sustained by the high-nutrient, low-chlorophyll waters of the ACC (Hebbeln et al. 2000a), variations in the latitudinal position of the ACC would result in latitudinal movements of the main nutrient source of the Chilean upwelling system. As surface sediment data show that nutrient availability and productivity decrease with increasing distance to the ACC, palaeo-productivity changes at a specific core site probably reflect increasing/decreasing distances to this source, resulting from latitudinal movements of the ACC.

Combining the available datasets, it can be deduced that palaeo-environmental variability across both onshore and offshore Chile mainly depends on large-scale oceanic/atmospheric circulation pattern in the SE Pacific, which in turn are primarily coupled to high southern latitude climate variability in the Southern Ocean and Antarctica. Tropical forcings are present but are much less important, although they may well play a more prominent role in short-term, i.e. centennial to decadal-scale, climate variability. The investigation of the impact and the behaviour of past long-term changes in modern climate modes such as ENSO, including those originating in the high latitudes (e.g. the Southern Annular Mode), is a major issue for future marine research along the Chilean continental margin.

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Late Quaternary environments and palaeoclimate

CLAUDIO LATORRE (coordinator), PATRICIO I. MORENO, GABRIEL VARGAS, ANTONIO MALDONADO, RODRIGO VILLA-MARTÍNEZ, JUAN J. ARMESTO, CAROLINA VILLAGRÁN, MARIO PINO, LAUTARO NÚÑEZ & MARTIN GROSJEAN

Chile possesses one of the most pronounced climate gradients in the world, extending from the world’s driest desert in the northern part of the country, where precipitation is measured in millimetres per decade, down to the channel and fiords region in southern Patagonia where rainfall can average up to 7 m per year or more. In contrast, thermal buffering by the Pacific Ocean contributes to ameliorating extreme temperatures, generating a latitudinal temperature gradient that is considerably less pronounced than across similar latitudinal ranges in other parts of the world (Miller 1976; Axelrod et al. 1991). Coupled with millions of years of geographic isolation induced by the massive barrier imposed by the Andean Cordillera, Chile today possesses a highly endemic fauna and flora whose distribution is tightly linked to these gradients (Arroyo et al. 1996; Hinojosa & Villagrán 1997).

Considering its geographic position and tectonic setting, it is hence not surprising that the geomorphology of Chile over the last two million years or so, i.e. the ‘Quaternary’ (see Gradstein et al. 2004), has been strongly influenced by climate along this broad latitudinal gradient. Whereas ancient landscapes preserved for millions of years exist in the hyperarid Atacama, repeatedly glaciated landscapes predominate in southern Chile. Elucidating the precise chronology of these Quaternary events affecting the western margin of southern South America is of great relevance to a number of scientific disciplines including ecology, palaeo-climatology, evolutionary biology, population genetics, phylogeography, biogeography and conservation.

Consequently, records of past climate and landscape change in Chile during the late Quaternary (an informal term generally spanning the last glacial–interglacial cycle) have been developed and described by an impressive number of researchers from many countries since the beginning of the twentieth century. The vast majority of such scientific endeavours have used either palaeo-ecological or geological/sedimentological evidence to describe palaeoclimate, geomorphology, neotectonics, biogeography and evolution in this region of South America. Over the past 20 years the pace of research has quickened, following a worldwide demand for ever increasing and more detailed palaeoclimate data. Recent developments in high-resolution ice core and marine records from the mid- and high latitudes of the southern hemisphere have led to new studies of climate linkages and controlling mechanisms at the interhemispheric level during and since the last ice age (Denton et al. 1999; Lowell et al. 1995; Kim et al. 2002; Mayewski et al. 2004).

To date, Quaternary research in Chile has produced a vast amount of data. Here, we provide an overview of the current knowledge based on work by active researchers (most reside in Chile) from a range of disciplines including Quaternary geology and stratigraphy, palaeoceanography, palaeo-ecology, biogeography and archaeology. Starting with a consideration of offshore influences (see also Chapter 11), we move onland through Chile from north to south, including special sections on the rich flora of central (‘mediterranean’) Chile and the Chiloé Archipelago and the prominent terraces attributed to the last interglacial period around the coastal city of Valdivia (see Fig. 12.1 for site locations). We finally attempt to integrate what we know about past climate change with the earliest colonization of Chile by humans.

For the sake of consistency, all radiocarbon ages younger than 26 000 14C yr BP in this chapter have been converted into calendar years before 1950 using CALIB 5.1 (http://www.calib.org). CALPAL using 2004 SFCP (www.calpal.de) was used for radiocarbon ages between 26 000 and 50 000 14C yr BP. Ages older than 50 000 were left uncalibrated. Except were noted, all ages in this chapter are expressed either in calendar years before 1950 (yr BP), thousands of calendar years before 1950 (ka BP), or millions of years ago (Ma).

Past Quaternary research in Chile

Ever since the discovery of Mylodon Cave in southernmost Patagonia during the late 1800s, the Quaternary period in Chile, and in particular the late Pleistocene, has received worldwide attention. Excellent summaries of past Quaternary research in Chile have already been published (Paskoff 1977; Ortleb 1995; Villagrán 1995b; Heusser 2003). Even though the Quaternary is a short ‘blip’ in the geological timescale, two million years is still quite a large chunk of time! We thus do not intend such a detailed synthesis in just one chapter of a book, but rather our aim here is to provide an overview of specific areas of ongoing research in Chile, especially focusing on late Quaternary palaeoclimates.

The first major review by Paskoff (1977) focused on several aspects of the Quaternary, with a strong emphasis on coastal geomorphology, together with his work on the geomorphology of semi-arid lands in north-central Chile (the ‘Norte Chico’) (see also Paskoff 1970). A synthesis of the palaeoclimate of northern Chile (the ‘Norte Grande’) was later provided by
Fig. 12.1. A physical map of Chile. All localities discussed in this chapter including offshore core sites are indicated. Key: 1, Nevado Sajama; 2, Pampa del Tamarugal; 3, Salar de Uyuñú; 4, Río Salado; 5, Calama; 6, Tuña; 7, Salar de Atacama; 8, Salar de Hombre Muerto; 9, Laguna Miscanti; 10, Salar de Punta Negra; 11, Quebrada del Chaco; 12, Marine core GeoB 3375-1; 13, Laguna Negro Francisco and Río La Gallina; 14, Valeriana and Encierro valleys; 15, Tongoy; 16, Fray Jorge and Talinay; 17, Nague I, II and Quereo; 18, Palo Colorado; 19, Marine core GiK 17748–2; 20, Marine core GeoB 3302-1; 21, Algarrobo; 22, Laguna Aculeo; 23, Laguna Tagua Tagua; 24, Marine core GeoB 3313–1; 25, Monte Verde; 26, Laguna Stibnite; 27, Mylodon Cave and Cueva del Medio; 28, Fell’s Cave; 29, Pali Aike; 30, Tres Arroyos; 31, Marazzi.
Ortlieb (1995), although a considerable amount of research has taken place since this publication (see a recent update by Latorre et al. 2005) (also see below). Several important syntheses of the impacts of Quaternary climate and vegetation change in southern Chile have been written by palynologists Carolina Villagráin Villagráin (1990, 1995a, b; Villagráin et al. 1998) and Calvin Heusser (2003). The latter publication includes not only a detailed historical narrative of past pioneers of the Quaternary in the southern Andes but also a summary of over 40 years of research including over 20 radiocarbon-dated palynological records. Quaternary impacts and distribution of past fauna were summarized in a review by Moreno et al. (1994) for northern and central Chile, by Latorre (1999) for Tierra del Fuego and more recently by Frassinetti & Alberdi (2000, 2001).

Late Pleistocene–Holocene palaeoceanography of the Peru–Chile Current (G.V.)

As detailed in the previous chapter, the Peru–Chile (or Humboldt) Current is a complex oceanographic system that modulates continental climate along the Chilean coastal margin. Off central and northern Chile, northward advection of cold Subantarctic Waters along its oceanic and coastal branches dominates the surface circulation, whereas a counter-current transports warm subtropical water masses poleward (Strub et al. 1998; Chaigneau & Pizarro 2005a, b, c). The subsurface Poleward Undercurrent transports nutrient-rich and oxygen-depleted Equatorial Subsurface Waters (Brandhorst 1971; Morales et al. 1996), and constitutes a main source of nutrients for the photic zone during upwelling processes, supporting not only primary production in the ocean surface layer, but also strong CO₂ outgassing particularly off central and northern Chile (Torres et al. 2002). South of approximately 42°S the West Wind Drift or Antarctic Circumpolar Current (ACC) intercepts the Chilean margin and flows southward into the Cape Horn Current (Strub et al. 1998; Chaigneau & Pizarro 2005a, b, c). The distribution of surface sediment properties off Chile confirms this oceanographic pattern, suggesting that upwelling processes are dominant as a key factor for primary production in central and northern Chile, whereas the ACC and nutrient fluvial input become dominant along the margin of southernmost Chile (e.g. south of 40°S) (Hebbeln et al. 2000a).

Detailed palaeoceanographic work on sediment cores collected between 27°S and 41°S all seem to indicate that oceanographic variations along coastal Chile during the late Pleistocene (Fig. 12.2), and particularly during the last glacial period, have been driven by high latitude climate change. These mostly include changes around Antarctica and implied latitudinal shifts of the Southern Westerlies and the ACC, following precession-driven changes in insolation, major changes in thermohaline circulation and global climate change.

Proxy analysis of terrigenous material present in sediment cores from central Chile (at 27°S and 33°S) (Lamy et al. 1998a, 1999, 2000) suggest strong precession-driven cyclic recurrences of more humid climates in both the High Andes, through increased snowfall, and in the Coastal Cordillera, with increased local rainfall during austral summer insolation maxima due to northward latitudinal shifts of the Southern Westerlies by c. 5°, coupled with increased influence of subpolar air masses. The intensified runoff in most of these records has been linked to an increase in fluvial input of continental iron together with heightened biological productivity (Dezileau et al. 2004).

During the Last Glacial Maximum (LGM: 23–19 ka BP), sedimentary records at 33°S and 41°S suggest sea surface temperatures (SSTs) of about 12°C and 9°C lower, respectively, than those during the Holocene (Kim et al. 2002; Lamy et al. 2002, 2004). This has been associated with a northward displacement of the position of the ACC and the Southern Westerlies, generating a considerably wetter climate in central Chile (Heusser 1990b; Lamy et al. 1999, 2000) as well as inducing high primary production rates off central Chile at 33°S (Hebbeln et al. 2002; Mohtadi & Hebbeln 2004) (also see below). According to these records, Termination 1 has been dated to 18.5 ka BP at 33°S and to 19.2 ka BP at 41°S, with a warm pulse, most evident at 33°S, occurring at c. 15 ka BP. These warming events culminated in maximum SSTs of 19°C at 13 ka BP and 15°C at 12.1 ka BP (41°S), respectively (Kim et al. 2002; Lamy et al. 2004).

The similar timing and direction present between the record of SSTs at this last site with the oxygen isotope (δ¹⁸O) record of the Byrd ice core in Antarctica (Blunier & Brook 2001), have led Lamy et al. (2004) to propose an Antarctic influence on the onset of the deglaciation and the maximum extension of glaciers in the Patagonian Ice Sheet (PIS), which occurred at 18–17.7 ka BP (Denton et al. 1999), some 1000 years after the increase of the SSTs at 41°S. A close dependence between ice sheet extent along the western slope of the Andes and offshore
SSTs during regional ice maxima has been suggested through model results by Hulton et al. (2002). The apparent inertia, however, of the PIS to ocean–climate forcing, which could have exerted a variable influence on continental environments, might explain the differences in timing between terrestrial records of the glacial and deglacial periods (Lowell et al. 1995; Denton et al. 1999; Moreno et al. 2001). According to Lamy et al. (1998a, 1999, 2000) increasing SSTs during deglaciation reflect a southward migration of the ACC and the Southern Westerlies, inducing overall more arid conditions at the Pleistocene–Holocene transition in central-southern Chile (Heusser 1990b; Villagrán & Varela 1990; Veit 1996; Lamy et al. 1998a, 1999, 2000). To date, marine records along coastal Chile have provided very little evidence for a Younger Dryas type event: only a weak signal is seen in the planktonic foraminifera record at 33°S (Marchant et al. 1999) which is absent in the alkenone-based SST reconstruction obtained from the same core (Kim et al. 2002).

The Holocene is characterized by lower primary production and increased climate variability from tropical sources such as El Niño events (Lamy et al. 1999, 2001, 2002; Marchant et al. 1999; Hebbeln et al. 2002; Mohladi & Hebbeln 2004). A strong cooling event characterized by SSTs 1°C lower than present at 33°S, occurred at c. 11 ka BP, a time of reduced summer insolation and sea-ice cover, followed by SST values similar to and slightly warmer than present (c. 17°C) until c. 8 ka BP (Kim et al. 2002). This may have supported more intense coastal fogs and wetter conditions in restricted water catchments along the semi-arid coast of Chile (e.g. Maldonado & Villagrán 2002). Afterwards, a strong warm pulse occurred between 8 and 7.5 ka BP, reaching a mid-Holocene maximum SST of about 19°C at c. 6 ka BP (Kim et al. 2002). This Holocene optimum closely matches the highest SST values of about 17°C at c. 5.5 ka BP recorded south at 41°S (Lamy et al. 2002). Although temperatures decreased gradually, high SSTs persisted during the mid-Holocene in both regions, probably associated with a southward shift of the ACC and the Southern Westerlies and strengthened SE Pacific Subtropical Anticyclone (SPA) (Kim et al. 2002; Lamy et al. 2002), roughly contemporaneously with warm phases and landward recession of ice sheets at coastal sites around Antarctica (Finocchio et al. 2005; Steig et al. 1998), and strong aridity in central Chile (Heusser 1990b; Villagrán & Varela 1990; Lamy et al. 1998a, 1999; Jenny et al. 2002b; Villa-Martínez et al. 2003). Increasing paleo-productivity and oceanographic variability during the late Holocene reported in sediment cores from central and southern Chile (Marchant et al. 1999; Lamy et al. 2001, 2002), occurs concomitantly with trends towards decreasing SSTs, particularly during the last four millennia at 33°S and 41°S (Kim et al. 2002; Lamy et al. 2002). This decreasing temperature trend is also coeval with more humid climate conditions in central Chile and with decreasing air temperatures, increased sea-ice formation and ice-rafted debris around Antarctica (Steig et al. 1998; Domack et al. 2001). One probable cause is the influence exerted by increased El Niño–Southern Oscillation (ENSO) activity beginning at c. 5 ka BP (Rodbell et al. 1999). This caused poleward advection of warm subtropical water along the Chilean margin (Marchant et al. 1999), increased soil formation, runoff and clastic sedimentation in central-northern Chile (Veit 1996; Lamy et al. 1999; Jenny et al. 2002a; Villa-Martínez et al. 2003), more frequent debris flow events due to heavy rainfall episodes along the coastal Atacama Desert in northern Chile and southern Peru (Vargas et al. 2000), as well as possibly controlling the extent of Antarctic sea-ice (Clement et al. 1999). Recent work on drilling sites off coastal northern and southern Chile suggests significant ocean–climate variations during the last few centuries, as demonstrated by increased primary production and interdecadal variability in the coastal upwelling system off northern Chile (Vargas et al. 2004) and in the Chilean fiords (Rebolledo et al. 2005; Sepúlveda et al. 2005).

A major unsolved problem for these records remains in the changes associated with the Regional Radiocarbon Reservoir Effect (∆R), especially during the last deglaciation and the Holocene, when intense global changes in the ventilation rates of seawater masses have been documented (Southon et al. 1990; Bard et al. 1994; Southon et al. 1998; Van Beek et al. 2002). This implies that significant, previously unrecognized, time lags may be present in chronologies of past ocean–climate change. This may also account for the apparent disagreement between ocean cores and continental records of climate change, in particular those from northern Patagonia (see below). For the Peru–Chile current system, recent work indicates that the equatorward advection of anomalously 14C-depleted water masses from the Southern Ocean during the early Holocene, combined with intensified coastal upwelling during this time, were probably responsible for higher ∆R values of up to around 1000 years, with a mean ∆R = 496 ± 304 years in northern Chile and southern Peru (Fontugne et al. 2004). In comparison, mid-Holocene values characterized by a mean ∆R = 175 ± 101 are of considerably less impact. This implies that the chronostratigraphy of some sedimentary records from both central-northern and southern Chile should be revised, at least for these specific time periods at millennial and submillennial timescales.

**Hyperarid northern Chile (Norte Grande) (18–27°S)**

Northern Chile is well known throughout the world as home to the Atacama Desert, the driest and perhaps the oldest of the major subtropical deserts found in the southern hemisphere (Dunai et al. 2005; Hartley et al. 2005). Cosmogenic 26Ne exposure dates on surfaces in Quebrada Tilivilche near Pisagua (19°35′S) indicate onset of extreme hyperaridity as early as 25–22 Ma (Dunai et al. 2005). The Atacama extends along the Pacific Andean slope from the southern border of Peru (18°S) to Copiapó, Chile (27°S) (Fig. 12.1). The region’s hyperaridity owes its existence to a combination of: (1) the extreme rainshadow effect of the High Andes, blocking advection of tropical/subtropical moisture from the east; (2) limited influence of winter storm tracks from the south owing to the presence of the semi-permanent South Pacific Anticyclone (SPA); and (3) the generation of a temperature inversion at c. 1000 m by the cold, north-flowing Humboldt Current that constrains inland (upslope) penetration of Pacific moisture. The SPA has been anchored against the westward bend in the South American continent probably throughout the Neogene. Uplift of the central Andes to their current elevation may have occurred as early as 15 Ma (Vandervoort et al. 1995; Alpers & Brimhall 1988), although palaeobotanical evidence suggests that the central Andes were only at half their modern elevation at 10 Ma (Gregory-Wodzicki 2000). Recent palaeothermometry work using stable C and O isotopes from soil carbonates along the Bolivian orinoco provides evidence for abrupt uplift (>3 km) of the Altiplano between 10.3 and 6.7 Ma (Ghosh et al. 2006). The Humboldt Current is thought to have been active since early Tertiary times (Keller et al. 1997), but may have reached its present intensity as a result of the major expansion of the Antarctic ice sheet between 15 and 12.5 Ma (Flower & Kennett 1993), or after the Central American Seaway closed between 3.5 and 3.0 Ma (Ibaraki 1997).

In both summer and winter, precipitation variability is primarily modulated by Pacific SST gradients and associated upper air circulation anomalies with seasonal and annual precipitation totals determined by the number of precipitation days (Vuille et al. 2000; Garreaud & Acéituno 2001; Garreaud et al. 2003). These anomalies can promote greater spillover of summer moisture (either from the Amazon to the NE or the Gran Chaco to the SE), or conversely, greater penetration of winter storm tracks from the SW.
At millennial to glacial–interglacial scales the role of the tropics in global change is still poorly understood. Is regional insolation a major forcing factor of the Bolivian High, or is the intensity of South American summer rainfall forced instead by changes in equatorial Pacific SST gradients with both tropical and extratropical teleconnections?

Late Quaternary millennial-scale changes in the frequency and seasonality of the scant rainfall are associated with shifts in plant and animal distributions with elevation in the Atacama (Fig. 12.3). Today, plant diversity and distribution follow a gradient of mean annual precipitation, which in the central Atacama (22–24ºS) increases from almost 0 mm/year at c. 2400 m to 200 mm/year at 4000 m. Along the Andean slope, a sparse vegetation zone known as Prepuna appears at 2900 m and gradually transitions into the Puna Belt (locally known as Tolar) between 3100 and 3900 m. This highly diverse vegetation belt is dominated by shrubs of the Asteraceae and Solanaceae families. The High Andean steppe is found above 3900 m, dominated by tussock grasses such as *Stipa*, *Nassella* and *Festuca*.

A large, ever-growing body of evidence has been obtained through vegetation changes analysed in rodent middens: bio-accumulations of plant and animal material made by rock-dwelling rodents (Betancourt et al. 1990; Betancourt & Saavedra 2002). Throughout the late Quaternary, midden (Betancourt et al. 2000; Latorre et al. 2002, 2003) and palaeo-wetland (Rech et al. 2002, 2003a) records from the central Atacama are in close agreement, in particular for a major wet phase between >15 and 9 ka BP. New evidence from midden research at Rio Salado in the Calama Basin (Latorre et al. 2006) has revealed that this wet phase may have begun as early as 17.4 ka BP, coinciding closely with the onset of deglaciation at higher latitudes (Lowell et al. 1995; Moreno 1997), lake transgression at Salar de Uyuni (Sylvestre et al. 1998; Placzek et al. 2006) and post-LGM globally warmer temperatures. Palaeo-lake Taucá collapsed at 14.1 ka BP, slightly reviving afterwards between 13 and 11 ka BP during the minor Coipasa phase (Placzek et al. 2006). This millennial-scale drought possibly occurred throughout the central Andes and is clearly evident in the Calama basin of northern Chile (Latorre et al. 2006).

A lesser wet phase during the mid-Holocene identified from preliminary evidence from rodent middens and palaeo-wetland stratigraphy in the central Atacama, was initially dated to between 7 and 3 ka BP (Betancourt et al. 2000). Recent work has replicated plant species migrations from upper into lower vegetation belts between 7.6–6.0 ka BP and 4.5–3.2 ka BP, with onset of drought at c. 5 ka BP (Latorre et al. 2003) (Fig. 12.3). Major regressive phases in high Andean lake records, however, also timing of human occupation of the Salar de Atacama area (‘silencio arqueológico’ = archaeological silence: Núñez et al. 2002) have been interpreted in terms of extreme aridity (see below). Among the explanations proposed for this discrepancy are different sensitivities of the proxies used (e.g. Grosjean et al. 2003), hardness contamination of ¹⁴C dates (Geyh et al. 1999) and regional impacts and interpretations (e.g. Grosjean 2001 versus Quade et al. 2001). As discussed in the last paragraph of this section, climate variability in the Atacama and the central Andes may be considerably more complex than previously thought.

The elaboration of preliminary new records from the southern Atacama at Quebrada del Chaco (25º30'S) constitute a step towards reconciling different palaeoclimatic interpretations (Latorre et al. 2005; Maldonado et al. 2005). The majority of wetland deposits in Quebrada Chaco and tributaries, as well as midden records from upper elevations (>3450 m) dating between 15.4 and 10.2 ka BP are in good agreement with a late Glacial/early Holocene wet phase indicated in the central Atacama records. Early onset of aggradation beginning at c. 20.8 ka BP, however, agrees with low elevation midden evidence (<3000 m) for an earlier pluvial between 23.8 and 18.1 ka BP. One source for this earlier pluvial could be the enhancement or northward migration of westerly storm tracks during the LGM (23–19 ka BP).

Fig. 12.3. Modern distribution of vegetation belts in northern Chile (after Villagrán et al. 1983). Grey arrows indicate plant species migrations during the late Quaternary (different shades are correlated with time of migration).
In contrast to the central Atacama, absence of wetland deposits along with impoverished midden floras implies widespread aridity during the Holocene, with a slight increase in moisture over the last 1500 years (Maldonado et al. 2005). This pattern is clearly different from that visible in midden and wetland records further north, and implies that mid-Holocene increases in summer precipitation were of insufficient magnitude to have reached Quebrada Chaco at 25°30'S latitude. Further work in the southern Atacama is needed to place both a maximum northward limit to the westerly excursion during the LGM as well as to establish the southern limit of increased summer moisture during the late glacial and middle Holocene.

A diverse array of records of continental climate change for the central Andes is now available (see special issues of Palaeogeography, Palaeoclimatology, Palaeoecology, 194, 2003 and Journal of Quaternary Science, 20(7-8), 2005). Notable among these are lake and salt cores from Lake Titicaca and Salar de Uyuni (Baker et al. 2001a, b; Fornari et al. 2001), chronostatigraphic work on shoreline deposits throughout the Altiplano (Clapperton et al. 1997; Clayton & Clapperton 1997; Sylvestre et al. 1999; Piacsek et al. 2006), ice cores from Nevado Sajama (Thompson et al. 1998) and Illimani (Ramirez et al. 2003), salt cores from Salar de Atacama (Bobst et al. 2001, Lowenstein et al. 2003) and Horcon Mierito (Gomzyey et al. 2003) and a lake core from Laguna Miscanti (Grosjean et al. 2001). Discrepancies in palaeoclimate interpretation are still apparent, such as the slightly wetter mid-Holocene, an event that is completely absent from the High Andean lake records (all of which indicate intense drought). Although part of the discrepancy arises owing to the different sensitivities of the parameters involved as reconstructed from such diverse palaeoclimate methods, substantial geographic variation in the sources and mechanisms of precipitation over the central Andes can help understand part of the problem. Summertime precipitation variability in the central Andes was largely determined by the upper-air zonal wind component aloft, with an easterly (westerly) flow favouring wet (dry) conditions. The influence of these upper-air circulation anomalies on precipitation becomes more dominant when longer moisture transport distances are involved and hence is more prominent along the western slope of the Altiplano (Vuille et al. 2000). Different source areas also underlie summer rainfall patterns in the central Andes. Whereas the eastern Cordillera and northern Altiplano receive most of their moisture from the Amazon Basin, the Gran Chaco becomes more prevalent for the southern Altiplano and adjacent Cordillera (Vuille & Keimig 2004). Consequently, past rainfall changes in this region could very well display both a temporal and spatial component depending on latitude and altitude (e.g. Maldonado et al. 2005).

The semi-arid Norte Chico (27°-32°S) (A.M.)

The Norte Chico lies along a transition between the hyperarid Atacama Desert to the north and the distinct mediterranean climate of central Chile to the south. The predominant blocking influence of the south Pacific Subtropical High is much stronger here than further south, causing occasional winter droughts that become more frequent with decreasing latitude (Van Husen 1967). Conversely, infrequent but heavy winter rainfalls associated with major El Niño events correlate with major animal population outbreaks and cycles of the flora and fauna (e.g. Lima et al. 1999; Gutierrez et al. 2000; Jaksic 2001; Gutierrez & Reserve 2003).

High resolution palaeoclimate archives from the Norte Chico are few, most not reaching even as far back as 6000 yr BP. Records of Pleistocene biotic and climate change are even less frequent. Despite this shortcoming, all published records (see Fig. 12.1 for locations) show an alternation of wet and dry periods, despite some disagreement regarding the exact timing and duration of these phases. To date, four palaeoclimate records have been obtained from local swamp forests along the coast at c. 32°S (Villagrán & Vareló 1990; Maldonado & Villagrán 2002, 2006), and three palaeopedological and limnogeological records from the Andean Cordillera (Grosjean et al. 1997, 1998; Earle et al. 2003). One study also establishes comparisons between palaeoclimates inferred from coastal and mountain palaeosols at the regional scale (27°-33°S) (Veit 1996).

Pollen records from swamp forests at Quereo and Palo Colorado indicate a climate wetter than today during the early Holocene. These forests persisted at Palo Colorado (32°05’S) between c. 10 000 and 8700 yr BP (Maldonado & Villagrán 2006). The pollen record at Quereo, on the other hand, indicates that swamp forests dried out some time after 9370 yr BP (Villagrán & Vareló 1990). Wetter conditions between 10 000 and 8300 yr BP have also been interpreted from palaeosols (Veit 1996).

Climate shifted towards extreme arid conditions after 9000 yr BP. The swamp forest at Palo Colorado disappeared and only matorral and herbaceous species persisted between 8700 and 7800 yr BP. Pollen-starved inorganic sediments suggest that these multimillennial-scale arid conditions persisted until 6000 yr BP. This also agrees with pollen evidence for a dry phase at Quereo (Villagrán & Vareló 1990) and palaeosol evidence for temperatures that may have been up to 3°C higher between 8100 and 5700 yr BP (Veit 1996).

A gradual, stepwise rather than abrupt shift characterized the return of more mesic conditions starting at 4200 yr BP. Onset of sedimentation begins at 6200 yr BP at the sites Nague I and III (31°50’S), with absence of the swamp forests and predominance of herbs and matorral shrubs until 4200 yr BP. Swamp forests began to colonize the sector at this time, associated with increased moisture (Maldonado & Villagrán 2002). Wetter conditions began at 5300 yr BP at Palo Colorado, as indicated by the appearance of Myrtaceae pollen, which gradually increased until 4200 yr BP. Palaeolimnological evidence for increased moisture during this interval also comes from Laguna del Negro Francisco in the High Andes near Copiapó (27ºS), where a saline lake that had existed since 6800 yr BP, shifted into a freshwater lake at c. 4200 yr BP (Grosjean et al. 1997).

High elevation records indicate an intensification of wetter conditions after 3200 yr BP when lake levels rose again at Laguna del Negro Francisco and glaciers advanced in the Valeriana and Encierro valleys at 29ºS (Grosjean et al. 1998). A down-elevation trend was observed for the Neogloboidea species began replacing herbaceous ones, although interrupted by several short drought episodes documented both at Nague (< 2500 yr BP) and Palo Colorado (3000-2200 yr BP). An arid phase occurs slightly later in the High Andes, dated between 1700 and 1100 yr BP at Río La Gallina (Earle et al. 2003) and < 1800 yr BP at Laguna del Negro Francisco. Wetter conditions during the late Holocene are also implied by palaeosol evidence across the region. Intense westerly activity may have reached as far north as Quebrada del Chaco at 25°30’S (see previous section) where pollen preserved in rodent middens indicates the altitudinal descent of hillslope species into the absolute desert at 1500 yr BP. Finally, the high variability seen in Myrtaceae pollen percentages at Palo Colorado together with relatively high percentages of annuals and geophytes between 2200 and 1300 yr BP (Maldonado & Villagrán 2006) may be related to increased frequency in ENSO events during the late Holocene, as reported by many other researchers for western South America (Rodbell et al. 1999; Moy et al. 2002; Riedinger et al. 2002; Jenny et al. 2003).

Central Chile (32°-35°S) (R.V.M.)

As with the Norte Chico, Central Chile (32°-35°S) occupies a transitional position between two major features that
characterize the western coast of southern South America, in this case the South Pacific Anticyclone (SPA) and the westerly wind belt. The variable climatic regime of this region is related to changes in the strength and position of the SPA. During the austral summer, this high pressure centre occupies a broad latitudinal range off South America that blocks the flow of humid air masses moving across the Pacific Ocean, leaving central Chile completely dry. During the austral winter, however, the anticyclone weakens and moves equatorward, thereby allowing cyclonic storms and related frontal systems to progress towards central Chile.

A major component of the interannual variability in precipitation in central Chile is linked to ENSO (Aceituno 1988, 1990; Rutland & Fuenzalida 1991; Aceituno et al. 1993). During El Niño years, weakening of the anticyclone and decreased upwelling of cool ocean waters along the west coast of South America are factors that lead to large increases in total rainfall across the entire mediterranean-climate region of Chile. During pronounced La Niña episodes when the SPA is stronger, rainfall decreases well below the average, particularly in the northern part of the mediterranean region. These strong episodes can extend even to the wetter southern limit of the region, causing droughts during summer (Holmgren et al. 2001).

Despite such a key geographical position for understanding subtropical palaeoclimatic changes, including the onset and subsequent evolution of ENSO, only a handful of palaeoenvironmental records are known for central Chile. For example, the only locality that has produced a pollen record that encompasses glacial times is from Laguna Tagua Tagua (34º30'S), a lake drained artificially in historic times, which spans the last 54 000 years (Heusser 1983, 1990b; Valero-Garcés et al. 2005). The broad-leaved sclerophyllous woodland vegetation that surrounds the site today was dominated by the southern beech Nothofagus and Prumnopitys conifers, indicating a colder and considerably wetter climate during the late Pleistocene. These wet and cold phases occurred between 50 000 and 40 100 yr BP, and 32 600 and 13 260 yr BP. Arid phases, dominated by grasses (Poaceae) and halophytes (Chenopodiaceae), occurred between 54 000 and 50 000 ¹⁴C yr BP and between 40 150 and 32 950 yr BP. The climate became drier after 17 500 yr BP, as indicated by the decline of arboreal taxa and concomitant increase in grasses and halophytes (Heusser 1990b).

New sedimentological, geochemical and palynological data from the Laguna de Tagua Tagua, published recently by Valero-Garcés et al. (2005) are used to infer much higher precipitation between 40 100 and 21 000 yr BP, followed by a general increase in aridity that was interrupted by two humid spells (19 500–17 000 and 13 500–11 500 yr BP). This new record clearly demonstrates that the early to mid-Holocene was probably the driest period for the last 46 000 years in central Chile.

The few Holocene palaeoclimatic records that exist document several major climate fluctuations, again in the form of alternating wet and dry phases, with a prevalence of warm, arid conditions during the early and middle Holocene. These studies include pollen records (Heusser 1983, 1990b; Villagrán & Varela 1990; Villa-Martínez & Villagrán 1997) and palaeosols (Veit 1996). The poorly constrained chronology and temporal resolution in most of these records, however, impedes further precision regarding our understanding of the timing and structure of these phases.

A more complete dataset of Holocene climate is now available from Laguna Aculeo (c. 34º S), a high-resolution multiproxy record from one of the few natural inland lakes present in the latitudinal range of South America that blocks the westerly flow of warm moist air that brings summer rainfall and thunderstorm activity. This is a feature of great significance for understanding the disturbance regime that has historically affected Chilean mediterranean ecosystems. It also accounts for the negligible importance of spontaneous fire in the vegetation of central Chile. The rainshadow effect of the Andes, responsible for maintaining hyperarid conditions in the Atacama Desert, also isolates the mediterranean region from the easterly flow of warm moist air that brings summer rainfall and thunderstorms to the subtropical Andes. Lightning storms are generally confined to elevations above 3000 m in the Andes (Arroyo et al. 1981).

Mediterranean-type ecosystems occupy a narrow band along the western margin of South America, from 30ºS to 36ºS in central Chile. Because of their position in the transition zone between the southern Atacama Desert and the mixed deciduous–evergreen temperate forests which occur south of 36ºS, they represent a highly heterogeneous vegetation mosaic. The major vegetation types are dry xerophytic thorn scrubs dominated by deciduous shrubs and succulents, mesic communities dominated by evergreen sclerophyllous trees in the coastal and Andean foothills, and forests dominated by winter deciduous trees in the south of the mediterranean area. Within the latitudinal range of the mediterranean climate, annual precipitation increases from less than 200 mm to more than 700 mm (Rundel 1981).

The mountainous topography of central Chile, contrasting radiation and moisture regimes, and strong rainshadow effects coupled with a mosaic of soil types and nutrient supplies, generate pronounced environmental gradients that have stimulated the evolutionary differentiation of the biota (Armesto & Martinez 1978; Rundel 1981; Rozzi et al. 1989). High floristic richness and a diversity of plant communities are a consequence of this environmental heterogeneity (Arroyo et al. 1993, 1995). In addition, strong climatic variability during the Quaternary (Jenny 1989; Varela 1990; Aceituno et al. 2001) probably promoted increased genetic variability among local populations. The mediterranean region is also a depository of ancient tropical lineages that found refuge in coastal valleys from the sustained drying trend initiated by Andean tectonics during the late
Neogene (Troncoso et al. 1980; Villagrán & Armesto 1980; Arroyo et al. 1995; Hinojosa & Villagrán 2005). Plant migrations associated with climatic fluctuations during the Quaternary (Heusser 1983; Villagrán 1995a) contributed greatly to increase species richness and the heterogeneity of the vegetation mosaic in central Chile (Arroyo et al. 1995; Villagrán 1995a). Late Neogene Andean uplift has also provided new opportunities for colonization and differentiation of local alpine floras, increasing floristic richness and diversity even further.

In addition, the strong influence of insect and bird pollinators in the evolution of plant-reproductive strategies in mediterranean-type ecosystems (Arroyo & Uslar 1993; Arroyo et al. 1993, 1995), and the role of animals in the dispersal of seeds of sclerophyllous tree species (Hoffmann & Armesto 1995), are important conditions favouring genetic variability and speciation in biotic communities within the region. High levels of self-incompatibility are known for the woody species in montane sclerophyllous vegetation (Arroyo & Uslar 1993). The historical phytogeography of central Chile presented below follows the general floristic scheme proposed by Arroyo et al. (1995). Although different and additional vegetation types are included here (Armesto et al. 2007). We briefly discuss the more important vegetation types in terms of their relative floristic richness, endemism, distribution and natural history.

A widely distributed vegetation type in central Chile is the coastal forest dominated by the monotypic and endemic Aextoxicon punctatum (‘olivillo’). These forests, described by Muñoz & Pisano (1947), Villagrán & Armesto (1980) and Pérez & Villagrán (1985), exhibit diverse species of climbers and epiphytes, including narrow endemics such as Peperomia coquimbensis (Piperaceae). Olivillo forests are generally immersed in a matrix of xerophytic vegetation, and reach as far north as La Serena (30°S) at Fray Jorge and Talinay (Troncoso et al. 1995). Although different and additional vegetation types are included here (Armesto et al. 2007). We briefly discuss the more important vegetation types in terms of their relative floristic richness, endemism, distribution and natural history.

The historical phytogeography of central Chile presented below follows the general floristic scheme proposed by Arroyo et al. (1995). Although different and additional vegetation types are included here (Armesto et al. 2007). We briefly discuss the more important vegetation types in terms of their relative floristic richness, endemism, distribution and natural history.

The regional extent of dry xerophytic vegetation versus mesic sclerophyllous tree communities in central Chile seems to be held in a fragile balance. This balance is usually tipped by climatic cycles between cool, wet conditions and hot, dry episodes. These in turn may be linked to varying frequencies of ENSO during the Holocene (Villagrán 1995a; Maldonado & Villagrán 2006). Large-scale regeneration of sclerophyllous woodland is normally limited by drought stress and the lack of seed banks (Fuentes et al. 1986; Jiménez & Armesto 1992), except during unusually wet periods or along water courses. It is likely that the strong El Niño events (warm ENSO phases) in central Chile promote significant regeneration of sclerophyllous tree species, especially in dry or disturbed habitats. Owing to recurrent anthropogenic disturbance through fire and permanent grazing pressure, however, vegetation cover may be presently shifting regionally towards dominance by dry xerophytic species (Arroyo & Martínez 1978), despite a regional trend towards increasing rainfall in the late Holocene (see previous section).

Xerophytic open woodlands, with the physiognomic aspect of a savanna (Fuentes et al. 1990), are currently widespread along the Central Valley in the mediterranean-climate region. These woodlands, dominated by leguminous trees, mainly Acacia caven and Prosopis chilensis, have a dense herb layer almost entirely made of introduced European annual herbs (e.g. Erodium spp.) and grasses, typically associated with grazing disturbance. Although some discussion has centred on the historical nature of this semi-arid formation (Arroyo et al. 1995), high percentages of introduced grasses found underneath Acacia trees, and the resprouting ability of persistent shrubs, suggests that this vegetation type has been shaped by human impact, especially through cattle grazing and fire.

Mixed sclerophyllous forest with abundant endemic palms Jubaea chilensis are a narrowly distributed and frequently overlooked component of the mosaic of vegetation types in coastal areas of central Chile. These evergreen forests are typical of very humid and mesic conditions found in deep canyons along the Coastal Cordillera (Arroyo et al. 1995). Only two small protected areas in central Chile contain significant palm populations at present, suggesting that mixed palm forests may have been more widespread, possibly in the recent past.
Finally, upland sclerophyllous woodland is found above 1500 m in the Andean mountains and differs both from the Andean foothill communities (Hoffmann & Hoffmann 1982; Arroyo et al. 1995) and from the upland Nothofagus communities of the Coastal Cordillera. This vegetation is dominated by Kagenckia angustifolia (Rosaceae), which constitutes a discontinuous treeline (Arroyo & Uslar 1993). Above the treeline, at 2300 m in central Chile, the montane woodland gives way to a low subalpine shrubland (Arroyo et al. 1981). Scattered patches of the long-lived mountain cypress Austrocedrus chilensis (Cupressaceae) may occur near the treeline at various locations between 32 and 36°S (Rundel 1981). The distribution of Austrocedrus is more continuous on the eastern side of the Andes south of 36°S. The presence in central Chile of these relict trees that possibly became established during wetter periods of the late Holocene, relying today on their sprouting ability to survive drought (C. Le-Quesne & J. C. Aravena, pers. comm.), further highlights the very sensitive nature of these vegetation formations to climatic variability over long timescales.

**Northern Patagonia (39–41°S) (P.I.M.)**

The study of interhemispheric climate linkages during and since the last ice age has benefited from the recent development of high-resolution ice core and marine records from the mid- and high latitudes of the southern hemisphere (Stenni et al. 2001; Lamy et al. 2004). Few palaeoclimatic records from terrestrial environments in these regions, however, have the temporal continuity, time resolution and adequate chronologic control to allow a detailed examination of the timing, rates, direction and phasing of climate change at millennial timescales.

Northwest Patagonia is ideal for the study of interhemispheric linkages throughout the Quaternary because this region has insolation regimes out-of-phase with northern mid-latitudes, is highly sensitive to variations in the Southern Westerlies, and is far removed from the direct influence of northern hemisphere ice sheets and deep water circulation (Moreno 2002). Stratigraphic, palynologic and charcoal records from small, high-sediment accumulating lakes in the Chilean Lake District (41°S) afford useful data for examining the interval from the LGM to the present.

A stacked palynological record (Fig. 12.4) from NW Patagonia that encompasses the last 24 000 years was developed to examine the timing, rates and direction of vegetation and climate change (Moreno et al. 1999; Moreno & León 2003; Moreno 2004a). The record shows extreme glacial climate (ΔT = −6.5°C, ΔP = c. 2000 mm/year) between 24 and 17.7 ka BP, followed by the abrupt expansion of north Patagonian rainforest taxa through successive warming events between 17.7 and 15.5 ka BP. Conditions approaching modern climate prevailed between 15.5 and 15 ka BP, followed by expansion of cold-resistant, hygrophilous north Patagonian rainforest trees at 15 and 13.5 ka BP, and subsequent warming pulses at 11.5 and 10.2 ka BP (Moreno et al. 2001; Hajdas et al. 2003). The earliest charcoal peaks, indicative of local fires between 15 and 14.5 ka BP, occurred despite the local dominance by temperate rainforests at the time under cool, temperate and wet conditions. Extreme warm, dry climate prevailed between 10.2 and 7.8 ka BP, as indicated by the predominance of the warmth-loving, drought-resistant Valdivian trees Eucryphia cordifolia and Caldcluvia paniculata, and lowered lake levels. Charcoal maxima and local vegetation disturbance are evident in several sites at different times between 13 and 8.5 ka BP. Cold-resistant north Patagonian rainforest trees underwent a step-like re-expansion between 7.5 and 5.5 ka BP, and reached peak abundance between 5 and 3 ka BP. Eucryphia and Caldcluvia re-expanded at 5 ka BP establishing a vegetation mosaic with podocarps and Nothofagus that persists until today. During the latter period, charcoal records indicate renewed fire activity with prominent increases at 5 and 3 ka BP. Modern vegetation and climatic conditions started at 2 ka BP, following a warm, dry phase between 3 and 2 ka BP.

Rainforest vegetation changed at ecological timescales (≤100 years) in response to climate forcing at millennial timescales since the last termination. Moreover, the timing and mean time spacing of events fall in the range of millennial-scale changes identified in the North Atlantic region (Bond et al. 1997, 1999; Bianchi & McCave 1999). Submillennial-scale variability is also evident: El Niño years in NW Patagonia typically exhibit below-normal summer precipitation (Montecinos et al. 2002).
2000), hence the co-occurrence of thermophilous, drought-resistant Valdivian elements and cold-resistant, hyrophilous north Patagonian rainforest trees since c. 5 ka bp might represent a vegetation response to the onset of El Niño-like variability in the mid- to late Holocene.

The record from NW Patagonia indicates intense fire activity during the warm, dry early Holocene and in some cases during the final portion of the late glacial (Moreno 2004a). It is likely that the primary control on fire occurrence in this region lies in changes in the position and strength of the Southern Westerlies at multilmillennial timescales, climate variability at subcentennial scales, as well as local (human?) ignition sources.

Biodiversity, glacial history and biogeography of the vegetation of Chiloé Archipelago (C.V.)

Biodiversity

The Chiloé Archipelago (41°47’–43°30’S; Fig. 12.5) possesses one of the most diverse and singular floras of Chile. Despite a lack of precise estimates for the total number of terrestrial plants in the Archipelago, it is probably over 1200 species, especially when the approximately 750 known species of vascular plants are considered (Ruthsatz & Villagrán 1991; Villagrán et al. 1986; Villagrán 2002) along with at least 360 known species of Bryophytes (Villagrán et al. 2003, 2005).

‘Ulmo’ (Eucryphia cordifolia) forests, one of the most important and diverse components of the Valdivian rainforest, reach their southern limits along the northern and central part of the Isla Grande (Schmithüsen 1956). North Patagonian rainforest communities are present along the Pichucho Cordillera, and along the southern portion of the Isla Grande and adjacent islets. These forests are dominated by ‘canelo’ (Drimys winteri), ‘tepá’ (Laureliopsis philippiana), myrtles such as ‘luma’ and ‘petagüa’ (Amomyrtus luma and Myrcceugenia ovata) and by Nothofagus nitida and the conifers Saxegothaea conspicua and Podocarpus nubigena (‘maños’) at higher elevations. A more complex vegetation mosaic covers the broad summits of the Pichucho cordillera and low-elevation wetlands. This mosaic comprises small stands of Magellan ‘coigue’ (Nothofagus betuloides) and ‘ñire’ (Nothofagus antarctica): magellanic moorlands (Astelia pumila, Donatia fascicularis and Gaimardia australis); ‘alerce’ (Fitzroya cupressoides); Guaietaceas’ cypress (Pilgerodendron uvifera) and ‘tepú’ (Tepualia stipularis). Many of these species, a true vanguard of the subantarctic flora, meet their southern limits along the northern and central part of the Isla Grande and adjacent islets.

Along the Pacific coast, diverse marsh, beach and intertidal rock communities are found adjacent to pristine forests of ‘arrayan’ (Luma apiculata) and ‘olivillo’ (Aextoxicon punctatum), one of the most unusual of Valdivian rainforest associations, with northern limits along the semi-arid coast of northern Chile (30°30’S) and southern limits at the Guapiquilán, Esmeralda and Guao islets at the SW end of the main island. The olivillo forests of Chiloé house numerous species of vines, epiphytes and rare cryptogams that are endemic to Chilean rainforests, many of which have disappeared from most of their original ranges and today exhibit pronounced disjunct distributions in remote areas (Villagrán & Armesto 2003; Villagrán et al. 2003, 2004a, b, 2005).

Glacial history

Considering the dramatic extent of past glaciations during the Pleistocene, how did Chiloé come to have such levels of biodiversity? Glacial geological records from the last glaciation (Fig. 12.6) (Hollin & Schilling 1981; Denton et al. 1999), show that most of the southern and eastern portions of the main island, as well as continental Chiloé and the islets, were heavily covered by ice during the last ice age, the late Llanquihue Glaciation (LLG), dated between c. 37 000 and 17 500 yr BP, with temperatures estimated to have been 6–8°C lower than today (C. J. Heusser et al. 1999). Periglacial processes such as solifluction, exerted considerable impact on the montane forests of the Pichué Cordillera right down to the foothills (Veit & Garlef 1996). Figure 12.6 indicates the location of the pollen records where these dramatic vegetation changes have been documented. Past changes in mean summer temperature during the LLG (Fig. 12.7) have been obtained from the pollen record at Taiquemó, the oldest such record in Chiloé (Heusser 1990a; Denton et al. 1999). Using this curve as reference, we can describe the following vegetation changes during the most extreme climate phases.

Interstadials during the early to middle LLG

Recent discoveries of in situ subfossil tree trunks of alerce (Fitzroya cupressoides) and Guaietaceas’ cypress (Pilgerodendron uviferum) in the Seno de Reloncavi and eastern margin of Chiloé Island (Fig. 12.6), dated between 45 470 and 51 050 yr BP, record the climate and glacial history of the north Patagonian rainforests as dominated by Nothofagus and conifers (Roig et al. 2001). Pollen analysed from three of these sites (Tenglo, Punta Pirquén and Molulco) evince dominance of arboreal taxa, mostly Nothofagus dombyei-type and conifers such as Fitzroya cupressoides, Pilgerodendron uviferum, Saxegothaea conspicua, Podocarpus nubigena and Lepidothamnus fonkii (Villagrán et al. 1995, 2004b). This assemblage suggests relatively warm and wet interstadials during the early to middle LLG. Conifer forests expanded considerably across the central graben from their disjunct ranges in modern montane forests. Other pollen records with subfossil wood of similar age and similar assemblages, stratigraphy and chronology are Punta Tentén, Chiloé (Heusser et al. 1995), Punta Pelluco, Punta Pesas and Canal Tenglo (Heusser 1981; L. E. Heusser et al. 1999).

Stadials of the late LLG

The dominant vegetation present during the stadials of the upper LLG, dated from c. 37 000 to 17 500 yr BP, has been described from three different pollen records taken from peatbogs in the NE part of Chiloé (Fig. 12.6): Taiquemó (Heusser 1990a), Loncomilla (Villagrán 1990) and Río Negro (Villagrán 1988a). Magellanic moorlands with grasses and composites alternated with small stands of Nothofagus and conifers. The middle to late LLG transition is documented by the discontinuous sections present at Pdipid, Tehuaco and Dalcahüe, along the eastern-central coast of the island (Fig. 12.6), (Villagrán 1985; Heusser 1990a; C. J. Heusser et al. 1999).

Late glacial

After piedmont glacier collapse at c. 17 000 yr BP, quick colonization and rapid expansion of a closed-canopy north Patagonian rainforest with Nothofagus, conifers and Myrtaceae ensued at three pollen records along the central to southern sectors of the eastern coast of Chiloé: Lagunas Pastahue, Tahuí and Mellí (Villagrán 1985; Abarzúa et al. 2004). Vegetation also began colonizing previously glaciated areas in the southern part of the Isla Grande (Puerto Carmen, Laguna Soledad and Chauguata (Villagrán 1988b)). Forest colonization also commenced in the Channel District (Bennett et al. 2000), nearly synchronous with moorland ascent to the summits of the Pichué cordillera and Andes (Villagrán 1991a), as evinced by pollen records from mainland Chiloé, Chaitén and Cuesta Moraga (Heusser et al. 1992).

The Holocene

A series of pollen records documents the transition from forests dominated by temperate–cold species to those dominated by more thermophilous species of the north Patagonian rainforest, such as ‘tineo’ (Weinmannia trichosperma) at the beginning of the Holocene (11 000 yr BP). This was followed by the
Fig. 12.5. Upper panel: Altitudinal and latitudinal vegetation zones along the western slope of the Andes of south-central Chile (Schmithüsen 1956). Note merger between the Valdivian and north Patagonian rainforests at the latitude of Chiloé (41–44°S) as well as conifer distribution among both forests. Lower panel: Frequency distribution along a west–east altitudinal survey of native forest communities across the Pucuhué Cordillera, NW Chiloé (Villagrán 1985).

widespread expansion of ‘ulmo’ (*Eucryphia cordifolia*) from c. 9200 to 6550 yr BP, dominating the vegetation along the SE sector of Isla Grande at Tahui, Melli and Puerto Carmen (Fig. 12.6). This event represents the largest southward expansion of the Valdivian rainforest over the last glacial cycle (Villagrán 1988b; Abarzúa et al. 2004).

**Biogeographic consequences**

An historical model based on the vegetation dynamics observed in the pollen records has been proposed by Villagrán (2001) (Fig. 12.8). This model hypothesizes that modern disjunct distributions of the conifer species in Chiloé, today found only in the coastal mountains and Andes (Villagrán et al. 1998; Villagrán & Armesto 2003), are the end result of forest recolonization of mountain habitats during deglaciations. The ranges occupied by these species would thus have been much more widespread and continuous during interstadials of the early to middle LLG along the low-lying intermediate valley of the Lake District and Chiloé. This hypothesis agrees well with recent phylogenetic molecular studies, which indicate high genetic diversity in all the isolated Chilean and Argentine conifer populations studied, as well as a high degree of divergence in the small stands of alerce (*Fitzroya cupressoides*) present in the Llanquihue central graben (Allnut et al. 1999, 2001, 2003; Premoli et al. 2000; Bekessy et al. 2002). These are most likely the last remnants of ancestral populations that occupied the entire valley during the aforementioned stages of the LLG.

The diversity of subantarctic flora present in Chiloé would thus be a consequence of the northward migration of Magellanic moorland mosaic during the stadials of LLG. Today, these moorland ‘islands’, present along the summits of the Piuchué Cordillera, mark the northernmost limits of *Euphrasia antarctica*, *Gunnera lobata*, *Prattia repens*, *Abrotanella linearisfolia*, *Perezia lactucoideas* spp. *palustris*, *Sisyrinchium patagonicum* and several species of *Cyperaceae* and *Juncaceae* (Ruthsatz & Villagrán 1991; Villagrán 2002). Some 229 species of liverworts have been recorded in Chiloé (Villagrán et al. 2005); most of these are endemic to southern Chile and Argentina with several endemic monotypic (*Roivainenia*, *Perdusenia*) or bitypic genera (*Nothostrepta*, *Arctoscyphus*) of Austral South America. Of these, 144 (63%) expand their ranges into the southern confines of the continent, south of 52°S (Hässel De Menéndez & Solari 1985). Fifty-seven of these (25%) reach their northern limits in the Chiloé Archipelago. The mosses of Chiloé also display a dominant subantarctic component (Villagrán et al. 2003).

In contrast, angiosperm species of the warm forest communities of southern Chile display maximum richness and restricted endemism between the Maule and Valdivia rivers (36–40°S), reaching their southern limits in Chiloé (Villagrán 1995a; Villagrán & Hinojosa 1997). This distribution, together with evidence from palaeopedological studies (Veit & Garleff 1996), the west–east direction of recolonization indicated by isopollen maps (Villagrán 1991b), and the southward recolonization of north Patagonian and Valdivian rainforests in the late glacial and early to mid-Holocene, has led researchers to hypothesize that the coast and western slopes of the Coastal Cordillera, between the Regions of Los Lagos and Bio Bio, have functioned as a major refuge for these temperate rainforests during past glaciations (Fig. 12.8).

Finally, the large disjunctions observed for Chilotan species of the Valdivian rainforest between central Chile (33°S) and semi-arid Chile (30°30'S) probably correspond to ancient
endemisms of previous widespread distributions along coastal Chile in the Tertiary (Villagrán et al. 2004a). This agrees with fossil evidence from the Neogene of central Chile (Hinojosa & Villagrán 1997) and with recent molecular evidence that shows strong genetic divergence between the northernmost populations of two of these taxa (Drimys (Jara et al. 2002) and Aextoxicon (Núñez 2004)). The persistence of these species in very restricted areas along the Pacific coast and islands is likely due to the highly oceanic character of these regions during Pleistocene glaciations (Villagrán 1991b). Among the Chilotan species of angiosperms with relict distributions in semi-arid Chile are Nertera granadensis, Mitratia coccinea, Sarmentia repens, Peperomia fernandeziana, Dysopsis glechomoides and Uncinia phleoides. Among the species of ferns with disjunct distributions between Chiloé, Valdivia and the Juan Fernandez islands are Blechnum corralense, Trichomanes exsectum, Hymenophyllum fuciforme, Histiopteris incisa and Gleichenia litoralis. The species with disjunct distributions in semi-arid Chile are Hymenophyllum peltatum, Asplenium dareoides, Runohra adiantiformis, Hypolepis poepigii, Polypodium feulliei var. feulliei and Megalastrium spectabile var. spectabile (Villagrán et al. 2004a). Of the mosses, Ptychomitrium falcatulum is found on Chiloé, Valdivia and Juan Fernandez, whereas Macromitrium longirostre, Fissidens berterii and Hennediella kanzeana are disjunct between Chiloé and central Chile (Villagrán et al. 2003). Perhaps the most surprising examples are the disjunct ranges exhibited by the Chilotan hepatics Colura calyptrofuria, Frullania fertilis, Microlejeunea ulicina, Monoclea gottschei ssp. gottschei and Plagioclista rufescens, all of which are also present in Fray Jorge–Talinaý in semi-arid Chile, a major gap spanning more than 1600 km (Villagrán et al. 2005).

### Interglacial terraces around Valdivia (M.P.)

Prominent terraces interpreted as either fluvial or marine in origin can be observed in the vicinity of the city of Valdivia (c. 38°30′–40°S) (Antinano & Mcdonough 1999; Brüggen 1944; Fuenzalida et al. 1965; Pino 1987, 1999; Rojas 1990). Their surface height varies from 10 m (in Poma, north of the Imperial River) to 67 m high (in Valdivia) (Fig. 12.9). According to the last interglacial sea-level reconstructions, these should not have been more than 8 m above the present level (Esat et al. 1999; Blum & Törnqvist 2000; Esat & Yokoyama 2000; Lambeck et al. 2002; Shackleton et al. 2003). Current thinking regarding these terrace surfaces is that they resulted from block neotectonics (Illies 1970). Faults that control the blocks (with north–south and NE–SW trending patterns) also control the orientation of the majority of local creeks (Grupo De Estudios Urbanos 1997). Relying on terrace altitude together with stratigraphical observations but lacking absolute dating, Lauer (1968) and Illies (1970) suggested a previous interglacial age, here termed the Valdivia Interglacial (VI). This age agrees with the degree of weathering present in these deposits.

Two different facies can clearly be recognized along the 160 km coastal zone. The first is composed of sediments derived from the local metamorphic basement (Metamorphic Complex of Mansa Bay: Duhart et al. 2001) associated with an accumulation of land-derived plant remains. This facies, defined as autochthonous, has surface development associated with present-day water courses. Coarse and fine alluvial gravels (including colluvial material) are interbedded within this facies, as well as several types of sand, silt and peat. Some of the gravel deposits consist mainly of quartz, while others are dominated by schist fragments. At least three levels of peat can be identified: the first is visible just above current sea-level (on the northern slope of Niebla’s Playa Grande), on Huapi Island in the mouth of Tornagaleones estuary and on the northern slope of Curíñanco Beach; the second is at approximately 34 m above present sea level; and the third lies some 6 m above the previous level (these latter two peat deposits are in the terrace directly north of Los Molinos creek). The peats include numerous logs and branches all extraordinarily well preserved. Among these have been identified as southern beech (Nothofagus sp.), two species of myrtles (Amomyrtus luma and Amomyrtus meli) and Guaiacées’ cypress (Pilgerodendron uviferum). Unexpectedly, small-size leaves of ‘ulmo’ (Eucryphia cordifolia) have also been collected. Two infinite 14C ages > 45 000 yr bp obtained from wood samples at Los Molinos and from a carbonized log from the upper Chan area provide the only chronology (Antinano & Mcdonough 1999).

The second facies, defined as allochthonous, is composed of medium to fine sandy sediments of volcanic origin, generally characterized by more than 15% ash matrix. Locally known as ‘caneaguá’, it is interbedded with silt- and clay-sized sediments,

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**Fig. 12.7.** A reconstruction of mean summer temperature of the Llanquihue Glaciation (LLG) indicating the presence of relatively warmer interstadials during the early to middle LLG (30 000–60 000 yr bp) based on past grass pollen fluctuations from the Taiquemó record (solid line). The dashed line corresponds to colder stadials during the late LLG (< 30 000 cal yr bp) inferred from the maximum extent of moraine deposition around Lago Llanquihue. The interval dated 14 600–10 000 14Cy r BP (17.5–11.0 ka BP) (upper solid line) was developed using pollen records from Taiquemó, Fundo Llanquihue, Canal de la Puntilla and Huelmo (Denton et al. 1999).
and is partially or totally reworked by weathering to kaolinite. The sandstone matrix has also undergone intense weathering, and has mostly converted into a sort of secondary cement. This second facies is abundant not only in the coastal area, but is also found along the most important estuaries and rivers, up to 15 km from their mouths. Outcrops in Playa Grande contain thin interbedded pumice ash layers. Two superimposed stratified flows, observable near the Niebla Spanish fort, dip towards the continent at an angle of c. 45°. The lower flow maintains this position throughout the whole outcrop, whereas the top flow is horizontal along the coastal cliff and starts to dip slowly towards the continent, eliminating the possibility of tectonic tilting. A similar situation is observed at La Misión Beach in Valdivia and along the coast of Lake Budi. This primary structure was first interpreted as aeolian in origin (Antinano & Mcdonough 1999), but given the grain size, presence of matrix, thickness of strata and the transition from flat-lying to a 45° tilt, it is more easily explained as the result of a small frontal or lateral deltaic deposit into a channel or small body of water (Brown 2001). The volume of volcanic sand present reaches at least 5 km³ along the coastal area. Associated with the finest-grained sediments, generally a clay–silt mixture produced
by weathering of volcanic ash, are estuarine and marine invertebrate fossil deposits (Estancilla, Niebla’s Playa Grande, Huapi Island), with well preserved shells and/or casts, ostracods and tree leaves. The shellfish *Crepidula dilatata*, *Crepidula fecunda*, *Caecum chilense*, *Nassarius* sp. and *Aulacomya ater* have all been identified from the Huapi Island outcrop (Pino et al. 2002). With the exception of *Nassarius*, located in the upper strata, the other invertebrates are intermixed with broken shells and in disorder, which suggests that their soft body parts might have been eaten after deposition by *Nassarius*. The only leaf found at Estancilla was lying in a vertical position, indicative of deposition under a non-laminar flow. Based on the spatial arrangement of the fossils, these deposits seem to have been retransported by a hyperconcentrated flow, forming a thanatocenosis. The coquina fossils from Huapi Island have been interpreted as forming a subtropical association (Alvarez & Gallardo 1996), although the invertebrates identified up to now are indicative of an environmental setting similar to the one in nearby Corral Bay. Volcanic ash content decreases from north to south in this facies. Near the mouth of the Imperial River estuary (northern limit of the outcrops) the volcanic ash is present not only in the matrix, but also forms layers of pure ash, together with a thanatocenosis of estuarine invertebrates.

Although the Sollipulli, Sierra Nevada and Llaima volcanoes (all post-glacial in age) are located in the present Imperial River basin, the probable origin of this hyperconcentrated flow of sand and volcanic ash seems to be related to a pre-Llanquihue (last glacial) caldera of Villarica volcano. The volcanic sandstone is composed of fragments of andesite, basalt, orthopyroxene, plagioclase, green hornblende, olivine and magnetite (Pino 1987). Most of the orthopyroxene crystals are not rounded, but rather have kept their euhedral shape, which supports the interpretation of a hyperconcentrated flow. Olivine and orthopyroxene are frequent products of post-glacial basaltic and dacitic volcanism (Naranjo & Moreno 1991), but the presence of green hornblende gives a measure of the importance of andesitic pyroclastic material, since it is not found in lavas (Smith & Lotosky 1995). Even today it has not been possible to associate the deposition of a hyperconcentrated sand flow to a particular volcanic event, as the pre-Llanquihue volcanic centres of the Andes have been buried or destroyed by successive glacier advances.

Both facies are vertically related in several ways. At Playa Grande and throughout the city of Valdivia, hyperconcentrated sand flows overlie peat with an undulating contact that corresponds to flame structures. In this particular case, evidence indicates transport and deposition of a viscous flow over a plastic material. At Los Molinos, allochthonous flows underlie autochthonous facies, whereas the opposite is true at Curiñanco.

If we accept that wavy-surface terrains in the Central Valley near Osorno are typically associated with lahar outcrops, such as at San Pablo (Corvalán 1974), then by analogy the same geomorphology in the Valdivia area indicates that the hyperconcentrated deposit of immature volcanic sand actually corresponds to a lahar. The Valdivia area and adjacent coastal zone exhibit this undulating pattern, alternating with the presence of terraces. The same wavy pattern can be seen along the
secondary road that links the main road (Ruta 5, Mafíl) to the highway between San José de la Mariquina and Valdivia. This road cuts almost perpendicularly through the undulations of a coarse sand deposit which is somewhat less cemented than the traditional 'cancagua', but contains numerous pumice stones. Soutiá (NW Argentine) structures can also be entered along the road towards Gorbea (Cuarta Faja), where very fine and immature volcanic sandstones appear in subsurface at the Santa Maria estate and correspond to what has been defined as 'cancagua'. Both outcrops lead towards the Villarica volcanic basin.

The reinterpretation of deposits containing marine or estua-rine fossils as thanatocenosis associations transported by ash flows and the existence of typically continental facies and fossils along the present Valdivia coastline support the notion that the entire above-mentioned deposits, previously described as marine, are in fact fluvial, lacustrine and paludal in origin. If sea level was somewhat higher than today, then active sedimentation took place as the coastline was several kilometres further to the west. This agrees with the probable existence of a fluvial plain several kilometres wide, where ancient rivers meandered in more or less parallel directions to the coast and where hyperconcentrated flow deposits would have moved southward from the mouth of the Imperial River.

The colonization of Chile by *Homo sapiens* at the Pleistocene–Holocene boundary (L.N., & M.G.)

In this section, we review key sites of the first human occupations from the extreme arid north (18°S) to subantarctic south (56°S) and document their response to palaeo-environmental variability. We focus on the Pleistocene–Holocene boundary, a specific time with very favourable environmental conditions for human colonization in southernmost South America.

Arid and semi-arid regions

Scarcie evidence exists for associations between human occupation and late Pleistocene faunas in the arid environments of Chile, despite widespread presence of the latter. For example, unpublished recent results from the Pleistocene lacustrine Calama basin document the presence of equids, camelds, megatheriids and macrauchenids (N. Hermosilla, pers. comm.). Other evidence for megaunal presence includes a milodentid unearthed in the highlands surrounding Antofagasta de la Sierra (C. Aschero, pers. comm.), an equid from Salar de Surire in the High Andes of Arica (C. Santoro, pers. comm.) and a megatherid from the *Prosopis tamunaro* forests in Pampa del Tamarugal (Casamiquela 1969–70). Cooler and moister climates around 14 000–11 000 yr BP are thought to have occurred at the Barro Negro site in the Argentine Punta, precisely the climate associated with extinct American species of equids, replaced by camelds c. 11 000 yr BP (Fernandez et al. 1991).

One of the few sites in northern Chile where associations between humans and megafauna have been established is along the pre-Andean range of the Tuina Hills in the vicinity of the Calama basin. Here, triangular projectile points (the ‘Tuina tradition’) are in the same stratigraphic position as American horse remains, with 14C dates between 11 700 and 11 200 yr BP. Of the 179 classifiable lithic artifacts discovered, two are bifacial triangular points made of allochthonous rocks together with 3879 camelid bone fragments. This find probably represents the last remnants of a Pleistocene fauna (Nunez et al. 2002).

The vast expanses of Puna and southern Altiplano (southern Bolivia, NW Argentina and northern Chile) were initially occu- pied by High Andean hunters of the Tuina tradition. Hunting sites were associated with modern camelds and were exclusively under rock shelters throughout the Andes between 12 050 and 11 000 yr BP, with strong evidence for a migrational regime across large areas (Aschero 1988; De Souza 2004; Nunez et al. 2002). Recent studies, however, have identified open-air camps at Punta Negra-1, contemporaneous with Tuina-1, on the southern margin of Salar de Punta Negra (2976 m) (Grosjean et al. 2005). A number of differentiated artifacts with a 'Fell' (from Fell's Cave in southern Chile) tradition, including one ‘fish tail’ point and bifacial lithics, probably formed part of a migratory population that took advantage of the palaeo-springs and wetlands (with Andean recharge) that cropped out here at < 3000 m (Grosjean et al. 2005). Fell points have also been found along the southern margin of the Argentine Puna (C. Aschero & A. Haber, pers. comm.) which implies that these peoples passed through the region at the end of the Pleistocene along the Andean highlands. At Punta Negra, lithic artifacts are intercalated with peat beads dated between 12 400 and 10 700 yr BP, including a 14C date on an adjacent hearth dated at 12 020 yr BP. Among a total of 964 unifacial lithic artifacts (scrapers, knives) only a single Fell point was found, along with four Punta Negra points similar to the Payjtan tradition, and three triangular points of the Tuina tradition (Fig. 12.10).

As the ‘Fell’ groups passed through the Atacama Desert, they overcame all the biological and extreme climate barriers of the continent (Villagran et al. 1983; Grosjean et al. 2005). After crossing over to the Pacific side of the Andes (Sandweiss et al. 1998), in a relatively short time these groups colonized palaeo-lacustrine basins rich in megafauna present in central-southern Chile (Nunez et al. 1994a, b). In all probability transients (Beaton 1991), they likely focused on assessing the availability of distant resources by passing through favourable sites along migratory paths. At Punta Negra, they made use of Punta Negra, they made use of Punta Negra, they made use of Punta Negra, they made use of Punta Negra wetlands rich in plant and animal (vicuñas) resources at a time when these were isolated from each other across the landscape. The record of large basaltic bifacial shards with expedite lateral retouches and nine bifacial points would imply, by and large, a group of low density, whose high rate of artifact output was enhanced by the close proximity of basalt outcrops (Grosjean et al. 2005).

The circum-Puna area experienced increased summer precipita-tion resulting in lake transgression throughout the region between 15 300 and 14 000 yr BP, with maximum levels present from 12 800 and 8900 yr BP, followed by lake collapse (Geyh et al. 1999; Grosjean et al. 2001). As previously stated, diverse interdisciplinary analyses have shown that the late Pleistocene of the Atacama Desert was a favourable environment for human settlement due to increased rainfall. It is thus almost beyond a doubt that the high abundance of Tuina tradition sites along the Andean front ranges resulted from the increased presence of wetlands adequate for the control of such a highly diverse territory.

Transitory Fell groups arrived in the semi-arid region further south by way of these same salar ‘corridors’ along the Andean front during the colder climates of the late glacial. This is indicated by the abundance of red soils, associated with wetter environments than today (Veit 1996), along the coast and lake shorelines, as well as the presence of relic forests and wetlands. These environments in particular would have concentrated resources such as megamammals, which further south occurred in association with *Nothofagus* parkland at the beginning of the Chilean Central Valley. An example is Quebrada Quereo, where shallow lake deposits contain an abundance of Pleis-tocene fauna (Nunez et al. 1994b; Jackson et al. 2004). This fauna disappeared with the abrupt onset of aridity at the beginning of the Holocene and coincided with the appearance of Fell point palaeo-indian predators (D. Jackson & C. Mendez, pers. comm.). The hunting of megamammals trapped among aquatic vegetation in these relict marshy areas was complemented by carrion feeding habits, eventually leading to extinction throughout the Central Valley in an amalgamation of climatic and cultural factors (Nunez et al. 1994b; Lopez et al. 2004).

Human populations that reached Quereo did so in two waves associated with the abrupt collapse of surrounding forests,
swamp and aquatic vegetation as well as encroaching semi-arid matorral in response to increasing aridity at the beginning of the Holocene (Heusser 1983; Villagrán & Varela 1990; Núñez et al. 1994b; also see section on Norte Chico). This ‘adverse’ scenario would have nevertheless allowed, and presumably even promoted, opportunistic strategies of megafauna hunting as resources became concentrated along remaining intermittent ‘ecorefugia’. The earliest sequence of Quereo I occurred between 14 000 and 13 250 yr BP. Among the megafauna present were equids, mylodontids, ‘paleo-llama’, swamp deer and gomphotheres (e.g. Stegomastodon and Cuvieronius, commonly referred to as ‘mastodonts’ in the literature, although they are not true mastodonts, i.e. Mammutidae). Smaller species of felids, canids, birds, rodents and amphibians have also been described. Among the species actually hunted are the remains of a horse skull with caved-in frontal nasal bones surrounded by lithics, cut bones, tapered artifacts, impacted, fractured and splinted long bones with polished and eroded ends, perforated horse vertebrae, microdiorite shards with natural edges (to expedite cutting), and planar blocks or platforms in situ surrounded by long bones and scarce local refuse (Núñez et al. 1994b; López et al. 2004).

The Quereo II occupation occurred between 13 100 and 10 600 yr BP with remains of human activity and waste deposited along the margins of a meandering river, under a warmer and drier climate, again repeating the strategic importance of these ecorefugia under drought conditions for both megafauna and their human hunters. American horse, deer, gomphotheres and mylodontids have all been identified at this site. Although Quereo is close to the coast, marine resources are rare and were probably gathered elsewhere; they include two shells of Concholepas concholepas (Chilean abalone) and a whale vertebra of Cetacea ballenidae. Human activity seems to have been transient and possibly tied to dry camps such as those present at Quebrada El Membrillo (Jackson et al. 2004). Cut bones and bone artifacts modified for polishing and hammering, bones fractured before fossilization, placement of flat slabs in situ around areas with highly concentrated equid bone remains, together with laminar lithics and shards with percussion waves and bulbs, including sharpened wood trunks, are all present at Quereo II and dated to 13 100 yr BP. The concentration of resources seen here, along with the species hunted, the lacustrine environment and the chronology, make this occupation contemporaneous with and similar in nature to the kill site located at Tagua Tagua (Montané 1968; Casamiquela 1969–70; Núñez et al. 1994a). Recent research along the upper reaches of the Quereo ravine has identified two additional occupations at Quebrada Membrielo, exposed by sediment deflation. The sites contain mylodontid bone with cut marks associated with surrounding flat slabs dated to 13 500 yr BP. A second layer, coeval with Quereo II, marks the presence of native horse and ‘palaeoliama’ associated with knives, scrapers, choppers and marginal scrapers, together with splintered and impacted bones.

As previously stated (see Norte Chico section), the climate became increasingly more arid after the Quereo II occupation, with 14C dates from immediately overlying peat layers dated to 10 600 yr BP. Overall, this would imply a collapse in the way of life of these ‘Fell’ groups, specialized as they were in the exploitation of these ecorefugia that began to disappear at a rapid pace throughout the interior basins of the central graben, as evidenced from Quereo and Tagua Tagua. This more than likely contributed to the general extinction of the megafauna at the end of this ‘post-glacial’ crisis (Varela 1976; Heusser 1983, 1990b; Villagrán & Varela 1990; Núñez et al. 1994a, b; Markgraf & Kenny 1997).

The fertile Central Valley

Exceptional lacustrine deposits from Lake Tagua Tagua (34°30’S) were discovered containing remains of megafauna in association with cultural remains, all initially dated to 13 250–12 800 yr BP (Montané 1968). More detailed studies eventually revealed a major (TT-1) location with five loci of skeletal remains, and discovered another location (TT-2) with nine loci including the remains of ten gomphotheres (juveniles and adults) all hunted and processed in situ, along with scarce horse and swamp deer remains. Many of these prey items were dismembered and piled onto ancient lacustrine beaches, along

Fig. 12.10. Projectile points representative of the Punta Negra site: (a, b) triangular points of the Tuina tradition; (c) a ‘fish-tail’ point of the Fell tradition.
the late Pleistocene–early Holocene transition (10,600 yr BP) loss of resources altering proboscidean behaviour, were also sites in North America between 27°N and 33°N. Hence, the synconcentration of these species in the few collapsing ecorefugia (Bryan 1975; Correal 1981; Haynes 2002). The eventual condition have been documented at numerous other ecorefugia as well for the abrupt and widespread extinction of the megafauna hunting and butchering at Tagua Tagua-2. The Tagua Tagua-2 ‘mastodont’ killing and butchering campsite. Inset: A ‘fish-tail’ Fell tradition point associated with megafaunal hunting and butchering at Tagua Tagua-2.

The latest Pleistocene marked the onset of palaeolake desiccation along the southern part of the central graben. Further south, as piedmont glaciers retreated up-valley, many of these regions were subsequently reoccupied by temperate rainforests (Moreno 2000; Moreno et al. 2001) and a new drainage network of rivers and lakes became established along the Patagonian steppe (Tatur et al. 2002). This scenario is associated with Monte Verde near Puerto Montt, one of the earliest human occupations in South America, with cultural layers spanning from 15,100 to 14,200 yr BP (Dillehay & Collins 1988). Vegetation at the time consisted of wet and cold southern rainforests rich in conifers, ‘avellano’ (Gevuina avellana) and southern beech (Nothofagus), and the site has a collection of seeds, tubers, nuts and wild berries, indicating proximity to a wetland (Dillehay 1989; Heusser 1990a). Located along the margin of a small tributary of the Maullín River, the Monte Verde campsite was composed of 12 cabin foundations, with rectangular floor plans fixed with stakes and associated with communal hearths and fireplaces. Monte Verde represents a semi-stable and diversified adaptive culture without dependence on seasonal fluctuations. A wide range of resources were utilized, including fish, shellfish and marine algae obtained from the coast c. 80 km distant from the site. Megafauna present were limited to gomphotheres and ‘palaeolama’ as well as other smaller game, with pre-Clovis and Fell bifacial foliaceous lithic points similar to the Jobo tradition of northern South America (Dillehay 1989). An earlier event, with formed lithic artifacts dated to 33,000 14C yr BP, located some 100 m from the classic site, is considered as a working hypothesis (Meltzer et al. 1997).

**Finis Terrae: colonization of Fuego–Patagonia**

Both small and large grazing herbivores found attractive habitats in the open and cold moist grasslands present at the threshold of human occupation of Fuego–Patagonia. Plant macrofossils analysed from ground sloth dung in Mylodon Cave dated to c. 14,400 yr BP indicate the presence of Cyperaceae, Juncaceae, grass and forbs (Moore 1978; Borroto et al. 1998). This moist grassland gave way, however, to a more xeric steppe between 13,000 and 11,000 yr BP with overall warmer temperatures (Markgraf 1988). Under environmental pressure, populations of large herbivores collapsed due to intense hunting by opportunistic hunters of the Fell tradition, which eventually became capable of colonizing the very southern tip of South America (Massone 1996; Borroto et al. 1998).

Optimum conditions existed for human existence at the end of the last glacial in Fuego–Patagonia: abundant game, both modern and extinct, extensive grasslands, water, prime material for lithics, and certainly the rich coastline. Fell’s Cave (at Rio Chico), surrounded by herbaceous steppe, is one of the most representative sequences of southermost hunter-gatherers (12,800–11,100 yr BP) (Bird 1988; Markgraf 1988). Underneath this rock shelter, American horses, mylodontids and modern camels have been preserved associated with excavated hearths and the remains of domestic artifact preparation typical of the Fell tradition, such as the ‘fish-tail’ points, long-frontal high-ridgeback scrapers, retouched unifacial shards, polishers, bone tools, knives, disc-shaped polished rocks and red pigments.

Other groups belonging to the Fell tradition came across a rock shelter at Tres Arroyos, in northern Tierra del Fuego. Both horse and mylodontids were consumed here between 12,720 and 11,750 yr BP in association with bifacial and unifacial artifacts, including several different scrapers with retouched edges (Jackson 1987; Massone 1987, 1996). Several hundred kilometres away at Cueva del Medio in the Ultima Esperanza area (Cerro Benitez), located only 1000 m to the shore of a palaeolake clearly sensitive to seasonal drought. An eventual return to wetter conditions (previously discussed) led to the burial and preservation of the evidence for human occupation at this site (Núñez et al. 1994a).

Judging by the lithics present at TT-1 (50 units) and TT-2 (79 units), both from open-air camps, and setting aside considerations regarding carrion feeding, the three Fell points (made of transparent quartz crystals) and numerous artifacts found provide unequivocal evidence for hunting and food processing (Fig. 12.11). Among these are unifacial implements (recoirs, side scrapers, knives, discoidal scrapers), hammering stones, an engraved gomphothere tusk dart and other bone artifacts, along with multiple cut and fractured bones, all indicative of processing in situ and accompanied by waste microdebitage (a by-product of on-site blade sharpening). All of these items were found in a single layer at TT-2 dated to 11,900–11,650 yr BP (Fig. 12.11). When compared to the earlier dates of TT-1 (13,250–12,800 yr BP) a pattern of lake regression becomes evident, with the earlier site located on a higher beach and the younger, lower site towards the middle, following overall increased aridity in the region. The dates are also in full agreement with the abrupt and widespread extinction of the megamammals at the end of the Pleistocene by c. 11,000 yr BP (Varela 1976; Núñez et al. 1994a, 2001).

Prolonged droughts during the Pleistocene–Holocene transition have been documented at numerous other ecorefugia associated with proboscideans in both North and South America (Bryan 1975; Correal 1981; Haynes 2002). The eventual concentration of these species in the few collapsing ecorefugia attracted palaeo-indian hunters to Tagua Tagua who specialized in large prey, in a similar fashion as at the classic mammoth sites in North America between 27°N and 33°N. Hence, the syngenetic events documented at Tagua Tagua, with the dramatic loss of resources altering proboscidean behaviour, were also synchronous with the North American Clovis groups during the late Pleistocene–early Holocene transition (10,600 yr BP) when abrupt onset of drought generated loss of wetlands and arboreal habitats (Haynes 1991, 2002). Thus, both Fell and Clovis cultures may have exploited similar ecocatastrophic scenarios, when opportunistic hunting strategies coupled with stressful palaeoclimate conditions between 12,800 and 10,900 yr BP quickened the demise of an already collapsed biomass of megamammals.

Fig. 12.11. The Tagua Tagua-2 ‘mastodont’ killing and butchering campsite. Inset: A ‘fish-tail’ Fell tradition point associated with megafaunal hunting and butchering at Tagua Tagua-2.
south of Myodon Cave, another group of people piled up, amid campfires, bones of both extinct and modern fauna between 13 100 and 10 750 yr BP, in association with cutting tools and scrapers including ‘fish-tail’ points of the Fell tradition (Nami & Case 1983; Miotti & Cattaneo 1997; Politis 2002; Jackson et al. 2004; López et al. 2004). The second well-represented wave of immigrants occurred between 13 000 and 11 000 yr BP, represented by typical Fell components and recognized throughout Chile and Argentina from 20°S to 56°S latitude. These occupations may have been more effective through the greater availability of materials (Gamble & Soffer 1990; Borrero et al. 1998) with the ‘popularization’ of Fell points and large unifacial shards throughout southern South America occurring over no more than a millennium.

By 13 000–11 000 yr BP all of the analysed sites indicate occupation by an ever-growing and more diverse assemblage of peoples from 18°S to 56°S latitude, giving way to the coexistence of different cultural traditions. Cultural responses to ever-increasing aridity are known from the Tuina tradition (triangular points) associated with modern faunas (camelids) and very scarce megafauna (equids). This culture formed an effective, stable and lasting presence under wetter climate regimes in the central Andes. Coeval with the Tuina tradition is the Fell tradition, which in contrast to Tuina, was formed by transient migratory groups exploiting point resources along the Andean piedmont (e.g. at Salar de Punta Negra) (Marshall 1993). The similarity of the points found at Punta Negra with the Payjan and classic Fell points of southernmost Chile (Fig. 12.10) would indicate that migration of the latter ensued from northern Peru, establishing contact with different occupations, such as Tuina, which were exclusively found only in the circum-Puna region (Dillehay et al. 2002; Chauzuch & Pellegrin 2004; Grosjean et al. 2005).

In a similar fashion to the way that resources today increase from north to south, a colonizing wave would have orientated itself towards more productive and wetter areas with better, more continuous resources. From the scarce presence of Fell cultures in the northern desert, to ecorefugia exploitation of semi-arid and central Chile, it was in the Central Valley where these cultures seem to have flourished the most, most likely due to increased availability of big game, both extinct and modern. As climate became drier at the end of the last glacial, mega-faunal resources became more concentrated but did not last very long. Exploitation eventually became untenable, making the paleo-indian way of life considerably more difficult. The identification of an intense Fell occupation at Lake Tagua Tagua affords an exemplary account of how proboscidean populations under paleo-environmental stress collapsed and became extinct across the threshold of the Pleistocene–Holocene boundary. Subantarctic steppe also became quickly colonized by Fell hunters, who then suffered abrupt collapse as climate became increasingly arid inducing an abrupt crisis in wild animal fodder.

Maritime cultures quickly became established between 13 000 and 11 000 yr BP along the entire length of the coast, evidenced by the Huentelauquen and Acha traditions (Muñoz & Chacama 1993; Llagostera et al. 1997). Along the Andean Precordillera, occupations of the Fell tradition applied Subandean cultural responses that were dependent on habitats found at lower altitudes (Saavedra & Cornejo 1995; Stehberg 1997). This included the Arica highlands, where a different response from the Tuina tradition was present (Santoro 1989). We thus propose that during the late glacial all the resources of the country were being used synchronously by hunter-gatherers (Muñez et al. 2001). The high variability inherent to abrupt climate changes at the Pleistocene–Holocene transition never constituted a barrier to colonization by the first Homo sapiens. By undergoing severe cultural adjustments, including applying opportunistic strategies, humans proved very capable of enduring colonization in the complex and dynamic post-glacial landscapes of southernmost South America.

Future challenges for Quaternary studies in Chile

Ongoing Quaternary research in Chile is highly diverse and cross-disciplinary. Among the major challenges posed for any overview of the subject is the need to provide a process-oriented integration of diverse settings, interpretations and scenarios that have been proposed previously, in the light of the ever-growing dataset of global palaeoclimate changes. A significant amount of interest is now invested in understanding the mechanisms involved in the generation and propagation of palaeoclimate signals at a global scale. Chile’s natural geography is key for testing the contribution of low, mid- and high-latitude palaeoclimate processes to global climate change. This is a major and ambitious enterprise, as palaeo-ecologists and palaeo-climatologists leave behind the purely descriptive phase and lead the Quaternary sciences into a more predictive stage. To accomplish this, the development of high-resolution ‘bullet-proof’ chronologies is of utmost importance. With the advent of AMS radiocarbon dating as well standardized cosmogenic dating techniques, many records are now datable with considerable precision. In the long run, the more precisely we know our recent past, the greater our confidence will be in model predictions and outcomes.

This chapter has also made evident some of the more poorly known regions of Chile in terms of past climate and vegetation change and in particular over the last glacial–interglacial transition (the last 25 000 years). Of these, the Norte Chico and central Chile stand out, especially the former. Hence, many of the scenarios for plant colonization and the origin of modern-day plant associations proposed by biogeographic evidence (e.g. see section on vegetation diversity and change in central Chile) will remain mostly circumstantial until falsified or confirmed by geohistorical records. Greater integration between marine and land records will be essential for palaeo-climatologists working
in northern/central Chile. The ideal situation is for data from one type of record to be used in developing working hypotheses for other kinds of records. Outcomes may not always be predictable, however, and previously unsuspected scenarios of either climate or vegetation change may arise.

Even more conspicuous is the lack of data from older portions of the Quaternary. Here, we have presented evidence of interstadial conifer forests and previous interglacial (although undated) deposits near Valdivia. In northern Chile, we note that marine records (Lamy et al. 2000) and records of salt facies changes at Salar de Atacama (Bobst et al. 2001; Lowenstein et al. 2003) provide evidence for climate change over the last 80 to 100 ka. No records of that antiquity exist in southern Chile: the oldest known date back to 50 ka BP (L. E. Heusser et al. 1999; Villagrán et al. 2004b). To date, there are no other precisely dated records of either past climate or vegetation change in Chile that cover older time periods.

An exciting new area of research has been the extraction of fossil DNA from animal coprolites preserved in middens and cave deposits (Hofreiter et al. 2000, 2003; Kuch et al. 2002; Hadly et al. 2004). By merging powerful laboratory DNA extraction techniques with well constructed records of past climate change, a strong potential exists for understanding not only present but also past population genetics and how modern biodiversity has arisen in the recent past.

Finally, Quaternary research is fast becoming a mature science discipline in Chile, with many new researchers devoted to developing precisely dated geohistorical records. This is perhaps the greatest difference from the situation almost 30 years ago, when Paskoff (1977) published his initial ‘state of research’.

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Field excursion from central Chile to the Atacama Desert

WES GIBBONS & TERESA MORENO

This final chapter is designed primarily for the foreign visitor to Chile who wishes to gain a broad overview of the field geology and scenery of central and northern Chile. We hope that it will also aid an appreciation of the nature of human interaction with this climatically and topographically challenging area, from the early settlers to the modern mining industry. This is a traverse along and across one of the world’s classic compressional ocean-continent subduction zones, where strong coupling between the oceanic and continental plates is linked to active mountain building, dramatic scenery, and frequent earthquake and volcanic activity. The drive moves north from Santiago, which lies in the forearc Central Valley west of the South American Southern Volcanic Zone, into one of the flat-slab Andean segments. Within this flat-slab zone we divert east to describe a west-to-east across-strike traverse from the coast into the High Andes where, because of the low dip of the subducting plate, volcanoes are absent (Chapters 4 and 5). Returning to the coast, the drive continues north into the latitude of the South American Central Volcanic Zone, with its superbly developed line of modern volcanoes in the hyperarid high Atacama Desert. A second west-to-east traverse from the coast to these volcanoes illustrates how the modern continental forearc region is segmented into a series of mountain belts and intervening basins, each with its own distinctive scenery and geology. Additional details on the rock units and places visited, and the tectonic settings of ancient and modern Chile, can be found by researching the previous chapters.

The route described runs from Zapallar, with its Mediterranean climate on the coast north of Santiago, to San Pedro de Atacama, which lies deep in the desert close to the Bolivian border (Figs. 13.1 & 13.2). A northern route has been chosen primarily because, given the dry climate, rock exposure is better and the spectacular landscapes of the Atacama Desert are a likely highlight of any visit to Chile. Those with additional time could head south to enjoy the dramatic geomorphology of the Central Valley and southern Andean volcanoes, as well as examine excellent coastal exposures such as the acretionary complex exposed south of Pichilemu, but anyone undertaking fieldwork in southern Chile must deal with thick vegetation and a damp climate.

Travel advice

Driving in Chile can be challenging. Firstly there are the distances: the length of time it is necessary to spend behind the wheel should not be underestimated. Although our one-way route crosses only around one-third of Chilean latitude, it includes detours away from the main south-north highway and involves over 2500 km of driving, excluding any travel back south. Driving the Panamerican Highway (the Panamericana) north of La Serena can be a dangerously somnolent experience, not least because high speeds on this truck-congested two-lane desert road are illegal, so particular care needs to be taken not to sleep at the wheel. Speeding is strictly enforced: if you drift above the speed limit during a long day’s travel you are very likely to be caught by police radar. Improvements are being made to the road, but these involve roadworks which can be epic in scale, involving long diversions on dusty dirt roads, and causing impressive tailbacks. Comfort and refuelling stops are few and far between, especially in the extreme desert north of Copiapó, so always carry plenty of water, food and fuel.

Our route mostly avoids unmetalled roads, with the notable exceptions of the tracks around San Pedro de Atacama and up the Hurtado Valley. The excursion over the mountains separating the Hurtado and Elqui valleys in particular is something of an adventure, and should not be attempted in bad weather, a small car, or by anyone with an excessively nervous disposition. We drove the entire route in a locally hired two-wheel-drive pick-up truck without incident, but confess that at times a four-wheel drive vehicle would have been more comforting. Although it is theoretically possible to make the journey in a saloon car, especially one with relatively high clearance, we would not recommend it. You are unlikely to encounter much rain most of the year but it is worth noting that the wettest month between Santiago and La Serena is usually June (the winter months have the highest rainfall), whereas in the far north around San Pedro de Atacama heavy rainstorms can occur between January and March (the summer months have the highest rainfall). You will almost certainly experience the Camanchaca coastal fog which can be a serious hazard on the Panamericana. If you are unlucky enough to coincide your visit with a significant earthquake, then disruption to the transport system is to be expected. These potential hazards, however, are no more than those met on any long distance drive and should not deter the traveller: the journey is unforgettable.

The route described starts from Zapallar, an attractive coastal village lying around 180 km NW of Santiago and a good place to rest after a long flight (Fig. 13.2). From here the itinerary follows the Panamerican Highway northwards, with diversions eastwards into the Hurtado and Elqui valleys, and westwards to the coast between Antofagasta and Tocopilla, finally traversing east to reach San Pedro de Atacama in the heart of the desert. The south to north journey can be done comfortably in seven or eight days, excluding the arrival day and the return journey from San Pedro. Recommended overnight stops are: 1. Zapallar (or Papudo); 2. Ovalle (or Termas de Soconos) – this night stop can be excluded if time is short (see below); 3. Hacienda los Andes; 4. Pisco Elqui.
Fig. 13.1. (a) Map of northern and central Chile showing regions I–X and main population centres (those with contributing authors are underlined, excluding Punta Arenas which lies in southern Chile). For southern Chile geography and geology (administrative regions XI and XII) see Fig. 2.1. (b) Plate tectonic setting of northern and central Chile. (c) Andean copper deposits illustrating the diachronous shift of mineralization through time. Boxed area refers to area relevant to the excursion described in this chapter.

(or Vicuña); 5, Bahía Inglesa (or Caldera or Coquimbo); 6, Mejillones (or Antofagasta); 7 and 8, San Pedro de Atacama. Those with more time should consider an extra night at the Hacienda los Andes (to allow a day exploring the hills), a night in Calama (to allow a visit to the Chuquicamata copper mine), and a further day or two at San Pedro (to allow further exploration of the high Atacama, such as El Tatio Geysers). Those with no desire to drive for hours up the Panamerican Highway can instead fly between Santiago and Antofagasta, the visit then being split into a southern loop (Santiago to Hacienda los Andes and back) and a northern loop (Antofagasta to San Pedro and back) (Fig. 13.2).
Fig. 13.2. Route map for excursion.
The most time-efficient way to make the full journey is to drive the outward journey and fly back, leaving the hire car at Antofagasta. If you choose to drive back from San Pedro to Santiago, give yourself at least two (extremely long) days. Finally, accommodation and car hire are best booked in advance, especially in peak season: the usual guidebooks (such as the Rough Guide, Lonely Planet and Let’s Go) or any internet search of the places where overnight stops are recommended will reveal the most obvious available options. Pre-booking at the Hacienda los Andes is essential, and San Pedro fills up regularly making a room commonly difficult or even impossible to find. In general, to best appreciate the serenity of the Chilean landscape and the clarity of the night skies, our preferred option is to stay in smaller places away from the main towns: but this is a personal choice.

To start the journey from Santiago international airport, drive east on the orbital motorway (the Avenida Circunvalación Améxico Vespucio) to join the Panamericana Norte and turn north to follow the Central Valley, with the Coastal Cordillera on the left and the northern end of the Southern (South American) Volcanic Zone forming the High Andes on the right. Out in the ocean to the west at this latitude the collision of the Juan Fernández Ridge with the modern Benioff zone has created a geotectonic boundary between normal subduction (to the south) and “flat-slab” subduction (to the north). Zapallar lies a couple of hours drive from Santiago on the coast west of the Panamericana, and can be reached either from the south, via Cachagua, or from the north via Papudo. If the arrival is early enough to allow examination of the coastal exposures around Zapallar on the same day, and an early energetic start can be made the following day, then one can save a day by driving from Zapallar to Hacienda los Andes, thus combining the first two days described below.

Zapallar to Ovalle (290 km)

The first part of the journey provides an introduction to Coastal Cordillera plutonic geology at Zapallar, followed by an easy 250 km drive to Termas de Sosoces. The route then turns eastwards away from the coast for 40 km to an overnight stop in the town of Ovalle, via the famous petrographic site of the Valle El Encanto. For those interested in experiencing the thermal baths of Termas de Socos, this is an alternative, unachronistic and peaceful place to spend the night, in which case either backtrack from the Valle El Encanto, or save that locality for the next morning (and instead consider a visit to the Fray Jorge National Park – see below – or just relax at Termas de Socos).

Zapallar coastline

Several hours can be spent admiring the coastal exposures of Mesozoic intermediate to acid plutonic arc rocks around Zapallar. Here, deep within the Coastal Cordillera (see Chapter 4), are some of the best, easily accessible examples of intrusive magma-mingling textures to be found in the world. Furthermore, the overprint of shearing deformation within the hot arc root has produced perfectly exposed transitions from undeformed plutons into orthogneisses. On the south side of the beach (Fig. 13.3a) there are abundant granitoid rocks intruding a diorite–granodiorite complex (DGC). Crossing to the NE side of the beach this DGC becomes the dominant unit, and contains late magmatic aplitic and pegmatitic pods with coarse feldspars and amphiboles. In many exposures there is also the obvious overprint of a prominent north–south steep foliation, commonly transposing the original plutonic textures into intensely deformed mylonitic orthogneisses. In areas of low strain are preserved classical magma-mingling textures, with the more mafic plutonic rocks (such as meladiorites) sometimes clearly chilled against more evolved lithologies (Fig. 13.3b). Follow the coast westwards to cross a mingled diorite–granodiorite complex cut by granitoid veins, dykes and larger bodies (this is the granite that is more prominent on the south side of the beach). The whole array of comagmatic and metamorphic textures can be examined by walking westwards from the north side of the beach towards Isla Seara, which comprises mostly pale granitic rocks. Further along the coast are more, equally spectacular, exposures of enclave-rich, mixed, mingled, and mostly more mafic components of the DGC.

Zapallar to Termas de Socos

The coastal road runs 25 km north from Zapallar past Papudo to the Panamericana Norte (PN). From here it is a 220 km drive on the PN to Termas de Socos, crossing from the semi-arid/subhumid climate of central Chile into the thinly populated distinctly more arid lands further north. In terms of plate tectonic setting the journey stays within the flat-slab segment of the Chilean subduction system, a zone marked by a non-volcanic gap in the Andes to the east and by the greatest number of epithermal precious metal deposits in the country.

The route initially passes huge spreads of Quaternary sands and gravels resting on plutonic rocks of the Coastal Cordillera (CC) where it intrudes early Mesozoic sedimentary and volcanic rocks. Representatives of these latter rocks are exposed on the coast at Las Molles (an optimal diversion off National Park) marine mudstones and sandstones deposited by contour and turbidity currents and belonging to the Los Molles Formation were deposited after the change from Triassic extension to the initiation of Andean subduction in Jurassic times.

North of Los Yiles are prominent dune fields and spreads of Quaternary alluvium against an inhospitable backdrop of Coastal Cordillera hills. The deeply weathered landscape becomes increasingly desert-like as one approaches Termas de Socos, with the highway roadcuts exposing heterogeneous net-veined and dyked plutonic igneous complexes. At km371 on the Panamericana the Termas de Socos and Ovalle turnoffs lie to the right. Termas de Socos is one of the better known of the 275 or so hot springs recorded in Chile, and, with potable waters, is one of the more successful of the 32 that are commercially exploited. The springs are hidden in a sheltered valley approached via a dirt road that winds through exposures of Quaternary gravels.

An early check-in at Hotel Termas de Socos (http://www. termasoces.cl/) may allow time to visit the relict Olivillo forest in the Coastal Cordillera at the Fray Jorge National Park (check opening hours first: they can be inconvenient). This curious, isolated woodland, which depends on the Camanchaca coastal fog for its continuing existence, lies 20 km further north on the Panamericana from Termas de Socos, then another 20 km on gravel tracks west to the park entrance. After paying the entrance fee, the sometimes steep and rough track continues for another 10 km to the woods, some of the plants in which are not found for another 1600 km to the south (see Chapter 12).

Termas de Socos to Valle El Encanto

From Termas de Socos the route turns east across the flat-slab ‘zone of transverse river valleys’ where no Central Depression/ Valley is present. The excursion initially follows the Limarí Valley, one of the most prominent of these ‘transverse valleys’, to traverse the central part of the flat-slab zone. Crossing the Jurassic then Cretaceous plutonic belts (and their Miocene–Quaternary sedimentary cover) that characterize much of the western half of Central Chile, the traverse moves into increasingly mountainous scenery to reach outcrops of rocks produced within the Tertiary magmatic arc. Due to the low-angle subduction of the underlying plate in this part of Chile, not only is there no Central Depression/Valley but neither is there a modern volcanic arc, unlike to the south (from
around Santiago southwards), and north (visited later in the excursion).

The Valle El Encanto (Enchanted Valley) is an important archaeological site that lies 20 km east of Termas de Socos. From the Panamericana take the paved road east towards Ovalle, following the Limari Valley. The road crosses a wide Tertiary pediplain now incised by river valleys containing Quaternary alluvium. Evidence for a long history of Tertiary continental deposition and erosion under semi-arid to hyper-arid conditions (Chapter 12) can be seen from around this latitude northwards all the way to southern Peru, a distance of over 2000 km.

At 15 km from the Panamericana turn right onto a dirt road south towards the signposted Valle El Encanto, passing vineyards (right) enclosed by walls built from rounded fluvial detritus. The wines grown on these fluvial soils in the Limari Valley, one of the most northerly vinecultures in Chile, have recently attracted much international attention. The Valle de Encanto (Fig. 13.3c) lies 5 km from the main road and offers good exposures of pale, homogeneous granite cut by thin felsic dykes and containing rare mafic enclaves. This intrusion is one of a series of batholiths that define the position of the magmatic arc in northern Chile during Early Cretaceous time. The main interest here, however, is the evidence for El Molle culture, the people of which lived in the period around 1900-1300 years ago in the valleys and coastline between the rivers Copiapó and Choapa. These people grew crops and herded animals from the mountains in the summer to the coastal valleys in the winter, with the permanent waters and distinctive ambience of the Valle El Encanto probably creating a major cultural focus for the tribe. Examination of the rock surfaces reveals 'Limari style' human petroglyphs (circles and lines for facial features, closed mouth), rare rock paintings in red (pictografías), and piedras tacitas or morteros where the granite has been carved into cup-shaped depressions thought to have been used to grind food, paints and/or hallucinogenic drugs for religious rituals (Fig. 13.3d).

Valle El Encanto to Ovalle

Return to the main road and continue 12 km east to descend into Ovalle, a town built on Miocene and Quaternary gravels within the River Limari Valley. Ovalle is the economic centre of the agriculturally rich province of Limari, a thriving, successful town despite the usual Chilean history of urban earthquake damage. Although twenty-first century earthquakes here have so far (2005) been relatively minor, a destructive event of magnitude 7.1 occurred during the evening of 14 October 1997 (local time) and was followed by around 150 aftershocks. This intraplate earthquake (Chapter 10), which occurred within the subducting Nazca Plate, was big enough to be felt as far away as Buenos Aires and was most damaging here in the provinces of Elqui, Limari and Choapa, where eight people were killed and thousands adversely affected, mainly by housing damage and landslides.

Ovalle to Hacienda los Andes (85 km)

Ovalle to Las Pintinacas

The route continues the across-strike traverse towards the High Andes begun previously at Termas de Socos, moving eastward from Cretaceous to Palaeocene exposures that emerge from beneath the Neogene and Quaternary cover in the valleys to form prominent mountain ranges. Leave Ovalle to the north on the La Serena road, climbing back onto the Miocene fluvial succession, then branching NE (right) onto the recently improved Samo Alto road. Some 18 km upstream from Ovalle the road reaches the Recoleta dam, beyond which it follows the Hurtado River valley. The road initially climbs and curves up through the conglomerate succession to reach extensive exposures of Lower Cretaceous volcanic rocks, such as those beneath the village of Tabiqueros (8 km after the dam). Beyond Baden Pangue the road once more climbs on to Miocene conglomerates before descending to Quaternary gravels and crossing the river.

A few kilometres further, on the approach to Samo Alto (20 km after dam), after passing the junction to the gold mining town of Andacollo (left), there are extensive views east of the volcanic succession overlain by thick conglomeratic sediments. A further 3 km beyond Samo Alto park (right) at the signposted stop of Las Pintinacas (on the tourist archaeological art route), follow the path up 100 m to view the valley exposing a microandesitic intrusion in which the artwork is preserved, with conglomerates exposed to the east (left). Continue climbing up the valley through the cactus desert for 10 minutes beyond the end of the path to find a prominent exposure of conglomerate cut by a dyke with chilled margins. The terrestrial detrital sediments here are mid-Cretaceous in age and record the infilling of the backarc basin (see next locality).

Las Pintinacas to Hacienda los Andes

Continue on the road eastwards for 5 km, passing La Aquala then turning left by the San Pedro de Pichasca sign onto the rough track to Monumento Natural Pichasca Pueblo de San Pedro (open 08:30-18:00 hours). Passing the riverside campsite the track heads right towards Al Monumento and ends at the car park from where a signposted path leads for 1-2 hours around the site. Although sometimes promoted as the 'Jurassic Park' of Chile, the terrestrial sediments of the Viñita Formation exposed here are actually Cretaceous in age, and preserve striking examples of silicified tree trunks (Fig. 13.3). The exposures comprise one of the most important palaeontological sites in the country, yielding the first fossils of the dinosaur Antarctosaurus wichmannianus. The sediments of the Viñita Fm comprise breccias, conglomerates, sandstones, siltstones and andesitic volcanic rocks deposited in a continental basin that developed to the east of the Coastal Cordillera between latitudes 22°S and 31°S (i.e. from south of Termas de Socos to Tocopilla, the most northerly point of this excursion). The exhumation of the earlier Mesozoic arc rocks to the west, and subsequent deposition of thick sedimentary successions in the former backarc region, took place in mid-Cretaceous times after a long period of extensional tectonics was terminated by a pulse of compressional deformation (see Chapter 3). The newly deposited Late Cretaceous successions were, however, themselves soon to be uplifted: subsequent latest Cretaceous and/or earliest Palaeocene inversion of the basins such as that containing the Viñita Fm shed detritus into new depocentres developing further east. Sedimentation in these more easterly basins, where thick early Palaeogene volcanic and volcanioclastic successions were deposited, was itself later terminated by a major Eocene deformation event (Chapter 3).

There are extensive views south (Fig. 13.3e) across hills of these continental successions (intruded by Palaeocene granites), as well as east up the Hurtado Valley to the snow-capped High Andes. Finally, the walking tour includes a visit to the cave known as La Casa de la Piedra where prehistoric remains have been discovered of a hunter-gatherer human group that existed here up to around 10,000 years ago. The cave lies in a prominent c. 20-m-thick feldspar porphyritic silt resting on red siltstones and sandstones.

Return to the Hurtado Valley road and continue east for 28 km. The gravel road passes through increasingly rough country as it climbs up-stratigraphy through the Late Cretaceous-Palaeogene detrital and magmatic succession, to reach the Hacienda los Andes (http://www.haciendalosandes. com/esp/hacienda.htm).
Hacienda los Andes to Pisco Elqui (90 km)

Warning: this 3-hour journey should be driven in a high-clearance vehicle, preferably four-wheel-drive, and in dry weather. Check with the owners of the Hacienda for advice. If in doubt, backtrack... 

The local Palaeogene geology is best visited on the network of walking trails behind the Hacienda los Andes on the south side of the valley; as the drive this day is short, there is enough time to spend at least the morning doing this. The dramatic contrast between the verdant Hurtado Valley and the surrounding deserts above the reach of irrigation by the river is strikingly appreciated by climbing these trails. A recommended walk runs south from the Hacienda up to the Cruz de Sur (Sierra Trail), passing through a sequence of volcaniclastic lavas and finer sediments, and porphyritic lavas, dykes and sills (Dónde Ana G.) Further up the dry stream valley is a former gold mine, whereas following a path to the east one enters the outcrop of a pale Tertiary granite intrusion with a porphyritic microgranite margin. The path eventually leads down to the river upstream of the Hacienda. On the return walk down the riverbank there are abundant boulders of various calcalkaline intrusive lithologies such as granitoids chocked with abundant comagmatic clutches. These are presumably derived from the Permian intrusive rocks of the High Andes Batholith which forms the mountains to the east (see Chapter 4).

From Hacienda los Andes continue east past Hurtado, then just beyond take a sharp left turn on the track to Vicuña (4 km); the easterly course of the Hurtado River here has been controlled southward by a former impositional mass of granitoid rocks to the east, in places rising to well over 4000 m a.s.l. The grinding, twisting climb out of the Hurtado Valley leads eventually to the high ground on the Hurtado-Elqui watershed, with the track staying on the same Palaeogene succession visited at Hacienda los Andes, with abundant granitoid boulders derived from the mountains rising to the east. To the west lies the Cerro Tololo Inter-American Observatory, situated at 2200 m and one of several astronomical observatories built in northern Chile to take advantage of the clarity of the night sky.

The subsequent long northward descent affords expansive views into the Elqui Valley (Fig. 13.3f), cut deeply into the Cretaceous and Tertiary sediments, volcanic and plutonic rocks of earlier, the outcrop of which runs north and south from the Hurtado Valley. Rising east and west of the town of Vicuña below are two outcrops of Early Jurassic granitoid rocks which represent some of the westernmost exposures of the High Andes Batholith (Chapter 4). After 43 km from the Hacienda los Andes, in the outskirts of Vicuña, the mountain track joins the main east-west road running along the Elqui Valley. Turn right (east) and follow this road over Quaternary gravels with views east behind a ridge of Mesozoic outcrops into the subduction-related Early Permian plutonic and metamafic rocks that form much of the High Andes Batholith. At 20 km from Vicuña one reaches Rivadavia: turn right (south), continuing for a further 4 km to locate the next stop: a parking space on the left below prominently bedded cliff exposures just before Tres Cruces.

At this locality the Mesozoic backarc deposits of the Early Jurassic Tres Cruces Formation can be examined (see Chapter 3): well bedded pink-red marine phricic limestone with biwal shell fragments, purple mudstones, and coarse feldspathic sandstones. Beds with erosive bases and coarsening-upward textures record turbiditic sedimentation in this backarc basin setting, with an increasing influence of volcanic detritus interfering with background limestone deposition. The sedimentary succession here is eventually overlain by volcanic rocks (Punta Blanca Member), produced during the resumption of oceanic plate subduction, a tectonic setting that has prevailed ever since (Chapter 4). The succession is intruded by dykes (a prominent example can be found 200 m uphill), and affected by copper mineralization (including malachite, azurite and tarnished chalcopyrite). There are views west across the valley where the continental Triassic (Las Brees Fm)-Jurassic Jurassic sedimentary succession rests on the Permian basement.

From here the road continues 10 km south to Pisco Elqui, through spectacular exposures of the granitoid basement rocks, criss-crossed by hypabyssal intrusives that form much of the High Andes between here and El Transito. 150 km to the north. The inhospitable desert mountain terrain, cut by the strikingly verdant ribbon of the Elqui Valley (Chapter 8), makes for memorable scenery, although it is difficult to get far away from the valley to examine the basement geology in any detail.

Pisco Elqui to Bahía Inglesa (520 km)

This long drive is essentially a linking day between the visits to the Chico Norte south of La Serena, and the extreme desert of the Atacama sensu stricto. The length of the drive can be ameliorated by exchanging a night in the tranquility of the Upper Elqui Valley for one in the town of Vicuña (or the even more urban setting of La Serena). North of La Serena the journey follows the Panamericana, which in this area mostly runs well away from the coast. In terms of geotectonic setting the route moves across the northern part of the flat-slab Andean segment and across the transition into the more typical ocean-continent subduction zone that characterizes northernmost Chile. Bahía Inglesa lies on the coast within the Coastal Cordillera, around 100 km east of the oceanic trench, and 250 km west of the volcanic arc which reaches nearly 7000 m a.s.l. at Ojos de Salado, the highest active volcano in the world (Chapter 3).

Pisco Elqui to La Serena

The 110 km drive down the Elqui Valley from Pisco Elqui to La Serena passes back through a succession similar to that seen over the previous two days: Palaeozoic basement and Mesozoic...
cover to Rivadavia. Tertiary granitoid rocks intruding Palaeogene sediments and volcanics around Vicuña, and then back onto Early Cretaceous volcanic rocks intruded by granitoids. There are particularly good exposures of the Early Cretaceous andesites around (and after) a prominent reservoir c. 50 km from Pisco Elqui. After 10 km from the reservoir one passes the turnoff for Talcuna and Marquesa, a mining district known for its Cu–Ag mineralization of Lower Cretaceous volcanic rocks. A further 10 km downstream, the road passes exposures of granitic rocks intruding the Lower Cretaceous andesitic succession, beyond which the river valley widens and extensive spreads of Quaternary alluvium increasingly obscure the underlying geology.

La Serena to Copiapó

Arriving back at the Panamericana, which runs to the west of La Serena, turn north and drive along-strike for 182 km to Vallenar, following the same sequence of granitoid and volcanic rocks seen in the drive through the Lower Elqui Valley. To the east lies the important Domeyko mining district where mid-Cretaceous porphyry copper and precious metals lie along the same belt as the Andacollo mining district SE of La Serena (Chapter 6).

In the Huasco Valley at Vallenar there are extensive deposits of the Miocene continental sediments (Atacama gravels) seen earlier, and these (along with spreads of Quaternary sediment) continue to dominate the geology and landscape for 145 km further north to Copiapó. This part of the route passes to the west of the ghost silver mining town of Chañarillo, situated in the northern part of the Lower Cretaceous backarc basin (silver was key to the growth of Copiapó). The Lower Cretaceous rocks in this northern part of the basin lack the abundant lavas seen further south in central Chile, and are instead dominated by limestones of the Chañarillo Group (Chapter 3). The eastward migration of the volcanic arc into the marine backarc basin occurred during mid-Cretaceous time (Chapter 4), and consequent intrusion of plutonic (mostly dianotic) rocks into the Lower Cretaceous succession generated significant mineralization to produce the high grade Zn–Pb Maria Cristina deposit, which lies some 10 km east of the Panamericana, 80 km south of Copiapó. Further north, the recently (1987) discovery of the Candelería gold–copper skarn deposit (20 km SSW of Copiapó) has stimulated new mining activity in this area (Chapter 6).

In Copiapó, if time allows, there is a good collection of Chilean rock, mineral and fossil specimens on display in the geology museum in Calle Atacama, a few minutes walk from the town centre (closed lunchtimes and Sunday afternoons) (http://plata u de. u c. cl/minas/apuntes/Geologia/Museovirtual/ tur012a.html).

Copiapó to Bahía Inglesa

The final part of the drive follows the Panamericana for 75 km west from Copiapó down the Copiapó River valley (back towards the Jurassic rocks of the Coastal Cordillera) then north to Bahía Inglesa (turnoff just south of Calédra). As the road turns north away from the verdant agricultural Copiapó Valley, the Atacama Desert begins in real earnest: this is the Norte Grande. There is a tremendous contrast between the popular coastal resort of Bahía Inglesa and the hyperarid, inhospitable country inland. The flat, low-lying ground between Bahía Inglesa and Calédra includes outcrops of Miocene–Plioene marine sediments (Bahia Inglesa Formation and Aqua Amarga Beds), underlain by coarse continental gravels and overlain by continental and littoral sediments (Calédra Beds), deposited on seven or eight uplifted marine terraces (Chapter 3). These young sediments rest unconformably on a basement of Coastal Cordillera lithologies, easily accessible exposures of which can be examined along the low coastline at Bahía Inglesa. At this locality they are mostly dioritic plutons cut by dykes, some of which are composite with a basic centre and felsic margins. The major earthquake of 1819 produced a powerful tsunami which hit the coast here, destroying the port of Calédra just to the north. There was another powerful event in 1922, causing severe destruction in Copiapó, Vallenar and La Serena, and a less damaging reactivation in 1983, with this time only a minor tsunami being produced (Chapter 10).

Bahía Inglesa to Mejillones (664 km)

This very long drive through the desert north to Antofagasta and then up the coast to Mejillones, is alleviated by visits to the spectacular orbicular diorites north of Caldera, the basement rocks exposed in the highly scenic Parc Nacional Pan de Azucar, and the Mejillones Peninsula. Mejillones is a small, quiet coastal town; those preferring a more urban setting should stay in Antofagasta.

Bahía Inglesa to Chañaral

At 20 km north of Bahía Inglesa, turn left off the Panamericana to visit the Santuario Nacional de la Naturaleza Granito Orbicular. Drive down the gravel track to the coast 1 km west, continuing until as close to the beach as you can get, and climb down to the clean, smooth rock surfaces on the beach. The best exposures lie just above high water mark where a 15-m-thick vertical dyke-like and highly orbicular granitoid intrusion intrudes a more mafic and homogeneous country rock pluton. The orbicular structures appear as globules mostly several centimetres in diameter and lying within a more felsic granitoid matrix surrounded by concentric rings of mafic minerals (Fig. 13.3g. h).

The drive continues north on the Panamericana for another 80 km to Chañaral, halfway to which the main road passes east of the coastal village of Balneario Flamenco. This is one of several places on the drive where there are good views of granitoid plutons intruding older rocks; Palaeozoic basement rocks crop out extensively both to the south and north of Chañaral (Chapter 2). The highway south of Chañaral winds through a prominent coastal wavecut platform, deeply dissected by erosion.

Chañaral itself is an important port serving both fishing and mining industries, most notably the Potrerillos and El Salvador mines which lie 120 km to the east. For much of the twentieth century porphyry copper mine tailings were dumped into the sea (over 280 million tonnes of waste), displacing the shoreline by around 1 km and creating a legacy of serious contamination, mainly by As, Mo, Cu and Zn.

Chañaral to Antofagasta

North of Chañaral, fork left off the Panamericana towards the Pan de Azucar National Park, 28 km to the north. The road, which is minor but in good condition, passes through attractive coastal scenery on the same deeply dissected wavecut platform, surrounded by exposures of Palaeozoic basement rocks intruded by granite and abundant dykes. Reaching the coastal settlement of Pan de Azucar (a popular spot for viewing marine wildlife), the road then turns NE, crossing similar geology, to rejoin the Panamericana after 23 km.

From here Antofagasta lies 344 km to the north. The Panamericana initially runs NW then NE, traversing a broad outcrop of Jurassic volcanic rocks overlain by extensive spreads of Quaternary desert deposits. Still south of the Tallal turnoff, the road then curves back NW as it enters the Atacama Fault
Zone, preserved within which at this latitude is an outlier of Lower Jurassic volcanic rocks which are more extensively preserved to the south and east. Atacama fault escarpments are particularly well observed as the main road curves eastwards in the vicinity of the Taltal turnoff (Chapter 9). From this point the Panamericana strikes far inland, crossing a desolate landscape underlain by Cretaceous granites and, further east still, Palaeogene volcanic rocks, for the most part much obscured by desert sediments. Finally, the route leaves the Panamericana to descend westwards down into Antofagasta, the main Chilean copper exporting port. The road cuts progressively down through the well exposed Jurassic volcanic succession (La Negra Formation of the Coastal Cordillera: Chapter 4) previously seen south of Taltal, to reach the flat coastal strip on which Antofagasta is built.

Antofagasta to Mejillones

Continue north through Antofagasta on Route 1, with views east to the escarpment of Jurassic volcanic rocks, west to the Mejillones Peninsula, and north to the Neogene sediments cropping out on the isthmus separating the peninsula from the mainland. Just over 10 km from Antofagasta turn left and drive 1 km to the viewpoint at La Portada. Here pale, shelved, mostly shallow marine conglomerates and sandstones belonging to the Miocene La Portada Formation are well exposed along the cliffline, and rest unconformably on a basement of Jurassic andesite (Fig 13.4a). ‘The Mejillones Peninsula represents a tectonically rotated block of the continental platform that has been collapsing since Miocene times, and it exposes Neogene deposits that in parts of the platform are found below sea level’ (Chapter 3).

From La Portada continue west for 10 km across the isthmus which is covered by a veneer of windblown sand, to reach the abrupt western boundary where Jurassic volcanic rocks reappear at a fault scarp capped by a peneplain (Chapter 9). The road climbs steeply up through andesitic breccias, and then on through the village of Juan Lopez. Park above the cliffline beyond the village to explore excellent exposures of the basement rocks exposed at the coast. The rocks form part of a maﬁc–felsic plutonic complex, with maﬁc enclave-rich granites, all highly deformed by Mesozoic shearing deformation within the Atacama Fault System (AFS). The AFS is the most important fault zone in the Coastal Cordillera, initiated as a dominantly strike-slip structure within the active magmatic arc, and traceable for over 1000 km on the west side of northern Chile between Iquique and La Serena. In this locality, deformation of complex magma-mingled and granite-vined relationships has produced banded diaclase and sometimes migmatisic orthogneisses, with a strong NNE-striking foliation. These rocks, produced deep within the arc at depths thought to be around 12–15 km but subsequently uplifted within the Atacama Fault Zone belong to the Bofín Complex and are probably Early Jurassic in age.

There are wide views back over Playa Colorado (coloured by the weathered Jurassic andesitic rocks), the Neogene cover of La Portada, and a prominent wavecut platform uplifted by tectonic activity. The most recent major earthquake here was the magnitude 8.1 Antofagasta earthquake which occurred on 30 July 1995. The earthquake was a classic low angle thrust movement produced by the Nazca Plate subducting beneath the South American Plate at a rate of around 60–70 mm/year with a dip of some 20° for the first 50 km. This recent earthquake ruptured the entire seismogenic subduction zone between the Mejillones Peninsula and Taltal, with damage being concentrated in Antofagasta, which was abruptly moved 80 cm westwards. Here, at the southern tip of the Mejillones Peninsula, intertidal red coralline algae were raised by nearly 1 m, turning white with the exposure and thus providing a graphic record of tectonic uplift (Chapter 10).

Return to Highway 1 and continue north, past the airport and across the Tropic of Capricorn which, at 23°30' south of the equator, marks the southern boundary of the tropics and is the farthest point south at which the sun can be seen directly overhead at noon (on 21/22 December). The town of Mejillones lies on the coast 42 km north of the airport. To the east rises the Coastal Cordillera, whereas to the west are the high hills of the Mejillones Peninsula, isolated within the Atacama Fault System and containing an exotic fragment of ancient metamorphic rocks that have yielded latest Proterozoic–Cambrian radiometric ages (Chapter 2). Prominent Pleistocene marine plantation surfaces are visible on the tops of many of the hills including Morro Mejillones, testifying to the rapid uplift that has affected the edge of the Coastal Cordillera during late Cenozoic times.

Mejillones to San Pedro (451 km)

This is another long and highly scenic drive, initially following the coast northwards to Tocopilla, the most northerly point reached on the journey. The route then turns east to cross the modern continental forearc: climbing over the Coastal Cordillera, to traverse the Atacama Desert to San Pedro, via the famous mining towns of Maria Elena (nitrates) and Chiquicamata (copper). This classic west to east traverse passes from the Coastal Cordillera through the Central Depression (Longitudinal Valley), over the Chilean Precordillera and down into the Precordillera Depression to arrive at the oasis town of San Pedro, situated on the northern end of the Salar de Atacama. From a neotectonic viewpoint the route moves from the Coastal Cordillera, which is currently undergoing extensional collapse, to the highly compressive eastern part of the subduction system: the Precordillera, forearc basin and the High Andes where the spectacular uplift of the Altiplano has taken place (e.g. see Chapter 9).

Mejillones to Maria Elena

The first part of this journey, which involves a 136 km drive along the formerly Bolivian coastline from Mejillones to Tocopilla, is especially extraordinary. Route 1 runs along a narrow coastal platform, completely isolated between the andesitic hills of the Coastal Cordillera to the east and the coast to the west. The scarp of the Coastal Cordillera is a Miocene cliffline that in places reaches 1000 m in height. Further west, the Chile Trench is close to reaching its maximum depth which, at over 8000 m, provides the sensation of driving near the summit of a huge mountain, most of which is covered by ocean water. The extreme tectonic instability and vulnerability of this region is emphasized 78 km and 86 km north of Mejillones where there are two abandoned population centres destroyed by catastrophic tsunamis (Fig. 13.4b; also see Chapter 10). The first of these is Cobi, Bolivia’s first and for a time most important port, and the second Gatico. A major earthquake hit the area around 17:00 hours on 13 August 1868. Although the most severe damage occurred further north in Arica (where the resulting tsunamis reached 18 m in height) incoming waves still reached 10 m high as far south as Mejillones, and Cobi was terribly damaged. This disaster was followed uneventfully by another tsunami nine years later at around 20:00 hours on 9 May 1877 (Chapter 10). Cobi, which was still recovering from the 1868 event as well as a subsequent epidemic of yellow fever, was practically wiped off the map. The survivors concluded that Cobi was cursed, the Bolivian government decided to move the port activities to Antofagasta, and the town was never repopulated (Mejillones also was nearly totally destroyed, but rebuilt). A cemetery wall by the roadside displays the mournful painted words ‘fiunmos personas’ (we were people). About 8 km further up the road another cemetery on
the left marks the former village of Gatipo, although here the place continued to function as a small port until as late as 1930. The desolation and isolation of these former coastal villages, and their raw scenic and tectonic setting, leaves a lasting impression on the transient visitor.

The geology of the Coastal Cordiller a along this stretch of coast is dominated by Early-Middle Jurassic calc-alkaline basaltic to andesitic volcanic rocks (La Negra Formation) intruded by Middle-Late Jurassic gabbroic to granitic rocks. The area is notable for Cu 

Ag ore deposits (both stratiform and breccia pipe bodies) mainly hosted by the volcanic rocks, and Cu-Fe-Au (chalcopyrite-magnetite-actinolite) vein deposits found in the intrusives (Chapter 6). The ore deposits, which are thought to be of hydrothermal origin and linked to the granitic intrusions, are exploited by several mines, the majority of which lie between Coquina and Tocopilla.

During the final 30 km of the journey the coastal plain narrows, producing more dramatic scenery with spreads of alluvial gravel deposits towering down from the mountains to the sea, and with several wavecut terraces raised above sea level by tectonic activity. An easterly dip to the La Negra volcanic rocks, which is particularly obvious here, is attributed to block rotation on normal faults during extensional collapse of the Coastal Cordillera.

The coastal corniche reaches the industrial port of Tocopilla, "la capital de la energia", initially passing a power station. The main functions of this town are to act as a railhead for the now much declined nitrate industry based at Santa Elena, and to feed energy to the copper mining town of Chuquicamata. Leaving the town and coastline, follow the road east (Maria Elena 78 km) climbing through the arid brown landscape created by erosion of the Jurassic igneous rocks many of which in this area are dioritic. After 13 km the road levels out as one enters the deeply weathered desert plain landscape of the Central Depression, which lies at around 1000 m a.s.l. along this western side and much of which is covered by a veneer of Neogene Atacama Gravels (Gravas de la Pampa) and Quaternary desert sediments. In this area the Salado del Carmen section of the Atacama Fault Zone runs along the boundary between the Coastal Cordillera (west) and the Central Depression ahead. The Central Depression was once the backarc basin to the Coastal Cordillera, becoming an interarc basin by mid-Cretaceous times as the focus of magmatic activity shifted eastwards. Today it lies in the uplifted continental forearc of the modern subduction system, bounded by the eroded remnants of the former Jurassic (Coastal Cordillera) and Late Cretaceous-Eocene (Chilean Precordillera) magmatic arcs.

60 km from Tocopilla turn right to Maria Elena, on a rough (but paved) road across the desert which crosses the railway then passes through a wasteland of material dumped by the nitrate industry (Chapter 7). The entry to Maria Elena is heralded by the huge nitrate plant (Fig. 13.4c), and the town itself has a museum dedicated to the history of this peculiar, now virtually defunct, industry, once of critical importance to the Chilean economy.

**Maria Elena to Chuquicamata**

Returning to the main Tocopilla-Calama road by the same route, turn right and continue east across the Gravas de Pampas plain, heading towards the hills of the Chilean Precordillera and the Domeyko Fault Zone, which form the eastern border to the Central Depression (see Fig. 3.5I). Crossing the Panamericana, continue east towards Chuquicamata and Calama, passing saline deposits of white, evaporitic (including nitrate) cemented gravels exposed in roadcuts. About 40 km from the Panamericana the road enters the hills of the Chilean Precordillera, with good views back westwards across the Central Depression to the Coastal Cordillera. The Chilean Precordillera has been produced by shortening which has uplifted basement and cover rocks in a series of broad anticlines, and forms a mountain range that separates the two basins of the Central Depression and the Preandean Depression.

At 53 km from the Panamericana a rough track leads off for 13 km to the Chuqu Chuq geophysics, an optional journey for which a four-wheel drive vehicle is recommended. Continuing on the main road, a further 12 km east from the Chuqu Chuq turnoff the road surmounts a col revealing views across the Calama forearc basin to the snow-capped volcanoes of the Andes. The road continues across a pale rocky landscape eroded into a granodiorite belonging to the late Eocene Fortuna Intrusive Complex which crops out west of Chuquicamata within the former Late Cretaceous-Eocene magmatic arc, which lay along the line of the Precordillera. These intrusions provide one of the best examples of the plutonic precursors to the copper mineralization: you are entering an area with the largest copper concentration in the world (Chapter 6).

The desert mining town of Chuquicamata lies a little further east and since 1913 has exploited a supergiant porphyry copper system, the outcrop of which is around 14 km long (Chapter 6). This is one of 16 or so giant porphyry copper deposits in the circum-Pacific belt, and one of the largest in the world. The metallogenic belts in Chile follow the same west-to-east diachronocity exhibited by the magmatic arc. Thus in northern Chile the oldest metal deposits are Mesozoic in age and occur

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**Fig. 13.4.** (a) Coastal cliffs exposing pale-shallow marine Neogene sediments at La Portada (looking SE), on the isthmus to the Mejillones Peninsula just north of Antofagasta. The sediments rest on a basement of Jurassic arc volcanic rocks of La Negra Formation, exposed at sea level. The hills behind forming the Coastal Cordillera, the eastern backdrop to Antofagasta, are also formed from La Negra Formation basaltic andesites (c. 180 Ma). (b) Deserted coastal cottage at an abandoned village along the coastal strip running north of Mejillones, looking south. Not far south of here was Bolivia’s first and, for a while, largest port, now also abandoned after a series of natural disasters (see text). The devastation of population centres by successive tsunamis produced by subduction earthquakes has left little evidence of human habitation along this vulnerable coastline, caught between the high hills of the Coastal Cordillera (left) and the open Pacific Ocean. The words painted on the wall are warning off with a curse anyone who dares to steal from or damage the sacred site. (c) The nitrate plant at Maria Elena, the last of the Olivares Saltires – the nitrate mining centres along this stretch were worked well for a hundred years during the height of the industry in the late nineteenth and early twentieth centuries. Olivares Saltires is the setting around which is grouped the administration, lodge camps, beneficiation plants and spoil dumps (the latter being referred to specifically as Torta). Looking east across the flat desert of the Central Depression towards the hills of the Chilean Precordillera.

(d) Roadcut during the descent from the anticline east of the Cordillera de la Sal into San Pedro de Atacama. Steeply east-dipping continental sediments of the Pacifca Group (Oligocene-early Miocene) are overlain with angular unconformity by Miocene-Pliocene continental sediments, within which are several prominent pale volcanic horizons which record eruptions during the early establishment of the volcano are in its current position in the Atacama around 20 million years ago. (e) View of llamas crossing the flat alluvial plain on the northeastern corner of the Salar de Atacama, overlooked by the Licancabur volcano. The Licancabur volcano has flowed over older ignimbrites (see Fig. 11). (f) Gorge NE of Tecana, incised through pale dactile ignimbrite produced during the earlier eruptive phase of the arc and subsequently overlain by andesitic lavas emanating from the modern volcanic centres. In the background are the twin volcanic cones of Lascar (that on the right is actively steaming), which rises to around 6000 m a.s.l. (g) Saline crust covering the Salar de Atacama, with views eastwards across the saline lake of the Lagunas Chaxa to the volcanic arc behind. (h) View north from Laguna Miscanti, which lies at 4350 m a.s.l. within the Reserva Los Flamencos towards the andesite volcanoes of Licancabur (left) and Lascar (double cone), and foreground andesite boulders.
closest to the coast, followed eastwards by north–south aligned belts of Palaeocene then Late Eocene–Oligocene deposits, to which Chiucoanama belongs (Figs 13.1 & 13.2). The main openpit is around 4 km × 2 km in size (and >750 m deep) and lies NE of the town, on the other side of the north–south striking West Fault. This latter structure is a major lineament over 1000 km long within the regional Domeyko Fault System and one that has strongly influenced mineralization. Movement on this fault system has been polyphase, involving both strike-slip and reverse movements, and with an aggregate sinistral transcurrent displacement of around 35 km. The ore deposit at Chiucoanama is virtually all hosted by, and related to, late Eocene porphyritic stocks (see Chapter 6), usually occurring in veins and veinlets. The low grade ore (only 1–2%) yields mostly copper, with smaller amounts of molybdenum and selenium. Tours of the mine are run daily, but usually only in the mornings (arrive before 09:45 hours) so an overnight stop at Calama will be necessary.

**Chiucoanama to San Pedro**

From Chiucoanama it is a 14 km drive south to Calama, a desert town that has existed for over 175 years, although destroyed by a major earthquake (on 22 April 1870) and shaken by many others (e.g. a magnitude 7.8 quake on 13 June 2005, which was felt in Calama with an intensity of 5 to 6). The road to San Pedro de Atacama runs SE from Calama, moving from Quaternary to Miocene continental sediments as it climbs the west flank of the anticlinal Domeyko Massif to the Paso Barros Arana (61 km from Calama). At the Paso Barros Arana sediments of the Cretaceous–Eocene Purilactis Group are thrust over gently westly dipping Miocene sediments and igmibrites. The thrust forms the western edge of a NNE–SSW trending anticline developed in response to the crustal shortening which produced the Chilean Precordillera. The Domeyko Massif, named after the Polish naturalist Ignacio Domeyko, forms an uplifted area some 600 km long and locally reaches well over 4000 m a.s.l. Further south in the Massif, basement rocks are exposed, although this is not the case here where exposures in the anticlinal core are continental sediments belonging to the latest Cretaceous to Eocene Purilactis Group. This latter stratigraphic unit is over 4 km thick, with playa lake red siltstones and evaporites at its base being overlain by coarse fluvial sandstones (seen here) capped by volcanic rocks. These deposits are coeval with the Late Cretaceous–Eocene Precordillera magmatic arc, infilling interarc depressions and the backarc basin.

After passing the turnoff to Rio Grande (65 km from Calama), with the volcanic arc of the High Andes now clearly in sight, the road turns SSE and begins a long curving descent into the Precordillera Depression towards the Salar de Atacama. The Purilactis Group is overlie by Miocene continental sediments which are equivalent to the Atacama Gravels, seen in places during the journey since Termas de Socos. After around 16 km of descent and just before the road passes through the Llana de la Paciencia, a broad shallow valley some 1 km wide, it takes an abrupt turn to the SE and traverses along the tip fold of a blind thrust that is actively propagating into the Llano. The road then climbs eastwards up the eastern side of the valley over an initially west- then steeply east-dipping succession in the anticlinal intrasial Cordillera de la Sal (to be examined the next day). There are glorious views as one descends to the oasis of San Pedro de Atacama in the Precordillera Depression, with the snow-capped volcanoes of the Western Cordillera behind defining the position of the present-day magmatic arc (Chapter 5). The volcanoes form the western edge of the Altiplano, a 150 km-wide plateau which extends into Argentina and Bolivia with an average elevation of 3700 m.

**San Pedro–Salar de Atacama circuit (330 km)**

This route initially makes a short (55 km) circular tour of the Valle de la Luna and Cordillera de la Sal immediately west of San Pedro, then heads SE to visit the spectacular Salar de Atacama and the geology of the western flank of the modern arc (Fig. 13.5; see also Fig. 3.51). The visit to Valle de la Luna involves driving on unsurfaced roads which, in good weather, are easily passable with a high-clearance vehicle.

**NW Salar de Atacama: Valle de la Luna and the Cordillera de la Sal**

Drive back west out of San Pedro de Atacama village and over the Puerto de San Pedro to turn left onto the bypass (signposted Tocanao). Pass a sign for the Valle de la Luna and in just over 3 km turn right onto a rough road which heads south on the west side of the San Pedro delta for 6 km to a sign: Bienvenidos a la Valle de la Luna. From here, on the Quaternary gravels, there are good views south across the Salar de Atacama, the southern end of which lies over 75 km away. Continue for another 300 m and park on the left in the valley just before the road begins its climb into the Cordillera de la Sal. Here you have crossed over west from the Salar de Atacama Quaternary sediments onto the hanging wall of the Cordillera de la Sal east-vergent fault system. This is an inverted Cretaceous extensional structure which has upthrust and exposed successions of folded Oligocene to Pleistocene sediments within the Precordillera Depression. Walk NW into the narrow canyon cut into the evaporitic Cordillera de la Sal. Underfoot are horizontal to gently east-dipping young (but reasonably well cemented) desert cross-bedded sands and gravels. These rest unconformably on and against the San Pedro Formation (Paciencia Group: Oligocene–early Miocene) which contains a translatent massive evaporite (halite) horizon tens of metres thick. The weathering of this evaporite unit into picturesque shapes and surfaces makes this locality a prime tourist attraction.

Continue driving west through the Cordillera de la Sal anticline, passing desert dune deposits and folded and faulted Paciencia Group sediments. The Llana de la Paciencia, the valley separating the intrasial Cordillera de la Sal from the main Domeyko Massif to the west, comes into view 6 km from the last stop. The road continues westwards for another 3 km, winding past dramatic exposures of huge sand dunes and steeply dipping evaporitic Paciencia Group sediments (locally overlain unconformably by horizontal gravels, at Cuesa de las Salinas). Reaching the valley bottom of the Llana de la Paciencia, turn right (north), with views NE towards Licancabur Volcano (5916 m) on the Chile–Bolivia border.

Follow this gravel road for 4 km to the main road, turn right and drive for 8 km, climbing back over the Cordillera de la Sal to park at a tourist viewpoint (with a cross) on the right. Here the thrust-generated folding of the Paciencia Group strata around the Cordillera de la Sal anticline can clearly be traced looking south over the exposures below. On the opposite side of the road, cliffs expose steeply east-dipping Paciencia Group reddish continental sediments overlain with angular unconformity by gently east-dipping Miocene and Pliocene continental sediments within which are several prominent pale ashly horizontal (Fig. 13.4b). These ashes, derived from the east and NE, record eruptions during the early establishment of the volcanic arc in its current position around 20 million years ago.

**NE Salar de Atacama: Tocanao, Laguna Chaxa and Peine**

Return east to the San Pedro bypass, avoiding the town, and take the road (right) SE to Tocanao. The road crosses the NE corner of the Salar de Atacama, the northern part of which is cultivated around the Rio San Pedro delta. There are extensive views of the modern volcanic arc to the east. Licancabur
Volcano rises to the NE (Fig. 13.4e), whereas to the SE is the elongate composite stratovolcanic complex of Lascar Volcano (Chapter 5), from which can commonly be seen rising from one of its vents. Lascar has been recently the most active volcano in northern Chile, erupting 15 times in the twentieth century alone (see 1.2). Don’t enter Toconao but park just after the bridge on the village bypass and walk eastwards up the gorge. The walls of the gorge expose pale Plio-Pleistocene dacitic ignimbrite, a pumiceous crystal tuff with abundant phenocrysts of biotite, quartz and feldspar and comagmatic enclaves of microgranodiorite. Extensive exposures of these ignimbrites, ranging back to Miocene in age, drape the western slope of the volcanic chain, and are over lain to the west by allu vial fan and playa deposits, and to the east by the andesites of the modern volcanoes. There are tremendous views SE to the double cone of Lascar Volcano (Fig. 13.4f).

Continue south on the main road to a signposted turnoff (right) to Laguna Chaxa (you are now 40 km from San Pedro). In 14 km turn right again, to arrive at Laguna Chaxa after a further 10 km. The Laguna is situated near the centre of the salar with wide views across this intermontane forearc basin to the volcanic arc rising to nearly 6000 m a.s.l. in the east, and the intrabasinal Cordillera de la Sal (with the main Domeyko Massif behind) to the west. Follow the guided walk clockwise around the site, with evaporitic deposits underfoot and a rich wildlife to enjoy (Fig. 13.4g). The Salar de Atacama lies at 2300 m a.s.l. and is the largest salar in Chile, a saline playa complex which rests upon a basin fill reaching nearly 9 km in thickness. This basin fill, which ranges back in age to Cretaceous times, is mostly siliciclastic, but the upper 1–1.6 km are highly evaporitic with thick halite deposits in the basin centre. The complex evaporitic chemistry, influenced by the nearby volcanism and high geothermal gradient, has produced borate and chloride-bearing (especially lithium) brines which have been exploited in the southern part of the salar (Chapter 7).

Drive back 10 km to the junction and turn right (south) towards Peine. The road, which can be rough in places, runs down the east side of the salar for 45 km then turns left (east) and climbs to the village of Peine. Drive through the village, cross a valley and park by the local football ground just beyond a water storage tank. The exposures here belong to the Permo-Triassic Peine Group, which form inliers of volcanic and continental siliciclastic lithologies and belong to the basement rocks exposed around the Salar de Atacama (Chapter 2). At this locality there are bimodal felsic (ash flow tuffs, breccias and porphyritic lavas) and mafic (feldspar porphyritic basalts)
volcanic rocks. All five of the main stratigraphic sequences in this area have now been visited: Peine Group (Permian-Triassic continental rifting), Puriñacutic Group (Cretaceous–Eocene backarc basin infill), Paciencia Group (Oligocene–Miocene transtensional basin infill), San Bartolo Group and associated ignimbrites (Miocene–Pleistocene forearc basin infill), and, most recently, the Holocene deposits of the modern forearc Salar de Atacama. Thus it can be appreciated how the diachronous movement of magmatic arc activity across Chile has changed the tectonic setting of the same basin through time.

Return through Peine, descend to the Salar, and turn back (right) towards the north, passing a sign for the lithium mine El Lithio. After 22 km from Peine turn right onto a link road that runs towards the main road to Socaire. Park 7 km after the junction (as the road turns a left bend and the main road junction ahead comes into view). Walk south for 10 minutes across the volcanic boulder-strewn desert surface, towards a prominent brown hill (Colina de Cas), to reach a wadi which runs below exposures of east-dipping volcaniclastic breccias belonging to the basement Peine Group. Exploration of the wadi area will reveal pale dacitic ignimbrite (similar to that seen at Toconao) resting upon the basement. The ignimbrite is in turn covered by poorly consolidated alluvial gravels. About 25 km to the ENE, Lascar Volcano dominates the scenery of the modern volcanic arc.

Return to the vehicle and continue for 1 km to the main road. Those wishing to undertake the challenging drive up into the oxygen-depleted higher ground south of Lascar Volcano need to divert SE (right) here via Socaire for 37 km to the rough track turnoff (left) to lagunas Minhiques and Miscanti (Fig. 13.4b). Be warned, however, that this involves more than 15 km driving over very rough rocky ground (and at Lago Miscanti you will be asked to pay a toll for your trouble) so a high-clearance vehicle is essential. A much easier alternative journey up into the volcanic arc can be made by turning left onto the main road and retracing the east side of the Salar de Atacama towards San Pedro (83 km) via Toconao. Just before San Pedro turn right onto the major road that runs east skirting Bolivia towards Argentina. The road climbs the east side of the forearc basin, affording spectacular views of Licancabur Volcano, with its black andesitic flows erupted over the pale ignimbritic sequence. Stop just east of roadmark km20 for views back west over the Salar de Atacama, the San Pedro delta, the Cordillera de la Sal and the Domeyko Massif: the whole day’s geology is spread out before you.

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The Geology of Chile

Edited by Teresa Moreno & Wes Gibbons

This book is the first comprehensive account in English of the geology of Chile, providing a key reference work that brings together many years of research, and written mostly by Chilean authors from various universities and other centres of research excellence. The 13 chapters begin with a general overview, followed by detailed accounts of Andean tectonostratigraphy and magmatism, the amazingly active volcanism, the world class ore deposits that have proven to be so critical to the welfare of the country, and Chilean water resources. The subject then turns to geophysics with an examination of neotectonics and earthquakes, the hazardous frequency of which is a daily fact of life for the Chilean population. There are chapters on the offshore geology and oceanography of the SE Pacific Ocean, subjects that continue to attract much research, not least from those seeking to understand world climatic variations, and on late Quaternary land environments, concluding with an account examining human colonization of southernmost America.

During his voyage on H.M.S. Beagle, an extended visit to Chile (1834-35) had a profound impact on Charles Darwin, especially on his understanding of volcanoes, earthquakes and tsunamis. Over more recent decades scientists have come to recognize the Chilean Andes as providing the classic example of a mountain belt produced by oceanic subduction beneath a continent, as well as some of the most dramatic scenic and climatic variations on Earth. In the final chapter, the editors offer a description of a drive from the Mediterranean landscapes of central Chile to the hyperarid Atacama Desert, a contribution designed to give visitors a chance to experience for themselves the geology and scenery of this extraordinary country.

Front cover photograph: The Chilean Andes are amongst the most active volcanic regions on Earth, due to the subduction of the Nazca plate beneath the continental margin. In northern Chile, northern Argentina and southwestern Bolivia, many active volcanoes are constructed on the uplifted Altiplano region and constitute the highest peaks, often reaching altitudes of over 5000 metres. The cover photograph shows a view across Laguna Miscanti (4,350m) on the High Chilean Altiplano northwards to the volcanoes of Lascar (5,916m; left) and Llaima (5,502m; right). The andesitic to rhyodacitic Llaima volcano is the most active in the northern Chilean Andes, with over 15 eruptions in the 20th century alone. Photograph by Wes Gibbons.