3 Tectonostratigraphic evolution of the Andean Orogen in Chile

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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 We present here an overview of this evolution considering the new information and interpretations, including some unpublished ideas of the authors.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 eastward 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 1973*a*; Aguirre *et al.* 1974; Frutos 1981; Jordan *et al.* 1983*a*, 1997; Malumián & Ramos 1984; Ramos *et al.* 1986; Isacks 1988; Ramos 1988*b*; 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 oroclinal bends around which are major changes in the orientation of the morphology and structure of the range (Fig. 3.1). Two oroclinal bends 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.* 27ºS to *c.* 33ºS) 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 (46–47ºS), 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 18º and 27ºS and the other between 33º and 46ºS, in which the Central Depression is well developed, and an intermediate segment which lacks a Central Depression (27–33ºS), 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.* (1973*b*) 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,

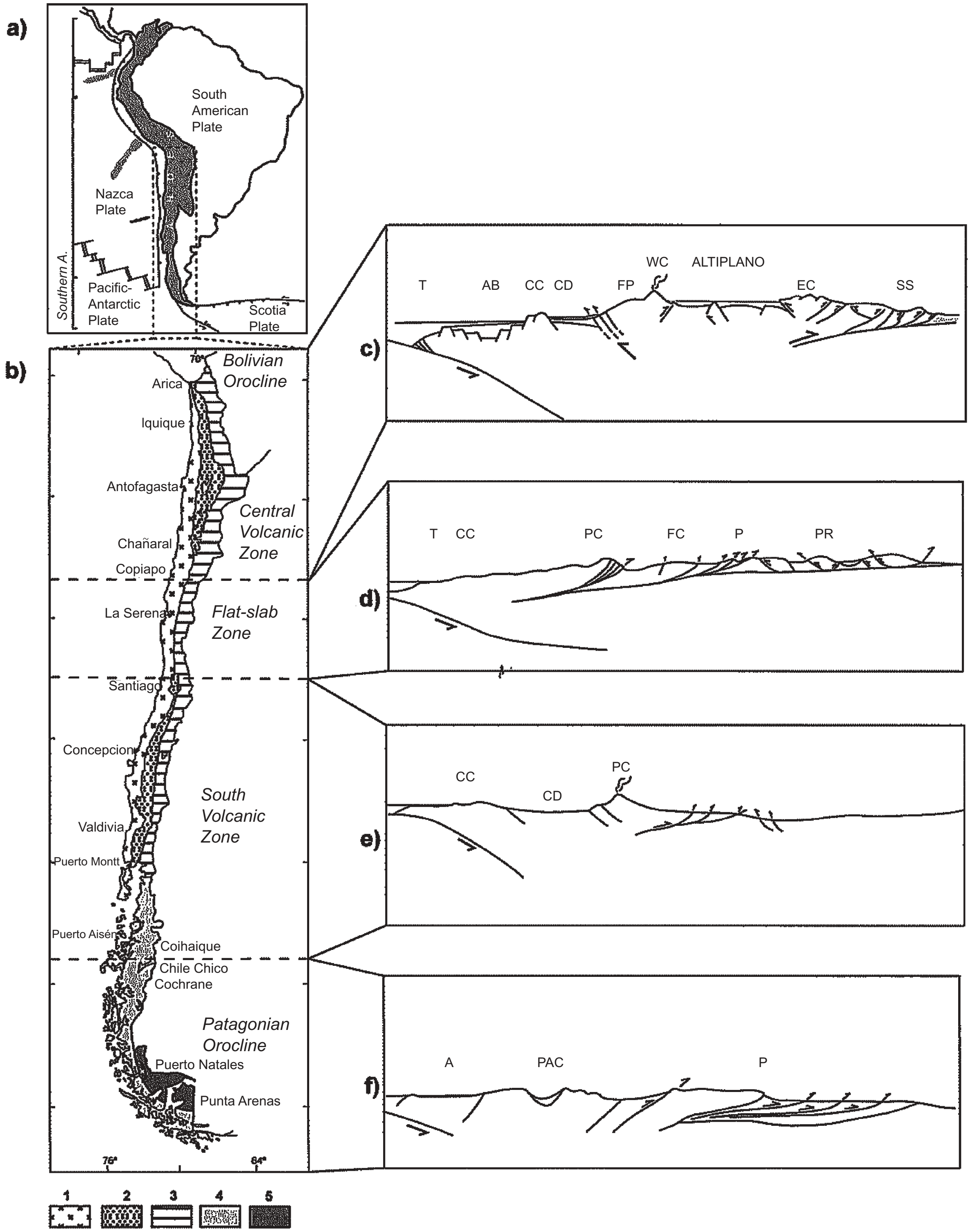


Fig. 3.1. Location of the Central and Southern Andes relative to South America, and major morphologic elements of the Andean Cordillera and oceanic plates facing the western margin of southern South America. (**a**) Location of the Andean Cordillera relative to South America, major subdivisions along 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 (17º45′ to 56ºS): Bolivian and Patagonian Oroclines, Andean segmentation, and morphostructural units: 1, Coastal Cordillera (Coastal Ranges, between 27º and 33ºS); 2, Central Depression; 3, Forearc Precordillera and Western Cordillera, between 18º and 27ºS, High Andean Range, between 27º and 33ºS (flat-slab subduction segment), Principal Cordillera, between 33ºS and *c*. 42ºS; 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 27º and 33ºS); PA, Patagonian Archipelago; PAC, Patagonian Cordillera; PC, Principal Cordillera; PR, Pampean Ranges; SS, Subandean Sierras; T, Trench; WC, Western Cordillera.

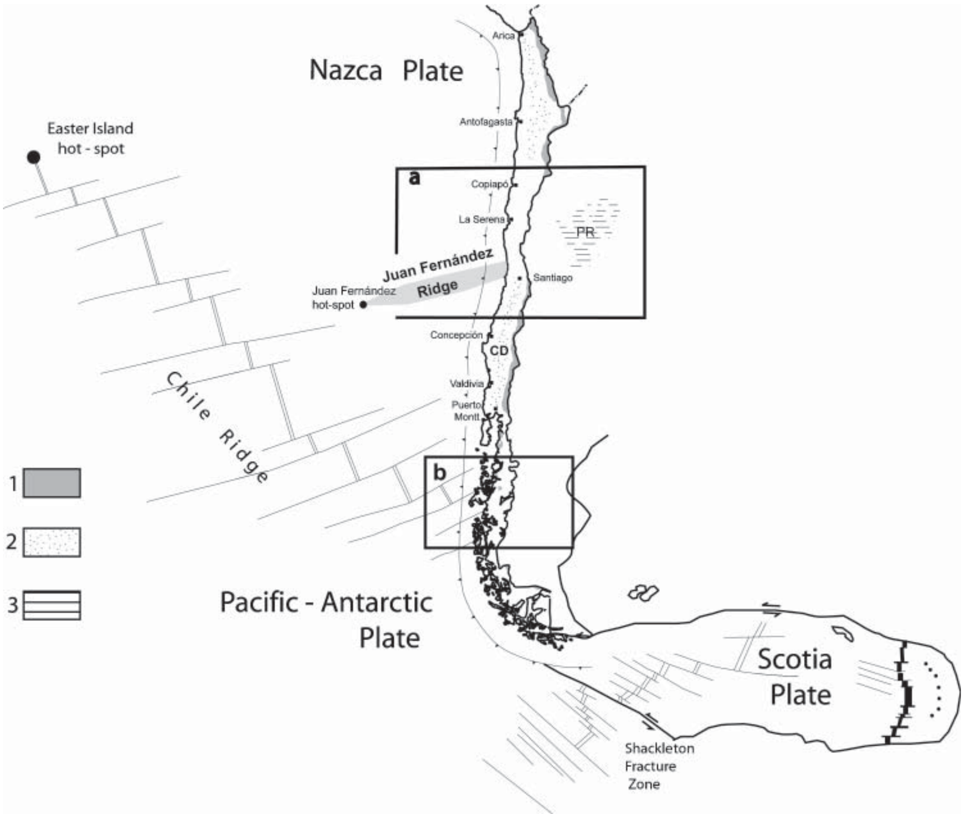


Fig. 3.2. Plate configuration of the SE Pacific Ocean with the Nazca, Pacific–Antarctic and Scotia plates. Chile Ridge (not to scale) separates the Nazca and the Pacific–Antarctic plates. The Scotia Plate is separated from the South America Plate by a left-lateral structure connected to the west with Magallanes Fault, the North Scotia Margin, and separated from the Pacific–Antarctic Plate by the Shackleton Fracture Zone. (**a**) Location of the Juan Fernández hot-spot and ridge, outlines of the Nazca Plate and segmentation of the orogen, with the flat-slab segment (Norte Chico region) without Central Depression and Plio-Pleistocene volcanic activity, and with development of the Pampean Ranges on the Andean foreland, in Argentina (inset corresponds to the region represented in Fig. 3.61). (**b**) Location of the triple-junction region (inset corresponds to a region illustrated in Figs 3.57 & 3.70). Key: 1, Plio-Pleistocene volcanism; 2, Central Depression; 3, Pampean Ranges (PR).

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; Thomas & Astini 2003). 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 occuring 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, ^{87}Rb–^{86}Sr whole-rock isochrons on migmatites and schists from the Sierra de Moreno gave Early Palaeozoic ages (511 $\pm$ 9 Ma, and 485 $\pm$ 12 Ma) (Skarmeta 1983). A somewhat older age has been obtained in the Mejillones Peninsula, at 23–23°30′S (*c*. 530 Ma; ^{87}Rb–^{86}Sr, 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 $\pm$ 7.5 Ma; concordia intercept on unabraded zircons) (Söllner *et al.* 2000*b*) (Fig. 3.3a).

The ^{87}Rb–^{86}Sr 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 radioisotopic dating has yielded ages of 544 $\pm$ 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

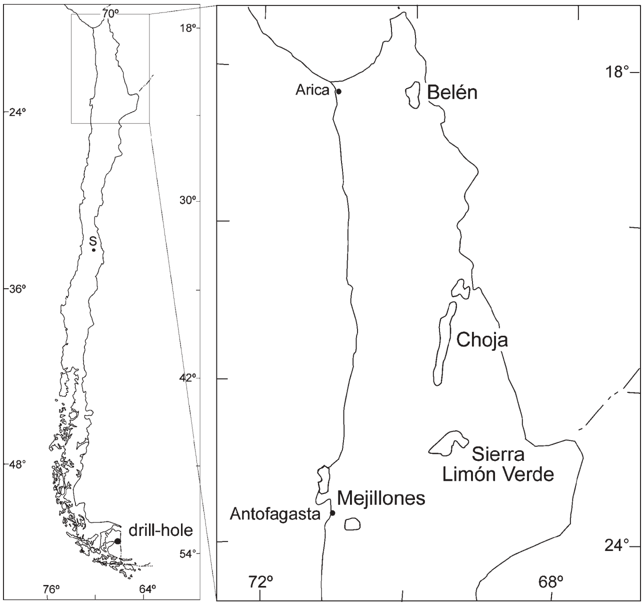


Fig. 3.3. Location map for units of the Pampean tectonic cycle in Chile. Localities in northern Chile (see inset top left) are figured on the right, based on Salas *et al.* (1966), García (1996, 2002), Vergara (1978*a*), Maksaev (1978), Damm *et al.* (1990), Skarmeta (1983), Díaz *et al.* (1985). Location 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.* 2000*b*), is indicated bottom left.

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 473 ± 2 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.* 2000*c*; 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 present-day location of the CMB.

The Choja Formation (Vergara 1978*a*) (Challo Formation of Maksaev 1978) or Choja Metamorphic Complex, exposed in the Altiplano and Precordillera (forearc Precordillera in northern Chile; see Fig. 3.1c) or western flank of the Altiplano, at 21ºS (Fig. 3.3b), consists of mica schists intruded by an Early Silurian granite dated at 431.05 ± 9.97 Ma (K–Ar on muscovite), and locally of migmatites formed by 'lit par lit' injection 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 1988*b*), 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 Puna (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 (39ºS) by Tavera (1979*a*, 1983) (Fig. 3.4).

The only stratified rocks of undisputed Ordovician age in Chile are exposed at Aguada de la Perdiz, in the Altiplano next to the international boundary (García *et al.* 1962; Ramírez & 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. *furcifera*: Breitkreutz 1985, 1986*b*), 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 (1986*b*) 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

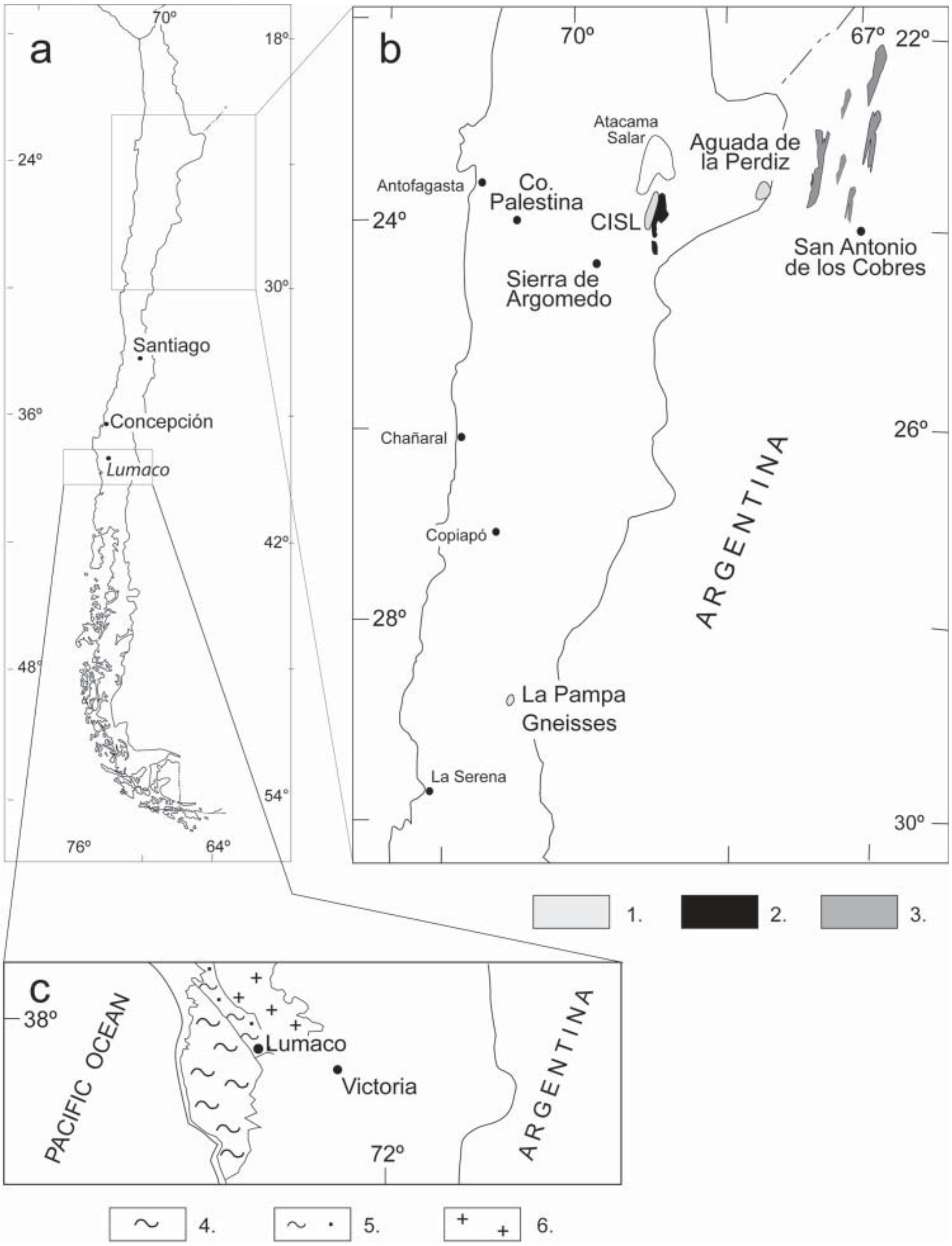


Fig. 3.4. (**a**) Location map for units assigned to the Famatinian tectonic cycle and localities cited in text in northern (b) and south-central Chile (c). (**b**) Location of dated and alleged Ordovician units between 23º30′ and 24ºS: Complejo Igneo y Sedimentario de Lila (CISL), Aguada de la Perdiz Formation, intrusive bodies of Late Ordovician and Early Silurian age, some of which are emplaced in the CISL, and Ordovician deposits from westernmost Argentina in the Puna (Coquena Formation and Calalaste Group), associated with the Aguada de la Perdiz Formation, and the La Pampa Gneisses, at 29ºS. Key: 1, Deposits in Chile; 2, Pingo-Pingo, Tilopozo, Tucúcaro, Tambillo, Alto del Inca and Choschas plutons; 3, Coquena Formation and Calalaste Group in Argentina. (**c**) Location of fossiliferous marine Silurian deposits forming less metamorphosed portions of the eastern series of the Metamorphic Complex at 38ºS, next to the locality of Lumaco (Tavera 1979*a*). Key: 4, Schists of the Western Series of the Metamorphic Complex; 5, slates, phyllites and metasandstones of the Eastern Series of the Metamorphic Complex; 6, Carboniferous to Permian granitoids of the Coastal Batholith.

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 1986*b*; 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 gabbric and rhyolitic subvolcanic plutons which are generally associated with lavas of similar composition (Niemeyer 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 1000-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 Llanvirnian trilobite has been found in the overlying Devonian–Early Carboniferous Lila Formation (Cecioni 1982). According to Niemeyer (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 6000-m-thick turbiditic 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 (Niemeyer 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 Ordovician times by thrusting of the island arc over the western border of the basin, initiating the development of a foreland basin. These authors consider that this tectonic event was caused by the collision of the AAT against the margin of Gondwana.

Silurian units

Silurian deposits have been reported by Tavera (1979*a*) from Lumaco in the Coastal Cordillera at 39ºS (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 24ºS) that do not intrude the CISL (Fig. 3.4), and are known as the Tambillo, Alto del Inca and Choschas plutons (Mopodozis *et al.* 1983). To the north of this locality, in the Precordillera or western flank of the Altiplano at 21ºS, a muscovite granite has been dated at 431.05 ± 9.97 Ma (K–Ar in muscovite) (Vergara 1978*b*). 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 29ºS (Fig. 3.4b), dated at 415 ± 4 Ma (^{87}Rb–^{86}Sr 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 1986*b*; Bahlburg *et al.* 1994; Bahlburg & Hervé 1997) has been attributed to the Ocloyic 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 Permian)

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 46ºS, rocks formed during this cycle are almost exclusively exposed along the westernmost part of the Coastal Cordillera, whereas south of 46ºS, 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.

Palaeo-magnetic 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,

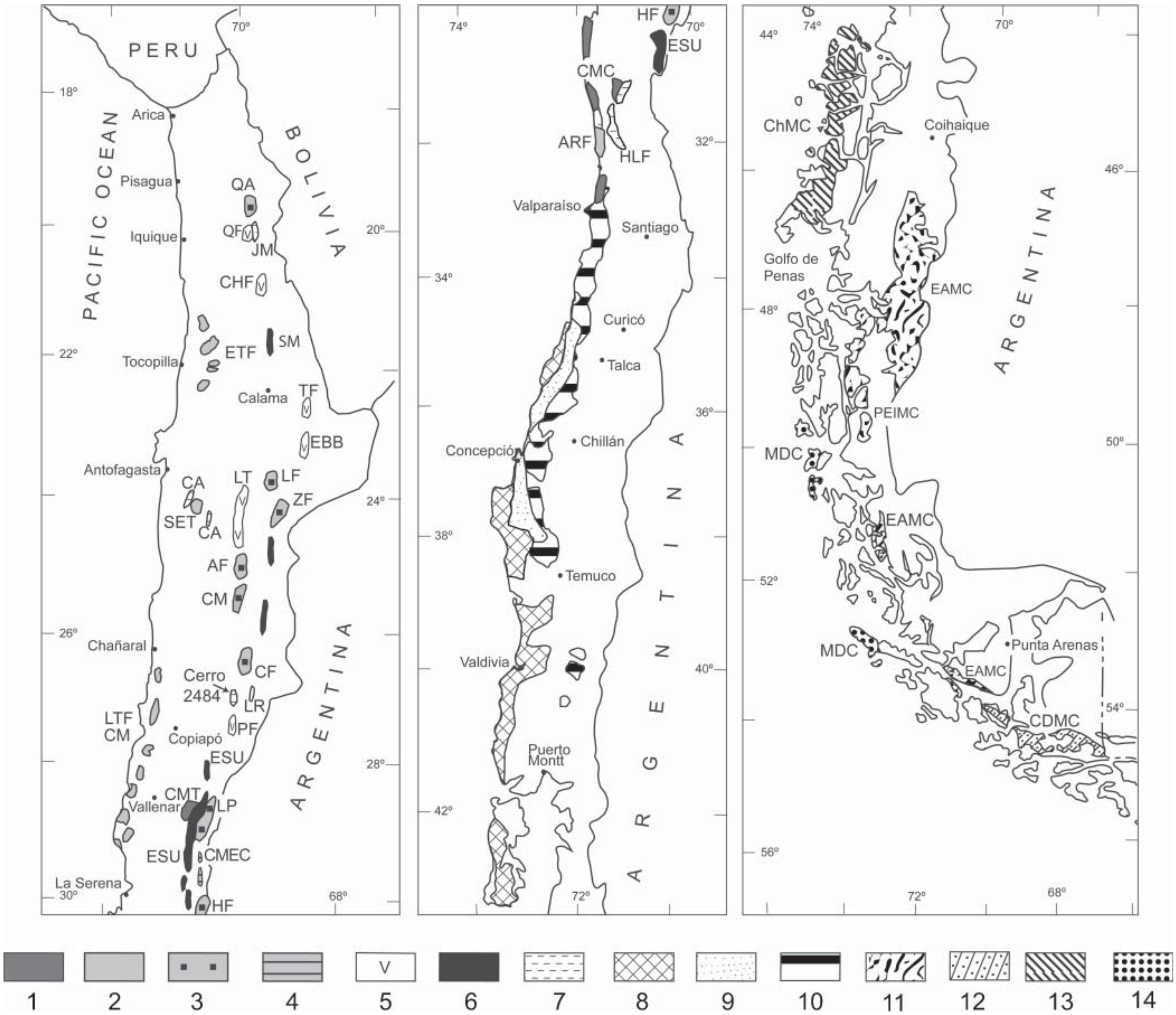


Fig. 3.5. Location map for units of the Gondwanan tectonic cycle. **North of 33ºS**, *First stage* (Middle/Late Devonian to Early Carboniferous). Key: 1, Metamorphic units: Complejo Metamórfico del Choapa (CMC) and Complejo Metamórfico El Tránsito (CMT). 2, Metasedimentary units and western sedimentary 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 Pantanoso Formation (PF). 6, Plutonic units: Sierra del Medio and Sierra de Moreno (SM) and further south along the Domeyko Range between 24º and 27ºS, Elqui Superunit (ESU). *Third stage* (Early Permian to Middle–Late? Permian). Key: 7, Juan de Morales Formation (JM), Cerro El Árbol Formation (CA), Las Represas Formation (LR), Cerro 2484 Beds, Huentelauquén Formation and Quebrada Mal Paso Beds (HLF). **Between 33º and 46ºS.** *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 46ºS.** 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).

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 radioisotopic 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; and (3) Early Permian to Middle–Late? Permian (Fig. 3.6). The main orientation of

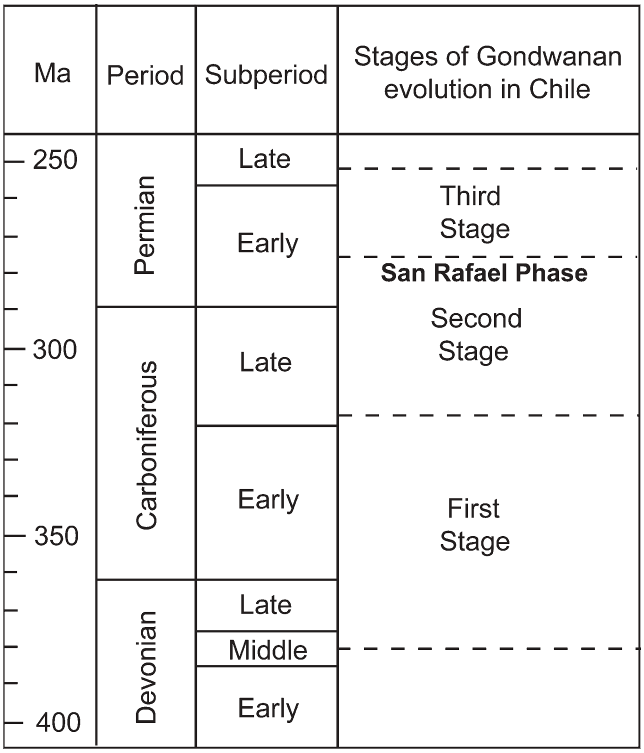


Fig. 3.6. Three-stage subdivision of the Gondwanan tectonic cycle and location of the San Rafael tectonic phase, based on the stratigraphic information available from the region north of 33ºS and applicable to the region between 33º and 46ºS. Subdivision of geological time is based on Harland *et al.*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

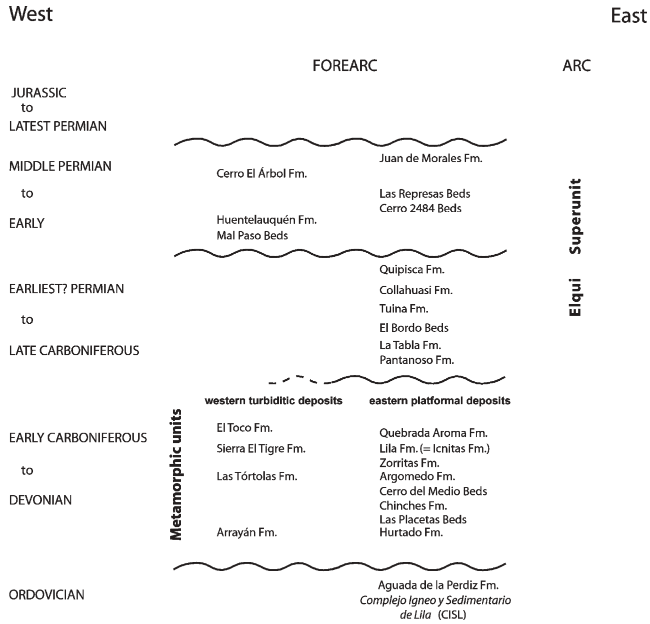


Fig. 3.7. Stratigraphic units of the three stages of the Gondwanan tectonic cycle in Chile mentioned in text. Within each of the stages eastern (right) and western (left) units are listed from north (above) to south (below) and face each other according to the latitudes of their occurrence.

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 to Early Carboniferous 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.* 1997*b*; J. Cortés 2000; Basso 2004; González & Niemeyer 2005); Las Tórtolas Formation (Uklricksen 1979; Bell 1982, 1984, 1987*a*, *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 1986*b*), the Sierra El Tigre Formation (Niemeyer *et al.* 1997*b*), 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.* 1997*b*), 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 Sierra El Tigre Formation contains a Devonian brachiopod (*Mucrospirifer* sp.) (Ferraris & Di Biase 1978; Niemeyer *et al.* 1997*b*; J. Cortés 2000; González & Niemeyer 2005). A Devonian–Carboniferous age has been assigned to the Las Tórtolas / Complejo Epimetamórfico Chañaral based on the presence of Ordovican to Devonian trace fossils (Naranjo & Puig 1984), and a Carboniferous conodont (Bahlburg *et al.* 1987). A ^{87}Rb–^{86}Sr ratio on the metasediments of this unit indicates a maximum age of deposition of 380 Ma (Middle Devonian). The Las Tórtolas Formation / Complejo Epimetamórfico Chañaral is intruded by Early Permian plutonic bodies (Berg & Baumann 1985). No fossils have been found in the Unidad Metasedimentaria de Agua Dulce. The Arrayán Formation has been assigned a Devonian to Carboniferous age based on badly preserved marine invertebrates (Muñoz Cristi 1942, 1968, 1973; Mundaca *et al.* 1979). Based on fossil plant remains, Cecioni (1962, 1974) suggested a Middle Devonian age, whereas Bernardes de Oliveira & Roesler (1980) favoured a Late Devonian to Early Carboniferous age. Although the fossiliferous content of these deposits indicates a Late Devonian–Early

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 mélange or broken formation, has been observed in the Sierra El Tigre (Niemeyer *et al.* 1997*b*) and Las Tórtolas (Mélange de Chañaral; Bell 1984, 1987*a*, *b*) formations, and in the Unidad Metasedimentaria de Agua 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, 1986*b*) (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.* 1997*a*). 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 1986*b*; Niemeyer *et al.* 1997*a*; 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 1986*b*; Niemeyer 1989). Breitkreuz (1986*b*) 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 (Isaackson *et al.* 1985; Dutro & Isaacson 1990; Boucot *et al.* 1995; Niemeyer *et al.* 1997*a*). The fossil content of the Argomedo Formation (several trace fosssils, and a specimen of Pholladomiidae indet., which indicates a maximum Early Devonian age for these deposits), its stratigraphic position below the Carboniferous–Permian 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 & Breitkreuz 1991, 1993; Breitkreuz 1986*b*; Niemeyer *et al.* 1997*a*), flanked to the east by a volcanic arc (Arco Puneño; Coira *et al.* 1982; Niemeyer *et al.* 1997*a*), which represented the source of the sediments (Fig. 3.8).

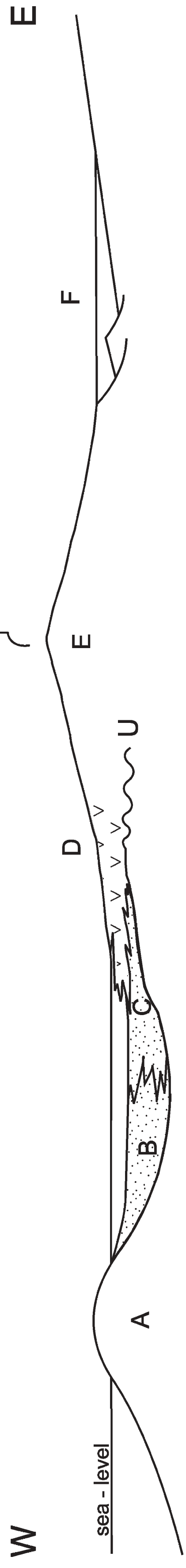


Fig. 3.8. Schematic east–west palaeogeographic section in Late Carboniferous–earliest Permian times in northern Chile (north of 33°S) across the emergent accretionary prism, the forearc basin with sediment supply from west and east, the arc and the backarc basin. A, Accretionary prism; B, western turbiditic deposits in the forearc basin; C, eastern platformal deposits with rare volcanic components in the forearc basin; D, volcanic and volcaniclastic deposits from the arc; E, Puneño Arc, partly located in western Argentina; F, Paganzo backarc basin in Argentina. B and C deposits correspond to sedimentary accumulations during Devonian to Early Carboniferous times (first stage of Gondwanan evolution) and D corresponds to the volcanic and volcaniclastic deposits accumulated in Late Carboniferous to earliest Permian times (second stage of Gondwanan evolution). U, Unconformity separating Famatinian from Gondwanan units.

Other deposits of probably similar age have been reported from the Precordillera, at 18°30′S (Quichoco 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 (Chinches 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 Chinches 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 comprising foliated shales, siltstones and very fine- to coarse-grained sandstones (Bell 1985) and a few tuff horizons (Sepúlveda & Naranjo 1982; 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 Lila and Zorritas formations.

Further south, two strongly folded marine units are exposed in the High Andes next to the international boundary, between *c.* 29°S and 31°S: 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* Frenguelli, *Sigillaria sauri Brant*?, *Plagiozamites*?). Badly preserved marine invertebrates indicate a marine environment. The so-far unfossiliferous 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.* 26° and 31°S with protoliths that tentatively can be assigned to the eastern platformal deposits, e.g. the Esquistos El Jardín (Muñoz 1986, *fide* Tomlinson *et al.* 1999) and the El Cepo Metamorphic Complex (Mpodozis & Cornejo 1988). The Esquistos 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 Cepo Metamorphic Complex (Complejo Metamórfico El Cepo: Mpodozis & Cornejo 1988), exposed in the High Andes between 29° and 31°S (CMEC in Fig. 3.5), forms small outcrops that correspond to roof-pendants in plutonic rocks of the Elqui-Limarí batholith. These rocks correspond to sandstones and pelitic deposits apparently affected by two episodes of metamorphism: an older dynamothermal 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 (27°S) and Los Vilos (32°S) (Fig. 3.5). In the high Andean cordillera, metamorphic outcrops between 28°30′S and 30°30′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 30°S 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; Rebolledo & 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 31°S 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; Rebolledo & 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.* 1981*a*). This view has been confirmed by subsequent authors (Thiele & Hervé 1984; Irwin *et al.* 1988; García 1991; Rivano & Sepúlveda 1991; Rebolledo & Charrier 1994; Bahlburg & Hervé 1997). ^{87}Rb–^{86}Sr 'errorchrons' for the metamorphic age of the CMT are 335 ± 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 quarz–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 Chinches formations, and Cerro del Medio Beds, all of which received their sedimentary supply from a volcanic source referred to as the Puneño Arc and located to the east and SE: see Niemeyer *et al.* 1997*b*).

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 Chinches Formation (Sepúlveda & Naranjo 1982; Bell 1985), volcanic quartz crystals present in the Zorritas Formation (Niemeyer *et al.* 1997*b*), and abundant volcanic sand grains in the Argomedo Formation (Marinovic *et al.* 1995). As indicated above, the source for these volcanic components most probably was the Puneño Arc located to the east of the forearc basin (Coira *et al.* 1983; Niemeyer *et al.* 1997*b*). The evidence for magmatic activity at this time would indicate that the Silurian–Early Carboniferous magmatic lull mentioned by Balhburg & 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 Puneño 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 20ºS (Galli 1968); the Collahuasi Formation, at 21º15′S (Vergara 1978*b*; Vergara & Thomas 1984; Ladino 1998); the La Tabla Formation, between 24ºS and 27ºS (García 1967; Marinovic *et al.* 1995; Cornejo *et al.* 1998, 2006; Tomlinson *et al.* 1999); the Tuina Formation (Raczynski 1963; Marinovic & Lahsen 1984; Mundaca 2002) or its equivalents the Agua Dulce Formation (García 1967; Marinovic & Lahsen 1984), the El Bordo Beds, at 22–24ºS (Ramírez & Gardeweg 1982), the Pantanoso Formation in the Copiapó region, at 27–28ºS (Mercado 1982; Iriarte *et al.* 1996) and the lower part of the Pastos Blancos Group (Guanaco Sonso sequence; 281–260 Ma) (Martin *et al.* 1999*a*) (Fig. 3.7); and other outcrops of Late Carboniferous–Early Permian age that have often been included in the Peine Formation (i.e. Marinovic & García 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, Imilac 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 (Huete *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 two 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 is 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 (Ingaguás 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, Cochiguás 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 ($\pm$hornblende) granodiorites that differ from the Guanta Unit in their higher SiO_2 content and lower K_2O content. The Cochiguás 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, coarse-grained biotite granites, and relative to the Cochiguás Unit it is richer in K_2O; 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 synmagmatic 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 (Llambías & Sato 1990). The age of the El Volcán Unit is younger than the Late Carboniferous (301 ± 4 Ma) age obtained from the Cochiguás 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 synmagmatic foliation in one granitoid association and not in the other synchronous association (see above) suggests that the synmagmatic 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.

Third stage: Early Permian to Middle/Late? Permian

The relatively uniform nature of the marine deposits in this third stage and their widespread outcrop for *c.* 1200 km along the country (Fig. 3.5) records a major palaeogeographic change from conditions prevailing during Late Devonian to earliest Permian (first and second Gondwanan stages) times.

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; García 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 Díaz-Martínez *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-Martínez *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; Harambour 1990; Díaz-Martínez *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 an abundant brachiopod, bivalve and gastropod fossil fauna (Marinovic *et al.* 1995; Niemeyer *et al.* 1997*b*). The formation includes the Cerros de Cuevitas Formation, in the Salar de Navidad region at 23º45′S (Niemeyer *et al.* 1997*b*) (the Cerros de Cuevitas 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 461-m-thick alternation of conglomerates, sandstones and shale intercalations; sandstones are often calcareous, and limestones predominate in the upper portion (Niemeyer *et al.* 1997*b*). 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 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 27ºS, and the Huentelauquén Formation in Chile, and the Arizaro Formation in the Argentinian Puna (Niemeyer *et al.* 1997*b*). Equivalents of the Cerros de Cuevitas Formation in the Antofagasta region are preserved as a limestone outcrop containing productid fossil remains (Niemeyer *et al.* 1997*b*), and the Cerro 1584 Beds.

Northeast of Copiapó, at 27ºS, 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 31ºS and 31º45′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 1991). The lower La Higuera Member represents a transgression–regression sequence that begins with a quartz-rich orthoconglomerate of variable thickness (10 to 80 m) followed by a 150–200-m-thick succession of black shales with a 15–20-m-thick turbiditic succession at its middle part, and a 50–60-m-thick succession of quartz-rich sandstones with pelitic intercalations that diminish towards the top. Coquina horizons intercalated in the dark, pelitic middle part of this member are interpreted as turbiditic deposits. The upper La Cantera Member consists predominantly of limestones (calcirudites, calcarenites and marls) with sandsone and conglomeratic 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) 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

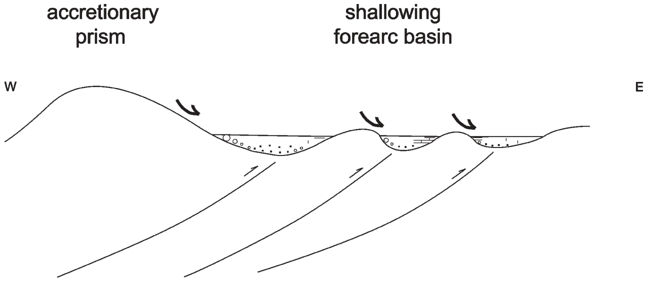


Fig. 3.9. Schematic section across the forearc basin during the third stage of the Gondwanan tectonic cycle and syntectonic sedimentation of the upper member of the Huentelauquén Formation. East-vergent thrusting, probably associated with final accretion of the prism to the continent, caused deformation of the lower deposits of the Huentelauquén Formation, shallowing of the forearc basin and favoured the development of a palaeo-topography of the basin bottom that permitted sedimentation all over the forearc basin of coarse clastic deposits with abundant limestone intercalations (see text). Thick arrows indicate main direction of syntectonic sediment supply.

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-wedging 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 backthrusted towards the continental margin of Gondwana and amalgamated to it. A third and local unconformity separates the two members of the Huentelauquén 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 Huentelauquén Formation at the end of the cycle. The erosional unconformity described by Rivano & Sepúlveda (1991) separating the lower from the upper deposits of the Huentelauquén 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 33°S 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 33°S, the Gondwanan metamorphic complex is almost continuously exposed along the Coastal Cordillera. Between 33°S and 38°S, 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 calcsilicate 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 Trafún 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 metabasites. Ultramafic serpentinized bodies were apparently tectonically emplaced (Godoy & Kato 1990; Kato & Godoy 1995). Metamorphism developed under high P/T ratios and increased westward. Crossite, glaucophane, 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*b*; 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 (^{87}Rb–^{86}Sr 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 (^{87}Rb–^{86}Sr on whole rock) of 311 ± 10 Ma (Late Carboniferous) for glaucophane 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., *Nereites? nahuelbutanus* and *Monograptus (Spirograptus) aff. spiralis spiralis (Gein)*) were found and determined by Tavera (1979*a*, 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.* (1966) 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 meta-ignimbrites (Söllner *et al.* 2000*a*; Duhart *et al.* 2001) and intercalations indicating exhalative origin, like iron formations, massive sulphides, tourmalinite and spessartine 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 (^{40}Ar/^{39}Ar) on muscovite of the same sample indicating a Triassic age for the metamorphic event (greenschist facies) (Antinao *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 (Antinao *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 ^{87}Rb–^{86}Sr 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.* 1997*b*), 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 46ºS

Several isolated exposures of metamorphic complexes occur south of *c.* 46ºS. 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 polydeformed and metamorphosed turbidites, limestones and pyroclastic rocks exposed east of the Patagonian Batholith between Lago General Carrera (46ºS) and the southern part of the Brunswick Peninsula (54ºS). 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 54ºS, consists (like 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 1955*b*; 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.* 2003*a*), 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.* 56º20′S), south of Cape Horn (Davidson *et al.* 1989), and is lithologically similar to the Diego de Almagro Complex. It consists mainly of polydeformed pyroxene-bearing metabasalts with pillow structures and pillow breccias, and minor mélange with inclusions of chert, greywacke, tuff, basalt and limestone. A ^{87}Rb–^{86}Sr 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 southernmost Chilean archipelago.

General discussion on the Gondwanan cycle

Considering the proposition made above in the Overview for the region north of 33ºS, 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. Huentelauquén 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 43º30′S and 47ºS, 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 (Latest Permian–earliest Jurassic)

Tectonic framework

We use the term ‘Pre-Andean’ for the cycle developed after the final phase of assembly of the Gondwana megacontinent 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 megacontinent, 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 megacontinent, 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 & Breitkreutz 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 (Llambías & Sato 1990; Llambías *et al.* 1993; Llambías 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 palaeogeographic 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 palaeogeographic 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 22ºS and 42ºS (Fig. 3.10), and in southern Chile, in the Chonos and Diego de Almagro archipelagos, between 45ºS and 52ºS (see Fig. 3.15). Accordingly, the two regions will be treated separately.

Evolution in northern and central Chile, between 22ºS and 42ºS

The extensional basins tentatively defined north of 42ºS 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–southeastward 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 Quereo–Los Molles; (E) Bío-Bío–Temuco.

Late Palaeozoic and Triassic volcanism is widely distributed (500 000 km^2) in Chile and Argentina between 21ºS and 44ºS and is generally referred to as the 'Permo-Triassic volcanism'. Formal names for the resulting deposits are: Choiyoi Group (Rolleri & Criado Roqué 1968) and Choiyoi Magmatic Province (Kay *et al.* 1989); however, it has received local names in each of the regions where it occurs. The Choiyoi Group has been correlated with the Mitu Group in Peru. This group can be subdivided into two volcanic portions with minor sedimentary intercalations (Llambías 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) (Llambías 1999); two age determinations from the Polvaredas locality in the Mendoza river valley (33ºS) 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. These 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 Choiyoi Group was developed by intense crustal melting under extensional tectonic conditions. For this reason, Llambías & 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 31ºS, extensive and thick volcanic and volcaniclastic deposits have been included in the Pastos Blancos Formation (Thiele 1964), elevated to group status by Martin *et al.* (1999*a*). 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.* 1999*a*). 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 crystalls 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 (Dicroïdium 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 Choiyoi 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 Choiyoi 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 Quereo–Los Molles basins. In the Vallenar region (28º30′ to 29ºS), 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 (32ºS), these deposits correspond to the Pichidangui Formation

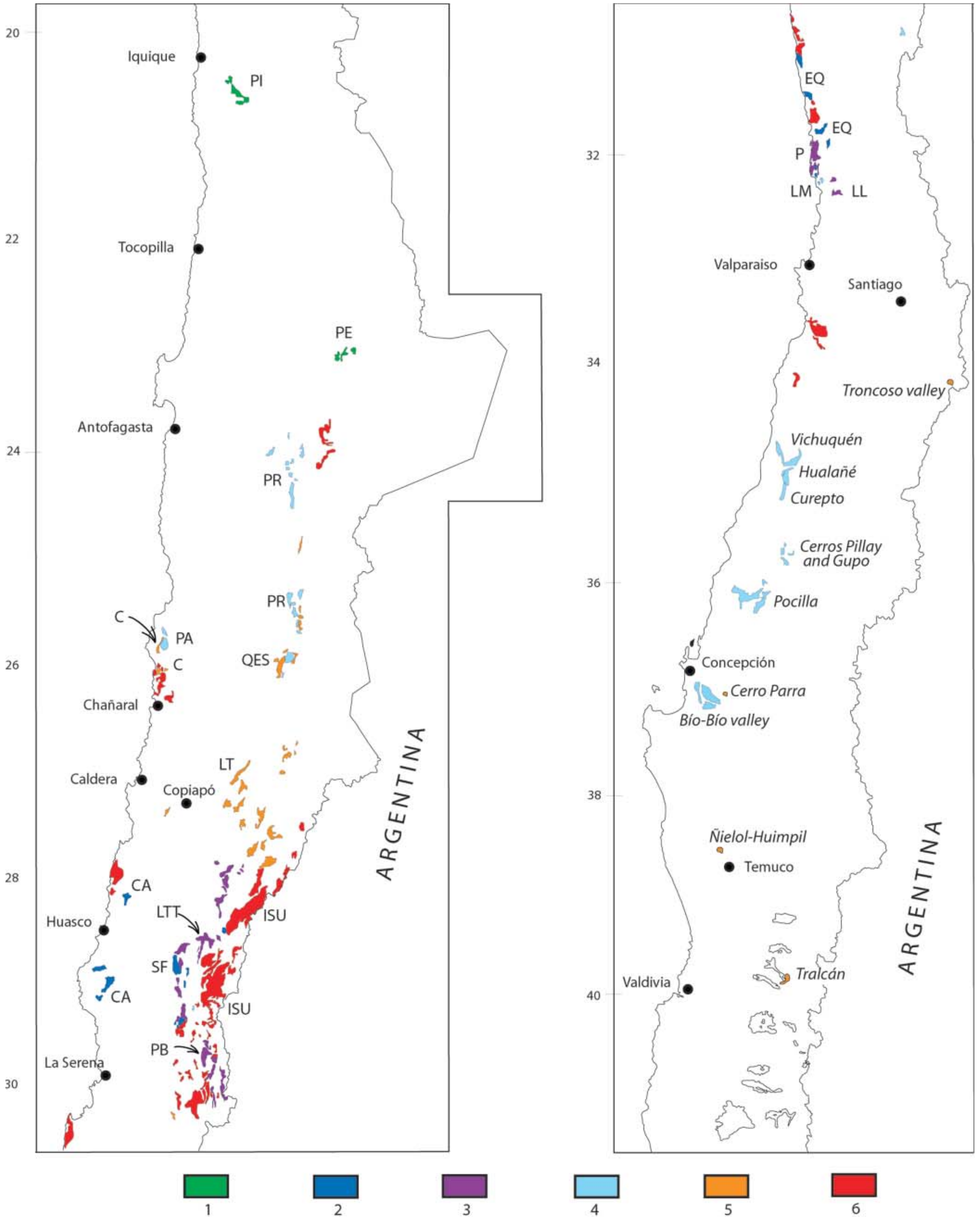


Fig. 3.10. Distribution of pre-Andean sedimentary, volcanic and intrusive units between 20ºS and 41ºS. Key: 1, Older siliceous volcanic and volcaniclastic deposits; 2, marine deposits of the First Stage; 3, siliceous volcanic and volcaniclastic deposits preceding the Second Stage (La Totora–Pichidangui 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.

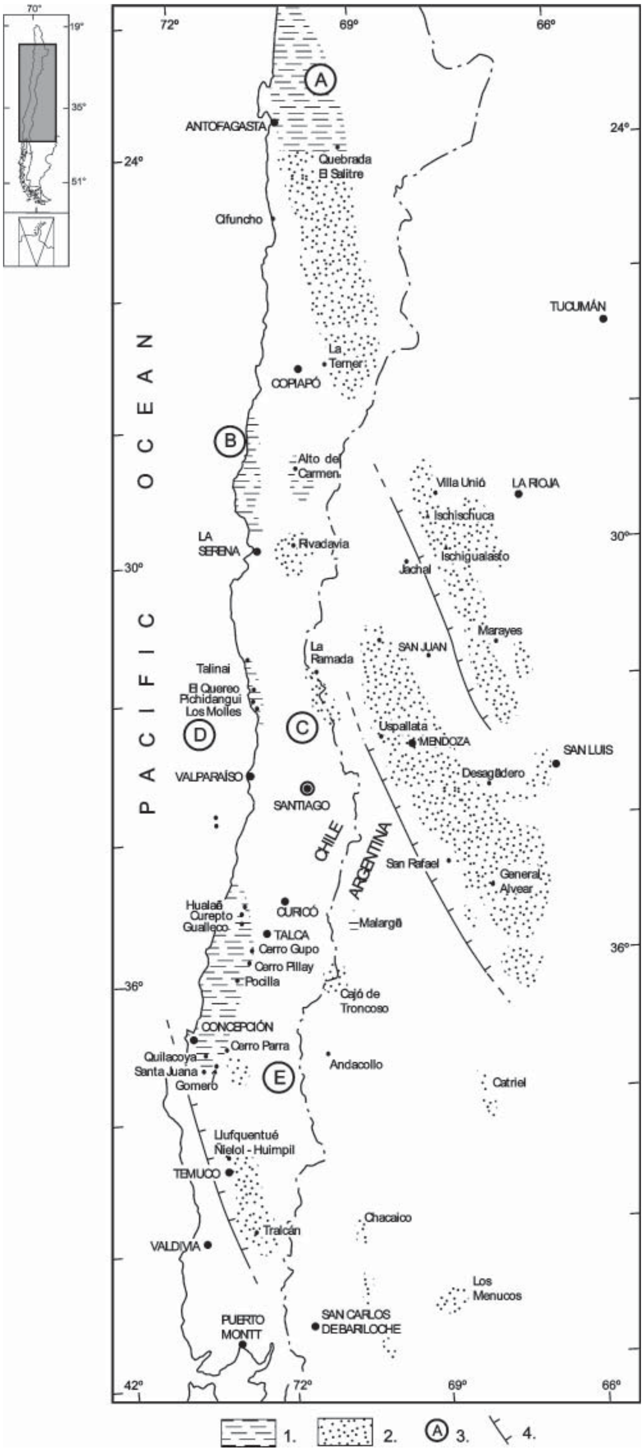


Fig. 3.11. Distribution of pre-Andean deposits in Chile (see text for relevant literature) and neighbouring regions of Argentina, based on Riccardi & Iglesia Llanos (1999), Spalletti (2001), and Stipanicic (2001), between 22ºS and 42ºS, and tentative assignment of these deposits to major pre-Andean extensional 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, Bío-Bío–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.

(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 (35ºS to 35º30′S), in the northern part of the Bío-Bío–Temuco Basin, silicic volcanic deposits are known at the base of the Late Triassic–Early Jurassic deposits in the Crucero de los Sauces Formation (Corvalán 1976), and La Patagua Formation (Hervé *et al.* 1976*b*). Further south, in the southern part of the Bío-Bío–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 & Mopodozis 1996; Iriarte *et al.* 1996, 1999; Arévalo 2005*b*) 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 volcaniclastic 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 Bío-Bío–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.

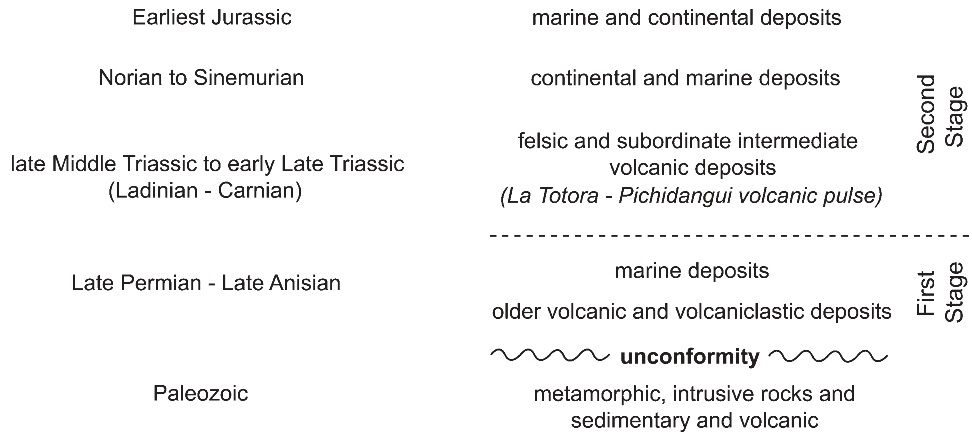


Fig. 3.12. Schematic representation of the two stages of depositional evolution in the Triassic basins in Chile. **First stage** (San Felix and El Quereo–Los Molles basins): rift phase, developed on Palaeozoic units, resulted in a transgression–regression sedimentary cycle (initially associated with local volcanism of Permian age), followed by thermal subsidence (sag) deposits. **Second stage** (Profeta-La Ternera, San Felix?, La Ramada, El Quereo-Los Molles, Bio-Bio–Temuco basins): rift phase, initially associated with intense felsic volcanism (La Totora–Pichidangui volcanic pulse), resulted in continental and marine deposits, followed by a thermal subsidence phase lasting until the earliest Jurassic with development of predominantly marine facies.

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 23ºS and 24ºS (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 Choiyoi Group, *sensu* Llambías & 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 rhyolitic Matahuaico Formation (Dedios 1967) exposed in Elqui river valley at 30ºS, which unconformably underlies the Late Triassic continental Las Breas Formation (Letelier 1977; Moscoso *et al.* 1982*a*). 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.* 1999*a*).

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.* 1982*a*) (Fig. 3.10). The San Félix Formation is exposed in the High Andean range in the El Tránsito drainage basin (28º30′ to 29ºS). 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 volcaniclatic 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 southeastward 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.* 1982*b*), 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 Pichidangui 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 26ºS (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 graben or half-graben (Suárez & Bell 1992). The Cifuncho Formation unconformably overlies the Las Tórtolas Formation, and the Pan de Azúcar Formation is conformably covered by the thick volcanic late Early to Late Jurassic La Negra Formation.

The Late Triassic to Early Jurassic Profeta Formation extends widely across the Domeyko Range (Chong 1977; Chong & von Hillebrandt 1985; von Hillebrandt *et al.* 1986; Suárez & Bell 1992) (Fig. 3.10). This formation transgressively overlies the 1000-m-thick, volcanic Carboniferous–Permian La Tabla Formation (von Hillebrandt *et al.* 1986) and extends its stratigraphic range up to the Tithonian stage. Short-distance facies variations have been reported in these deposits (von Hillebrandt *et al.* 1986) which comprise a shallow marine succession of conglomerates and sandstones with thin fossiliferous intercalations followed by turbiditic intervals, siltstones, limestones, and some conglomeratic intercalations. According to von Hillebrandt *et al.* (1986) the deeper facies occur in the westernmost outcrops. This suggests the proximity to the outlet of the Triassic basin into the open sea.

Further south in the Domeyko Range, the 650–1000-m-thick Quebrada del Salitre Formation, Late Triassic to Sinemurian in age (Naranjo & Puig 1984; Tomlinson *et al.* 1999; Cornejo & Mpodozis 1996, 1997), represents a lateral equivalent of the Profeta Formation and corresponds to the most southeastern marine facies in the Profeta–La Ternera Basin (up to 26º20′S in the Salar de Pedernales region) (Fig. 3.13). This formation comprises two members: a lower volcanic and sedimentary member, and an upper marine sedimentary member (Naranjo & Puig 1984; Tomlinson *et al.* 1999; Cornejo & Mpodozis 1996, 1997). 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.

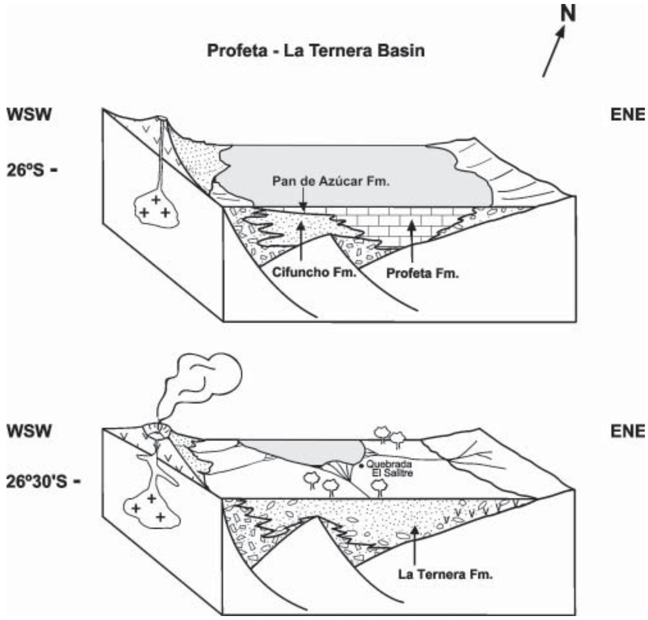


Fig. 3.13. Palaeogeographic and tectonic sketch illustrating the relationships between the Late Triassic–earliest Jurassic deposits of the younger or second stage of pre-Andean evolution in the Profeta–La Ternera Basin in northern Chile: Cifuncho, El Profeta and La Ternera formations (synrift deposits) and the Pan de Azúcar (sag deposit). Type locality of the Quebrada del Salitre Formation (mentioned in the text) is indicated to facilitate comprehension of the palaeogeographic reconstruction. Volcanic rocks in the lower (southern) block diagram in the continental deposits would correspond to the Basaltos de Sierra Doña Inés Chica.

Immediately to the south of the Salar de Pedernales region, at 26º30′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 (28ºS) (Fig. 3.10), exposes a $>$1800-m-thick succession of clastic sediments and an upper 300-m-thick succession of andesitic and basaltic lavas (Segerstrom 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 limestones of the Lautaro Formation, a Late Triassic–Hettangian? age has been assigned to this formation (Solms Laubach & Steinmann 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, mudstones, 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 accumulation 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 Azúcar 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 Azúcar 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 Totora Formation, indicating that some deformation and erosion occurred in Late Triassic times (e.g. Cerro 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 lacustrine environments. Basaltic, andesitic and rhyolitic lavas are also included in this unit (Mpodozis & Cornejo 1988; Pineda & Emparan 2006), and plant remains (Dicroidium Flora) indicate a Late Triassic age. These deposits overlie the silicic volcanic Matahuaico Formation, a local representative of the Choiyoi Group, which is a probable equivalent of the Guanaco Sonso unit of the Pastos Blancos Group (Martin *et al.* 1999*a*) exposed to the NW in the High Andes at 29º30′S. The Tres Cruces Formation consists of a strongly variable succession of conglomerates, 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.* 1982*a*) 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 Azúcar 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 Dicroidium Flora and microflora) unit overlies rhyolitic deposits of the Choiyoi 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 & Iglesia 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) (Raiz 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 times before development of the volcanic arc, which in this region is represented by the Ajial Formation (Thomas 1958).

Further south, at 36ºS, close to the international boundary, the 110-m-thick Cajón de Troncoso Beds correspond to sandstone 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 **Bío-Bío–Temuco Basin** (Fig. 3.11) are located in the Coastal Cordillera, whereas the continental ones are located in the Central Depression and eastern 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 1965*a*, *b*, 1976; Escobar 1976; Hervé *et al.* 1976*b*; Thiele & Morel 1981; Spichiger 1993) that we assign to the La Totora–Pichidangui 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 Bío-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 laterally 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 invertebrates, 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 Bío-Bío–Temuco Basin correspond to the following localities: Cerro Parra, immediately east of the described outcrops of the Bío-Bío river valley, in the western margin of the Central Depression; Llufquentué and Ñielol-Huimpil, next to Temuco; and Tralcán, next to Panguipulli, at 40ºS (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 to the Panguipulli and Tralcán formations of middle to Late Triassic ages (Rodríguez *et al.* 1999). The slightly metamorphosed Panguipulli Formation forms a 750-m-thick succession of rhythmic shales and sandstones and quartz-rich conglomerates deposited in a lacustrine environment, representing distal fan and talus channelized facies. The >800-m-thick Tralcán Formation consists of fluvial and lacustrine conglomerates, sandstones and mudstones, with coal seams being known from the Temuco region (Fritsche 1921; Brüggen 1950; Hauser 1970; Hervé *et al.* 1976*a*; Parada & Moreno 1980; Rodríguez *et al.* 1999). The Tralcán deposits (Aguirre & Levi 1964) unconformably overlie 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 1950; Hauser 1970; Hervé *et al.* 1976*b*; Arrondo *et al.* 1988; Troncoso *et al.* 2004; Herbst *et al.* 2005) and palynomorph content (Askin *et al.* 1981; Zavatierri *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 24ºS and 31ºS (Mpodozis *et al.* 1983; Nasi *et al.* 1985, 1990; Mpodozis & Kay 1990), and along the coastal region between Chañaral (26º30′S) and San Antonio (34ºS), 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.* 27–31ºS) 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 hypersilicic, calc-alkaline 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 Llambías (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, Chollay, El León, El Colorado and La Laguna (Nasi *et al.* 1985, 1990; Mpodozis & Cornejo 1988; Mpodozis & Kay 1990, 1992). These units, except for the La Laguna gabbros, consist dominantly of biotite–hornblende granodiorites and monzogranites, and syenogranites; graphic granites are known from the youngest El Colorado units. Radioisotopic age determinations (K–Ar and ^{87}Rb–^{86}Sr) in the Ingaguás Superunit yielded ages between 276 and 192 Ma that fall in the interval between Early Permian and Early Jurassic (Brook *et al.* 1986; Rex 1987; Mpodozis & Cornejo 1988; Nasi *et al.* 1990). More recently, Martin *et al.* (1999*a*) 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 (muscovite) 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 (26–27ºS), 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 Vetado and Capitana plutons that correspond to leucocratic S-type granitoids with high ^{87}Sr/^{86}Sr 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 (26º20′S) with ages between 202 Ma and 186 Ma and a calcalkaline character (Berg & Breitkreuz 1983; Breitkreuz 1986*a*; 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 28ºS and 30ºS (Moscoso *et al.* 1982*a*), 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 NNW–SSE 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 (33–34ºS), 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.* 31º30′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 (Profeta–La Ternera and Bío-Bío–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. During the second rifting stage only the El Quereo Basin 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

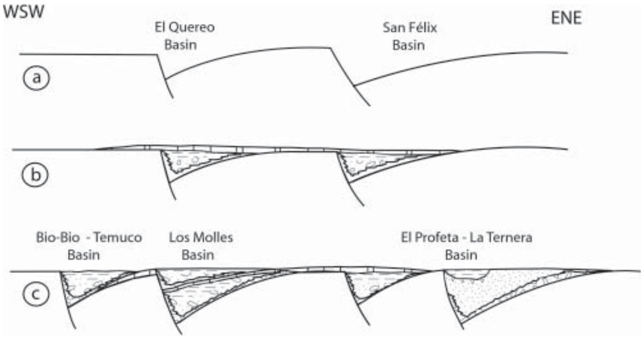


Fig. 3.14. Tentative WSW to ENE section showing rift progression during the pre-Andean tectonic cycle in Chile, between 22º and 42ºS. (**a**) Older or first rifting stage in Late Permian–late Anisian times with development of the El Quereo and San Félix basins, with older volcanic deposits of Late Permian age in the base of the basin fill (synrift and thermal subsidence). (**b**) The La Totora–Pichidangui felsic volcanic episode developed in late Middle Triassic to early Late Triassic (Ladinian–Carnian) on the areas of the previous El Quereo and San Félix basins. (**c**) Younger or second rifting stage in Norian to Sinemurian–Pliensbachian times (synrift and thermal subsidence) with development of the external basins: the Profeta–La Ternera Basin located east-northeastward of the previous San Félix Basin, and the Bío-Bío–Temuco Basin located west-southwestward of the previous El Quereo Basin. Note that the felsic volcanic episode only partially reached the Bío-Bío–Temuco basin area.

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 & Alcober (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 (26º30′S to 27ºS), and by Suárez & Bell (1992) for the Profeta–La Ternera Basin, and Bell & Suárez (1995) for the Los Molles Formation. The La Totora–Pichigangui 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 Bío-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 43º30′S and 47ºS

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 45º25′S, and Hervé & Fanning (2003) discovered Triassic zircons in the Diego de Almagro Metamorphic Complex, at 51º30′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.* 1981*b*; 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 (i.e. La Negra and Ajial formations).

The Diego de Almagro is an accretionary 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

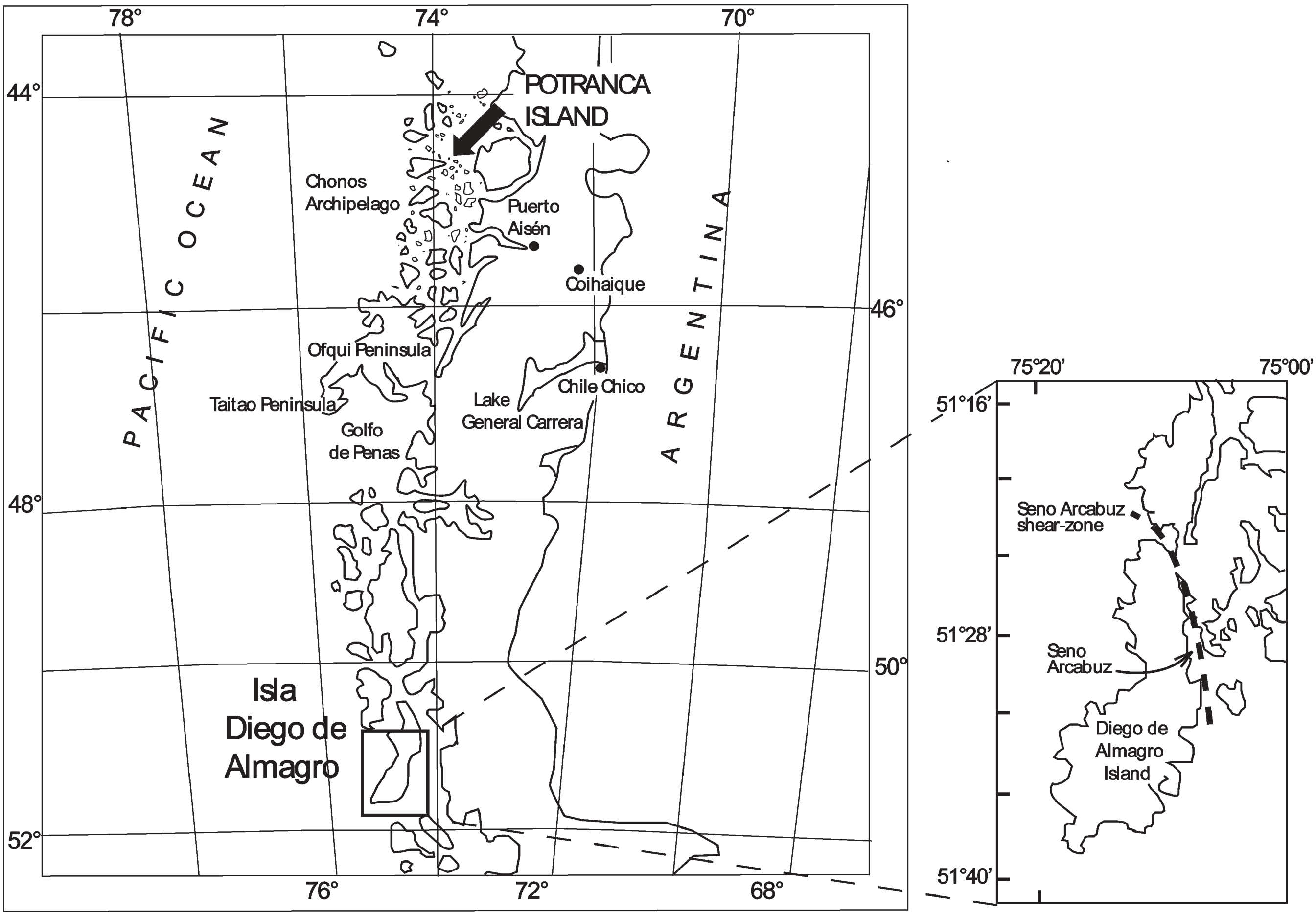


Fig. 3.15. Location of the Potranca and Diego de Almagro islands in the Patagonian Archipelago. Triassic marine fossils from the Potranca Island, on the eastern side of the Chonos Metamorphic Complex, U–Pb SHRIMP and fission track age determinations reveal a Late Triassic to Early Jurassic age for this subduction complex (Hervé *et al.* 1981*b*; Fang *et al.* 1998; Hervé & Fanning 2001; Thompson & Hervé 2002). In the Diego de Almagro Island, the Seno Arcabuz Shear Zone consisting of mylonitized granitic rocks of Middle Jurassic age separate the Diego de Almagro accretionary complex, to the SW, from the Madre de Dios Complex, to the NE. This situation suggests that the Middle–Late Jurassic anatectic magmatism extended as far as the margin of the continent at that time.

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.* 1990), 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 39ºS; and (2) southern Chile (Patagonia), south of 42ºS. The area between these two regions represented some kind of *c.* NNW–SSE orientated swell separating the basins developed north of 40ºS and south of 42ºS. 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 39ºS) 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

PERIODS*	STAGES**	SUBSTAGES**	AGE
Late Period	Third Stage		Late Paleogene to Present
	Second Stage	Second Substage	Early Paleogene (Paleocene-Early Eocene)
		First Substage	late Early Cretaceous to Late Cretaceous
Early Period	First Stage	Second Substage	Kimmeridgian-Tithonian to Albian
		First Substage	late Early Jurassic to Kimmeridgian

* 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.

described by Coira *et al.* (1982) for this tectonic cycle. However, each of these periods can be subdivided into shorter stages, which can be differentiated from each other by major palaeogeographic changes (Fig. 3.16). These changes are the consequence of modifications of the convergence and subduction pattern in this region. An important feature is the development of major trench-parallel shear zones developed along the axes of the successive magmatic arcs, i.e. the Atacama, Domeyko, and the Liquiñe–Ofqui fault zones, all of which can be traced for several hundreds of kilometres along the mountain range.

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.

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.

Only the eastern portions of this Jurassic arc are preserved in northernmost Chile, whereas towards the south exposures of the roots of the arc and the volcanic deposits are more extensive, suggesting that the arc was slightly orientated to the NW, and that subduction erosion of the continental margin has probably destroyed more of this continental margin in northern Chile than further south.

In this first Andean stage it is generally possible to separate two substages (Fig. 3.16). This separation is clear for the backarc basin deposits, although it is less evident for the arc deposits and very difficult to establish for the plutonic units. The first substage (late Early Jurassic to Kimmeridgian) is characterized by intense activity in the arc and development of a transgressive–regressive marine cycle in the backarc basin. The second substage (Kimmeridgian to Aptian–Albian) is characterized by apparently less activity in the arc in some regions, and by a second transgressive–regressive marine cycle in the backarc basin, except in the Iquique and Antofagasta regions where continental conditions were maintained in the backarc. The existence in the backarc basin of regions that were covered by the sea and others that remained under a continental regime indicates considerable palaeogeographic variations along the basin. At the end of the first substage the sea progressively retreated from the backarc basin and gave rise in Oxfordian and Kimmeridgian times to the deposition in most regions of thick gypsum levels. The transition between the backarc deposits of both substages is generally continuous.

The following description of the evolution of the first stage of Andean evolution north of 39ºS is subdivided into three

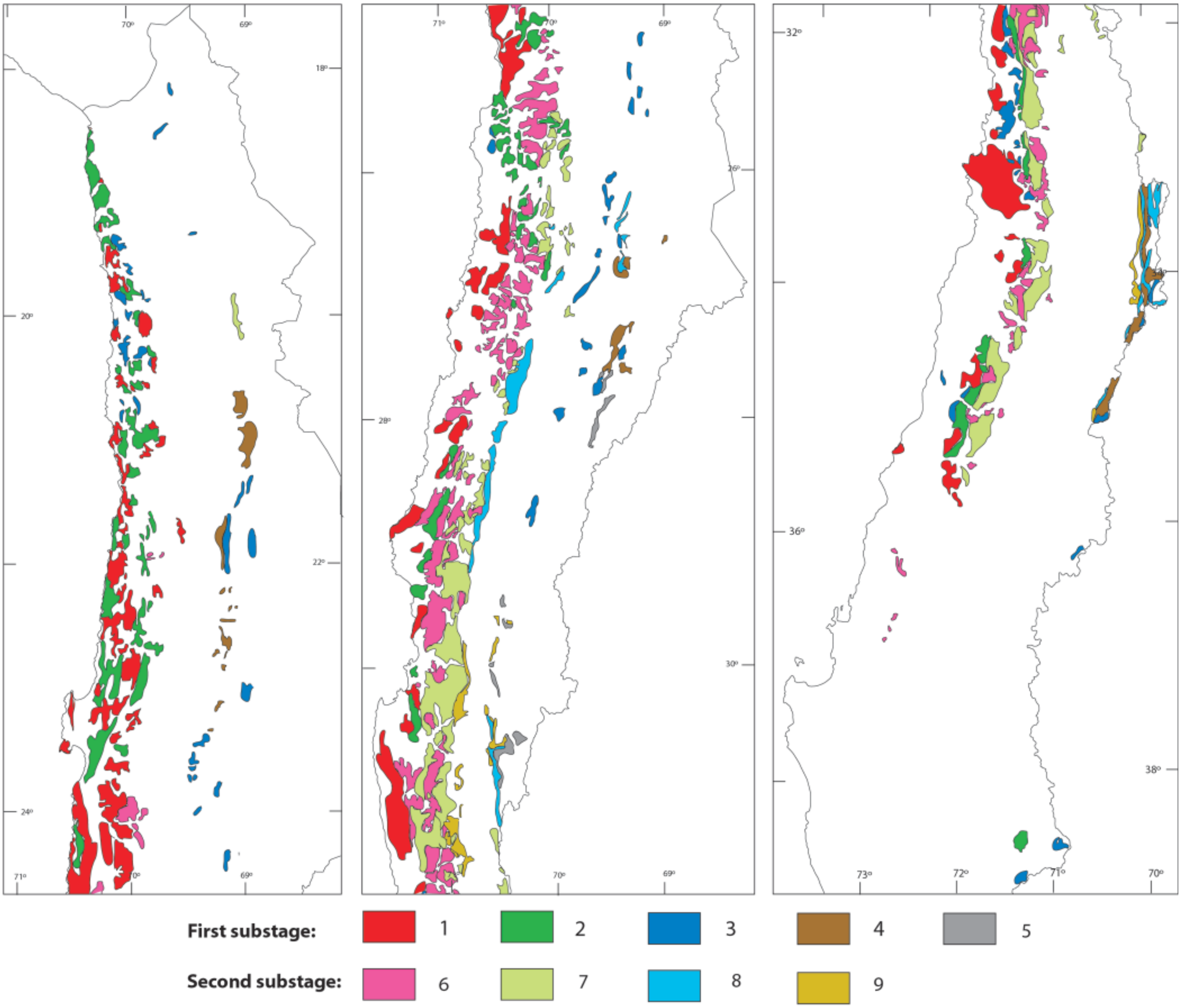


Fig. 3.17. Distribution of units of the first (late Early Jurassic to Kimmeridgian) and second (Kimmeridgian to Aptian–Albian) substages of the first stage of Andean evolution (late Early Jurassic to Aptian–Albian) (see Fig. 3.16). Note distribution of arc rocks close to the present-day coastline and narrowing of the outcrops of plutonic rocks toward the north. First substage: 1, intrusive rocks; 2, arc deposits; 3, backarc marine deposits; 4, backarc continental deposits; 5, Kimmeridgian deposits (red continental clastic successionsand lavas). Second substage; 6, intrusive rocks; 7, arc deposits; 8, backarc marine deposits; 9, backarc continental deposits.

regions: Arica to Chañaral (18ºS to 26ºS), Chañaral to La Serena (26ºS to 30ºS) and La Serena to Valdivia (30ºS to 39ºS).

Northernmost Chile between Arica and Chañaral (18ºS to 26ºS)
Arc 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; Tobar *et al.* 1968; Vogel & Vila 1980; García *et al.* 2004), in the Arica region (18–19ºS); 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.* 26ºS). 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 (Muñoz 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; Muñoz 1989; Muñoz *et al.* 1988*b*; Godoy & Lara 1998; González & Niemeyer 2005). Intensive $^{40}Ar/^{39}Ar$ dating of the La Negra Formation between 18º30′S and 24ºS 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.* 1988*b*; 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 $^{87}Sr/^{86}Sr$ 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 transtensional 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.

Arc 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 Arica region, the Late Jurassic–Early Cretaceous Atajaña 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 Atajaña Formation is conformably covered by the marine Neocomian Blanco Formation (Cecioni & García 1960). Furthermore, pyroclastic components in the upper Neocomian part of the Livilcar Formation (Muñoz *et al.* 1988*a*) confirm the existence of volcanic activity at this time. A similar situation exists in the Antofagasta region; here, the stratigraphic equivalent of the Atajaña Formation is the Caleta Coloso 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 Taltal–Chañaral region, volcanic rocks of this second substage correspond to the >1000-m-thick Aeropuerto Formation (Ulricksen 1979; Naranjo & Puig 1984; Marinovic *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 Gabbroic Complex, Middle Jurassic Cerro Bolfín Complex, and Late Jurassic Cerro Cristales Pluton (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 Mackenna Batholith that forms the bulk of the Coastal Cordillera in this region, and consists of several units of Jurassic and Early Cretaceous ages (Marinovic *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 ^{87}Rb–^{86}Sr 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 Mackenna Batholith (Marinovic *et al.* 1995), and the two younger units of this batholith, Remiendos and Herradura. All these units have ages (K–Ar and ^{87}Rb–^{86}Sr 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 (18–19°S), it corresponds to the *c.* 1700-m-thick, Sinemurian to Early Cretaceous (Neocomian) Livilcar Formation (Muñoz *et al.* 1988*a*). 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.* 1988*a*;

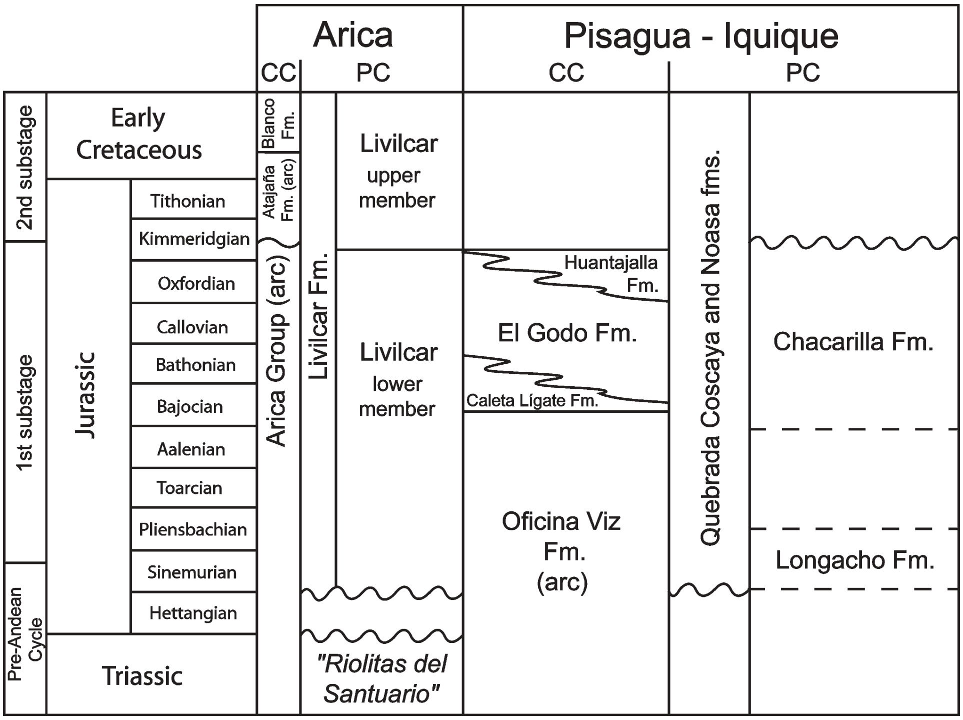


Fig. 3.18. Chronostratigraphic relations betwen arc and backarc units during the first substage of the first stage of the Andean tectonic cycle for the Coastal Cordillera (CC) and Precordillera (PC) in the Arica and Pisagua–Iquique regions (18° to 20°30′S).

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.* 1988*a*).

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 Lígate 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 19°30′S to 20°S, the backarc deposits representing the two substages correspond to the Sinemurian to Neocomian Quebrada Coscaya and Noasa formations (Harambour 1990), which are the southern equivalents of the Livilcar Formation, although with a continental upper part. Immediately to the south, between 20°S and 20°45′S, 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, ornitopods) 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 ornitopod 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. Comformably 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 gypsum intercalations of Oxfordian age (Maksaev 1978; Ladino 1998) that mark the end of the marine cycle and separate the lower from the upper member. A continental, upward-coarsening upper member comprises a *c.* 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 1978*a*; Vergara & Thomas 1984) and the Capella Beds (Vergara 1978*b*), exposed towards the NE in the eastern Precordillera.

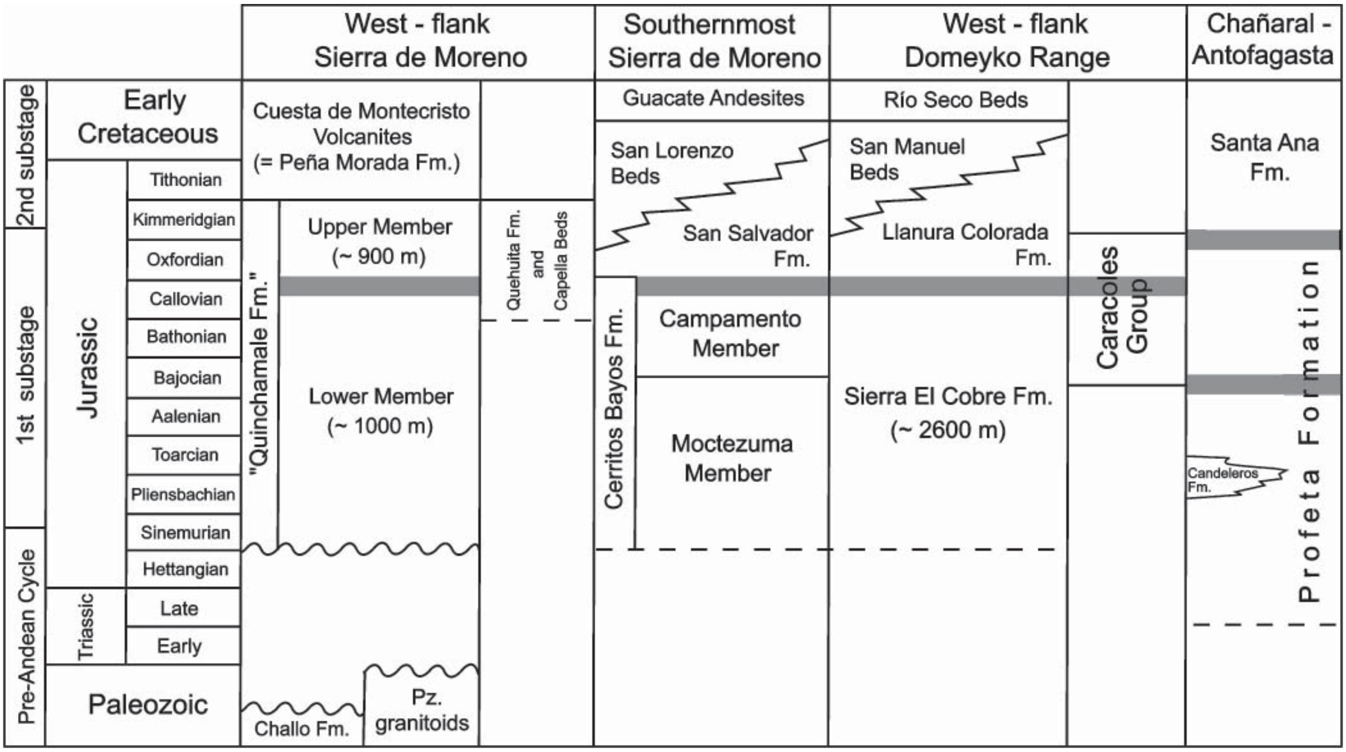


Fig. 3.19. Chronostratigraphic relations between units of the first substage of the first Andean stage along the west flanks of the Sierra Moreno and Domeyko Range, in the Antofagasta region (22º to 24ºS). Dark bands represent gypsum intercalations indicating the first regression in Late Jurassic times.

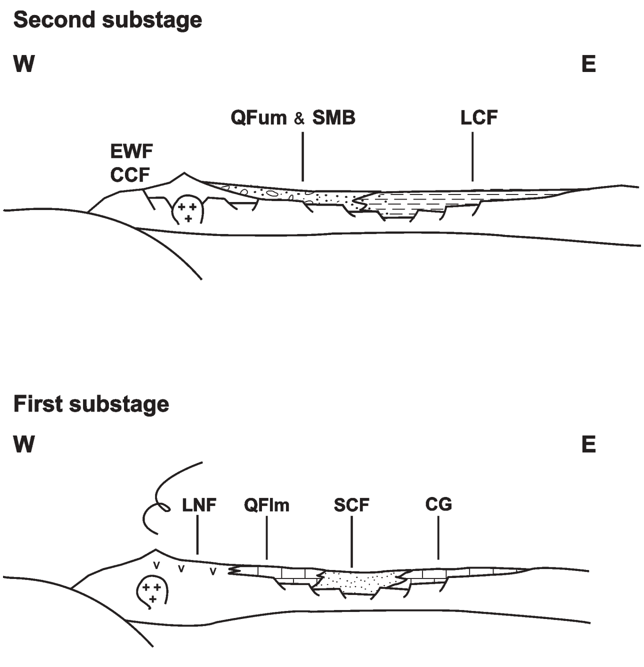


Fig. 3.20. Schematic palaeogeographic sections for the first and second substages of the first Andean stage in northernmost Chile, based on the region of Antofagasta. Note the forearc deposits in the second substage. Abbreviations for the first substage: LNF, la Negra Formation; QFlm, 'Quinchamale Formation', lower member; SCF, Sierra El Cobre Formation; CG, Caracoles Group. Abbreviations for the second substage; CCF, Caleta Coloso Formation; EWF, El Way Formation; QFum, 'Quinchamale Formation', upper member; SMB, San Manuel Beds; LCF, Llanura Colorada Formation.

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 Sierra El Cobre Formation corresponds to the deepest deposits of the backarc basin, and consists of a 2600-m-thick succession of calcarenites, limestones, turbidites and shales, with gypsum intercalations at its upper part. The Bajocian to Kimmeridgian Caracoles Group is composed of five formations: from base to top, the Torcazas, Caracoles, Doralisa, Millonaria and Honda formations (García 1967; Ramírez & Gardeweg 1982; Marinovic & García

1999). This group comprises a >1300-m-thick fossiliferous succession of tuffs, 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 fluvial 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 Montecristo Volcanites on the western flank of the Domeyko Range (Ladino 1998) and the Guacate Andesites 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 corresponds 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 (26ºS to 30ºS)

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 Fomation (Jensen 1976; Soffia 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 (Soffia 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 (Reutter 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.* 1982*a*) or Formation (Segerstrom 1960; Arévalo 2005*a*) 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 2005*a*), and underlies the marine backarc deposits of the Chañarcillo 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 (Algarrobos) (Marschik & Fontboté 2001*b*) 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 2001*b*; 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 Chañarcillo Group (Segerstrom 1960; Segerstrom & Ruiz 1962; Jurgan 1977*a*, *b*; Arévalo 1995, 2005*a*). 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.* 1982*a*). In this region, the Cerrillos Formation unconformably overlies the transitional arc–backarc Bandurrias Group and the backarc Chañarcillo Group (Arévalo 2005*a*, *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 Ovalle Group by H. Thomas (1967). This unit forms a series 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 tuffs with limestone

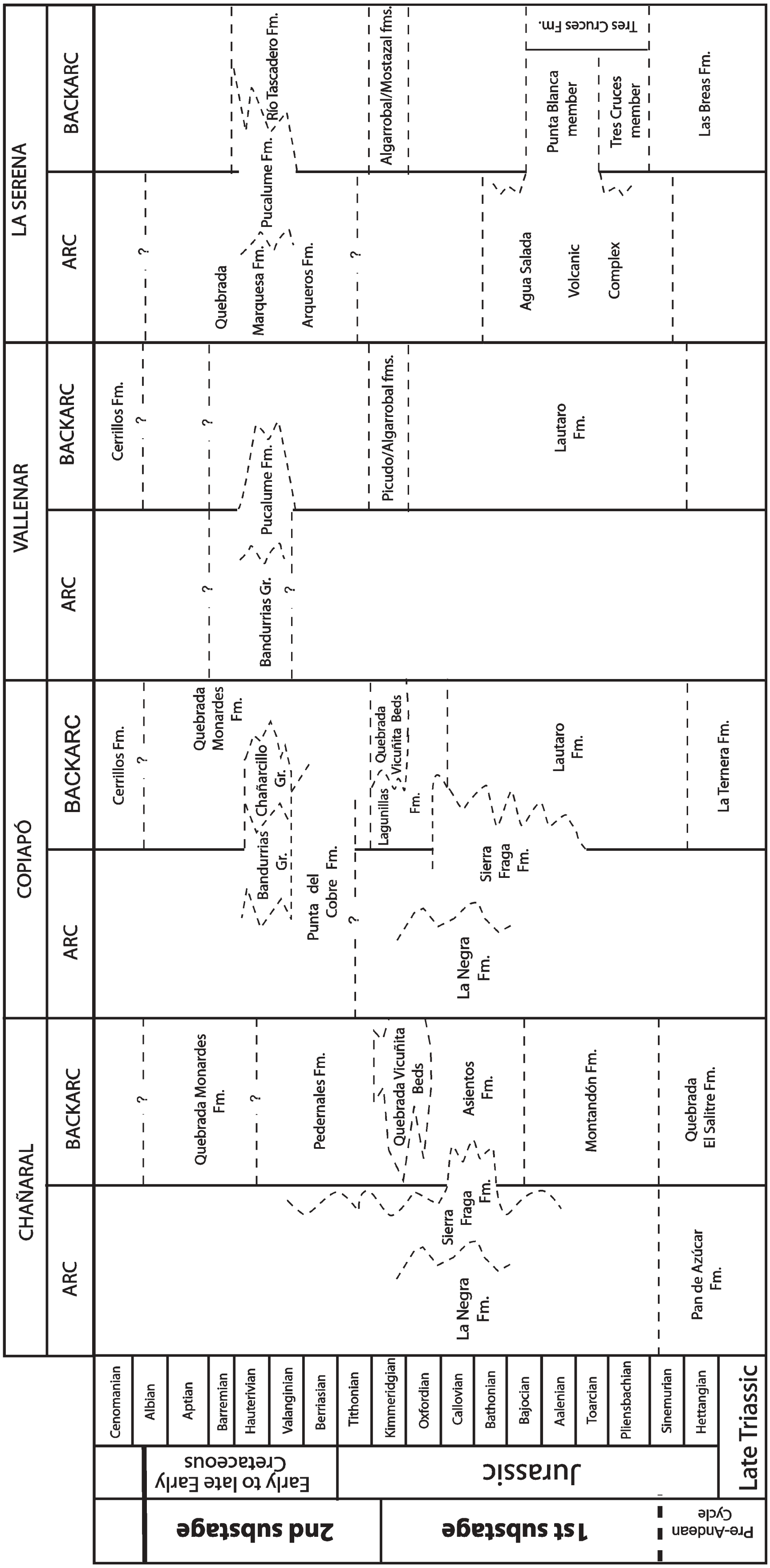


Fig. 3.21. Chronostratigraphic relations between arc and backarc during the first stage of the Andean tectonic cycle in the Chañaral, Copiapó, Vallenar and La Serena regions.

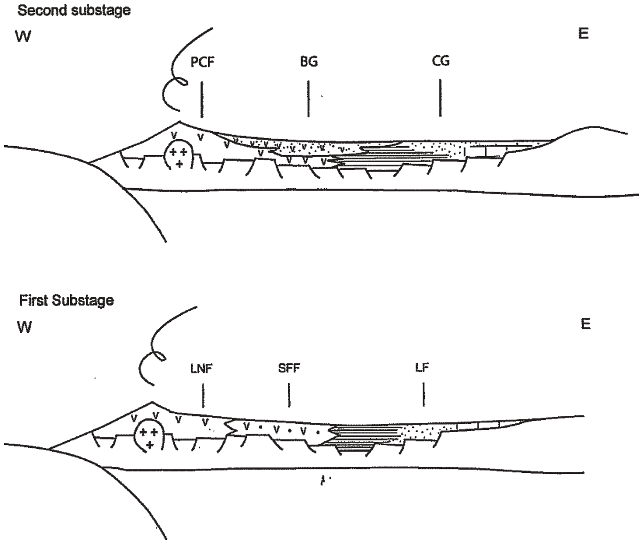


Fig. 3.22. Schematic palaeogeographic cross-sections for the first and second substages of the first Andean stage in the Copiapó region. Abbreviations of stratigraphic units: BG; Bandurrias Group; CG, Chañarcillo Formation; LF, Lautaro Formation; LNF, La Negra Formation; PCF, Punta del Cobre Formation;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

intercalations, ignimbrites with intercalations of andesitic lavas and subvolcanic porphyric andesitic, and dioritic stocks, sills and dykes. The enormous thickness of this complex and the presence of calcareous deposits indicate 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

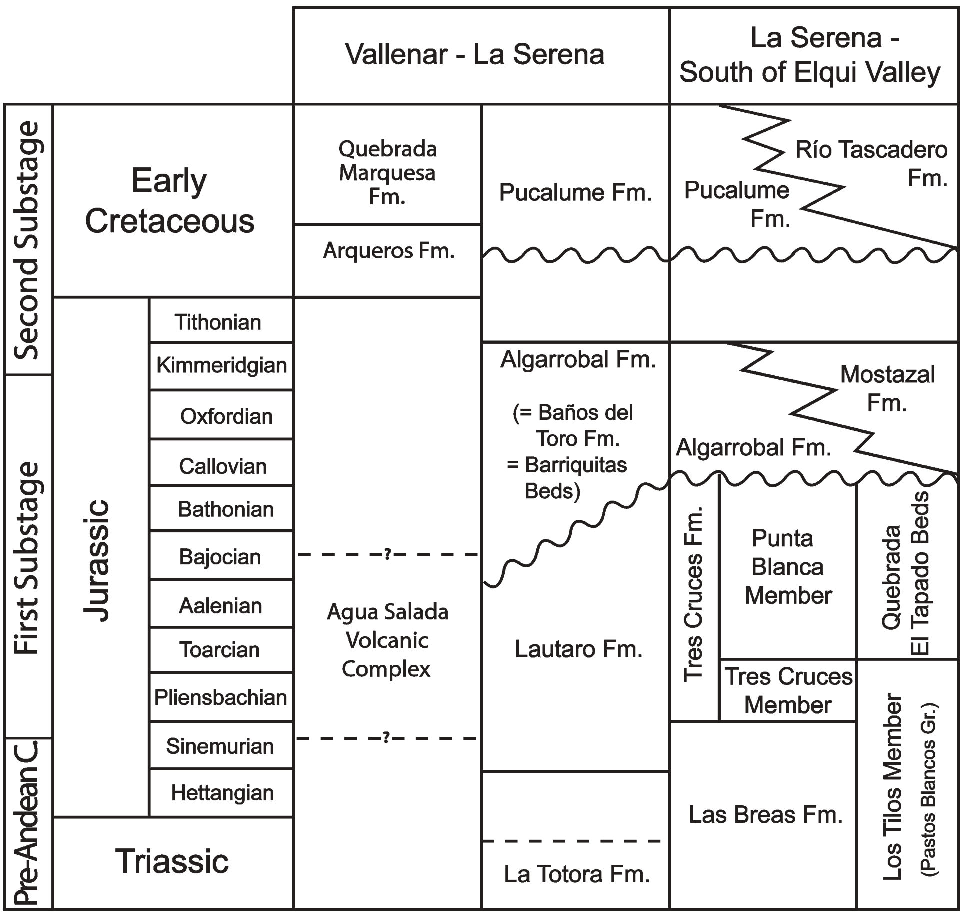


Fig. 3.23. Chronostratigraphic relations between arc and backarc deposits of the first and second substages of the first stage of the Andean tectonic cycle in the regions between Vallenar and La Serena (28°30′S to 30°S), and south of the Elqui River valley (30° to 31°30′S).

major structural feature, the Romeral Fault Zone (Emparan & Pineda 1999, 2000, 2005; Pineda & Emparan 2006), and comprise thick volcaniclastic deposits, and andesitic, and subordinately rhyolitic, 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 (ocoites) containing bitumen in amiydules 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 $^{40}Ar/^{39}Ar$ plateau age on amphibole of 129 ± 3.3 Ma. The eastern equivalent of the Arqueros Formation is the marine Río 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 also 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 Río Tascadero Formation further east. The Quebrada Marquesa Formation is overlain by the Late Cretaceous Viñita 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 Río Tascadero formations, and containing Hauterivian 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 radioisotopic age determination from a higher level in this unit yielded an age of 107.0 ± 0.6 Ma (U–Pb on zircons), indicating that the age range of the Quebrada Marquesa Formation is Hauterivian to middle Albian. According to Morata & Aguirre (2003), the high Al_2O_3 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 coeval 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 Chañaral (Mercado 1978; Breitkreuz 1986*a*; 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 Ánimas, with ages between 160 Ma and 150 Ma (Lara & Godoy 1998). These subduction-related plutons have $^{87}Sr/^{86}Sr$ 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 Áspera 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 Brea (123–117 Ma) plutons (Lara & Godoy 1998; Godoy *et al.* 2003; Arévalo 2005*a*, *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; Soffia 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 (Segerstrom 1959, 1968) further south, up to 30ºS (Nasi *et al.* 1990) (Fig. 3.21). These successions are equivalent to the southward prolongation of the Profeta Formation. The Montandón Formation (Pliensbachian 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, green 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 Vicuñita 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 Vicuñita 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.

Adjacent to the city of Copiapó, deposits of the first substage and the beginning of the second substage are not exposed. To the east of Copiapó, in the Precordillera, the Lautaro, Quebrada Vicuñita and Lagunillas formations are conformably overlain by the regressive, continental red sandstones of the late Neocomian to late Early Cretaceous Quebrada Monardes Formation (Muzzio 1980; Mercado 1982; Iriarte *et al.* 1999); the Pedernales Formation is not exposed in this region. The Lautaro Formation, an equivalent of the Montandón and Asientos formations (Fig. 3.21), consists essentially of marine calcareous sandstones, marls and limestones with sporadic volcanic intercalations. The thickness is variable, usually being some hundreds of metres (thinning to the east), but occasionally can reach over 1000 m. The rich fossil content places this formation as Sinemurian to Bajocian in age, possibly also to Callovian in the eastern exposures (von Hillebrandt 1970, 1971, 1973, 1981*a*, *b*, 2002; Reutter 1974; Jensen 1976; Jensen & Vicente 1976; von Hillebrandt & Schmidt-Effing 1981; Mercado 1982; Sepúlveda & Naranjo 1982; Soffia 1989; Nasi *et al.* 1990; Iriarte *et al.* 1999). The Lautaro Formation is overlain by the Quebrada Vicuñita (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 Vicuñita 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 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 1977*a*, *b*; Arévalo 1995, 2005*a*, *b*). The lower 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 & Ruiz 1962; Moscoso *et al.* 1982*a*). In the Copiapó region (Copiapó Basin), the richly fossiliferous deposits of the Chañarcillo Group have been subdivided into four formations: from base to top, Abundancia, Nantoco, Totoralillo and Pabellón (Tavera 1956; Segerstrom 1960; Segerstrom & Parker 1959; Corvalán 1974; Jurgan 1977*a*, *b*; Pérez *et al.* 1990; Mourges 2004; Arévalo 2005*a*, *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 Totoralillo 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 west-vergent thrust (Cerrillos Thrust in the Copiapó region; Arévalo 1995, 2005*a*, *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, 2005*a*, *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 1973; Reutter 1974; Nasi *et al.* 1990; Martin *et al.* 1999*a*) (Fig. 3.22). The distribution along north–south orientated outcrops, controlled by mainly east-vergent thrust faults, and the main lithological features of these deposits, do not differ essentially from the ones described for the Copiapó region further north (see above). Here, the grain size of detrital components and their abundance increases eastward, and the thickness of the deposits decreases in the same direction, indicating proximity to the eastern margin of the basin (Nasi *et al.* 1990). In the northern part of this region, the Lautaro Formation is overlain by the Picudo Formation which comprises lavas, tuffs, volcanic breccias, conglomerates and sandstones (Reutter 1974) (Fig. 3.21). Its southern equivalents (south of 30ºS), namely the Tres Cruces Formation and the Quebrada El Tapado Beds (Mpodozis & Cornejo 1988), are unconformably overlain by the Algarrobal Formation (Nasi *et al.* 1990; Mpodozis & Cornejo 1988) (Fig. 3.23).

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.* 30ºS and 31ºS (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 Río 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 (30ºS), 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 Vicuñita 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 Late Jurassic–Early Cretaceous transgression–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 34ºS.

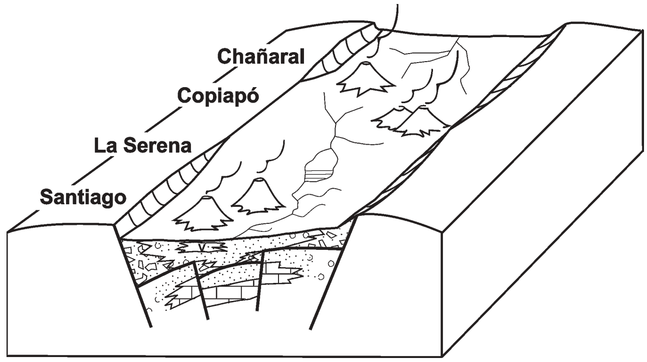


Fig. 3.24. Palaeogeographic model for the Kimmeridgian extensional event, between 26ºS and 36ºS, explaining the existence in the backarc domain of thick continental successions consisting mainly of coarse red detrital deposits locally with thick andesitic intercalations, and the unconformable relationship in some regions between these deposits and the Early to Middle Jurassic marine successions.

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 Río 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 (Río 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 Río 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 (30ºS to 39ºS): 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.* 42ºS). 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

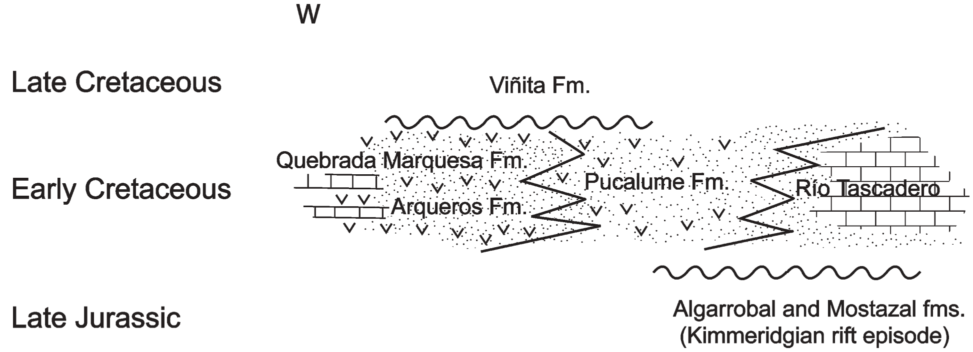


Fig. 3.25. East–west relationship between latest Jurassic–Early Cretaceous arc and backarc deposits at the latitude of La Serena. Arc deposits consist of the Arqueros and Quebrada Marquesa formations and the backarc deposits correspond to the Río Tascadero Formation; the calcareous marine intercalations of the Arqueros and Marquesa formations possibly correspond to deposits accumulated on a west-facing coast separated from the backarc by the relief formed by the arc. The Pucalume Formation represents the transitional deposits between the arc and the backarc.

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? fossiliferous 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-Andean 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 continental volcaniclastic deposits, but also contains deltaic, volcaniclastic turbiditic 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 & Tosdal (1996) in the Coastal Cordillera west of Santiago (see Plutonic activity below).

In the Curicó región (35ºS), the Jurassic arc activity is represented by the Middle Jurassic Altos de Hualmapu 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-Andean cycle. South of 35º30′S, these deposits have fewer exposures along the Coastal Cordillera, and disappear altogether south of 36ºS (Sernageomin 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 Oscuro and Cavilolén 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 Huentelauquén Formation and the pre-Andean 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-kinematic). 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 Oscuro and Cavilolén (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, tonalities, granodiorites and granites (Gana & Tosdal 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 both of the Algarrobal and Mostazal formations,

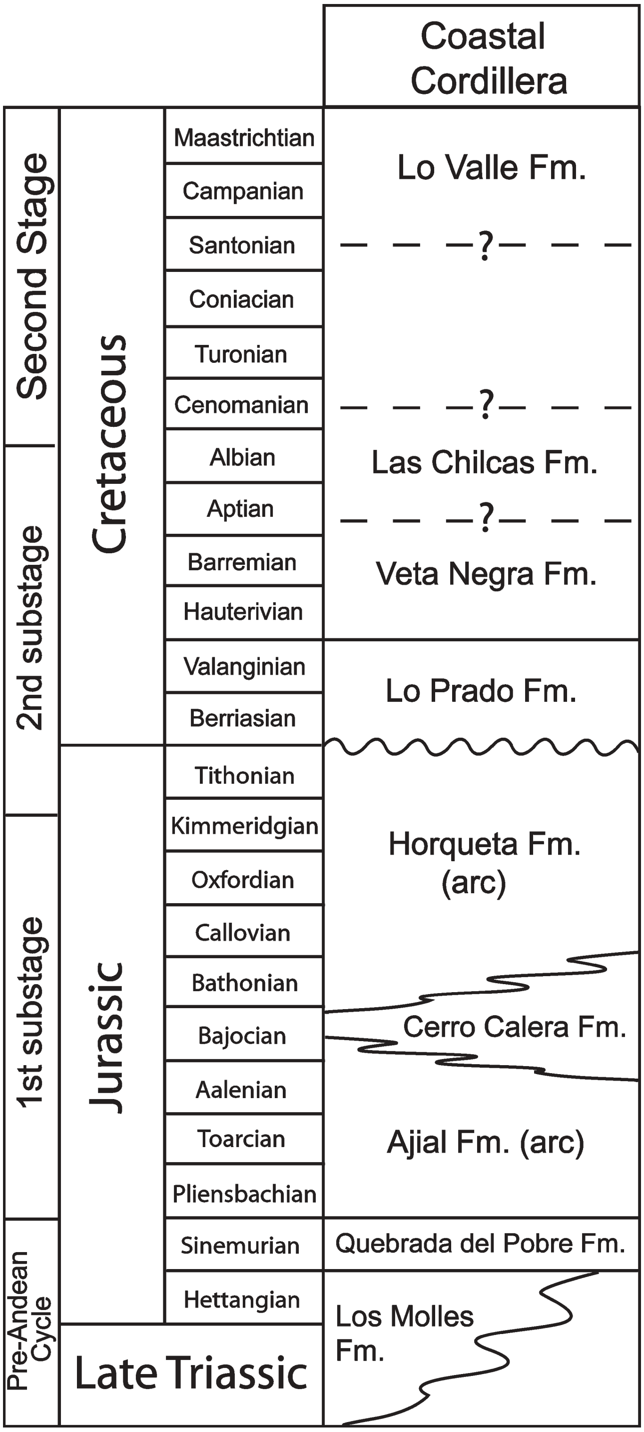


Fig. 3.26. Late Triassic to Cretaceous stratigraphic succession for the Coastal Cordillera in central Chile, between 32ºS and 34ºS.

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 33ºS and 39ºS 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 Rio Colina Formation (Thiele 1980), the Nieves Negras Formation (Alvarez *et al.* 1997; Charrier *et al.* 2002*a*) (formerly Leñas–Espinoza Formation, defined by Klohn (1960) and redefined by Charrier (1982)), and further south at 35ºS to the Nacientes del Teno Formation (Klohn 1960; Muñoz & Niemeyer 1984), Valle Grande Formation of González & Vergara (1962) at 35º30′S, and the Nacientes del Biobío Formation (De la Cruz & Suárez 1997; Suárez & Emparán 1997) at 38º30′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 Bío-Bío 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 rhyolitic rocks of possible Triassic age at 35ºS (Davidson 1971; Davidson & Vicente 1973), and of confirmed Triassic age (Cajón de Troncoso Beds) between 36ºS and 37ºS (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 Principal' (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 33ºS (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 35ºS), the Río Damas Formation consists of a *c.* 3000-m-thick red continental, detrital succession, with coarse 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 Rio 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 cyle 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

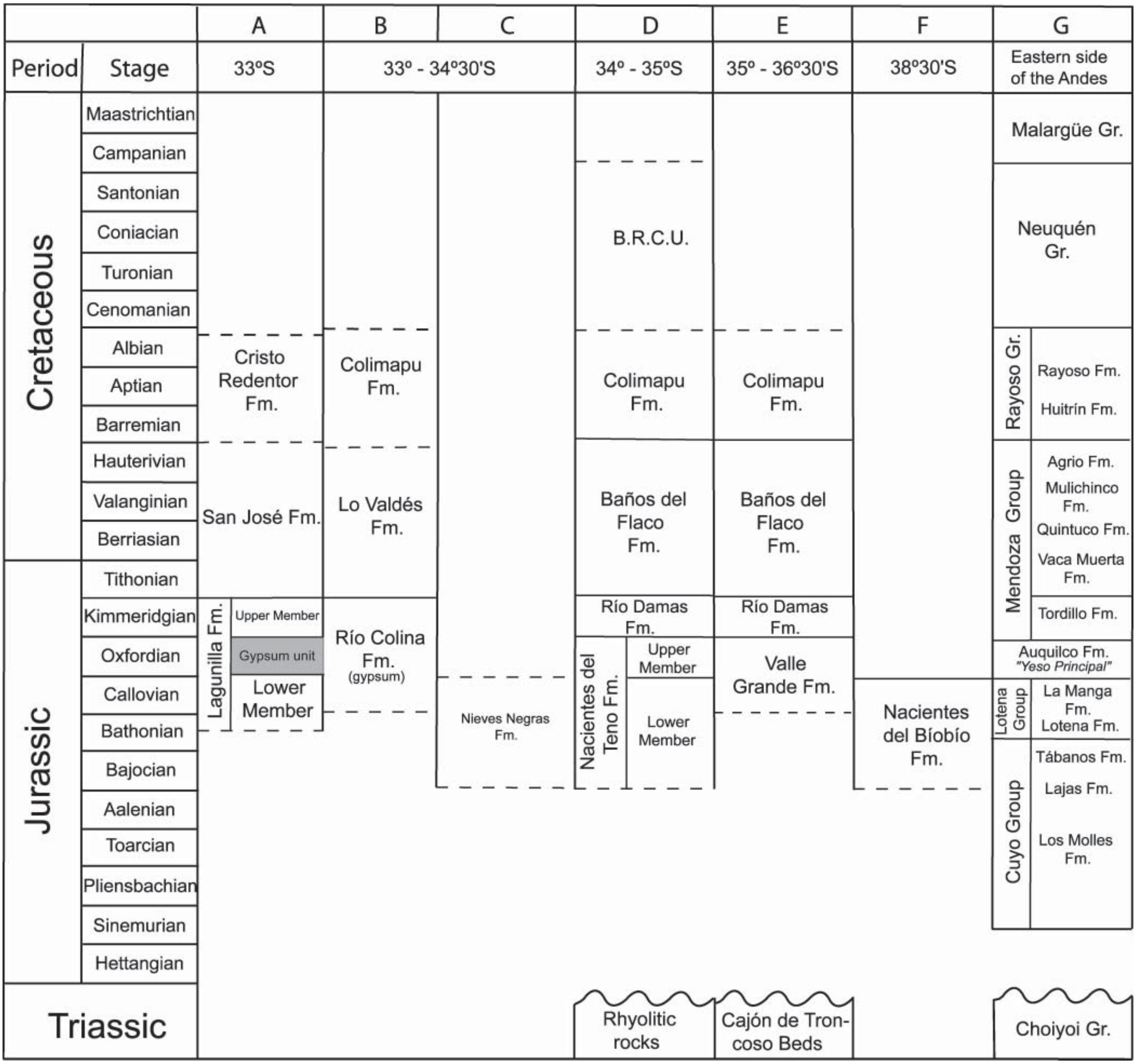


Fig. 3.27. Stratigraphic succession for Jurassic to Cretaceous deposits in the Principal Cordillera in central Chile and Argentina, between 32ºS and 39ºS, based on: (A) Aguirre (1960); (B) González (1963), Thiele (1980); (C) Álvarez *et al.* (1997); (D) Klohn (1960), Davidson (1971), Davidson & Vicente (1973), Charrier (1973*b*, 1981*b*), Charrier *et al.* (2002*a*); (E) González & Vergara (1962), Muñoz & Niemeyer (1984); (F) De la Cruz & Suárez (1997), Suárez & Emparán (1997); (G) Legarreta & Gulisano (1989).

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 33ºS, 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 (33º30′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.* 35ºS. 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))

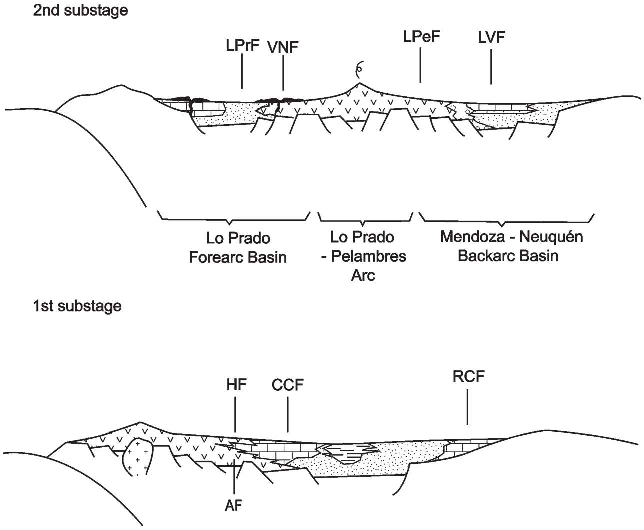


Fig. 3.28. Schematic palaeogeographic cross-section for the first and second substages (late Early Jurassic to late Early Cretaceous) of the first stage of Andean evolution in central Chile between 32ºS and 33ºS. Because the western deposits represented in the two sections are exposed in the Coastal Cordillera and the eastern ones crop out in the Principal Cordillera, the middle parts of the sections presently located in the Central Depression are inferred. Note the different palaeogeographic organization developed during the second substage with, from west to east, a forearc basin, an arc domain and a backarc basin. Abbreviations of stratigraphic units: AF, Ajial Formation; CCF, Cerro Calera Formation; HF, Horqueta Formation; LPeF, Los Pelambres Formation; LPrF, Lo Prado Formation; LVF, Lo Valdés Formation; RCF, Río Colina Formation; VNF, Veta Negra Formation.

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_2O_3 and low MgO high-K to shoshonitic porphyric basaltic andesites and 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.* 2005*a*).

The Las Chilcas Formation (Thomas 1958) consists of a 3500-m-thick succession of mostly coarse alluvial and volcaniclastic 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,

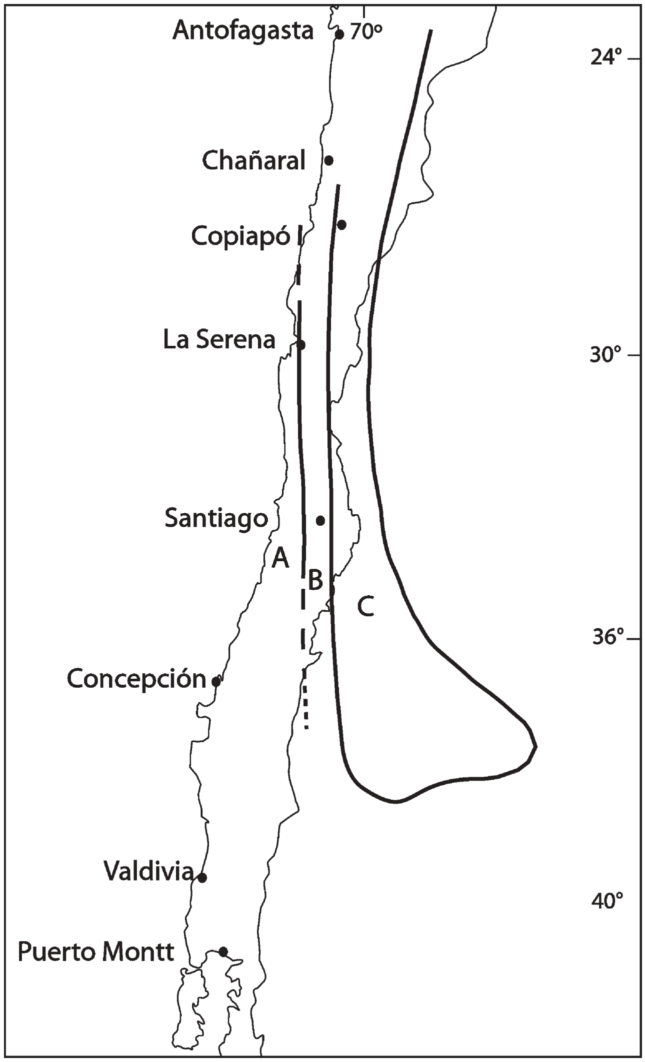


Fig. 3.29. Late Jurassic–Early Cretaceous (second substage of the first Andean stage) palaeogeographic distribution for the forearc and backarc basins separated from each other by the Pelambres arc in central Chile. The slight NNW orientation of the palaeogeographic elements is possibly controlled by pre-existent weakness zones. A, Lo Prado Forearc Basin; B, Lo Prado–Pelambres Volcanic Arc; C, Mendoza–Neuquén Basin. Because of the oblique orientation of the described palaeogeographic units, the Lo Prado Forearc Basin is exposed to the north up to the latitude of La Serena and the Lo Prado–Pelambres Volcanic Arc is located along the coast north of this latitude.

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 volcaniclastic Pucalume Formation (see Fig. 3.25), can be extended for some distance southward (Rivano & Sepúlveda 1991). South of *c*. 32ºS 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 international boundary and extending eastward into western Argentina. Arc volcanism in the coastal region is represented by volcanic intercalations in the Late Jurassic–Early Cretaceous marine and continental forearc basin deposits of the Lo Prado (Thomas 1958) and La Lajuela (Vegara 1969; Bravo 2001) formations. The eastern belt represents volcanic intercalations in the westernmost marine deposits of the Mendoza–Neuquén Backarc Basin, and forms the >2000-m-thick Los Pelambres Formation in the High Andes of Illapel (31º30′S) (Rivano & Sepúlveda 1991) (Fig. 3.28), the partially equivalent Juncal Formation on the eastern flank of the Principal Cordillera in western Argentina (32–33ºS) (Ramos *et al.* 1990; Aguirre-Urreta & Lo Forte 1996; Cristallini & Ramos 1996) and volcanic intercalations in the Lo Valdés (González 1963; Biro-Bagoczky 1964) and Baños del Flaco formations (Charrier 1981*b*) in Chile, between 33º30′S and 34º15′S. In these formations the volcanic intercalations rapidly disappear eastward indicating that the source of the lavas and coarse volcaniclastic sediments was located to the west.

The volcanic domain from which these deposits originated must have been of considerable width, extending westward over the present-day Central Depression as far as the eastern flank of the Coastal Cordillera, as indicated by the location of the western belt and that of the Cretaceous intrusives (Gana & Tosdal 1996; Gana *et al.* 1996). The Los Pelambres Formation consists of andesitic lavas and subordinate tuffs, with coarse- and fine-grained sedimentary intercalations, and marine limestones. The Early Cretaceous age of this formation is deduced on the basis of its stratigraphic conformable position below the Pucalume Formation. According to Rivano & Sepúlveda (1991), this formation represents the easternmost deposits of a volcanic zone that grade eastward into a marine sedimentary succession devoid of volcanic intercalations (Fig. 3.28). The lavas in the Lo Valdés Formation form a thin intercalation in its lower portion (Biro-Bagoczky 1964), and volcanic intercalations in the Baños del Flaco Formation form a 440-m-thick succession of volcanic breccias and silicic lavas between marine fossiliferous sediments of Tithonian–Neocomian age (Charrier 1981*b*).

Plutonic activity. The plutonic activity at this time in the northern part of the region is represented by the Illapel Superunit, which is exposed across an area of 3000 km^2. It is made up of two main units: the Chalinga unit is mainly tonalitic and quartz-dioritic and contains within its central part an elongated pluton of leucodiorites and trondhjemites known as the Limahuida unit (Parada *et al.* 1988). K–Ar ages from the Chalinga unit range from 134 to 86 Ma. The rather low Sr initial ratios from the Early Cretaceous Illapel granitoids indicate direct derivation of the majority of the parent magmas from the upper mantle, with virtually no continental crust involvement (Parada *et al.* 1988).

At the latitude of Santiago, the Early Cretaceous granitoids in the Coastal Batholith that intrude Palaeozoic and Jurassic rocks are exposed on the eastern flank of the Coastal Cordillera (Gana & Tosdal 1996). These plutonic bodies are exposed further south along the eastern border of the Coastal Cordillera, at least as far as 35º30′S (the latitude of Talca), where a monzogranite yielded a K–Ar age on biotite of 100 ± 3 Ma (Bravo 2001). South of 38ºS, these plutons are exposed in the Principal Cordillera.

Backarc. The backarc basin deposits in the La Serena region correspond to the Río Tascadero Formation, which grades westwards and upwards into the Pucalume Formation as indicated in Figures 3.23 and 3.25. Further south, the backarc deposits are still represented by the Río Tascadero Formation, but here they grade westwards into the mainly volcanic Los Pelambres Formation (Rivano & Sepúlveda 1991). The latter formation represents the easternmost reaches of the volcanic arc which was located further west in what is now the Central Depression and the Coastal Cordillera (Fig. 3.28). No backarc deposits of the second substage are exposed in the Chilean cordillera south of 36ºS.

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), Lo Valdés (González 1963; Hallam *et al.* 1986) and Baños del Flaco (Klohn 1960; González & Vergara 1962; Covacevich *et al.* 1976; Charrier 1981*b*; 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 Colimapu 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 Colimapu Formation (Klohn 1960; González & Vergara 1962; González 1963; Charrier 1981*b*), 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 underneath 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 Chilean Andes and which first developed during this stage. This 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 (20ºS) and south of La Serena (Los Vilos and Los Molles at 32ºS). 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 (29º30′ to 30º30′S) (Emparan & Pineda 2000, 2005) and between Los Vilos and Los Molles (32ºS) (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 1987*a*, *b*; Naranjo *et al.* 1984; Scheuber 1987, 1994; Scheuber & Andriessen 1990; Brown *et al.* 1993; Scheuber *et al.* 1994; Arévalo 1995, 2005*a*; 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 1987*b*; 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 1987*a*; Naranjo 1987; Armijo &

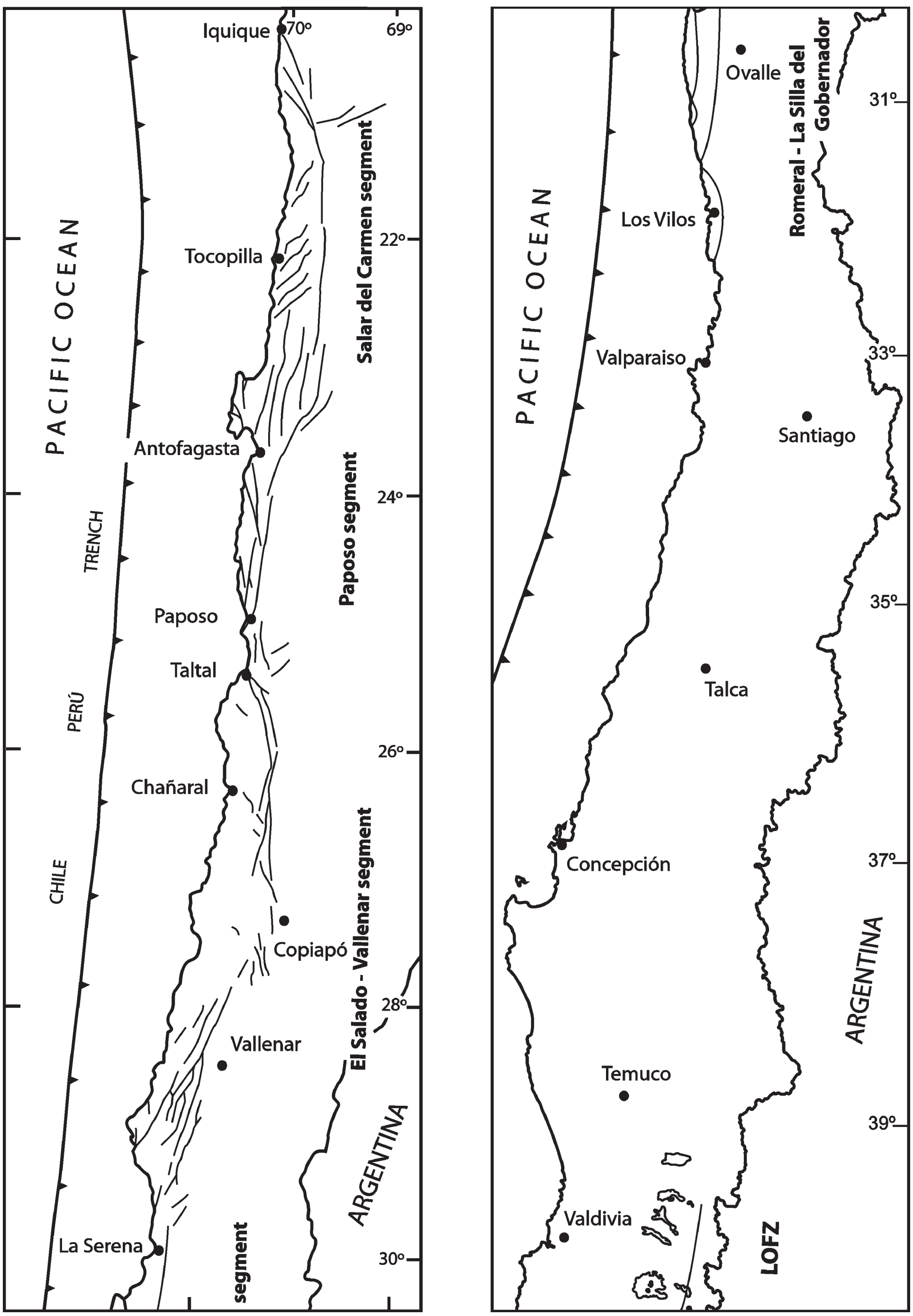


Fig. 3.30. Traces of the different segments of the Atacama Fault Zone, possibly reaching as far south as 31º30′S, from north to south: Salar del Carmen, Paposo, El Salado–Vallenar and Romeral–La Silla del Gobernador segments. Based on Thiele & Hervé (1984), García (1991), Brown *et al.* (1993), Scheuber & González (1999), Emparan & Pineda (1999, 2005), Arancibia (2004) and others (see text).

Thiele 1990; Marinovic *et al*. 1995; González & Carrizo 2000, 2003).

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 25ºS, 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; $^{40}Ar/^{39}Ar$ 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 27ºS and 36º30′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.* 1973*b*). In particular, Rivano & Mpodozis (1976) and Mpodozis & Cornejo (1988) deduced for the Ovalle region (30º30′S) the

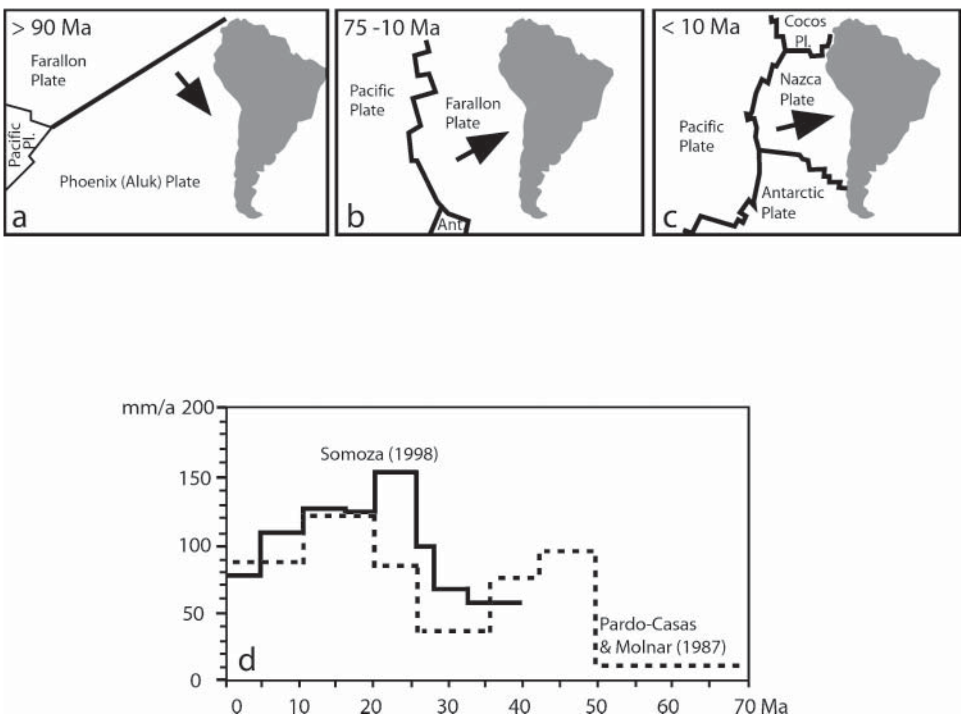


Fig. 3.31. Plate geometries for (**a**) Early Late Cretaceous; (**b**) Late Cretaceous to late Miocene, and (**c**) Late Miocene to Present, taken from Scheuber *et al.* (1994) and Reutter (2001). Different obliquity between (a) and (b) explains different movements along the Atacama Fault Zone in Jurassic 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).

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 Kimmeridgian deposits correspond to an extensional rather than to a compressional event, and that the stratigraphy in the Ovalle region reflects a palaeogeography 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*a*), 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 (32ºS) (Arancibia 2004). These faults include cataclastic as well as mylonitic zones. Radioisotopic determination on the fault rock of the Romeral Fault Zone (29º30′ to 30º30′S) yielded several ages (K–Ar on minerals and whole rock and one $^{40}Ar/^{39}Ar$ 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, ductile shear zone interpreted as an inverted normal fault with a slight dextral component originally associated with the extension of the backarc basin (Arancibia 2004). Geochronological studies performed by the same author on this structure indicate a maximum $^{40}Ar/^{39}Ar$ age of 109 ± 11 Ma and a minimum age of 97.8 ± 1.5 Ma for the reverse ductile 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 (30–31ºS) (Emparan & Pineda 2005) and the Quebrada Teniente Fault (31–31º30′S) (Irwin *et al.* 1988; García 1991) and the ductile El Cajón Fault at 33ºS (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 cyles separated from each other by an episode of abundant continental sedimentation and in some regions intense volcanic activity.

South of 30ºS, 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, 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 palaeogeography (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 1991*b*).

As already indicated, statigraphic, 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; Äberg *et al.* 1984; Mpodozis & Ramos 1989). Based on palaeogeographic 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 coincided 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 Äberg *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 plutonic 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 Damas Extensional Basin (rift phase) corresponds to one or more pull-apart basins.

In the Arica region, the western position of the Atajaña (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 Subhercynian or Peruvian; see Charrier & Vicente 1972; Aubouin *et al.* 1973*b*; 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 palaeogeographic 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–south-ward, 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 palaeogeographic 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 same age that hosted the Hornitos, Viñita and Salamanca 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 volcaniclastic successions were deposited. The end

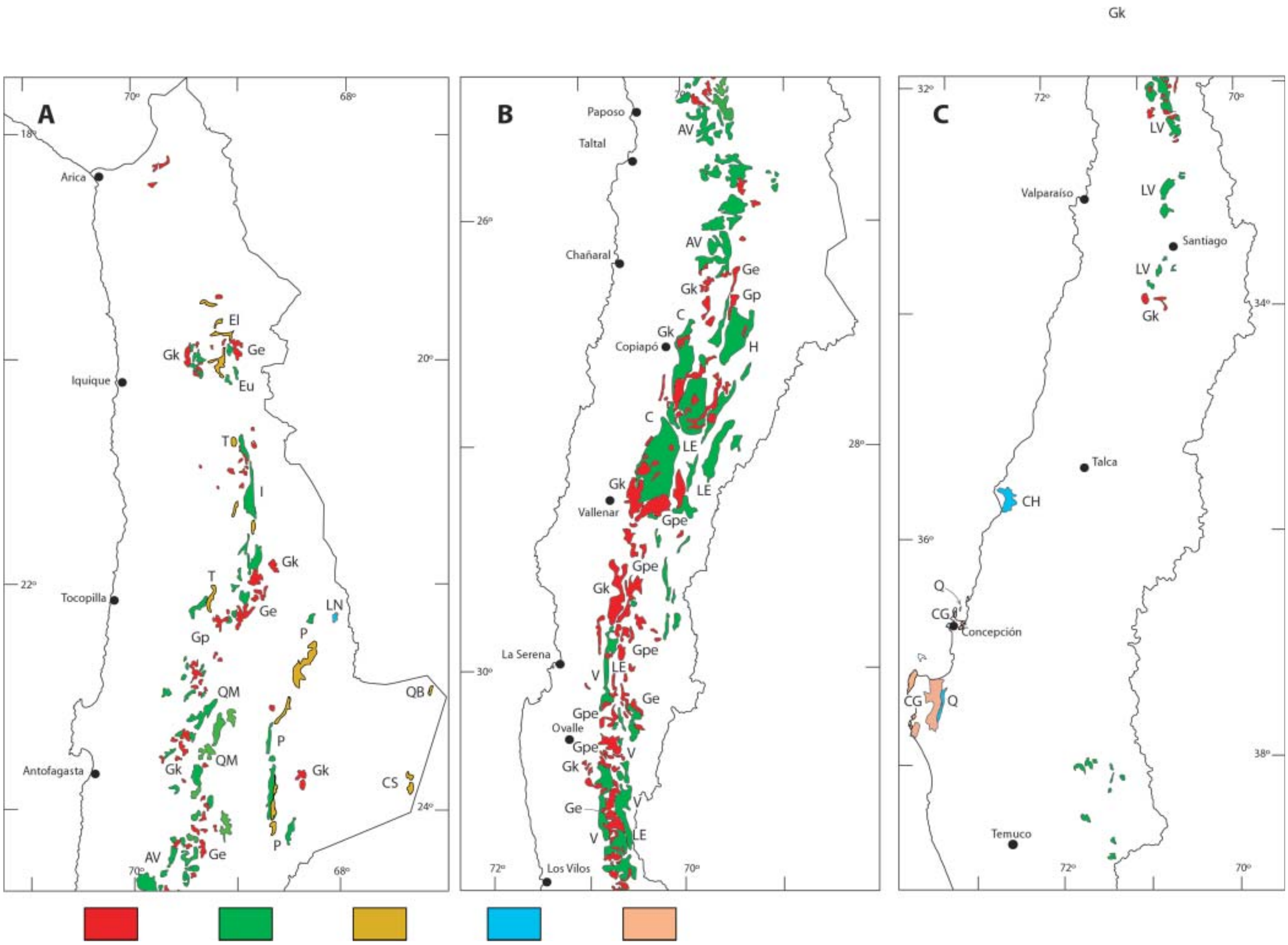


Fig. 3.32. Distribution of units developed during the second stage of Andean evolution between 18ºS and 39ºS. Location of intrusives shifted to the east of the ones formed during the first stage (compare with Fig. 3.17). Late Cretaceous granitoids are generally located west of the early Cenozoic granitoids. Colour scheme (from left to right): Red, Granitoid rocks; Green, volcanic rocks; orange-brown, clastic continental deposits; blue, marine deposits; orange-pink, continental/marine deposits. Abbreviations for formation and group names: AV, Augusta Victoria; C, Cerrillos; CF, Chanco; CG, Concepción; CH, Chanco; CS, Chojfias and Siglia; El, Lower Empexa /Tambillos; Eu, Upper Empexa; H, Hornitos; I, Icanche; LE, Los Elquinos; LN, Lomas Negras; LV, Lo Valle; Q, Quiriquina; QB, Quebrada Blanca de Poquis; QM, Quebrada Mala; P, Purilactis and Tonel; T, Tolar; V, Viñita. Abbreviations for age of granitoids: Ge, Eocene; Gk, Late Cretaceous; Gp, Palaeocene; Gpe, Palaeocene–Eocene.

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 a second period during which extension was reactivated in 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 (18ºS to 26ºS)

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 sedimentary accumulations. Late Cretaceous, mainly volcanic deposits correspond to the following formations, from north to south: Panjuacha (Harambour 1990), Cerro Empexa (Galli 1968; Galli & Dingman 1962), Quebrada Mala (Montaño 1976; Muñoz 1989; Marinovic & García 1999) and Augusta Victoria (García 1967). In addition, the volcanic Lomas Negras Formation, with marine intercalations containing Miliolid foraminifera of undetermined age (Marinovic & Lahsen 1984) and dated at 66 Ma (Hammerschmidt *et al.* 1992), represents the Late Cretaceous arc volcanism in this region. Other volcanic 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 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 Permian to 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.* 20ºS, 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 Altiplano (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 overlying strongly deformed Early Cretaceous and older formations, and uncomformably 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 Argomedo–West Fissure (Falla Oeste) Fault developed along the Domeyko Range and its northern prolongation, the Sierra del Medio. The Calama Formation deposits

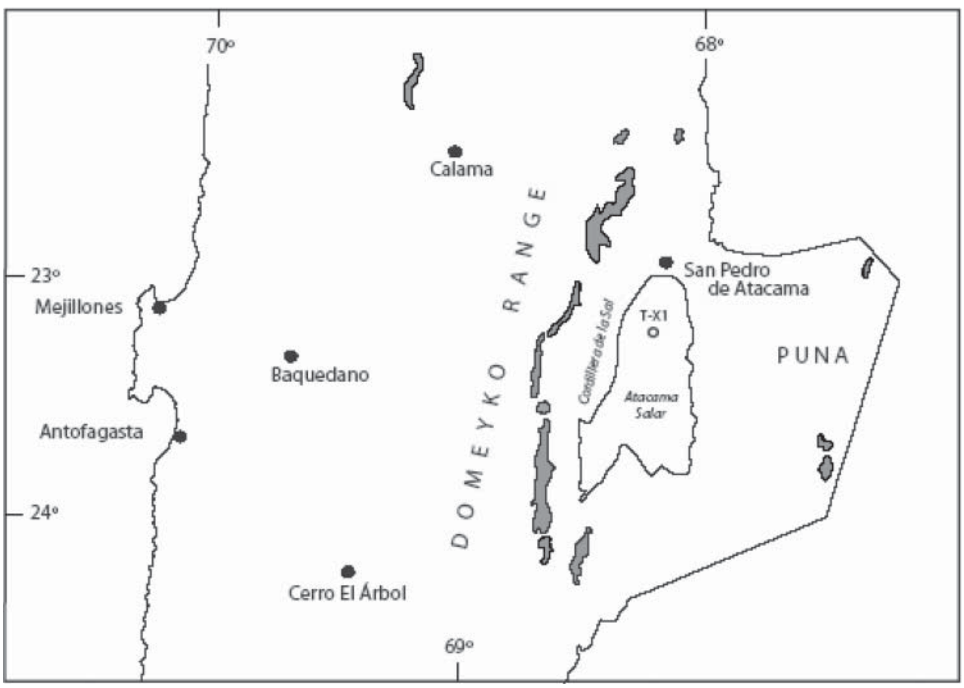


Fig. 3.33. Distribution of the foreland deposits of the second stage of Andean evolution in the Antofagasta–Salar de Atacama region with location of the Toconao-X1 hydrocarbon exploration well. Outcrops NW of Calama correspond to the Tolar Formation, an age equivalent of the deposits east of the Domeyko 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 age (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 early-middle Eocene. This age coincides with the age of the Icanche 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 volcaniclastic deposits, and subvolcanic intrusive bodies have gabbroic to dacitic compositions. Still recognizable volcanic structures, such as the nested Cachinal calderas of Palaeocene age with large areas affected by hydrothermal alteration, are emplaced in rocks of the Augusta Victoria Formation. The Augusta Victoria Formation unconformably overlies the Late Jurassic–Early Cretaceous arc deposits of the Aeropuerto Formation and the marine backarc deposits of the Santa Ana Formation (Naranjo & Puig 1984; Marinovic *et al.* 1995). It unconformably underlies the Cenozoic Pampa de Mulas Formation (Chong 1973; Marinovic *et al.* 1995) and the Atacama Gravels (Naranjo & Puig 1984). Its age is well constrained by several radioisotopic age determinations (K–Ar) that yielded Late Cretaceous to Eocene ages (between 66.6 ± 2.2 Ma and 41.2 ± 2.2 Ma) (Naranjo & Puig 1984; Marinovic *et al.* 1995); an older $^{87}Ar/^{86}Ar$ age of 71.7 ± 2.9 Ma has been obtained by Hermann & Zeil (1989). The geochemical composition of these rocks indicates a K-rich calcalkaline character (Naranjo & Puig 1984; Marinovic *et al.* 1995).

The wide outcrop of the arc deposits of the two substages (late Early to Late Cretaceous and early Palaeogene), such as across the region between 24°S and 25°S, from the Central Depression to the western Altiplano (Marinovic *et al.* 1995), indicates that the width of the volcanic arc was probably over 100 km (Scheuber *et al.* 1994).

Finally, with regard to plutonic arc activity, Late Cretaceous to Eocene intrusions are found in the Central Depression, along the Precordillera and western Altiplano, and the western flanks of the Sierra de Moreno and the Domeyko Range, although they form considerably smaller outcrops than the older Mesozoic plutons of the previous stage (see Marinovic *et al.* 1995; García *et al.* 2004). These younger plutons are emplaced in older Mesozoic units and into the Augusta Victoria Formation itself (Fig. 3.32). Geochemically they are similar to the Augusta Victoria Formation, being alkali-rich subalkaline in composition (Marinovic *et al.* 1995).

Foreland evolution. Terrigenous deposits of Late Cretaceous to Early Cenozoic age that can be correlated with the Tolar and Tambillo formations are exposed on the eastern flank of the Domeyko Range and have been recognized in drill cores from the Toconao-X1 hydrocarbon exploration well (Fig. 3.33). The surface deposits correspond to the Purilactis Formation (Brüggen 1934, 1942), or the Tonel and Purilactis Formations of Dingman (1963) and to further subdivisions and modifications of this succession by Hartley *et al*. (1992), Flint *et al.* (1993), Charrier & Reutter (1994), Arriagada (1999) and Mpodozis *et al*. (1999, 2005). The Tonel Formation unconformably overlies the Late Carboniferous–Early Permian Agua Dulce Formation, and forms a 1300–1400-m-thick succession of basal limestones, fine-grained sandstones, and mudstones with frequent evaporitic intercalations. There is a transitional contact with the overlying Purilactis Formation, which consists of several essentially conglomeratic and coarse sandstone members of different colour, grain size and outcrop extent exposed along the eastern border of the Domeyko Range, with intercalations of basaltic lavas and tuffs. Radioisotopic age determinations yielded Late Cretaceous to middle Eocene ages between 68 Ma to 39.9 Ma (Hartley *et al*. 1992; Flint *et al*. 1993; Charrier & Reutter 1994; Arriagada 1999; Mpodozis *et al*. 1999, 2005). This unit corresponds to syntectonic alluvial and coarse fluvial deposits associated with the eastward thrusting of the Domeyko Range (Muñoz *et al*. 2002). Analysis of seismic reflection data and drill-cores from the Toconao-X1 well allowed recognition of a stratigraphic succession in this basin fill comparable to the one described for the eastern flank of the Domeyko Range (Muñoz *et al*. 2002). The lower basin fill deposits contain thin limestone beds similar to the basal limestones of the Tonel Formation, with a poor microfaunal planktonic association of probable Senonian age (Muñoz *et al*. 2002). These marine levels are correlated with the Late Cretaceous, foraminifera-bearing Lomas Negras Formation (Marinovic & Lahsen 1984), and with calcareous outcrops of Late Cretaceous age exposed further east in the Puna (Altiplano) next to the Chile–Argentina border (Ramírez & 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 (Salfity *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 the basin, whereas Charrier & Reutter (1994), Muñoz *et al.* (1997, 2000, 2002), Arriagada (1999) and Mpodozis *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; Salfity & Marquillas 1994). Normal faults, originally formed during the extensional stage of the Salar de Atacama Basin, were inverted 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 (Cordón 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 Perú and Bolivia in Cenomanian–Santonian times and a southward connection with the opening Atlantic Ocean in Campanian–Maastrictian times (Marquillas & Salfity 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 ingression 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 (Salfity *et al.* 1985). These deposits have been correlated with parts of the Purilactis Formation (Ramírez & Gardeweg 1982; Donato & Vergani 1987) (Fig. 3.33).

Southern north Chile: region between Chañaral and La Serena (26ºS to 30ºS)

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 (26–26º30′S), the 3500-m-thick volcanic and volcaniclastic Llanta Formation represents the Late Cretaceous deposits of this stage. This unit is exposed east of the Sierra Castillo Fault and is unconformably overlain by the Palaeocene volcanic products of the Salvador Caldera and the Cerro Valiente Beds (Tomlinson *et al.* 1999; Cornejo *et al.* 1997, 1998) (Fig. 3.34). Age determinations with the K–Ar, ^{40}Ar/^{39}Ar and U–Pb methods performed in the Llanta Formation yielded several ages within the Campanian–Maastrichtian (80.2 $\pm$ 0.3 Ma with U–Pb and 65.7 $\pm$ 0.9 Ma with ^{40}Ar/^{39}Ar) (Tomlinson *et al.* 1999; Cornejo *et al.* 2003). Further south between the latitudes of Chañaral and Copiapó (26º and 27º30′S), the Llanta Formation is unconformably overlain by the Venado Formation (Sepúlveda & Naranjo 1982; Iriarte *et al.* 1996) of Palaeocene age. The latter formation consists of a 1500-m-thick succession of trachyandesitic and dacitic lavas, andesitic volcanic breccias, a 10-m-thick ignimbritic intercalation, and sandstones and sedimentary breccias. Other Early Palaeogene deposits in this region are the Quebrada Vásquez Beds, and the Cerro Vicuña, La Banderita and San Pedro de Cachiyuyo calderas. Frequent intrusive bodies of the same age are emplaced in the above mentioned deposits and older units (Iriarte *et al.* 1996; Tomlinson *et al.* 1999; Cornejo *et al.* 2003).

Recent mapping in the Copiapó region has resulted in the recognition of (1) an unconformity separating late Early to early Late Cretaceous from Late Cretaceous deposits, and (2) several new formations and informal stratigraphic units that represent the different ages and facies of the Late Cretaceous and Early Cenozoic deposits, as well as several plutons associated with the volcanic activity (Arévalo 2005*a*, *b*; Iriarte *et al.* 1996, 1999). The accumulation space of these deposits, and the emplacement of volcanic activity and plutons being controlled by systems of faults that were later reactivated, explains the existence of the many different stratified and plutonic units at this time. In this region the sediments of the 4000-m-thick continental Cerrillos Formation (Albian–Santonian) rest unconformably on the Bandurrias (continental) and the Chañarcillo (marine backarc) groups of the preceding stage. This succession consists of a lower part with 800 m of sandstones followed upwards by breccias and conglomerates with basaltic andesite intercalations, which were deposited in a strongly subsiding extensional basin. The Cerrillos Formation is unconformably overlain by the Hornitos Formation of Campanian–Maastrichtian age, which was deposited during a later extensional episode controlled by faults that cut into the Cerrillos Formation (Hornitos Extensional Basin) (Arévalo 1995, 2005*a*) (Figs 3.32 & 3.34). The 2200-m-thick continental deposits of the Hornitos Formation consist of breccias and conglomerates with coarse sandstone matrix and lenticular intercalations of sandstone, calcareous mudstone and limestone. Minor ignimbritic intercalations are also present. At its top, a continuous member of basaltic and trachybasaltic lavas is developed. The Hornitos Formation can be correlated with the Llanta Formation further north and is unconformably overlain by the Venado Formation (Sepúlveda & Naranjo 1982) of Early Palaeocene age (Arévalo 2005*a*).

Immediately to the east and SE, Iriarte *et al.* (1996, 1999) and Arévalo (2005*a*, *b*) differentiated several thick stratified volcaniclastic deposits and plutonic units of Late Cretaceous to middle Eocene age, formed during Hornitos extension and after contractional deformation of the Hornitos deposits. Stratified units equivalent to the Hornitos Formation (i.e. Late Cretaceous in age) include La Puerta Beds, Quebrada La Higuera Beds, Quebrada Seca Formation, Sierra Los Leones Beds, Sierra La Dichosa Lavas and the Cerro Los Carneros Beds. In this region, some of the stratified units and volcanic structures formed in Early Palaeozoic times after deformation of the Hornitos Formation, include the Venado Formation, Cerro Valiente Beds, Quebrada del Medio Beds, Quebrada Romero Beds, Sierra La Peineta Ignimbrites and Pintadas–Esquivel Beds, and calderas such as the Cerro Blanco Caldera and the Carrizalillo Megacaldera (Rivera & Mpodozis 1994), as well as associated nested calderas (Lomas Bayas, El Durazno, Agua Nueva and Bellavista). Some of these units have been included in Figure 3.34.

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 26ºS. In the Copiapó region (27–27º30′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 Vaca (Arévalo 1995, 2005*a*; Iriarte *et al.* 1996).

In the northern part of the region, between Vallenar and La Serena (29–30ºS), the late Early to early 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

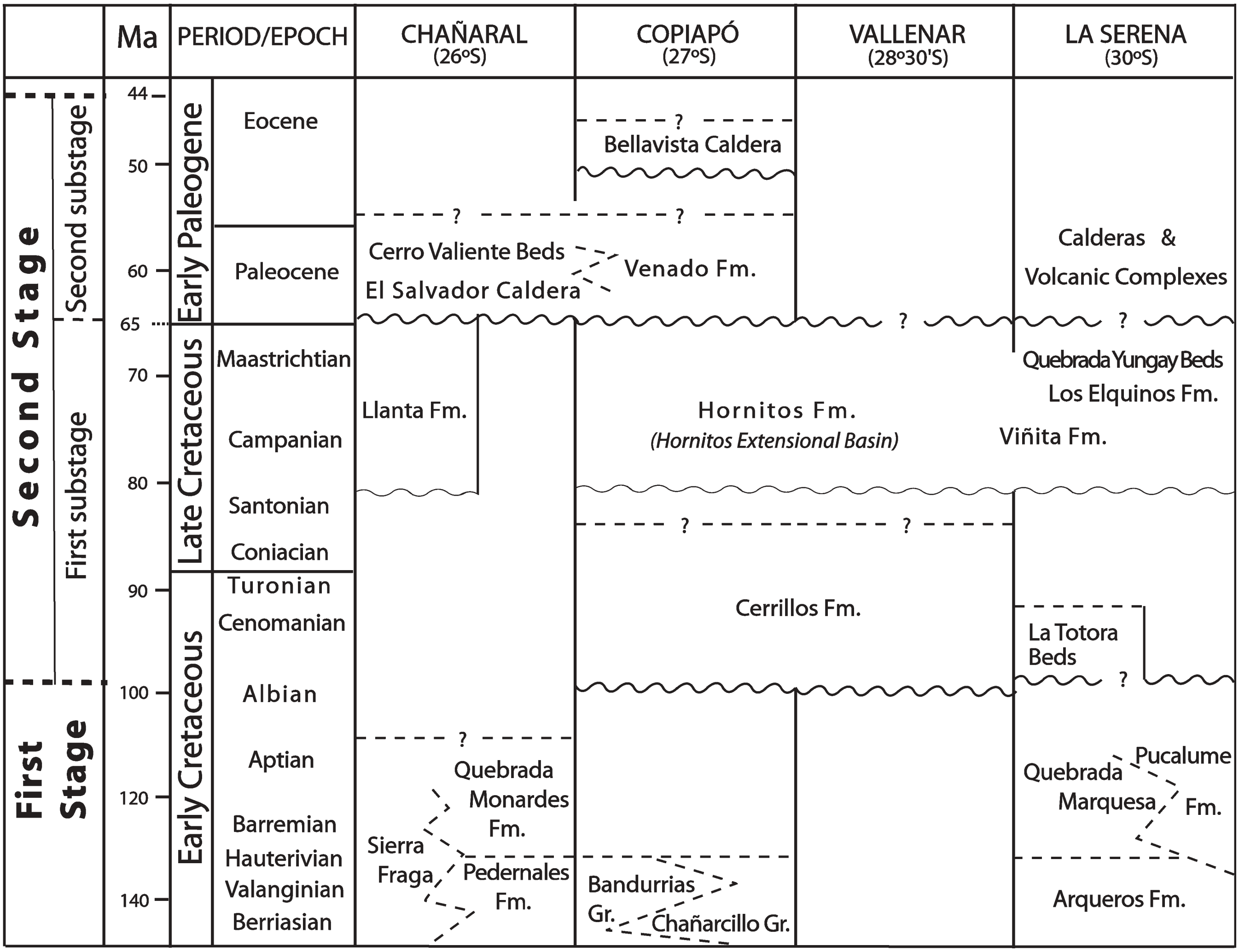


Fig. 3.34. Stratigraphic succession for the first (upper part) and second stages of Andean evolution, from Early Cretaceous to Early Palaeogene in the Chañaral, Copiapó, Vallenar and La Serena regions (see Fig. 3.21), based on Nasi *et al.* (1990), Arévalo (1995, 2005*a*, *b*), Cornejo *et al.* (1998, 2003), Emparan & Pineda (1999, 2005), Iriarte *et al.* (1996, 1999), Tomlinson *et al.* (1999), Pineda & Emparan (2006). The unconformity of late Early Cretaceous age at the base of the second stage of Andean evolution corresponds to the so-called Subhercynian tectonic phase; the unconformity of middle Eocene age separating the second from the third stage corresponds to the main Incaic tectonic phase (not represented). A generalized deformation occurred during this stage next to the Cretaceous–Tertiary boundary (Incaic phase I or 'K-T' tectonic event) that separates two substages in the second stage of Andean evolution.

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, 2005*a*; 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 (30ºS to 39ºS)

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 30ºS, in the Illapel region, between 31ºS and 32ºS, 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 volcaniclastic 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 then active from Iquique southwards.

Finally, plutonic activity in this northern part of central Chile (30–33ºS) 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 33ºS and 39ºS, 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 are 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 fossiliferous marine deposits related to the eustatic high stand developed at this time, and are exposed in the following localities from north

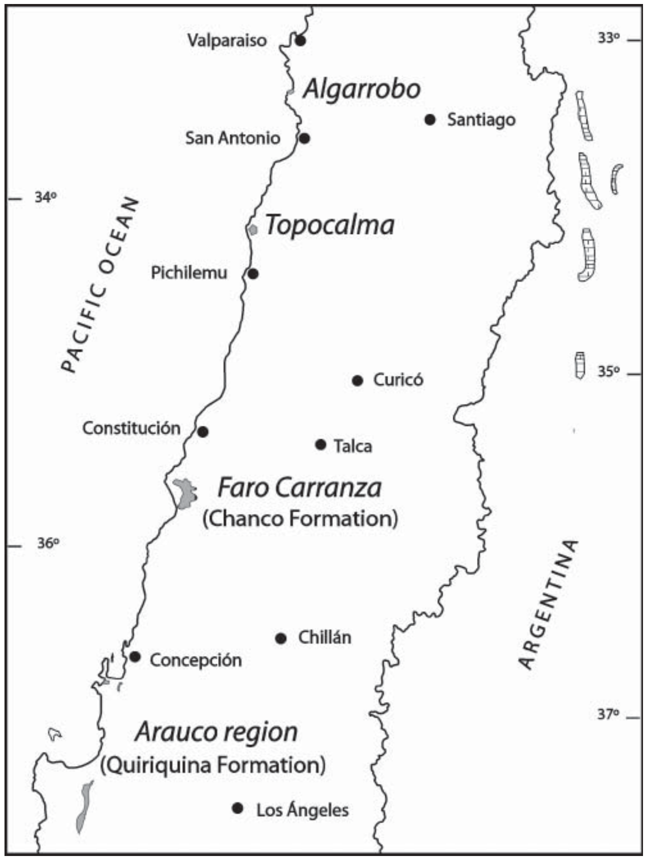


Fig. 3.35. Distribution of outcrops of Late Cretaceous to Palaeocene deposits related to the eustatic highstand developed in Chile and Argentina. Outcrops in Chile are found at Algarrobo, Topocalma, Faro Carranza (Chanco Formation) and in the Arauco region (Quiriquina Formation) (based on Sernageomin 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).

to south: Algarrobo (Levi & Aguirre 1960), Topocalma (Tavera 1979*b*; Charrier 1973*b*), Faro Carranza, south of Constitución (Chanco Formation; Cecioni 1983), and in the Arauco region, at the latitude of Concepción (*c.* 37ºS) (Quiriquina Formation; Steinmann *et al.* 1895; Muñoz Cristi 1946, 1956; Stinnesbeck 1986) (Fig. 3.35). The deposits in Algarrobo, Topocalma and Faro Carranza record marine platformal (littoral and transitional) environments and comprise sandstones with conglomeratic intercalations, siltstones and limestones (Levi & Aguirre 1960; Charrier 1973b; Cecioni 1983). In Las Tablas Bay (Arauco region) the Quiriquina Formation consists of a *c.* 85-m-thick Late Cretaceous to earliest Cenozoic succession of conglomerates at the base, sandstones, coquinas, and calcareous sandstones with concretions (Stinnesbeck 1986). The Cretaceous portion of this formation contains abundant invertebrate and vertebrate faunal remains, notably ammonites and saurians, whereas the Cenozoic part is poor in fossil remains (Steinmann *et al.* 1895; Stinnesbeck 1986). It overlies the Late Palaeozoic metamorphic complex and is unconformably overlain by the Late Palaeocene? to Eocene Concepción Group.

In the coastal region at the latitude of Concepción, Late Palaeocene?–Eocene deposits crop out as the Concepción Group, which unconformably overlies Quiriquina Formation deposits. This sedimentary group comprises repeated alternations of continental and marine deposits accumulated in extensional basins formed along the coast in Late Palaeocene?

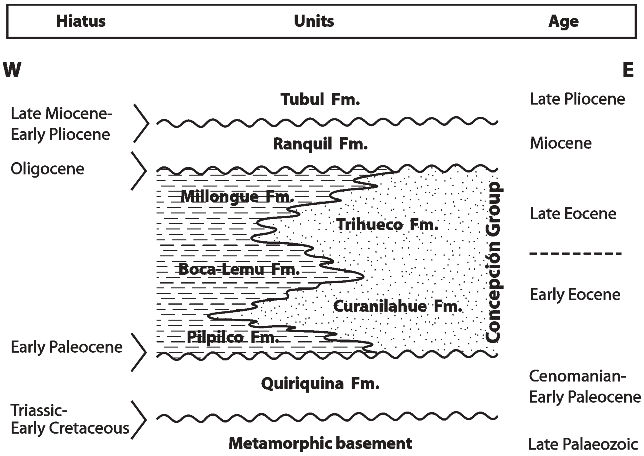


Fig. 3.36. Late Mesozoic and Cenozoic sedimentary deposits in the Arauco region on the western side of the Coastal Cordillera. Continental (right-hand side) and marine (left-hand side) deposits build up the Concepción Group. Major hiatuses separating the Cenozoic deposits are indicated.

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 1983*a*, *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 32ºS and 35º15′S are represented by the Lo Valle Formation (Thomas 1958; Godoy 1982; Moscoso *et al.* 1982*b*; 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 33ºS yielded 70.5 ± 2.5 Ma and 64.6 ± 5 Ma (Vergara & Drake 1978; Drake *et al.* 1976), and $^{40}Ar/^{39}Ar$ 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 35ºS, a small outcrop of upward fining and thinning 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 36ºS 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 33ºS and 35ºS 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 33ºS and 39ºS is concentrated along the eastern flank of the Coastal Cordillera where it forms discontinuous outcrops intruding Mesozoic units (Bravo 2001).

Second Andean stage: summary and discussion

The palaeogeographic setting during this stage was characterized by a wide forearc, a magmatic arc/intra-arc (located several tens to about 100 kilometres from the present-day coastline) and a foreland basin. The forearc domain mostly comprised igneous rocks formed in the previous Jurassic and Early Cretaceous magmatic arcs, although, depending on the location, Late Palaeozoic metamorphic rocks and sedimentary and volcanic Triassic rocks were also exposed. All these older rocks were at this time subjected to erosion. In contrast, across the westernmost parts of the forearc domain (coastal region), deposition of alternating marine and continental successions occurred.

Deposition in the arc domain during this stage was continuous in northernmost Chile, though including thick conglomeratic deposits which suggest important tectonic activity at this time. Plutonic activity apparently was continuous throughout the entire stage, although it is possible that the age of this activity is still insufficiently constrained. Further south, in the Chañaral and Copiapó regions, angular unconformities separate late Early and Late Cretaceous successions from each other and the latter from the Early Palaeogene deposits. The unconformity separating the Cerrillos from the Hornitos formations, in the Copiapó region, and the Quebrada La Totora Beds from the Viñita Formation, in the La Serena region, located approximately at the boundary between Early and Late Cretaceous (Fig. 3.34), is probably a local effect resulting from the reactivation of tectonic extension at the beginning of deposition of the Llanta, Hornitos and Viñita formations. This deformation event can tentatively be named the ‘Intersenonian’ phase, following Groeber (1951). The unconformity separating Late Cretaceous from Early Palaeogene deposits is well developed in the arc domain between Antofagasta and Illapel and further south in the forearc domain in the coastal region of Concepción. This regionally extended unconformity separating Cretaceous from Palaeocene rocks indicates existence of a major tectonic compressive event near the Cretaceous–Tertiary boundary that we name, following Cornejo *et al.* (2003), the ‘K-T’ tectonic event or Incaic phase I. After this event, extension resumed, though with less intensity compared with the late Early to Late Cretaceous extension. This Early Palaeogene extension was associated with abundant volcanic and shallow plutonic activity. Finally, a new compressive deformation event occurred during the Eocene epoch, the main Incaic phase or Incaic phase II, that ended the second stage of Andean evolution.

The great thickness of the volcanic and associated coarse sedimentary deposits during late Early to Late Cretaceous 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–Cenomanian times in the Copiapó region by Mpodozis & Allmendinger (1992, 1993). According to these and other authors (Arévalo 1995, 2005*a*; 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 'Puquios Caos'. As a consequence of this extensional event were deposited the Cerrillos Formation (Arévalo 1995, 2005*a*; 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 Äberg *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, Hornitos 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 Valparaiso (33ºS). North of this latitude, the batholith's exposures are located along the High Andes (ESU in Fig. 3.5), whereas south of Valparaiso they are abruptly shifted to the west and located along the Coastal Cordillera (Coastal Batholith south of 33ºS in Fig. 3.5). Based on evidence indicating that the batholith, though not exposed, exists further north of Valparaiso underneath the Choapa Metamorphic Complex, Rebolledo & Charrier (1994) suggested the possibility that the two parts of the batholith (one to the north and the other to the south of Valparaiso) were separated by an event of extensional tectonics orientated slightly oblique (NNW–SSE) to the main batholith trend and probably controlled by pre-existing weakness zones. The major Aptian–Cenomanian 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 apatite fission track ages from the Coastal Cordillera at *c.* 33ºS 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 ± 5.4 Ma and 82.0 ± 5.6 Ma (Parada *et al.* 2005*a*)) 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 (1997*a*), and over a longer time span according to Maksaev & 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 generation of accommodation space in the Atacama 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 resulted in considerable crustal thickening. Exhumation of the uplifted blocks occurred according to Maksaev & 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; Cornejo *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; Maksaev 1978; Cornejo *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; Somoza 1998; see also Reutter 2001, Fig. 5).

The 20-million-year long exhumation period detected in the Domeyko Range by Maksaev & 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 backarc 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

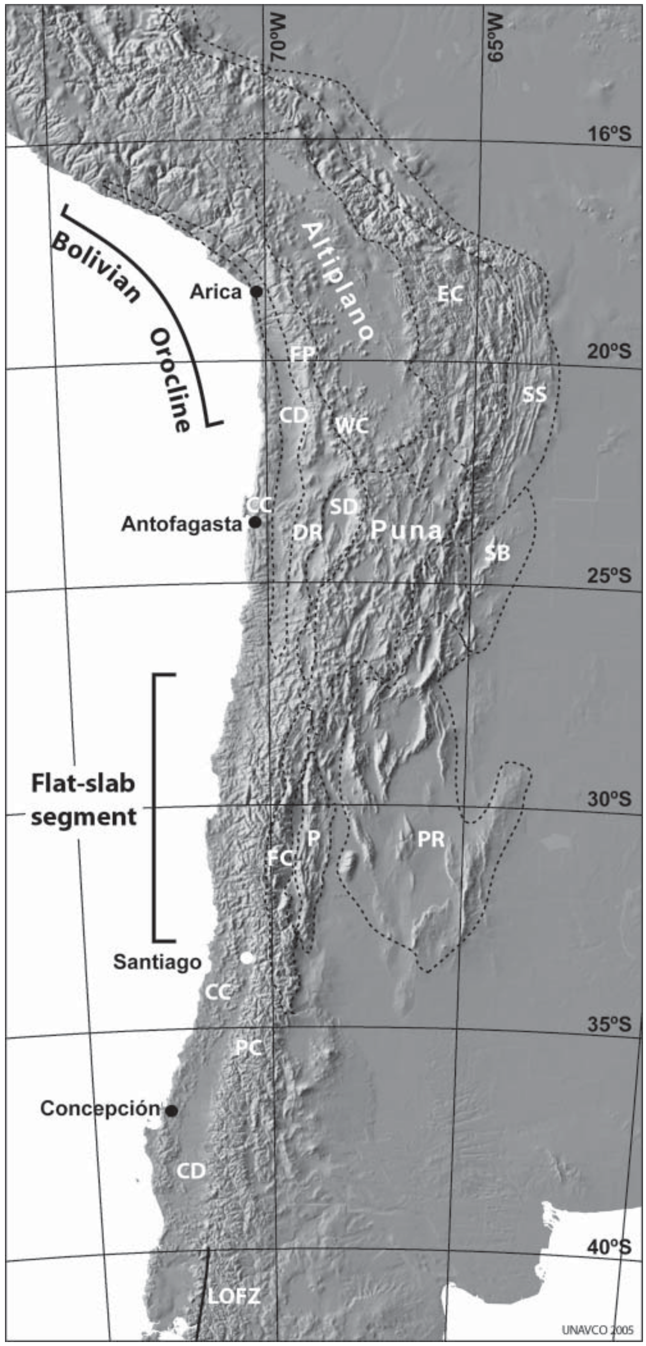


Fig. 3.37. Digital elevation model of the Andes between 16ºS and 40ºS with indication of the main geographical, tectonic and morphostructural features. Compare with Figure 3.2a. Abbreviations: CC, Coastal Cordillera; CD, Central Depression; EC, Eastern Cordillera; FC, Frontal Cordillera; FP, Forearc Precordillera (western flank of the Altiplano) and, further south, Sierra de Moreno; LOFZ, Liquiñe–Ofqui Fault Zone; P, Precordillera in Argentina; PC, Principal Cordillera; PR, Pampean Ranges; SB, Santa Barbara System; SD, Salars Depression; SS, Subandean Sierras; WC, Western Cordillera.

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;

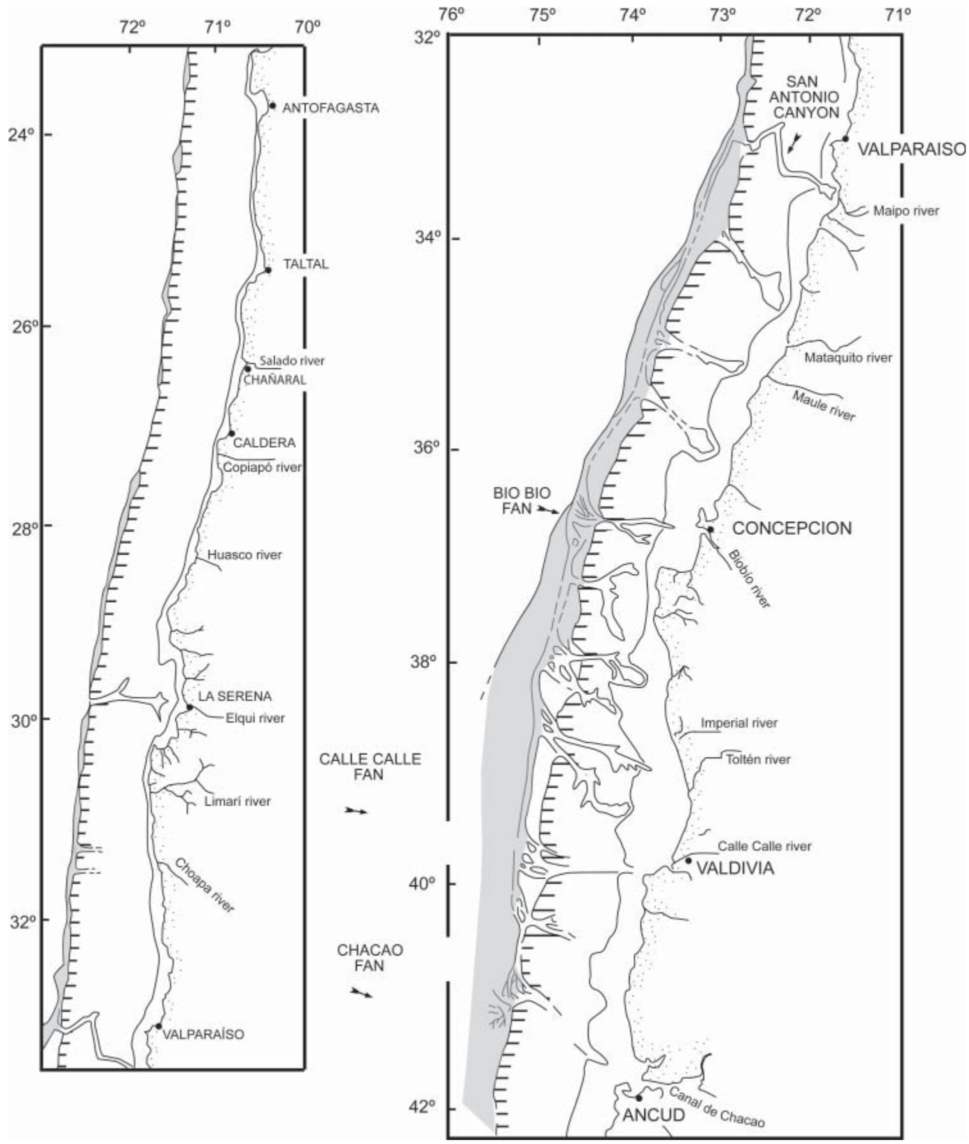


Fig. 3.38. Morphology of the Continental Platform and Chile trench between 24ºS and 42ºS (Antofagasta to Chiloé). Note northward diminution of the trench filling and close association between the main rivers and the development of marine canyons.

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 1987*a*, *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).

Northernmost Chile (region between Arica and Iquique, 18–20ºS) In this region, the morphostructural units are, from west to east: the Coastal Cordillera, the Central Depression, the Forearc Precordillera and the Altiplano. The Precordillera corresponds to the western flank of the Altiplano. The volcanic arc or Western Cordillera is developed to the east of the Precordillera, on the western side of the Altiplano (Figs. 3.1c & 3.37). Apart from the extensional Arica Basin in the Continental Platform (Muñoz & Fuenzalida 1997), Cenozoic deposits are mostly developed in the Central Depression, Forearc Precordillera and Altiplano.

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; Farías 2003; Pinto *et al.* 2004*a*, *b*; Victor *et al.* 2004; Farías *et al.* 2005*a*); 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.* 2005*b*) (Fig. 3.39). The features described for the

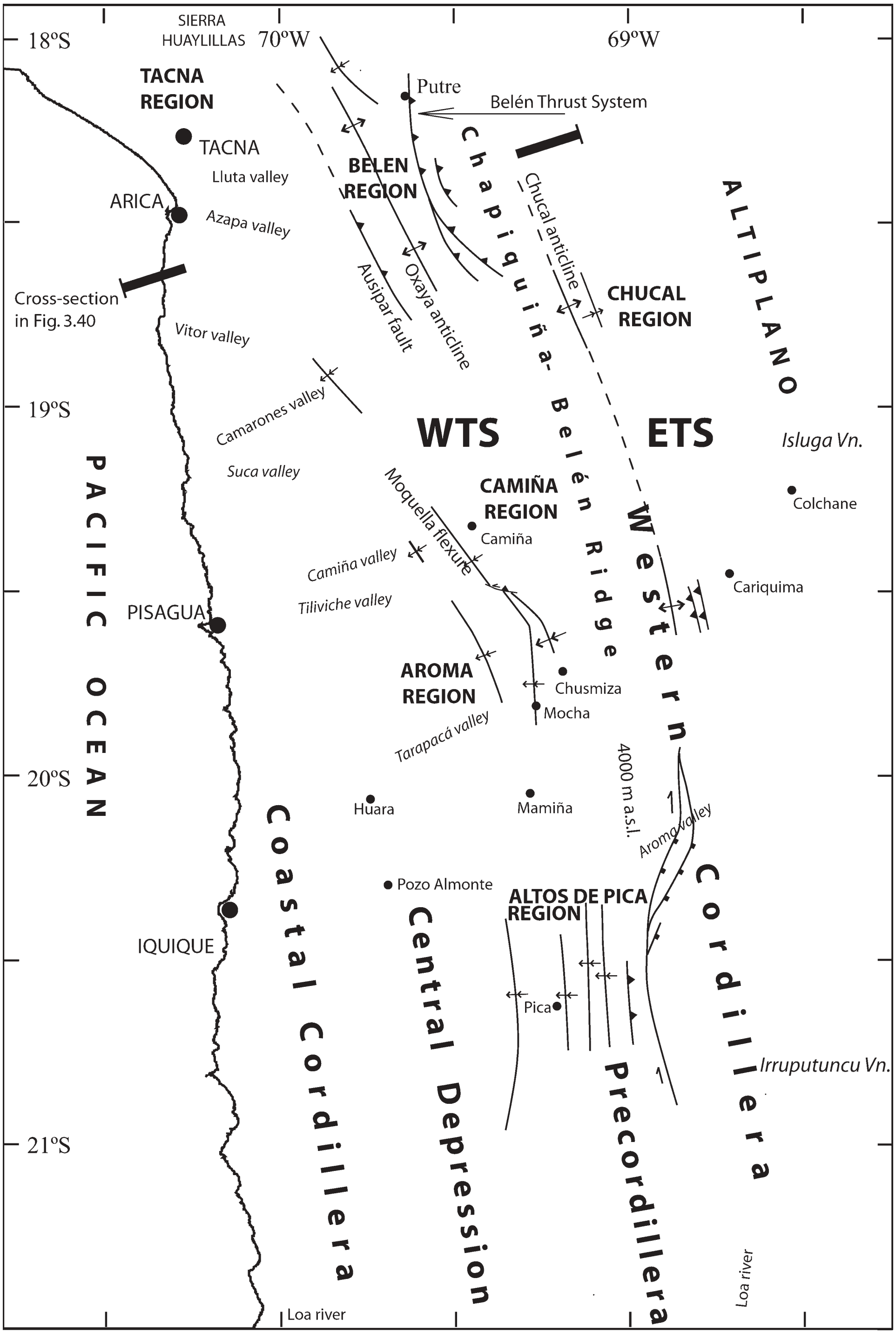


Fig. 3.39. Structural map for the west-vergent thrust system (WTS) and the east-vergent thrust system (ETS) in the Arica–Iquique region, based on García (1996, 2002), Muñoz & Charrier (1996), Pinto (1999), Victor (2000), Victor *et al.* (2004) and Farías *et al.* (2005*a*).

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 (18–19ºS) (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 1996; Parraguez 1997; García 2001, 2002; García *et al.* 1996, 2004; Victor 2000; Farías 2003; Victor *et al.* 2004). Locally they correspond to lacustrine and playa environments. Coarseness of the sedimentary deposits diminishes westward. Sedimentary and intercalated pyroclastic deposits wedge toward the west and their more distal parts onlap over the Mesozoic rocks of the Coastal Cordillera (Salas

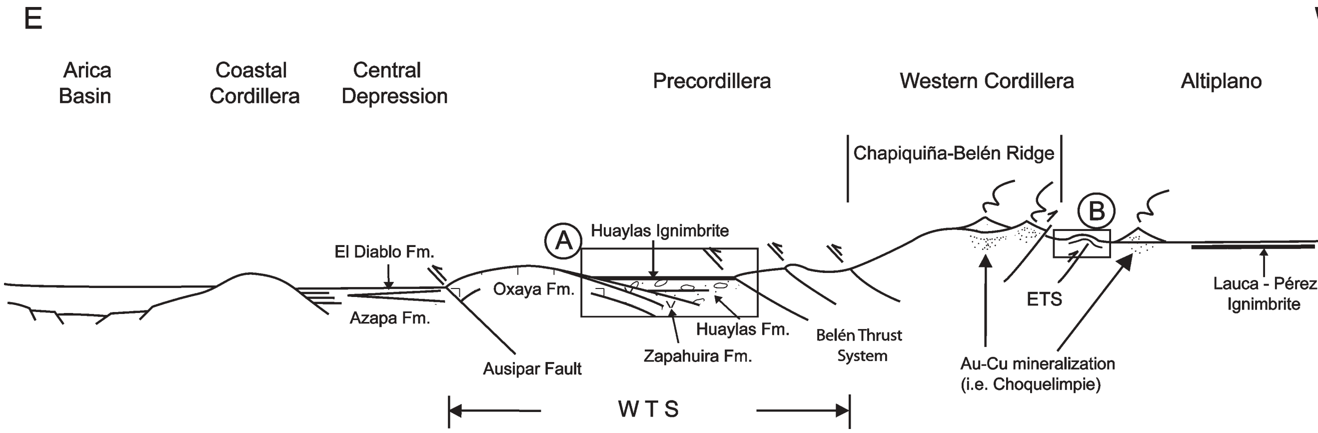


Fig. 3.40. Cross-section in the Arica section, in northernmost Chile at 18º45′S (not to scale); for location see Figure 3.39. The main stratigraphic and mophostructural units, and major tectonic features are indicated: Ausipar Fault and Belén thrust system (west-vergent thrust system; WTS), east-vergent thrust system (ETS) or Chucal thrust system, and the Chapiquiña–Belén Ridge, bounded by the two thrust systems. A, Syntectonic deposits on the back-limb of the Oxaya Anticline and on the front side of the eastern WTS; B, syntectonic deposits and growth unconformities in the ETS.

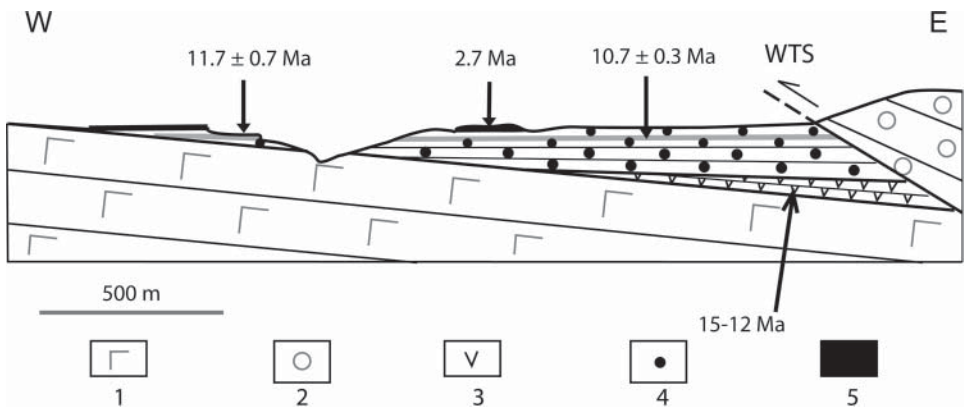


Fig. 3.41. Detail of the stratigraphic relationship between the Oxaya and Zapahuira formations onlapped by the syntectonic Huaylas Formation on the back-limb of the Oxaya Anticline in the eastern Precordillera (A in Fig. 3.40). The Zapahuira Formation yielded K–Ar ages between 15 and 12 Ma and the overlying syntectonic deposits have been dated at 10 and 12 Ma and contain late Middle Miocene mammal remains (see text). The onlapping syntectonic deposits of the Huaylas Formation resulted from the activity along the Belén thrust system, in the eastern west-vergent thrust system (WTS). Based on the angularity between the Huaylas Formation and the underlying units it has been deduced that tilting of the back-limb occurred between *c.* 12 and 11 Ma. Key: 1, Oxaya Formation; 2, Joracane Formation; 3, Zapahuira Formation; 4, Huaylas Formation; 5, Huaylas Ignimbrite. Horizontal and vertical scales are the same.

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 Copaquilla–Tignámar 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.* 2002*c*, 2005*b*) (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

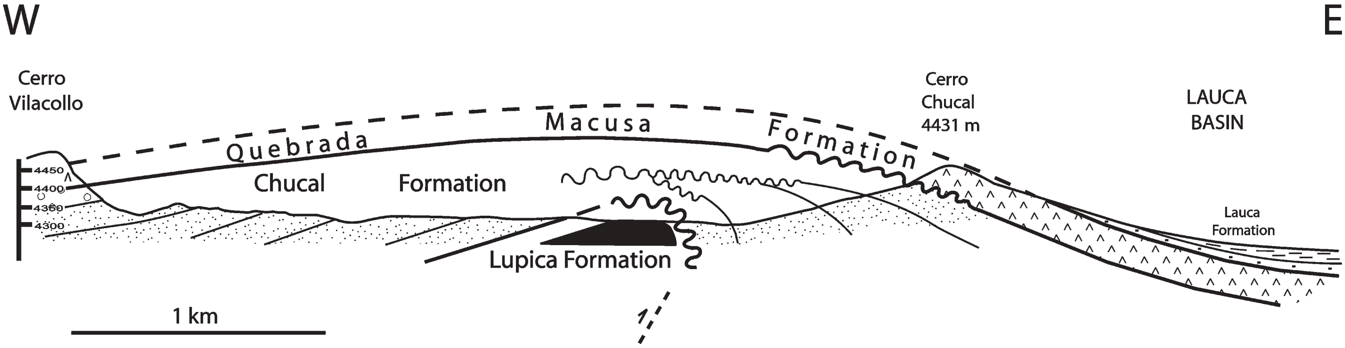


Fig. 3.42. Section across the Chucal fault propagation anticline that forms part of the Chucal thrust system in the east-vergent thrust system (ETS) at 18°45′S (Arica cross-section) showing progressive deformation (see B in Fig. 3.40). Progressive unconformities separate in the east flank of the fold the Chucal Formation from the underlying Lupica (21.7 ± 0.8 Ma) and from the overlying Quebrada Macusa (17.5 ± 0.4 Ma) formations, and include a 4.2 Ma time span during which syntectonic deposition took place. The Lauca Formation is developed on the eastern flank of the fold.

1997; García *et al.* 1996, 1999*a*, *b*, 2000, 2004). The Azapa and El Diablo formations can be followed further north into southern Peru, where they have been named Moquegua Inferior (Marocco *et al.* 1985) and Magollo Unit (Flores 2004). The outcrops in the Central Depression are separated from those in the Precordillera by the high-angle Ausipar Thrust (Fig. 3.40).

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}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 Livilcar Formation, exposed in deeply incised valleys (Muñoz *et al.* 1988*a*), 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}Ar/^{39}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.* 2000*b*). These ages are in good agreement with magnetostratigraphic data obtained in this formation (Parraguez 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.* 2000*b*). Similar deposits are also known in southern Perú 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 southward relaying flexures along the Precordillera formed major, essentially north–south orientated 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.* 2004*a*). 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 1999; Pinto *et al.* 2004*a*) 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.

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.* 2000*b*, *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 silicic 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 Copaquilla–Tingnámar 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

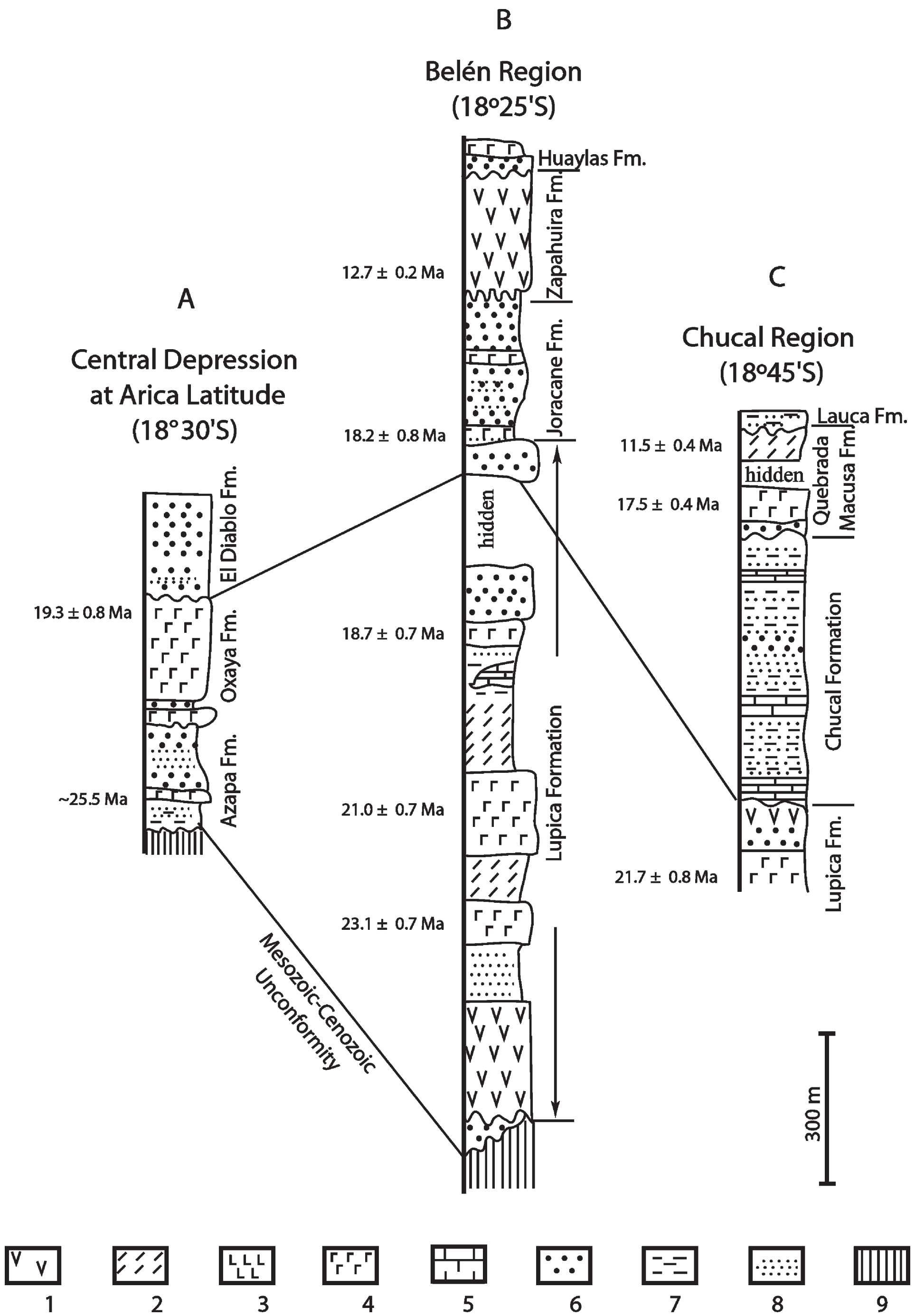


Fig. 3.43. Correlation of Cenozoic deposits between the Central Depression (A), the eastern Precordillera (B) and the Chucal region in the western (Chilean) Altiplano (C) in the Arica region, based on Farías *et al.* (2005*a*). For location of morphostructural units see Figure 3.39. Key: 1, Basaltic andesite; 2, Andesitic tuff; 3, Rhyodacitic lava; 4, Ignimbrite; 5, Limestone; 6, Conglomerate; 7, Mudstone; 8, Sandstone; 9, Pre-Oligocene substratum.

Chucal Anticline, and with associated syntectonic sedimentary deposits that contain an abundant fossil mammal fauna (Charrier *et al.* 1999, 2000, 2002*c*; Chávez 2001; Flynn *et al.* 2002*a*; 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.* 2002*c*, 2005*b*) (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 Lauca Formation in the Lauca Basin (Lauca Ignimbrite of Kött *et al.* (1995), and Muñoz & Charrier (1996); Lauca–Pérez Ignimbrite of Wörner *et al.* (2000*b*)) 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.* 2000*b*), 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 Ignimbrite by Kött *et al.* (1995), Riquelme (1998) and Wörner *et al.* (2000*b*). The Pérez Ignimbrite 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 basaltic–andesite cinder cone (El Rojo Norte) gave a $^{40}K–^{40}Ar$ date of 3.1 ± 0.2 Ma (Kött *et al.* 1995) (Fig. 3.40).

In addition to the deposition of the above-mentioned formations, volcanic arc activity in the eastern Precordillera and western Altiplano produced a middle Miocene to Quaternary succession of basaltic and andesitic lavas, laharic deposits and ignimbrites. These include the already-mentioned Zapahuira Formation (García 1996) or Zapahuira Layers (García 2001) and the deposits of the Mamuta and El Marquez volcanoes as well as several other arc volcanoes. Associated with this volcanic activity are exhalative gold deposits (e.g. Choquelimpie; Aguirre 1990), which have been included in the Altiplano–western Cordillera polymetallic metallogenic 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 Huaylillas formations, and Calientes Units respectively (Marocco *et al.* 1985; Flores 2004). The Lauca (or Lauca–Pérez or Huaylas) Ignimbrite in northern Chile is correlative with the Pachía Ignimbrite 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 Lauca Formation with undetermined 'lacustrine deposits' (Lavenu *et al.* 1989). A southern equivalent of the Azapa Formation is the Sichal Formation exposed further south in the Precordillera between 21º30′ and 22ºS (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 & Hérail 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 (19–20ºS)

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 19ºS and 20ºS, 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.* 2004*a*), 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 & Oncken 1999; Victor 2000; Farías 2003; Victor *et al.* 2004; Farías *et al.* 2005*a*) (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 (Farías 2003; Farías *et al.* 2005*a*), already described for the Arica cross-section. These deposits consist of thick alluvial westward-fining 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 from now partly eroded stratovolcanoes (Cerros 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 (Farías 2003; Farías *et al.* 2005*a*). The Camiña Ignimbrite (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 19ºS, the Moquella flexure appears to be the prolongation of the Ausipar Fault (Muñoz & Sepúlveda 1992; Pinto 1999). Further south, at 19º30′S, three westward-propagating flexures are well developed; from west to east these are Calacala, Aroma and Soga (Farías 2003; Farías *et al.* 2003), with the Aroma Flexure being connected with the Moquella Flexure by a NW-striking thrust with left-lateral component. At 20º30′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.* 2004*a*). Farías (2003) and Farías *et al.* (2005*a*) calculated a total relative surface uplift for the Calacala, Aroma and Soga flexures of 2100 m. If we consider that the altitude of the Central Depression in this region is *c.* 1000–1100 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. Farías (2003) and Farías *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; Sheffels 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 Ignimbrite, 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.

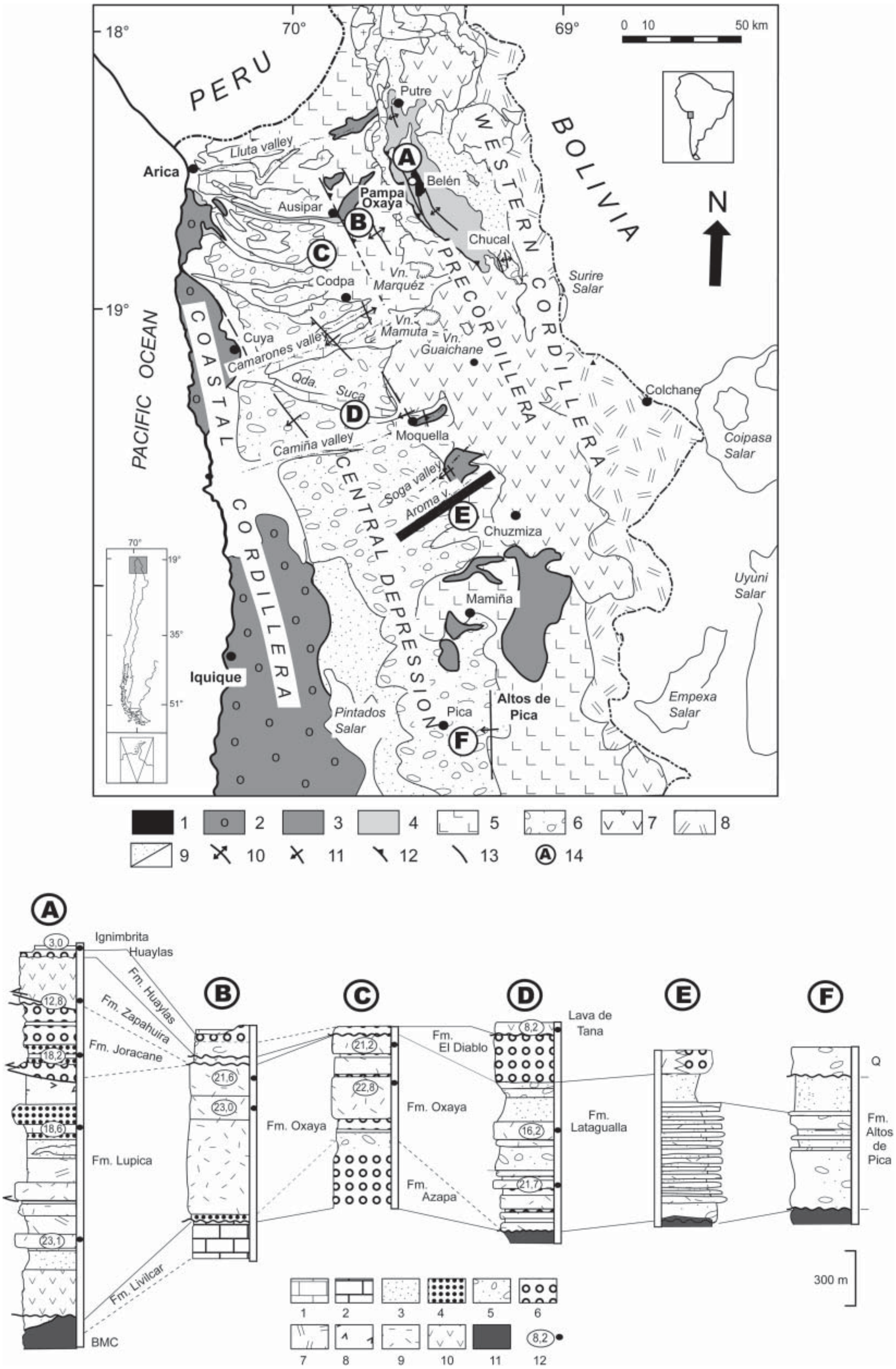


Fig. 3.44. Geological map for northernmost Chile between Arica and Iquique and correlation of Cenozoic successions along the Precordillera and Central Depression based on Pinto *et al.* (2004). Geological map: 1, Belén Metamorphic Basement; 2, Mesozoic units in the Coastal Cordillera; 3, Mesozoic units 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), García (2002), Farías *et al.* (2005*a*), 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.

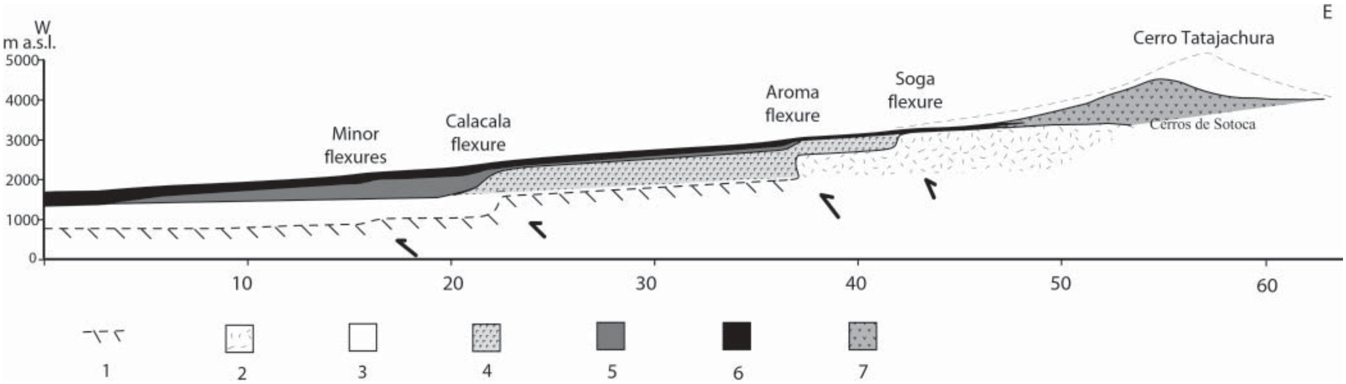


Fig. 3.45. Structural cross-section in the Aroma region with west-vergent flexures and associated syntectonic deposits (for location see Fig. 3.39 and Fig. 3.44) based on Farías *et al.* (2005*a*). Key: 1, base of the Cenozoic cover; 2, hidden; 3, Palaeozoic and Mesozoic substratum; 4, Altos de Pica Formation; 5, El Diablo Formation, lower member; 6, El Diablo Formation, upper member; 7, andesitic and basaltic lavas that interfinger with the El Diablo Formation.

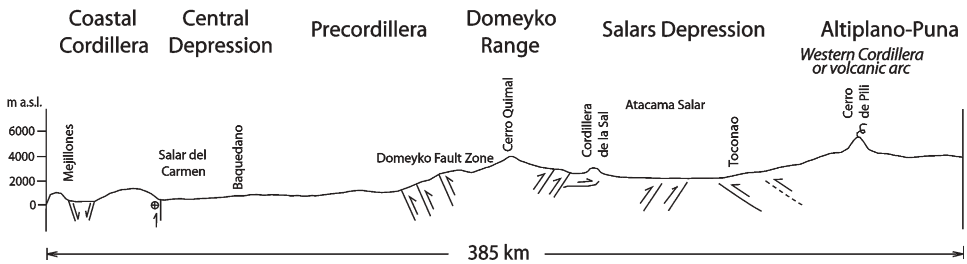


Fig. 3.46. East–west schematic cross-section at the latitude of Mejillones (Antofagasta region) showing morphostructural units and main structural elements that define the main relief features. Vertical scale approximately five times horizontal scale.

Altiplano. In this region, the western Altiplano is mostly covered by volcanic arc deposits.

Iquique–Chañaral (20ºS to 26ºS)

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 21ºS to 27º30′S, at the latitude of Copiapó. North of 21ºS 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, Maricunga and Negro Francisco, among others. South of *c.* 22ºS, 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.* 1996*b*, 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.* 1996*a*; 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 Maksaev & 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). $^{40}Ar/^{39}Ar$ 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

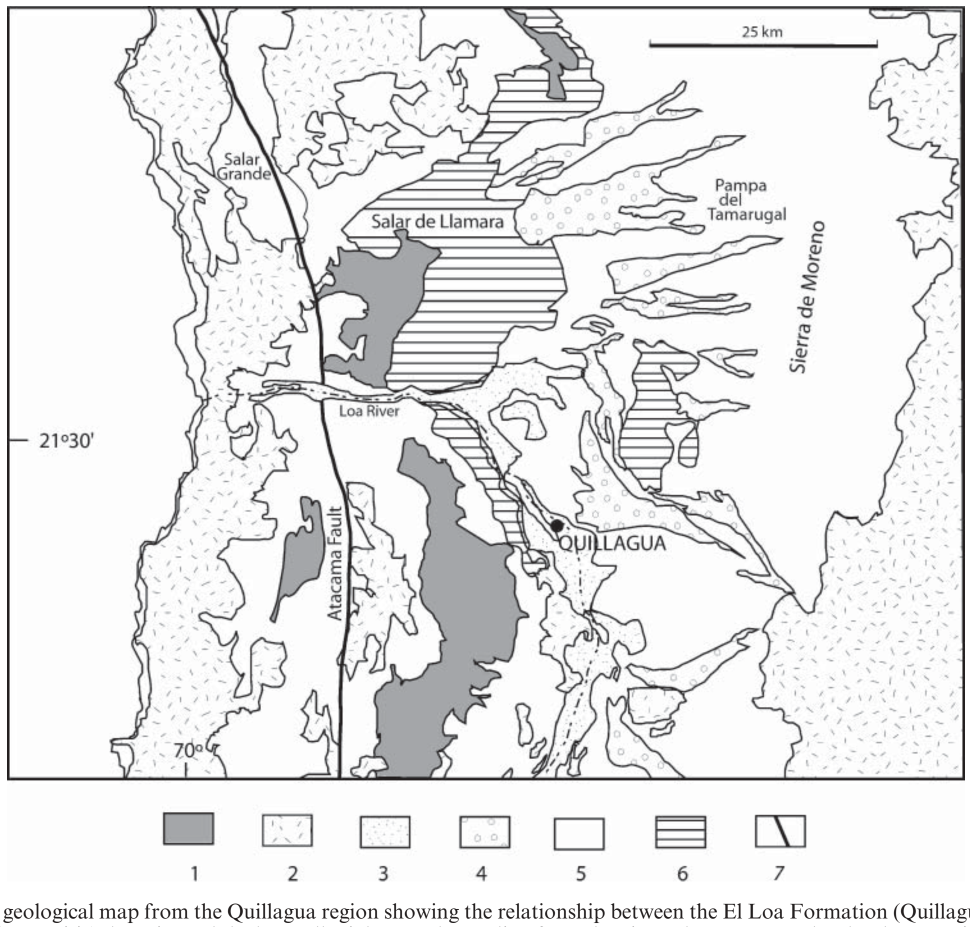


Fig. 3.47. Simplified geological map from the Quillagua region showing the relationship between the El Loa Formation (Quillagua lacustrine deposits), the Salar (evaporitic) deposits and the huge alluvial cones descending from the Sierra de Moreno. Salar development is controlled in this region by uplift 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).

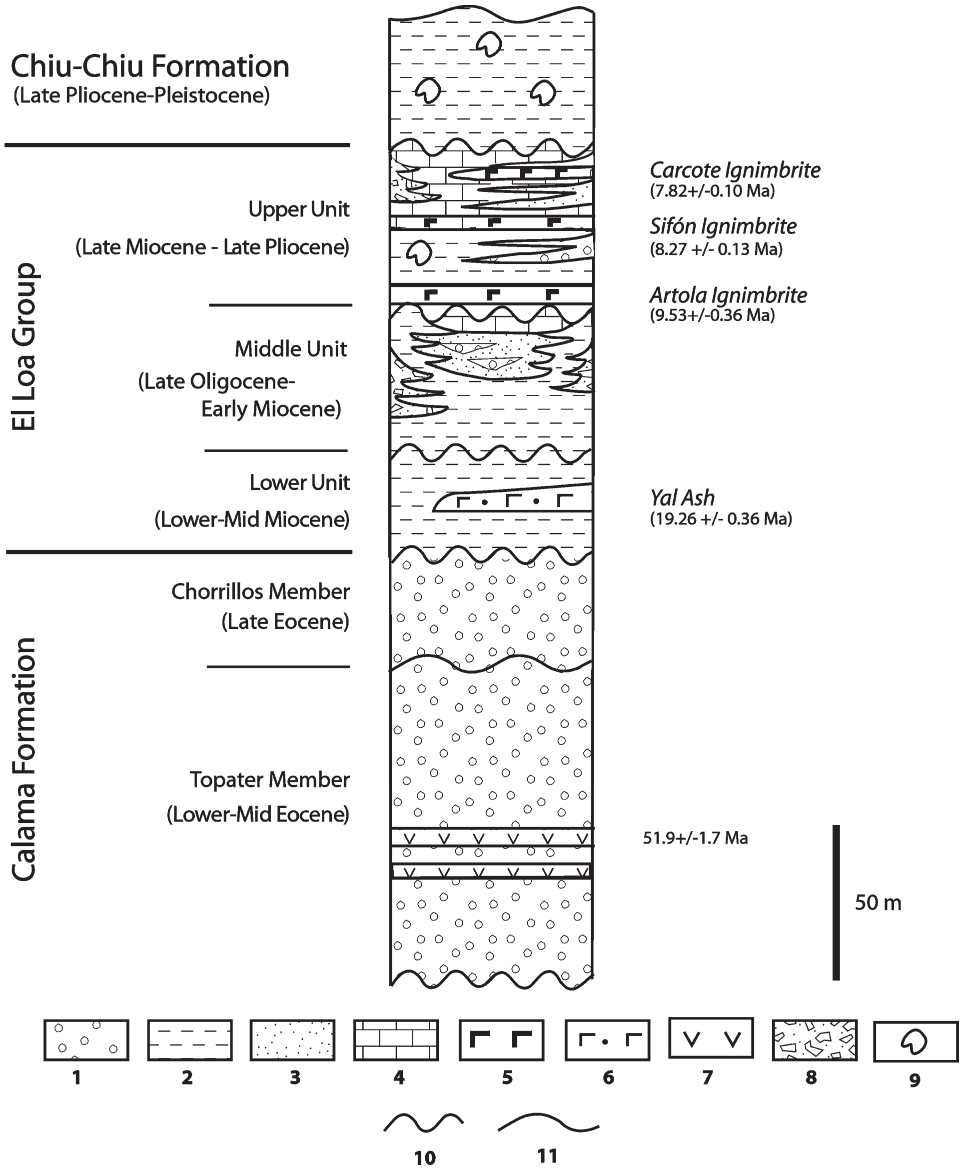


Fig. 3.48. Stratigraphic succession in the Calama Basin in the Calama region, adapted from May *et al.* (1999, 2005), Blanco *et al.* (2003) and Blanco (pers. comm.). Lower part of figure includes deposits of the second stage of Andean evolution in the lower part of the coarse clastic Calama Formation. These deposits are separated by an unconformity from the younger deposits of the third stage. The two upper units of the El Loa Group consist of several formations that represent different environments in the basin; in general, coarser materials are exposed on the sides of the basin and finer materials, like mudstones and limestones, form the inner deposits, e.g. the Jalquinche and Lasana formations in the middle unit respectively and the Chiquinaputo and Opache Formations in the upper unit. Key: 1, conglomerate and coarse sandstone; 2, siltstone and fine sandstone; 3, sandstone; 4, limestone; 5, ignimbrite; 6, ash deposits; 7, lava; 8, breccia; 9, mudstones with diatoms; 10, unconformity; 11, sedimentary discontinuity.

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).

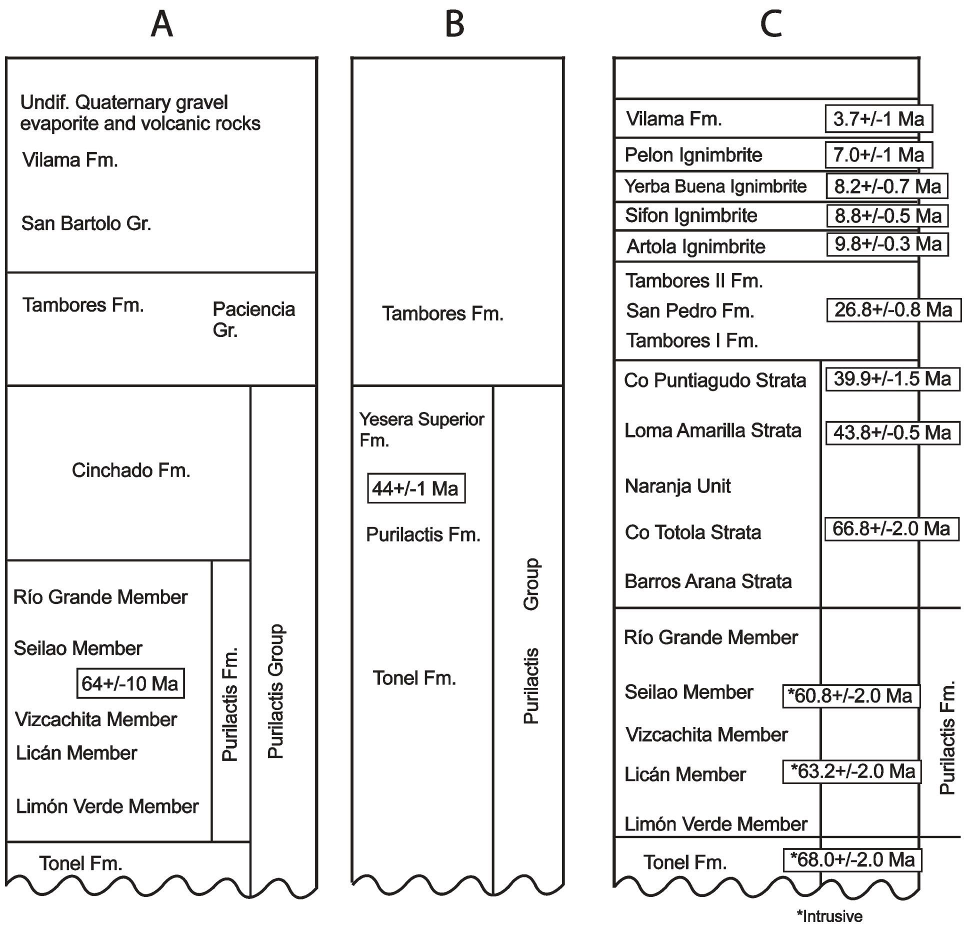


Fig. 3.49. Stratigraphic succession on the eastern flank of the Domeyko Range and western Salar de Atacama with radioisotopic age determinations, after Hartley *et al.* (1992) and Flint *et al.* (1993) (A), Charrier & Reutter (1994) (B) and Arriagada (1999) and Mpodozis *et al.* (2000) (C). Modified from Muñoz *et al.* (2002).

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; Ramírez & Gardeweg 1982; Marinovic & Lahsen 1984; Wilkes & Görler 1988; Hartley *et al.* 1992; Flint *et al.* 1993; Naranjo *et al.* 1994*b*; Mpodozis *et al.* 2000; Muñoz *et al.* 2000) (Fig. 3.49). Intercalated tuffs 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 ($^{40}Ar/^{39}Ar$ inverse isochron on hornblende) yielded an age of 7.65 ± 0.10 Ma age for the Sifón Ignimbrite.

In addition there are coarse to fine detrital and evaporitic deposits such as the Pliocene 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 1978*a*). 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 (Ramírez & Gardeweg 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 1987*a*; 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 orientated 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 1997*a*, *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, Zaldívar, La Escondida, and, further south, Exploradora and Potrerillos) (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 Palaeozoic granite intruded by porphyries (dated at 32–30 Ma) and their alteration products (Maksaev *et al.* 1988*b*; 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°S to 30°S)

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.*

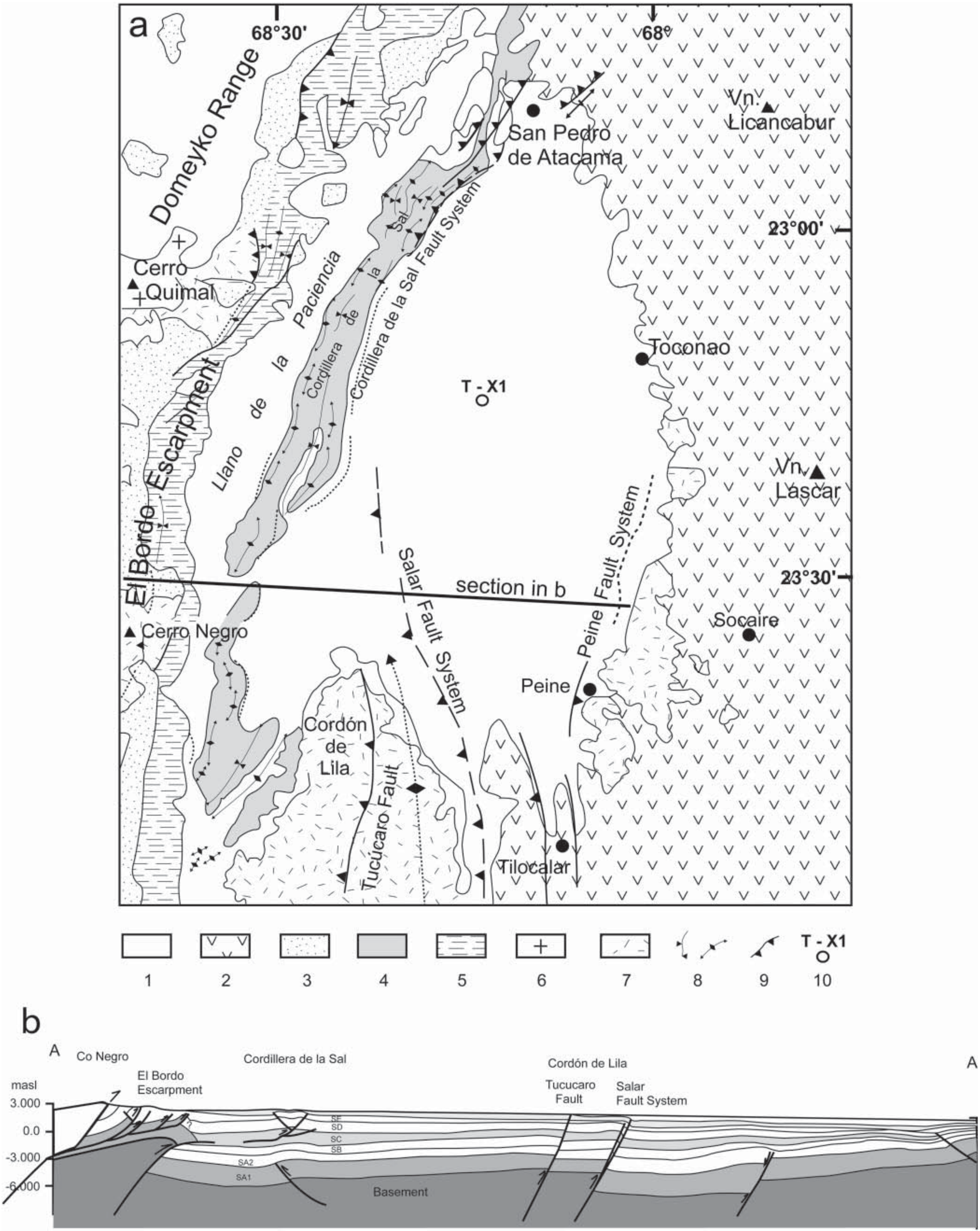


Fig. 3.50. Geological and structural setting of the Atacama Salar area, based on Muñoz *et al.* (2002). (**a**) Geological map with location of structural section. Key: 1, Late Miocene to Quaternary deposits; 2, Pliocene to Quaternary ignimbritic flows; 3, Miocene deposits; 4, Oligocene–Miocene deposits; 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.

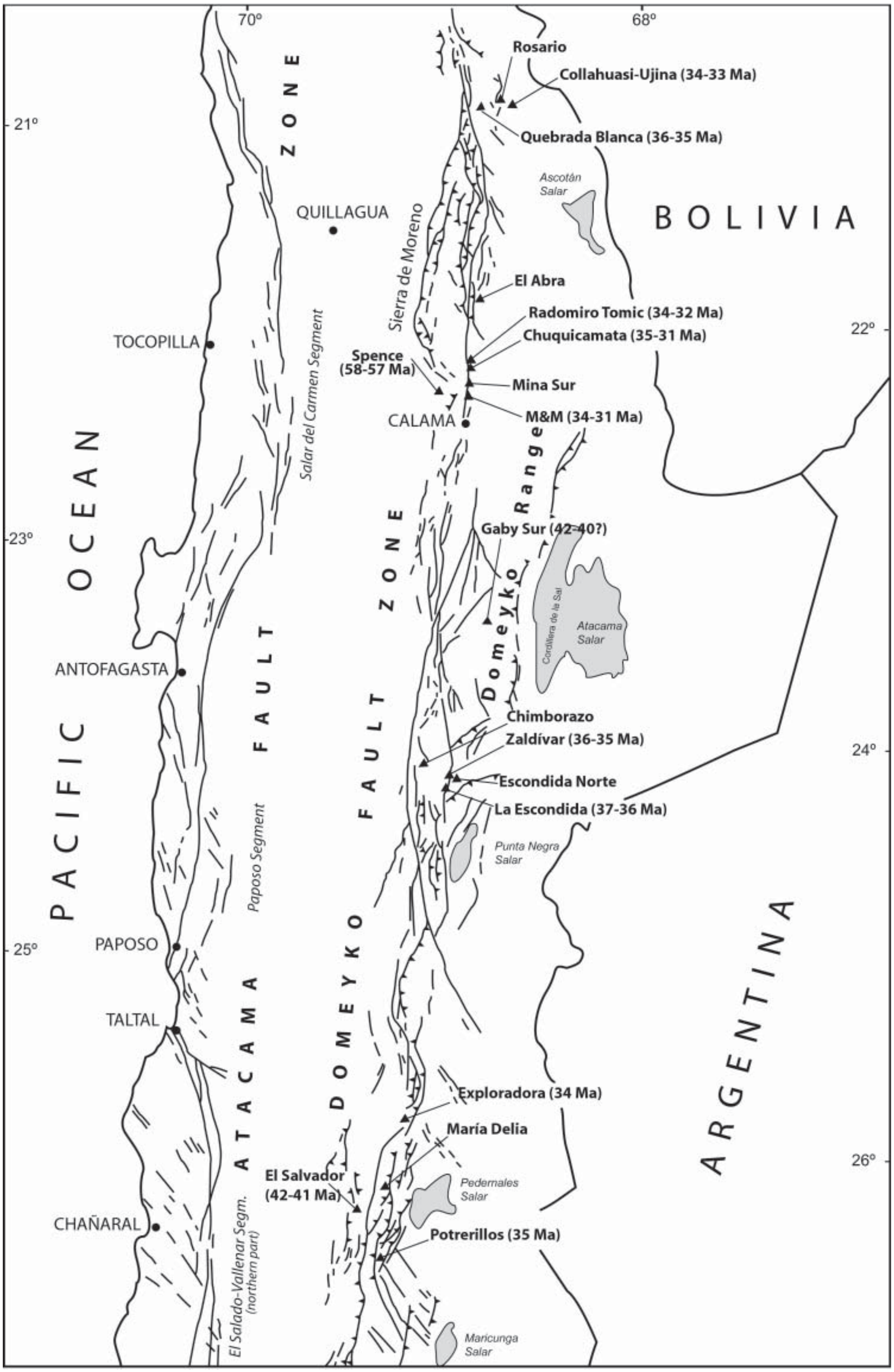


Fig. 3.51. Schematic structural map for northern Chile between 21° and 26°S with the Atacama Fault Zone and its different segments, and the Domeyko Fault Zone with the associated porphyry ore deposits and their emplacement ages. Porphyry ore deposits on the western side of the Domeyo Fault Zone are emplaced in Late Cretaceous deposits and have somewhat older ages than those on the eastern side, which were emplaced mainly in Palaeozoic rocks. Based on Cornejo *et al.* (1997), Camus (2003) and Cornejo (2005).

2003; Gómez 2003). 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

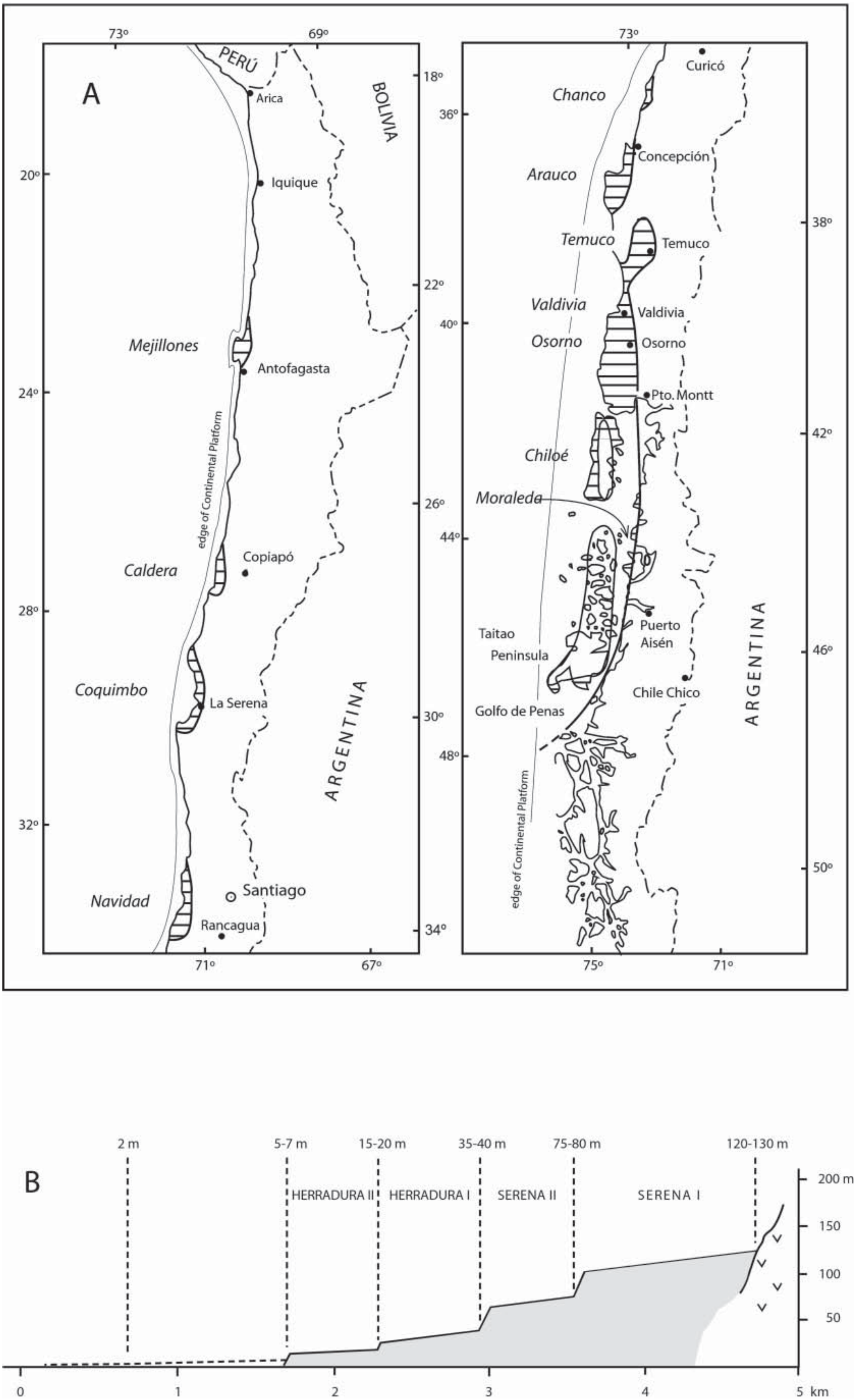


Fig. 3.52. Marine Neogene coastal deposits and morphological features on the western side of the Coastal Range in northern, central and southern Chile up to the triple-junction at 47ºS. (**A**) Distribution of coastal Neogene basin deposits, based on Martínez-Pardo (1990) and Sernageomin (2003). (**B**) 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 ingression occurred in the middle Miocene coinciding with a marine highstand and that extensional deformation occurred during late Miocene times. This middle Miocene marine ingression might be correlated with the similar ingression 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–Talinay, 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 1987*a*; Armijo & Thiele 1990; Delouis *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 north of this region (immediately north of the flat-slab segment), the Precordillera is again covered by coarse and fine detrital deposits, although with fewer ignimbritic intercalations than further north (Fig. 3.53). These deposits (mantling gravels of Willis (1929) or Atacama Gravels of Sillitoe *et al.* (1968)) consist of a poorly consolidated, up to 800-m-thick succession of gravels that fill a previously (late Palaeogene) deeply incised relief cut in the Intermediate (Sillitoe *et al.* 1968) or Sierra Checo del Cobre Surface (Mortimer 1973). This old drainage system, developed on the relief formed during the Incaic phase of deformation, crossed the entire forearc and reached the coast. After vertical movements along the Atacama Fault System that uplifted the western side of the coastal range, this connection with the ocean ceased. Depositional aggradation began 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 & Mpodozis 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 & Mpodozis 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 Precordillera to the ocean by uplift of the western Coastal Cordillera; (2) consequent abundant gravel sedimentation on the west-dipping Precordillera; 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 Precordillera 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 Farías (2003) and Farías *et al.* (2005*a*) for the incision of the Altos de Pica and El Diablo formations, and equivalent deposits, in the Precordillera 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-Wodzicky 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 Altiplano or Puna (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) (Mpodozis *et al.* 1995). The climatic hyperaridization has allowed preservation of the original shape of the oldest late Oligocene volcanoes and domes, and five volcanic 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 (Mpodozis *et al.* 1995).

In the southern part of this region, Reutter (1974) described, on both sides of the water divide at *c.* 29ºS, the Río 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 evaporitic deposits (gypsum and intercalated halite), and the latter comprises <500 m of gravels forming a 1° to 2° westward-dipping surface ('mantling gravels') with intercalated tuffs. The Río de la Sal deposits were later assigned

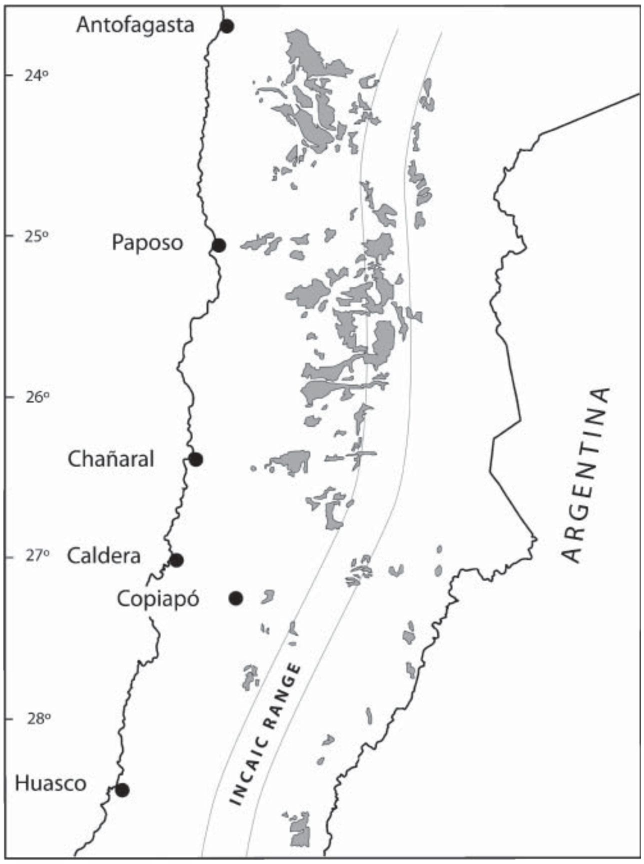


Fig. 3.53. Distribution of Miocene gravels between Antofagasta (24°S) and Huasco (28°30′S). Major developments of gravels is located between 25°S and 27°S and corresponds to the Atacama Gravels, based on Sernageomin (2003). Approximate distribution of the Late Cretaceous and Early Palaeogene granitoids (second stage of Andean evolution) is given to tentatively show distribution of the Incaic or 'proto-Domeyko' Range from which the gravels originated.

to the Escabroso Member of the Doña Ana Formation by Nasi *et al.* (1990). The overlying Potrerillos Formation (exposed north of 29°S) is correlated with the Miocene Atacama Gravels (see above) and represents the southernmost outcrops of these gravels (see Fig. 3.53). Between 29°S and 30°S, 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 29° and 30°S, 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, 1999*a*).

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 Plio-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.* 28°45′S and 30°30′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.* 28°45′ and 29°15′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, 1997*b*; 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.* 2005*a*).

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.* 2005*a*). These units have late Eocene to Miocene ages, and are known as El Maitén–Junquillar, Bocatoma (^{40}K–^{40}Ar, ^{40}Ar/^{39}Ar and one U–Pb ages range from 39.5 ± 1.3 Ma to 31.1 ± 1.2 Ma), Río Grande and Infiernillo (16.7 ± 0.6 Ma), and young to the east (Mpodozis & Cornejo 1988; Martin *et al.* 1995, 1997*b*; Bissig *et al.* 2001). The age of the intrusive rocks matches the age of the Abanico and Farellones formations between 33°S and 36°S; this is older than those emplaced in these formations, as well as in Mesozoic deposits forming the eastern Principal Cordillera (between 33°S and 36°S), 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.* 2002*c*).

Central Chile between Santiago and Concepción

South of 33°S both the Central Depression and basalt–andesitic 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 33°40′S and 34°15′S, are extensive exposures of late Miocene marine sediments known as the Navidad Formation (Darwin 1846; Tavera 1979*b*; Encinas *et al.* 2003; Finger *et al.* 2003) (Fig. 3.52). This formation is overlain by the

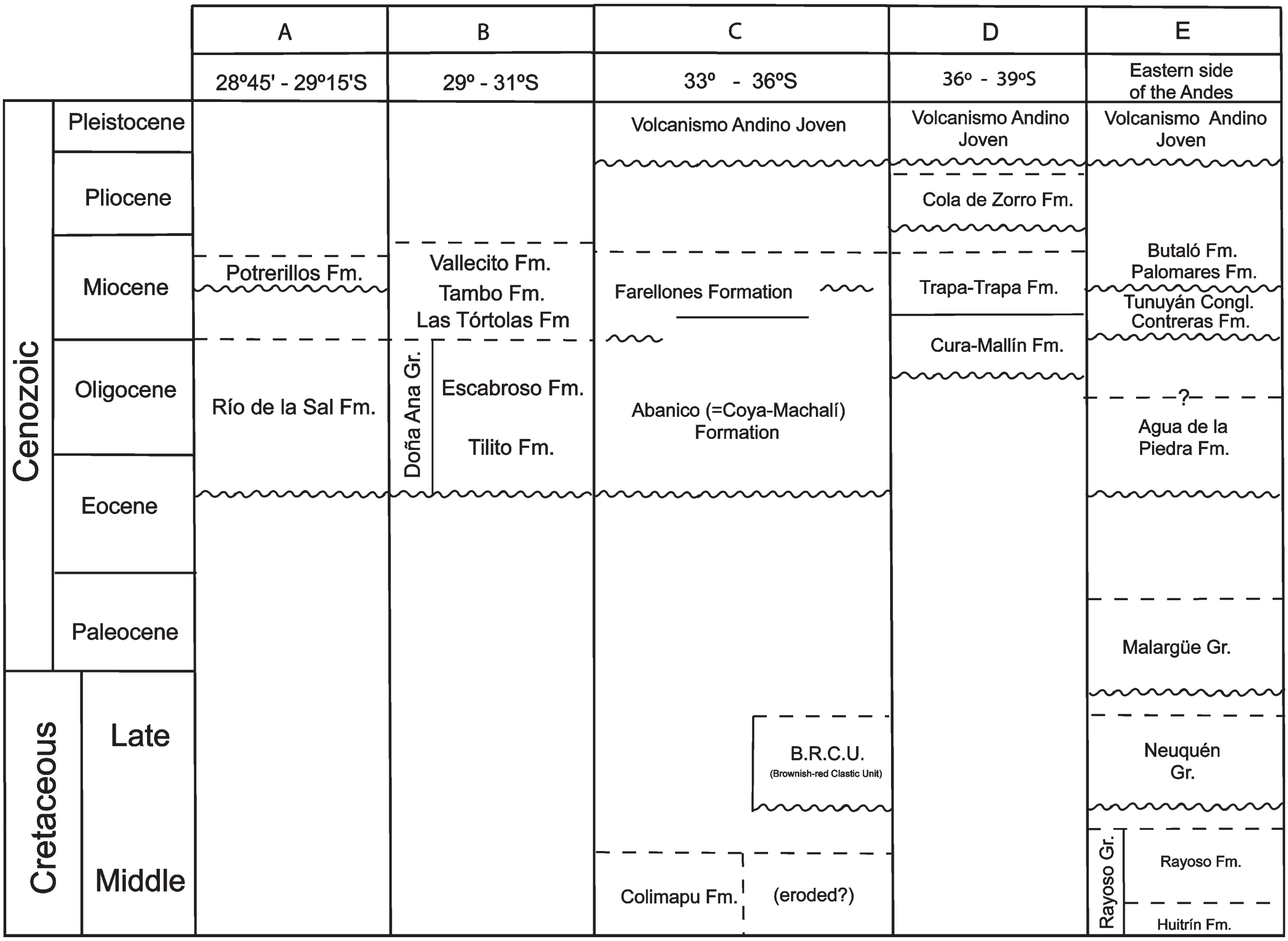


Fig. 3.54. Stratigraphic relations between Cenozoic units along the Principal Cordillera between 28°45′S and 39°S, and their relation with Mesozoic underlying units and equivalent deposits on the east side of the Andes, based on: (A) Reutter (1974); (B) Mpodozis & Cornejo (1988), Nasi *et al.* (1990), Bissig *et al.* (2001); (C) Klohn (1960), Thiele (1980), Charrier (1981*b*), Charrier *et al.* (2002*b*, 2005*a*); (D) González & Vergara (1962), Muñoz & Niemeyer (1984), Niemeyer & Muñoz (1983); (E) Legarreta & Gulisano (1989), Giambiagi (1999), Ramos (1999), Giambiagi *et al.* (2001).

Pliocene, transitional marine to continental La Cueva Formation (Tavera 1979*b*) 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 33°S 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 33°30′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 lahar and volcanic avalanche deposits such as La Cueva Fm that reached the Central Depression and even the coast. Between 33°30′S and 34°30′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.* 1984*a*). The origin of these tuff deposits is apparently associated with the Maipo Caldera which is situated at 34°S (Stern *et al.* 1984*a*). The wide distribution of such tuff deposits suggest that this pyroclastic flow covered most of the Central Depression around 34°S, 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

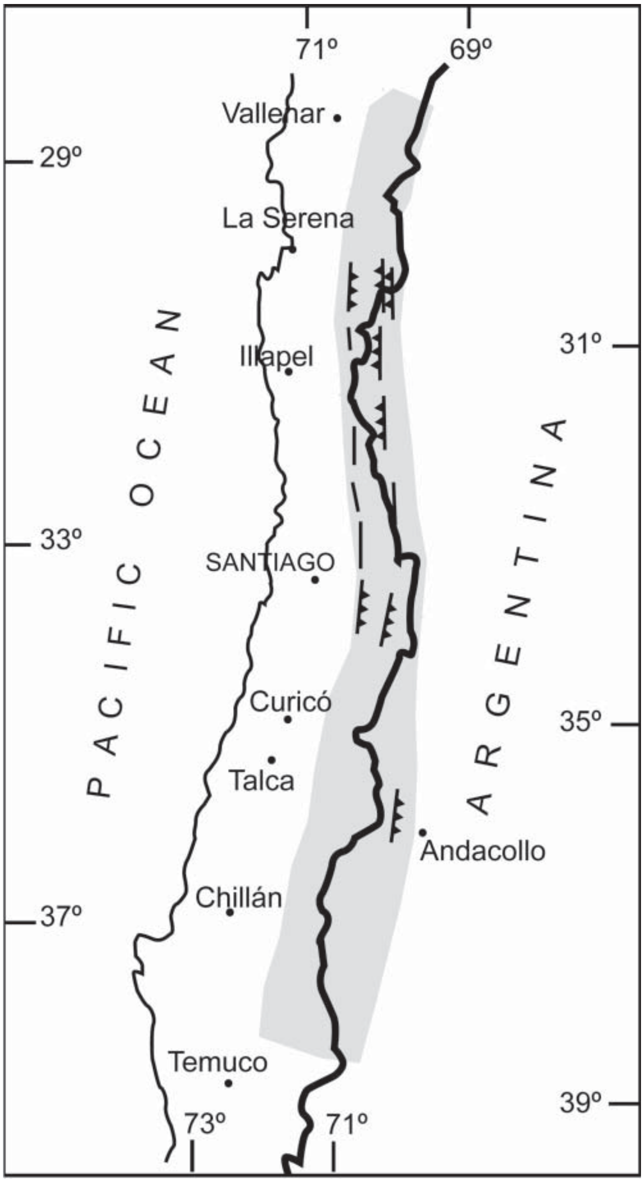


Fig. 3.55. Tentative suggested extent of the Cenozoic continental Abanico Extensional Basin along the Principal Cordillera, between 29ºS and 39ºS, and trace of major faults that probably controlled basin development and tectonic inversion. Triangles along the faults point towards the hanging wall of the first normal faults and then reversed faults.

Cordillera that were channelized along the main river valleys (Marangunic *et al.* 1979).

Principal Cordillera. Cenozoic deposits in the Principal Cordillera in this region correspond to the southward prolongation of the coeval units described in the La Serena region and accumulated in the Abanico Extensional Basin (Fig. 3.55). The dominantly volcanic, middle?–late Eocene to early Miocene Abanico Formation, and its local equivalents, the Coya-Machalí (Klohn 1960) and the Colbún formations (Karzulovic *et al.* 1979), and the early–late Miocene Farellones Formation make up the pre-Pliocene Cenozoic deposits in the Principal Cordillera of Central Chile (31–36ºS) (Aguirre 1960; Klohn 1960; González & Vegara 1962; Charrier 1973*b*, 1981*b*; Thiele 1980; Drake *et al.* 1982*a*; Charrier *et al.* 2002*b*; Fuentes 2004). The Abanico Formation consists of a locally strongly folded, *c.* 2000 m thick succession of volcanic, pyroclastic volcaniclastic and sedimentary deposits including abundant subvolcanic intrusions of the same age (Vergara *et al.* 2004), with a well developed paragenesis of low grade metamorphic minerals (Levi *et al.* 1989; Aguirre *et al.* 2000; Fuentes *et al.* 2002; Bevins *et al.* 2003; Padilla & Vergara 1985; Fuentes *et al.* 2002; Fuentes 2004). The outcrops of this formation form two north–south orientated belts separated by the Farellones Formation (Sernageomin 2003). At its western side the 34.3 ± 2.2 Ma old basal deposits of the Abanico Formation unconformably overlie the Late Cretaceous (72.4 ± 1.4 and 71.4 ± 1.4 Ma) Lo Valle Formation (Gana & Wall 1997). The hiatus between both formations covers a time span of 37 million years.

The younger Farellones Formation is a thick, gently folded, almost entirely volcanic unit forming a north–south trending zone of outcrops between approximately 32º and 35ºS (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 1981*b*; 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 compressional 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 (>38ºS) in Chile is reflected by the presence of thick volcanic series equivalent in age to the ones exposed between 33ºS and 36ºS. 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 30ºS (see Charrier *et al.* 2005*a*), 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 basement 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 44º S. The finding of Eocene fossil mammal remains (El Tapado Fauna) within the Abanico Formation, in the Tinguiririca river valley at 35ºS west of Termas del Flaco (Wyss *et al.* 1994), indicates a late Eocene onset for deposition (Charrier *et al.* 2002*b*). This fauna is considerably older than the one collected at Termas del Flaco next to the basal unconformity (Tinguiririca Fauna; Wyss *et al.* 1990, 1993, 1994; Flynn *et al.* 1995, 2003; Flynn & Wyss 1999).

The deposits of the Farellones Formation typically cover the Abanico Formation. The contact has generally been reported as unconformable (Aguirre 1960; Klohn 1960; Jaros & Zelman 1967; Charrier 1973*b*, 1981*b*; Thiele 1980; Moscoso *et al.* 1982*b*), but other authors (Godoy 1988, 1991; Godoy & Lara 1994; Godoy *et al.* 1999) view the contact as (1) being conformable or pseudo-conformable, or (2) corresponding to a regional low-angle thrust that they connect with the El Fierro Fault at Termas del Flaco described by Charrier *et al.* (2002*b*). The age of the basal Farellones deposits that cover the Abanico Formation and the contact between these units varies considerably at different localities. Considering that the youngest ages obtained for the Farellones Formation in Central Chile (33–36ºS) correspond 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.* 2002*b*).

The Abanico Formation is correlated southward with the Cura-Mallín Formation (Charrier *et al.* 2002*b*). The outcrops of the Farellones Formation end at 35ºS; to the east it can be correlated 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 35ºS, 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 34º30′S occurred between 25.2 ± 0.2 Ma and 16.1 ± 0.5 Ma (Charrier *et al.* 2002*c*). In Miocene times, during the inversion of the Abanico Extensional Basin, movement along the San Ramón Fault that constitutes 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 1981*b*; 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, developed within hydrothermal alteration zones linked to multiphase 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 34ºS (Camus 1977; Charrier & Munizaga 1979; Charrier 1981*b*; Cuadra 1986; Godoy *et al.* 1994; Gómez 2001), Sierras de Bellavista at 34º45′S (Klohn 1960; Vergara 1969; Charrier 1973*b*; Malbran 1986), and possibly also at 35ºS (Corona del Fraile Formation; González & Vergara 1962). These areas form a north–south alignment suggesting a tectonic control for this activity. Finally, the volcanoes 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.* 1995*a*). 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, Longaví, Chillán, Lonquimay and Llaima (see Chapter 5).

South-central Chile (Concepción–Lonquimay region, 37ºS to 39ºS)

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 (Quiriquina 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 following Neogene to Pliocene formations have been recognized: Ranquil (Miocene, marine sequence), and Tubul (Pliocene, marine sequence) (Muñoz Cristi 1946, 1973; Wenzel *et al.* 1975; Pineda 1983*a*, *b*; Arévalo 1984). The Ranquil and Tubul formations can be correlated to the north with the Navidad and Coquimbo formations respectively. The evolution of this outstanding sedimentary sequence, which contains hydrocarbon and important coal reserves, was characterized by an alternation of transgressive and regressive episodes, controlled by eustatic changes, local subsidence and uplift of tectonic blocks (Wenzel *et al.* 1975; Pineda 1983*a*, *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.* 2003*a*). 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 (36–43º30′S) formed when the locus of Andean magmatic activity 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 generated sedimentary basins west, within, and east of the Main Cordillera (Muñoz *et al.* 2000). Continuing extension accompanied 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.* 2005*a*), 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 36ºS, the Cura-Mallín Formation represents the prolongation of the younger part of the Abanico Formation down to 39ºS (Niemeyer & Muñoz 1983; Muñoz & Niemeyer 1984; Suárez & Emparan 1995, 1997; Radic *et al.* 2002) (see Fig. 3.54). Between 36ºS and 37ºS, 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 Malla-Malla Member; Niemeyer & Muñoz 1983; Muñoz & Niemeyer 1984). Between 37ºS and 39ºS, 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 36ºS and 38º 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 37ºS (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.* 2002*b*; 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 38ºS 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 ^{40}K–^{40}Ar 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 ^{40}Ar/^{39}Ar 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 37º10′S. In the Lonquimay region between 38°S and 39°S, Suárez & Emparan (1997) report several ^{40}K–^{40}Ar ages between 19.9 ± 1.4 Ma and 7.8 ± 0.8 Ma (this youngest from an altered sample) for the Cura-Mallín Formation. The upper fluvial and lacustrine Río Pedregoso Member yielded ^{40}K–^{40}Ar ages of 17.5 ± 0.6 Ma and 13.0 ± 1.6 Ma (Suárez & Emparan 1997). With regard to palaeontological dating of the formation, the remains of a variety of vertebrates (fishes, mammals and a bird; Marshall *et al.* 1990; Suárez *et al.* 1990; Wall *et al.* 1991; Rubilar 1994; Azpelicueta & Rubilar 1998) and a specimen of toxodontid notoungulate *Nesodon conspurcatus* (Croft *et al.* 2003) are referable to the Santacrucian South American Land Mammal 'Age' (late early Miocene, 17.5–16.3 Ma, *sensu* Flynn & Swisher 1995). The Cura-Mallín Formation is conformably overlain by the Late Miocene Trapa-Trapa Formation, an equivalent to the younger parts of the Farellones Formation.

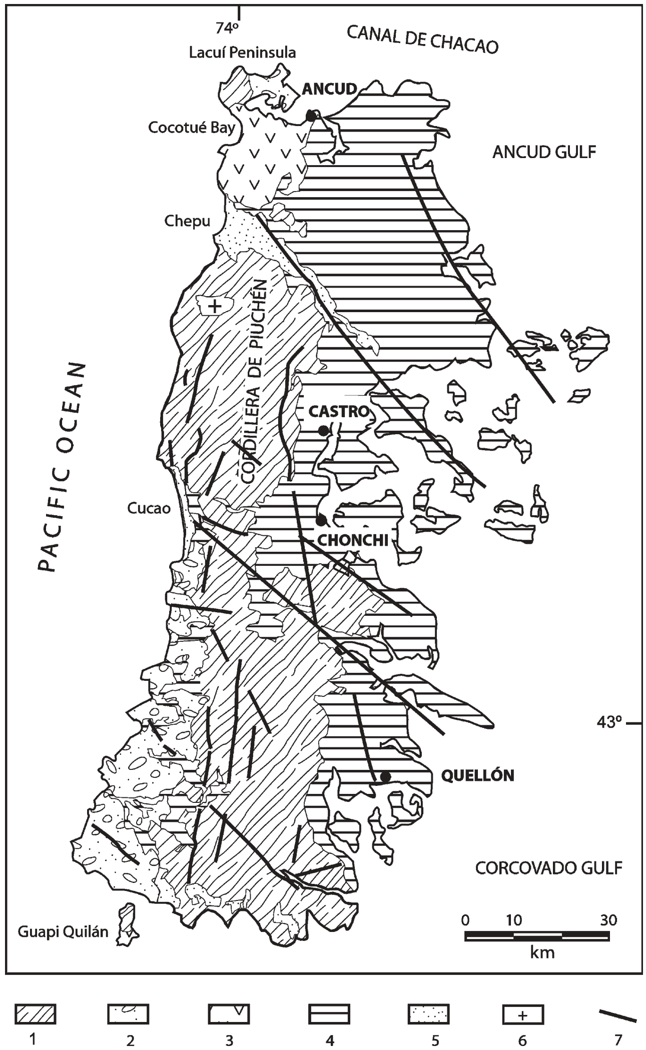


Fig. 3.56. Geological map of Chiloé Island (41–43º30′S) showing the main stratigraphic units, based on Muñoz *et al.* (1999) and Antinao *et al.* (2000). The island that forms the southward prolongation of the Coastal Cordillera south of Canal de Chacao consists of a north–south orientated core of metamorphic rocks with Cenozoic deposits on both sides. Key: 1, Palaeozoic–Triassic metamorphic basement; 2, latest Palaeogene–Neogene sedimentary deposits; 3, volcanic complexes (Ancud and Guapi Quilán); 4, glacial deposits; 5, beach, estuarine and fluvial deposits; 6, Eocene Metalqui Pluton; 7, 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).

Lonquimay to Chiloé (39ºS to 43ºS)

Coastal Cordillera. Cenozoic deposits in Chiloé Island (41–43º30′S) (see Fig. 3.52B) are exposed on both sides of a north–south orientated axis of metamorphic rocks (Cordillera del Piuchén) (Fig. 3.56) which have been correlated with the Bahía Mansa Metamorphic Complex exposed in the Coastal Cordillera west of Osorno (40º45′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 (Antinao *et al.* 2000). The stratigraphic succession continues with the Chonchi Beds (Quiroz *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 Lacui Formation (Valenzuela 1982) and the Cucao Beds on the western side (Quiroz *et al.* 2003), which are correlative with the Cheuquemó 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 northwestern (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 (Quiroz *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-orientated 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.* 1979*a*; López-Escobar *et al.* 1995*a*; Cembrano *et al.* 1996, 2000; Lavenu & Cembrano 1999*b*; Folguera *et al.* 2001; Melnick *et al.* 2002, 2003*b*; Rosenau *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 (Traiguén and Ayacara formations), the LOFZ has been active at least since the Eocene epoch (Hervé 1994).

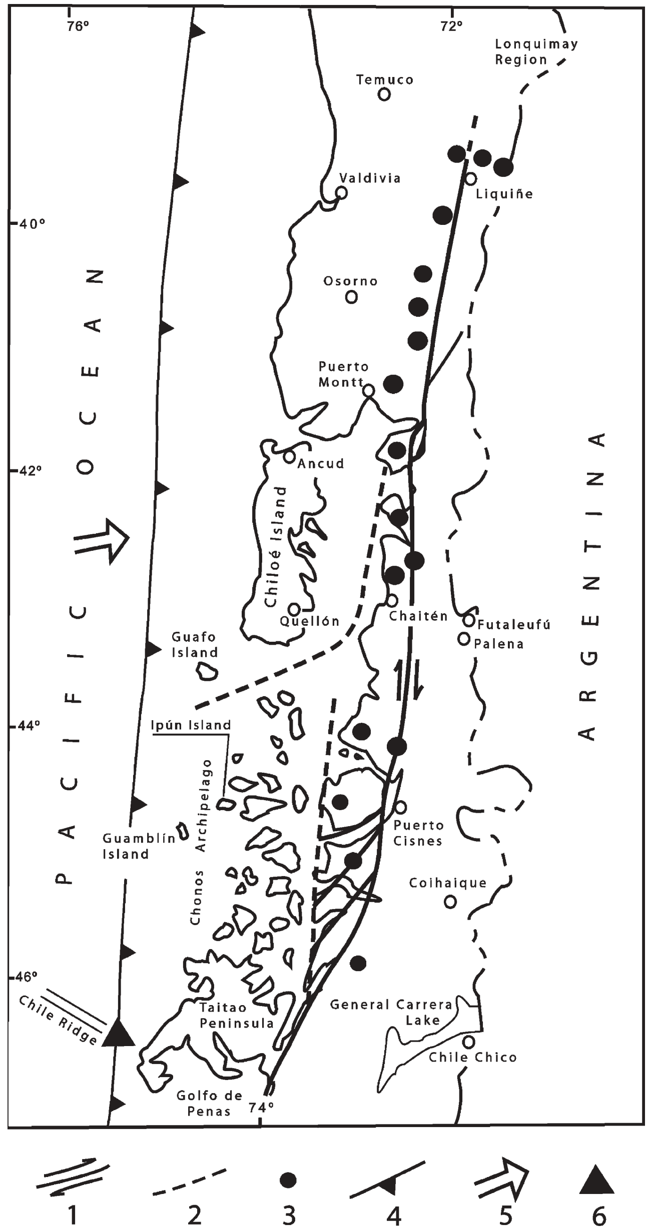


Fig. 3.57. Liquiñe–Ofqui Fault Zone and associated structures and volcanic centres (see Figs 3.30 & 3.51 for a regional overview of the major fault systems in Chile). The Golfo de Penas pull-apart basin is indicated. Based on López-Escobar *et al.* (1995*a*), Cembrano *et al.* (1996) and Silva (2003). Key: 1, main trace of the Liquiñe–Ofqui Fault Zone; 2, inferred fault traces; 3, Quaternary volcanic centres; 4, approximate location of the Chile trench; 5, subduction vector; 6, Nazca–Antarctica–South America triple-junction.

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 as well as 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.* 1995*a*). 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 Río de la Sal Formation, the Doña Ana Group and the Las Tórtolas, Abanico and Cura-Mallin 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 splays of the Liquiñe–Ofqui Fault Zone. The latter lineament not only influenced 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 30°S, i.e. within and south of the flat-slab segment, the existence of the Incaic 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 35°15′S (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. 3.53). 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 Puna or southern Altiplano below the Plio-Quaternary volcanic cover.

To complete the palaeogeographic scenario, the broad western flank of the Incaic Range in central and northern Chile (now the Precordillera) was separated from the Coastal Cordillera to the west by a narrow depression (the proto-Central Depression) (Fig. 3.58). West of the Coastal Cordillera lay the continental platform and the coastal area where alternating marine and continental sediments, interspersed with erosional unconformities, were deposited.

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 Potoco 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.

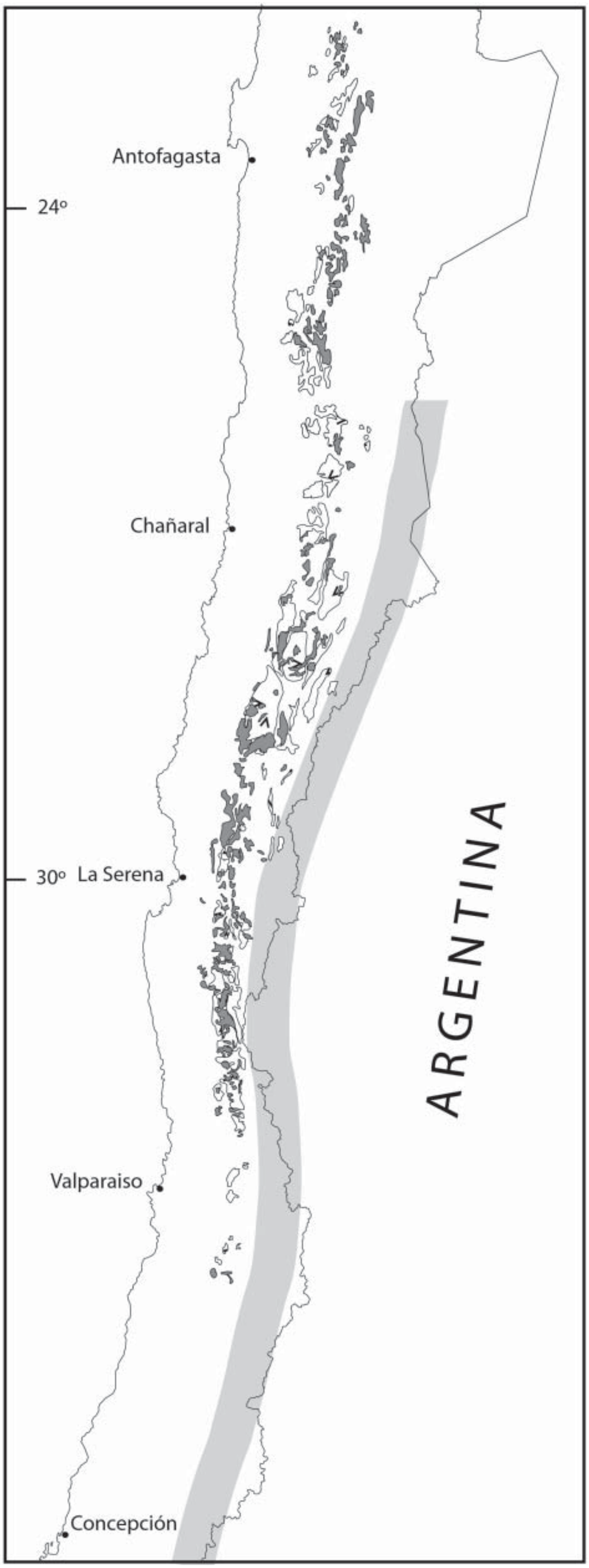
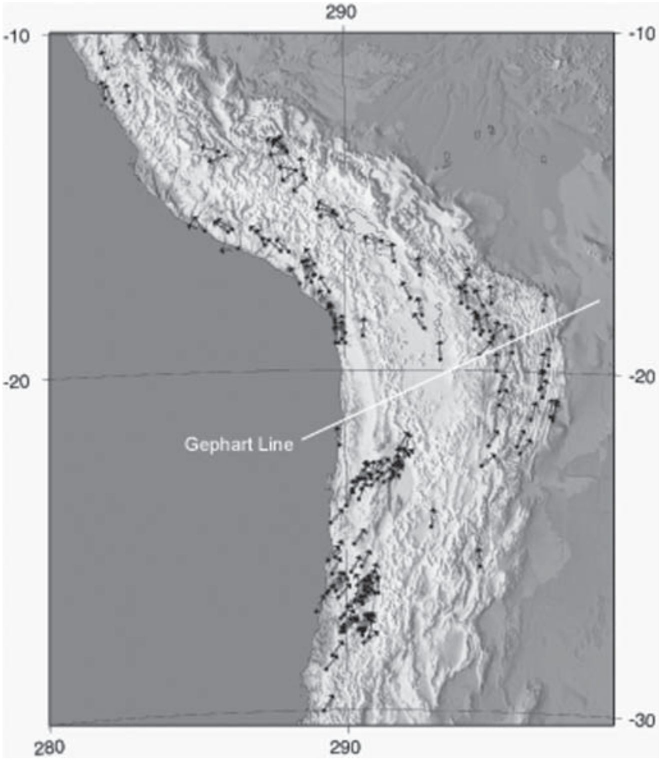


Fig. 3.58. Tentative palaeogeographic organization during Late Palaeogene (third stage of Andean evolution) with the Incaic relief (Proto-Domeyko Range) indicated by the outcrop distribution of the Late Cretaceous and early Palaeogene intrusive (dark areas) and volcanic rocks (white areas) and the Abanico Extensional Basin (light grey areas) to the east between 26ºS and 37ºS. Somewhat later, during early Neogene times, deposition of the Atacama Gravels was occurring on the western flank of the Incaic Relief (see Fig. 3.53) while on its eastern side, on the inverted Abanico Basin, volcanic activity gave rise to the Farellones Formation.

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 14ºS, in southern Peru to 27ºS), the Southern Volcanic Zone (33ºS to 47ºS) and the Austral Volcanic Zone (south of 52ºS) (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.* 90º. 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). Palaeo-magnetic 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 Cenozoic 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 ocurred during uplift in Late Neogene time. Absence of post-Miocene rotation across the area suggests that other



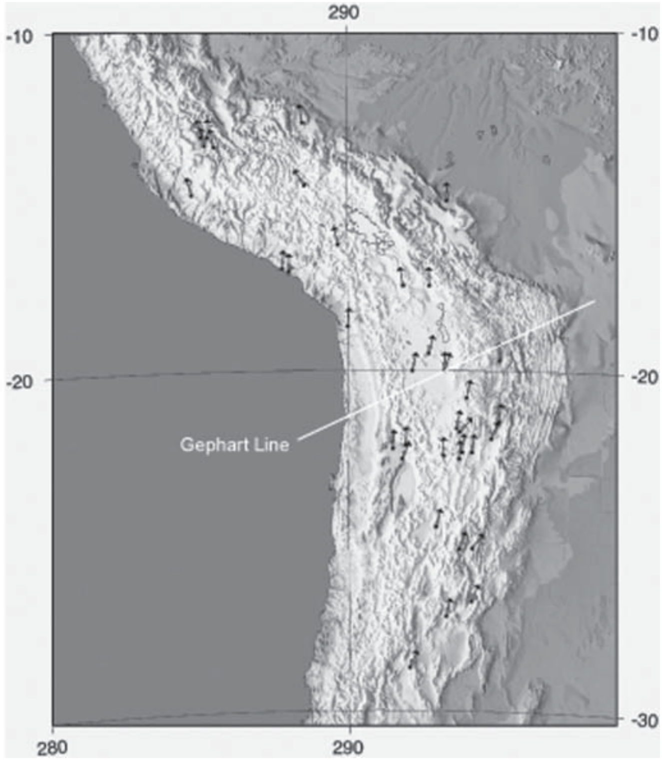


Fig. 3.59. Bolivian Orocline and Gephart Line of symmetry with palaeo-magnetic data. (**Left**) Considerable pre-Neogene counterclockwise rotation in the Andean forearc north of the Gephard Line in southern Peru and clockwise rotation south of this line in northern Chile, and minor rotation east of the arc, in the Altiplano, Eastern Cordillera and Subandean Sierras. (**Right**) Neogene palaeo-magnetic data showing minor deformation at this time. Observe slight rotation in the Altiplano. Based on Arriagada (2003), Arriagada *et al.* (2003) and Roperch *et al.* (2006).

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 ocurred 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 Antactica, 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.* 27ºS and 33ºS (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, N76ºE 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, N76ºE-orientated segment, began to form 25 million years ago, and intersected the continental margin *c.* 12 million years ago at 33º 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.* 2002*a*). 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.* 1983*a*; 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

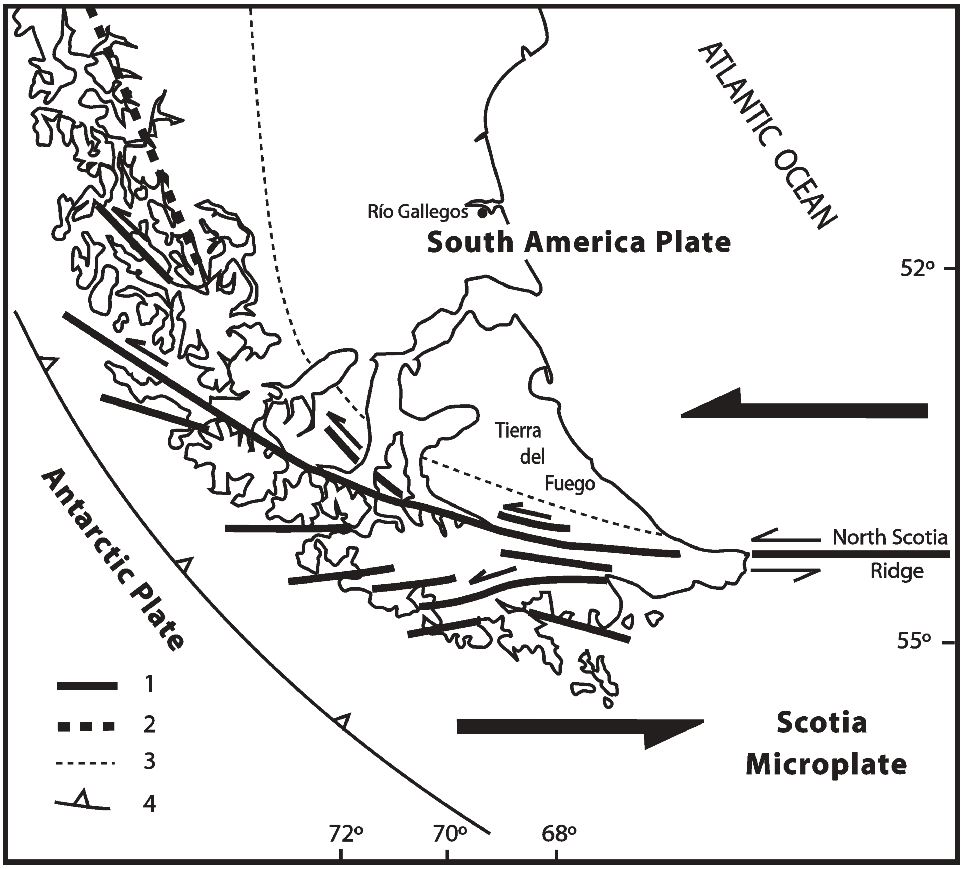


Fig. 3.60. Patagonian Orocline. Left-lateral movement along shear zones located along the weakness zone represented by the suture of the 'Rocas Verdes basin' caused bending of the mountain range (Cunningham *et al.* 1991). The major left-lateral Magallanes–Fagnano Fault Zone separates the South American Plate from the Scotia Microplate. If the same tectonic condition continues in that region, strain accommodation will require development of new shear zones further north along the same suture zone (see broken line). Key: 1, main faults associated with the orocline bend; 2, suture zone (apparently not yet active); 3, boundary between the Patagonian Andes and the foreland; 4, trench.

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.* 2002*b*) (Fig. 3.62). 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.* 1994*c*, 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.* 1994*a*; 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.* 2004*a*; Farías *et al.* 2005*a*) 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)

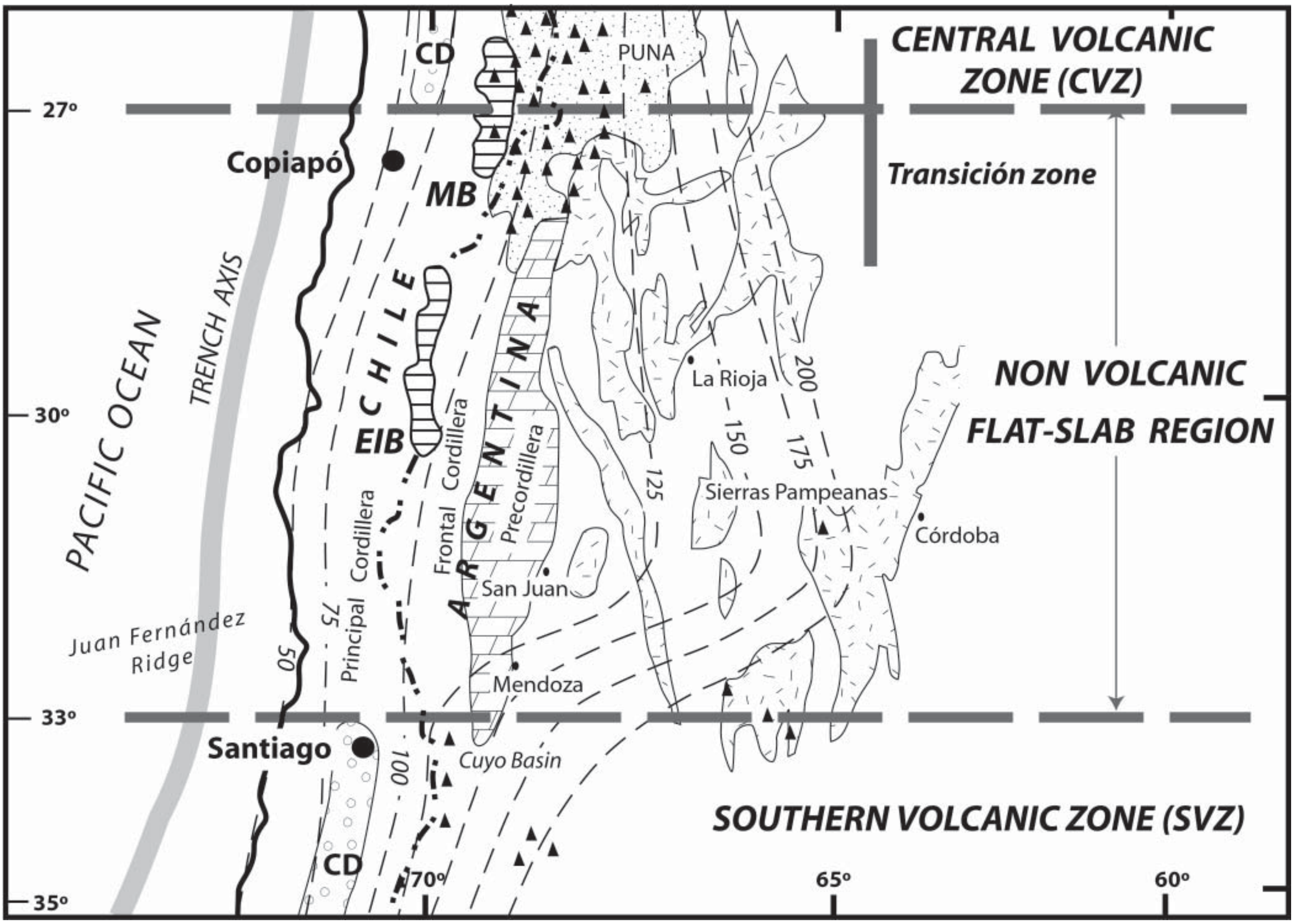


Fig. 3.61. Morphological features associated with the flat-slab subduction segment (zone), distribution of volcanoes (triangles) and contours of depth of Wadati–Benioff plane and (for location along the Chilean continental margin, see Fig. 3.1b). Central Depression (CD) is developed only north and south 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.

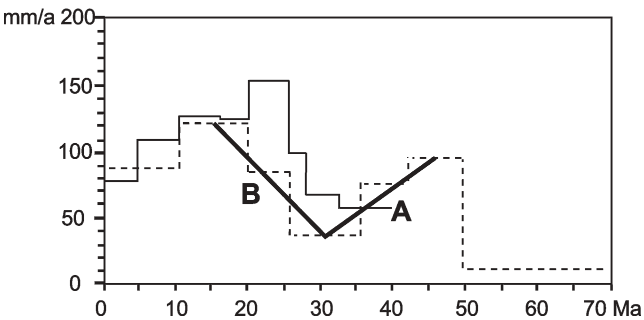


Fig. 3.62. Correlation between periods of decreasing and increasing convergence rates, according to Pardo-Casas & Molnar (1987) (discontinuous line) and Somoza (1998) (continuous line), and periods of extension and compression in the central part of the Southern Andes, based on Charrier *et al.* (2002*b*).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

and Maksaev *et al.* (2003). Kurtz *et al.* (1997) modelled the exhumation of plutons on the base of $^{40}Ar/^{39}Ar$ 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 on apatites 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 $^{40}Ar/^{39}Ar$ ages in the same bodies, suggesting an episode of accelerated denudation at the end of Miocene and beginning of Pliocene times (Maksaev *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 Maksaev *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 Río 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

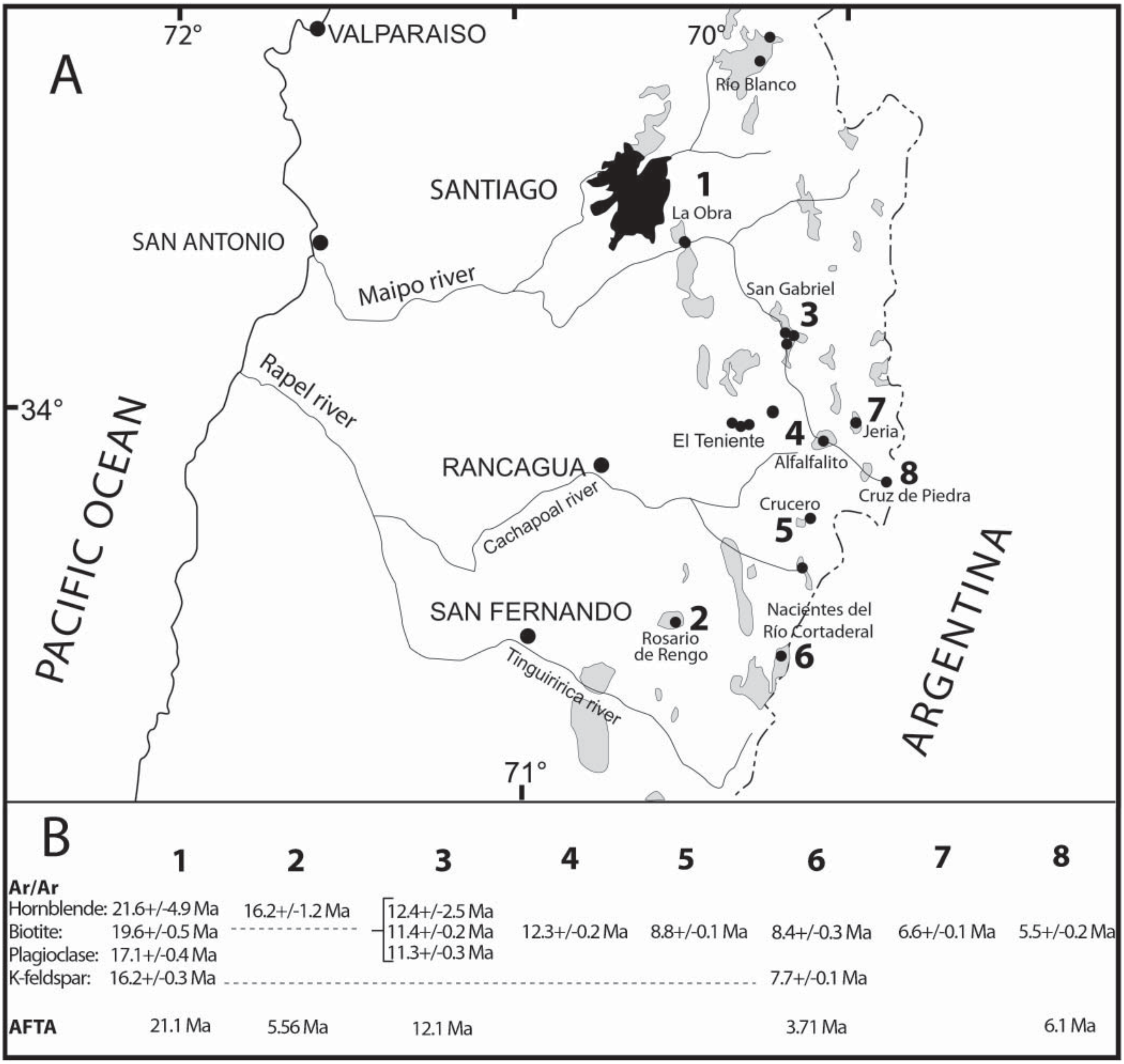


Fig. 3.63. Late Cenozoic plutons in the Principal Cordillera between 33ºS and 35ºS, based on Maksaev *et al.* (2003). (**A**) Distribution of plutons and outcrop localities (for numbers see lower part of figure). (**B**) Radioisotopic (Ar/Ar) and apatite fission track ages (AFTA) are given to show eastward younging of plutons and the young exhumation age of the Andes in this region.

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.* 2004*a*; Charrier *et al.* 2004, 2005*a*; Farías *et al.* 2005*a*; 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 Marquartdt *et al.* 2003). This time coincidence stresses the regional extent of the Andean uplift phase.

Evolution in southern Chile (Patagonia), south of 42ºS

The evolution of the Patagonian Cordillera, located south of *c.* 42ºS, occurred on the west side of the Patagonia Terrane (Ramos 1984, 1988*b*; Ramos *et al.* 1986) (that forms the eastern, non-orogenic side of Patagonia south of *c.* 39ºS) 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 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 generalized extensional tectonic event followed by a Middle to Late Jurassic silicic volcanic event, both affecting the entire Patagonian realm (i.e. not only the Andean domain, but also the foreland region); (2) the development south of 50ºS of an Early Cretaceous ocean-floored extensional basin; (3) Cenozoic subduction of the Chile Ridge under the southern tip of South America and the northward shift of the triple junction, which is now located at 47ºS. These three distinctions allow the region south of 42ºS to be treated separately from the rest of the Chilean Andes.

The southernmost Chilean territory (south of *c.* 50ºS) extends gradually towards the Andean foreland (Figs 3.1b & 3.64), providing the opportunity to examine more external features of the orogen than is possible further north, such as the thrust-fold belt and parts of the foreland basin, where there has been major drilling by the hydrocarbon industry. For this reason, some stratigraphic units in this region are known from both direct observation (Patagonian Cordillera, proximal thrust-fold belt and foreland) and drill cores (Springhill or Patagonian Platform). Commonly these units have been assigned different names. A great amount of geological information has been obtained from the southern Chilean Patagonia or Magallanes region, but much of it remains in company reports, and has only been partially cited in publications by staff members and graduate students financed by the Chilean petroleum company (SIPETROL-ENAP). For the purposes of this book we emphasize information made available in journals and theses.

In the Patagonian Andes three stages of tectonic evolution have been distinguished (Fig. 3.65): Middle to Late Jurassic regional extension, Early to mid-Cretaceous thermal subsidence, and Late Cretaceous to Present tectonic inversion, with foreland basin and thrust-fold belt development (Biddle *et al.* 1986; Harambour & Soffia 1988; Soffia & Harambour 1989; Skarmeta & Castelli 1997; Mella 2001).

First Stage: Middle to Late Jurassic. The oldest Mesozoic rocks known in this region correspond to the Late Triassic Potranca Formation exposed in the Chonos Archipelago (see Pre-Andean Cycle; Fig. 3.15). From then until Middle Jurassic times there is no lithological record for deducing the geological evolution at the beginning of the Andean cycle in this region. Considering that further north, during the pre-Andean cycle, the continental margin was passive until approximately middle Late Jurassic times, it is possible that in this region conditions were similar.

In Middle to Late Jurassic times a second Mesozoic regional extensional event, probably caused by a major diapiric mantle uplift that culminated with the opening of the Atlantic Ocean, affected 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 extensive felsic 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 & Rolleri 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). Volcanic activity produced thick deposits included under the general term of 'Complejo Porfirítico de la Patagonia' (Quensel 1913). These rocks have been assigned to the Ibañez 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), and to the Quemado Group or Complex in Argentina (Riccardi 1971) and Magallanes region in southernmost Chile (Katz 1963, 1964). The term Tobífera Series or Formation is the name given to equivalent deposits in the subsurface (Thomas 1949). These deposits covered Palaeozoic metamorphic rocks and, on the westernmost part of the continental margin, Late Triassic deposits of the Potranca Formation. The essentially rhyolitic volcanic activity formed deposits of variable thickness mainly consisting of ignimbrites and tuffs, with minor andesites and intercalations of tuffaceous sandstones and pelites. To the west, in the present-day Patagonian Archipelago, slices of granitic bodies that probably belong to this acidic magmatic event apparently have been affected by tectonic erosion and became involved in the subduction process (Seno Arcabuz Shear-zone; Fig. 3.15) (Hervé & Fanning 2003). In the present-day Patagonian Cordillera, these volcanic materials were deposited in deep marine waters, whereas towards the foreland deposition occurred essentially in subaerial environments on the Patagonian (Springhill) Platform (Bruhn *et al.* 1978; Allen 1982; Fuenzalida 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 & Rolleri 1980; Fuenzalida & Covacevich 1988; Covacevich *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 swathes strongly affected by east-vergent thrusts (Dalziel *et al.* 1974; Stern *et al.* 1976*a*; De Wit & Stern 1978; Calderón 2006). Basalt geochemistry indicates typical oceanic tholeiites to transitional-type basalts and associated differentiates (Stern *et al.* 1976*a*). 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 Porfirítico 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;

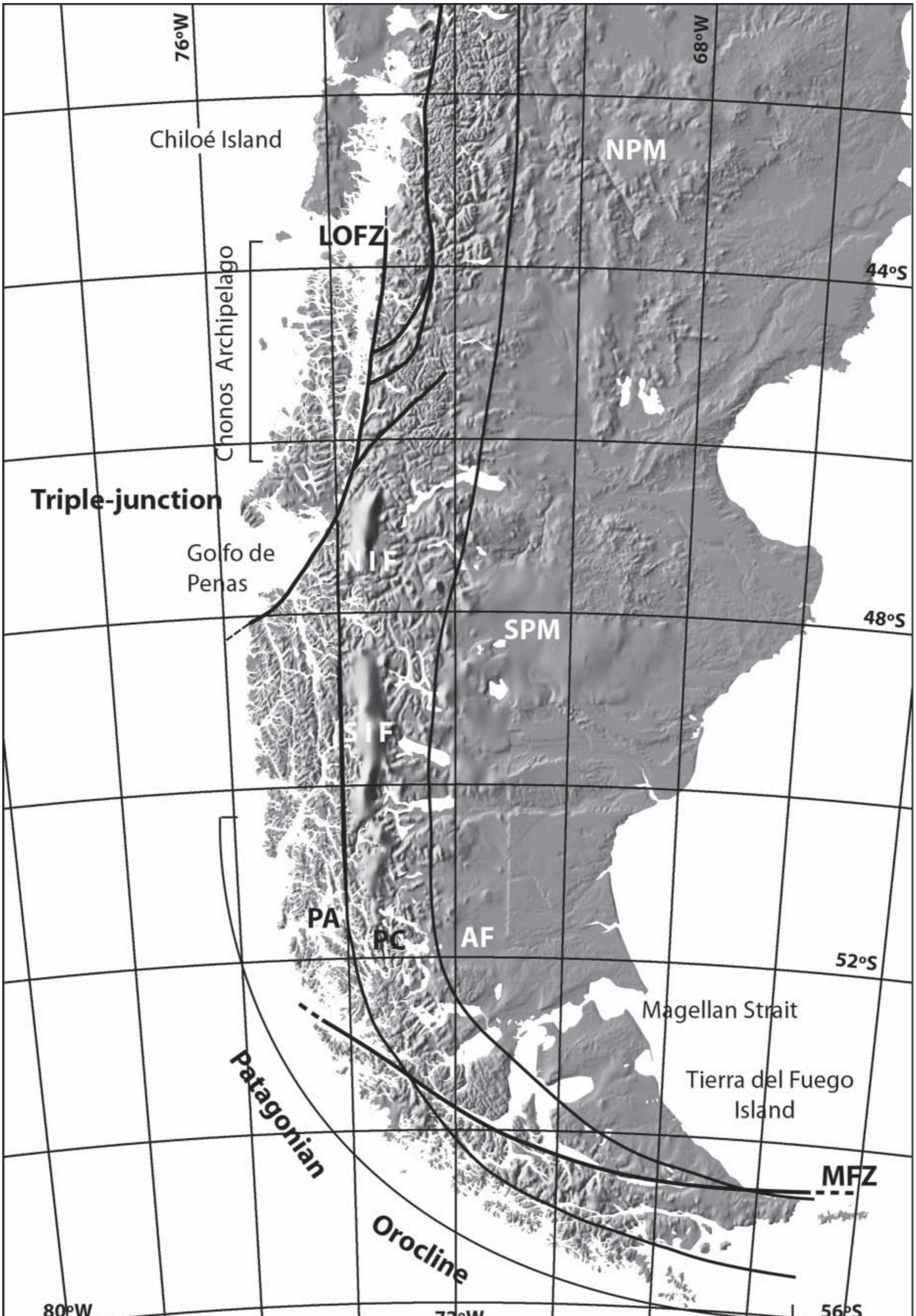


Fig. 3.64. Digital elevation model of the Chilean Andes south of 42°S with indication of the main geographic, tectonic and morphostructural features. Abbreviations: AF, Andean Foreland; LOFZ, Liquiñe–Ofqui Fault Zone; MFZ, Magallanes Fault Zone (main trace) ; NIF, northern ice-field; NPM, North-Patagonia Massif; PA, Patagonian Archipelago; PC, Patagonian Cordillera; SIF, southern ice-field; South-Patagonia Massif.

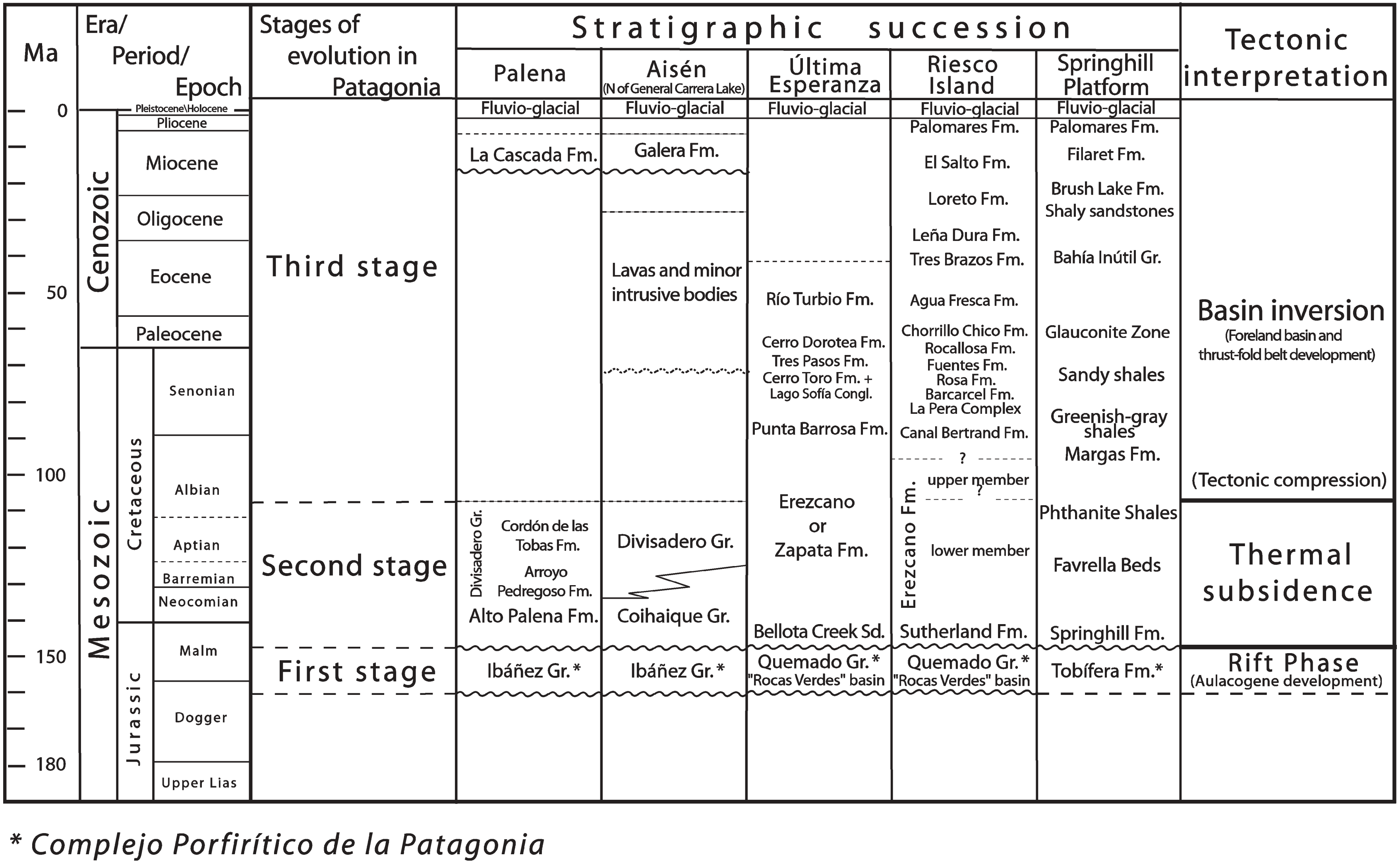


Fig. 3.65. Stages of geological evolution in Patagonia, stratigraphic successions exposed in surface in the Palena, Aisén and Magallanes regions (Última Esperanza and Riesco Island sections and Springhill Platform) and tectonic interpretation, based on several authors (see text).

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.* 50ºS and by oceanic crust south of that latitude), reaching from the Palena region at 43ºS 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** (43–44ºS), the backarc basin deposits consist of several marine formations of Tithonian to Hauterivian age (Fuenzalida 1968) which are included in the

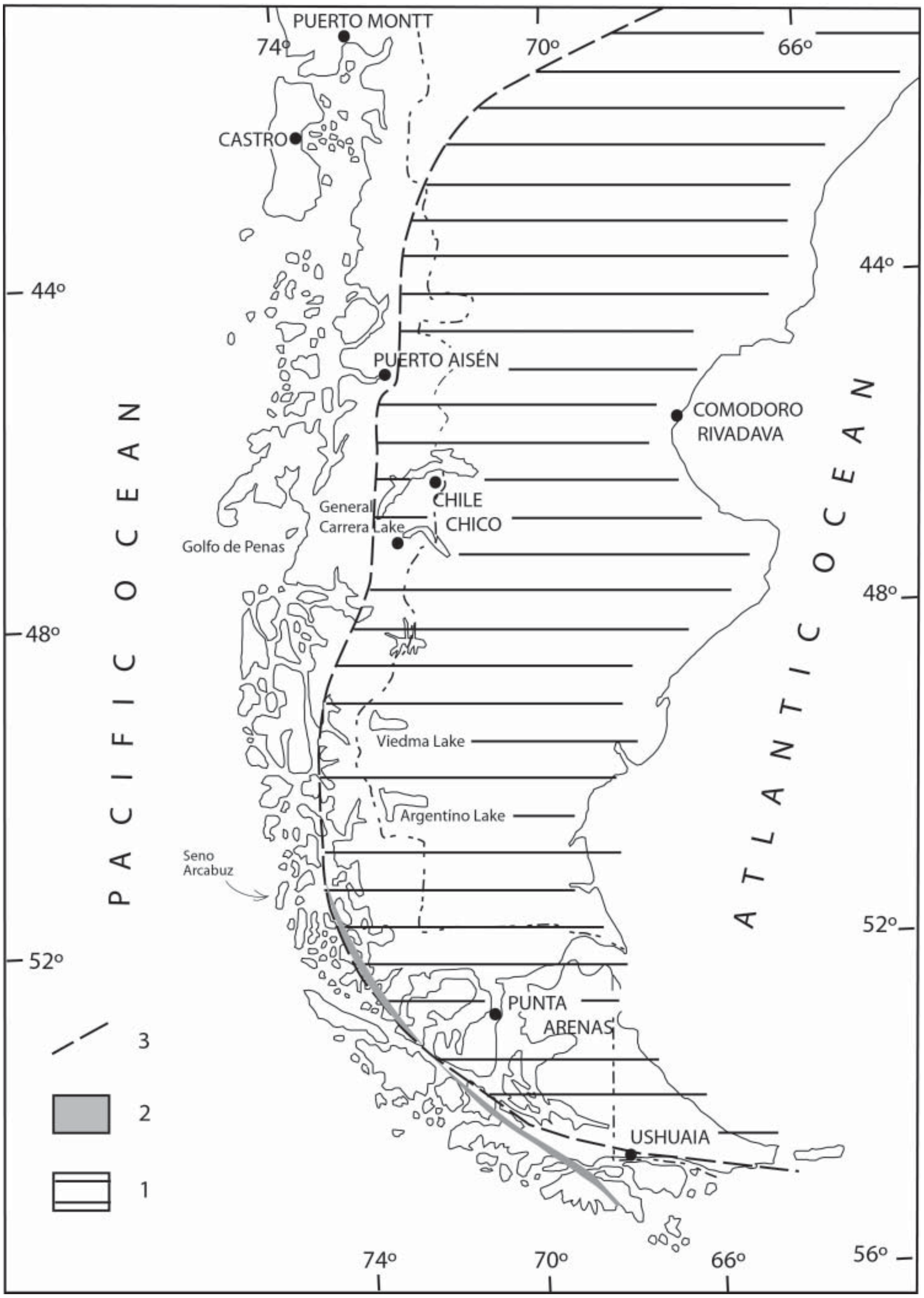


Fig. 3.66. Distribution of the deposits of the first stage of Patagonian evolution. Key: 1, Complejo Porfirítico de la Patagonia (Ibañez Group, Quemado Group, Tobífera Formation) or Chon Aike magmatic province (see Mpodozis & Kay 1990). 2, Ophiolitic suite of the 'Rocas Verdes Basin', south of 50ºS. 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).

Alto Palena Formation of Thiele *et al.* (1978). More recently, 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 (Haller & 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 La Cascada 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** (44–46ºS), the backarc Aisén Basin (=Río Mayo Embayment) contains the Coihaique 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 Cotidiano 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 Ibañez Group and consists of calcareous and volcaniclastic sediments deposited on the high-energy shorelines of active andesitic volcanoes formed in the eastern magmatic arc. The Katterfeld Formation conformably overlies the former and consists of thick black shales deposited under anoxic conditions during Valanginian to early Hauterivian times. Environmental conditions changed in middle to late Hauterivian times (*Favrella americana*) to open shelf on which the sandstones (some are tuffaceous) and mudstones of the Barremian Apeleg Formation were deposited (González-Bonorino & Suárez 1995; Bell & Suárez 1997). Marine sedimentation in the Aisén Basin ended in late? Aptian times, when a regional andesitic to dacitic volcanic event covered wide areas of northern Patagonia with thick deposits of lavas, tuffs and volcaniclastic deposits assigned to the <1000-m-thick Divisadero Group (late Aptian to Albian) (Niemeyer *et al.* 1984; De la Cruz *et al.* 2003). These subaerial deposits unconformably cover the Coihaique Group. Because of rather deep erosion that completely eliminated the Apeleg Formation they locally directly overlie the Katterfeld Formation (Bell & Suárez 1997). In Argentinian territory they lie directly on the Quemado Group (Riccardi & Rolleri 1980). The Divisadero Formation can be correlated northward with the Cordón de las Tobas Formation of the Divisadero Group, in the Palena region (Fig. 3.65).

South of General Carrera Lake at 46º30′S, Early Cretaceous deposits correspond to the >100-m-thick Cerro Colorado Formation (Suárez & De la Cruz 2000; Suárez *et al.* 2000) which consist of fossiliferous shallow marine clastic deposits. These correspond to the 'Primer nivel marino con Ostraea' of Charrier *et al.* (1978, 1979) and Niemeyer *et al.* (1984), and are conformably overlain by the subaerial Flamencos Tuffs consisting of a 140-m-thick succession of tuff, tuffites and ignimbrites. K–Ar ages on biotite in these deposits yielded ages of 128 ± 3 Ma, 125 ± 3 Ma and 123 ± 3 Ma (Barremian) (Suárez & De la Cruz 2000, 2001; Suárez *et al.* 2000), which permits correlation of the Cerro Colorado and the Flamencos Tuffs formations with the Apeleg Formation, or alternatively the Cerro Colorado with the Apeleg Formation and the Flamencos Tuffs with the Divisadero Group (probably basal) in the Coihaique region. Further south, north of Cochrane (for location see Fig. 3.68), De la Cruz *et al.* (2004) report the existence of marine deposits that they assign to the Coihaique Group, which are conformably overlain by the Flamencos Tuffs.

South of 50ºS, in the backarc basin in the **Magallanes region**, Late Jurassic to latest Early Cretaceous sedimentation formed a thick succession of marine deposits with rapid east–west facies variations that permit reconstruction of environmental conditions across the basin. Facies changes across and along the basin gave rise to a complex stratigraphic nomenclature for these sediments. Because of this we have selected two approximately east–west orientated classic sections that serve to illustrate basin evolution: Última Esperanza, and Riesco Island (see Fig. 3.67).

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 Calcareous Sandstones of the Bellota Creek; Cortés 1965), which have been correlated with the Springhill Formation in the Springhill Platform (Harambour & Soffia 1988; Soffia & Harambour 1989). These transgressive deposits are overlain in turn by the Early Cretaceous Erezcano Formation (Cecioni 1955*a*, 1956, 1957*a*, 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 silicic volcanic deposits in the Patagonian Cordillera (Seno Rodríguez Formation; Cecioni 1955*a*, 1958) comprise the Sutherland and Erezcano 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 1955*a*; Riccardi 1976, 1977; Riccardi & Rolleri 1980) in the Springhill Platform. According to Cecioni (1955*a*, 1957*a*, 1958), the Erezcano 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 (Erezcano) member to the greywacke-bearing part of Cecioni's (1956) definition. The lower member is correlated to the east with the Favrella 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. Contractional deformation, uplift and erosion began in the arc and internal backarc, causing closure (obduction) of the ocean floor south

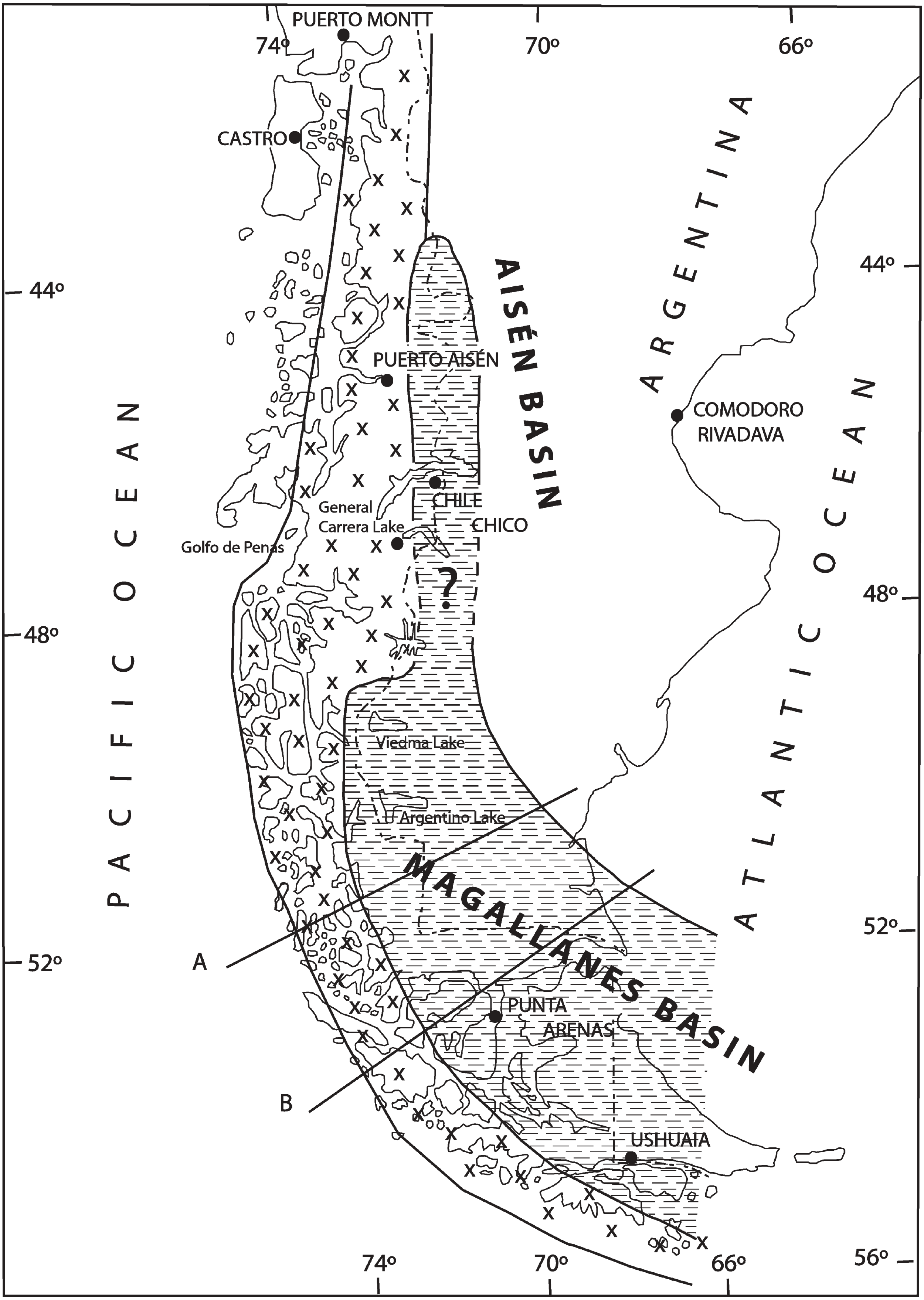


Fig. 3.67. Second stage of Andean evolution in Patagonia. Distribution of the Aisén and Magallanes basins and the Patagonian Batholith. Connection between the two basins is uncertain because of lack of outcrops. Última Esperanza section (A) Riesco Island section (B) correspond to the regions described in text. Figure is based on González *et al.* (1965), Bell & Suárez (1997) and Sernageomin (2003).

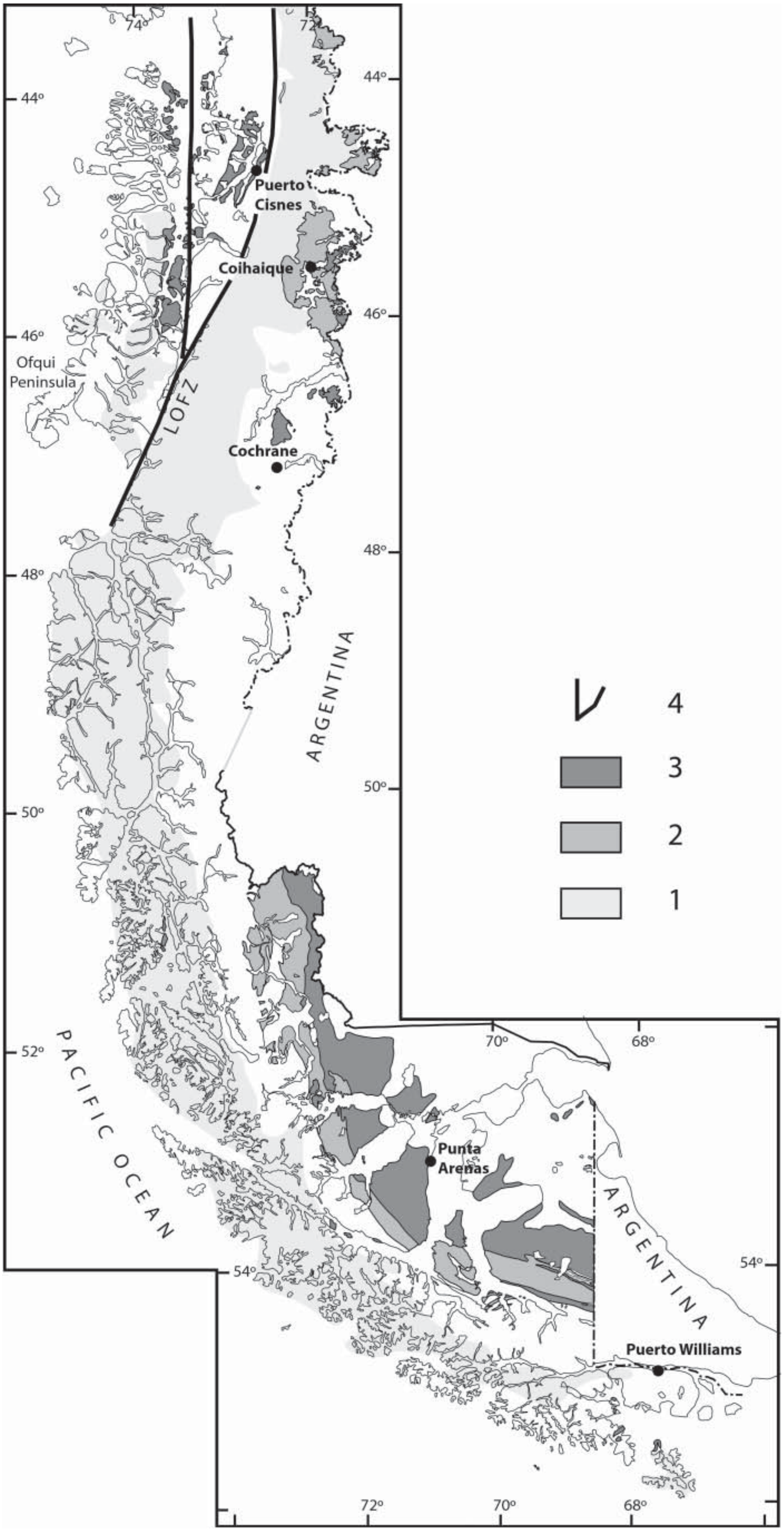


Fig. 3.68. Distribution of the Patagonian Batholith and relative position of the deposits pertaining to the first and second stages of Andean evolution in the Aisén and Magallanes backarc basins (see Figs 3.66 & 3.67) and in pull-apart basins associated with the Liquiñe–Ofqui Fault Zone (LOFZ) (see Figs 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 50°S (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 50°S 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 (39–43°S) 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 (47°S) 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

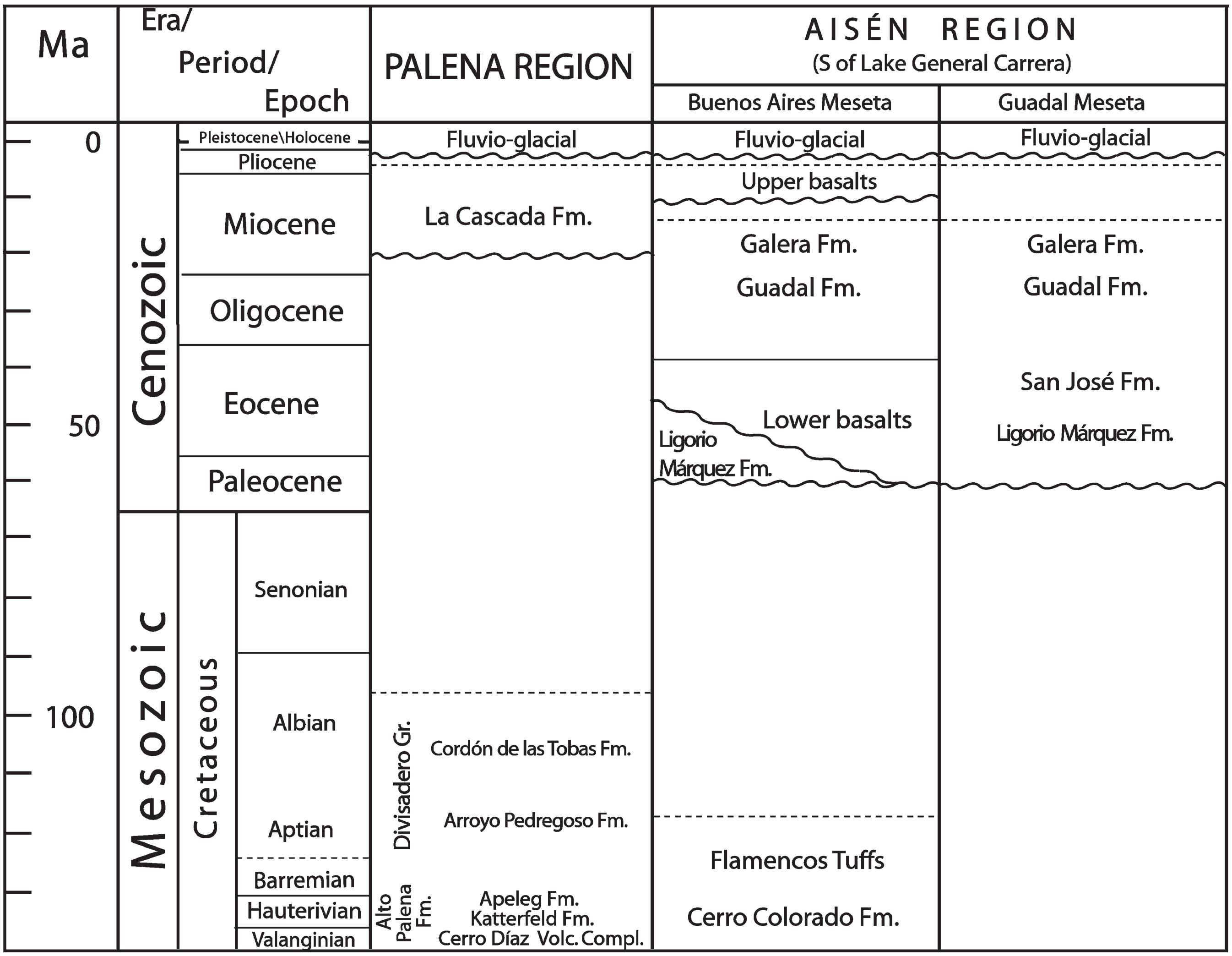


Fig. 3.69. Stratigraphic succession and correlation of Mesozoic and Cenozoic deposits in the Palena and Aisén (south of Lake General Carrera) regions. For the Aisén region, stratigraphic succession is given for the eastern region next to the borderline (Buenos Aires Meseta) and for the western region (Guadal Meseta) (compare with stratigraphy north of Lake General Carrera in Fig. 3.65).

Cretaceous Cerro Colorado Formation and the Flamencos Tuffs (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 2003*a*) 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 Busteros & 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 (Niemeyer *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.* 2002*b*).

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 2003*a*). Recent magmatic arc activity in this region is, however, hardly developed, with just the Macá (45°S) and Hudson (46ºS) volcanoes testifying to the subduction of the southernmost Nazca Plate underneath the South American continent. In the region between 46ºS and 49ºS, south of Aisén, there is a complete gap in recent volcanic activity.

In the **Ultima Esperanza section of the Magallanes region**, the deposits that record the initiation of basin inversion correspond to the deep-water turbiditic lobes of the Cenomanian Punta Barrosa Formation (Cecioni 1956, 1958; Katz 1963, 1964; Cortés 1965; Scott 1966; Riccardi & Rolleri 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, 1957*a*, 1958; Katz 1963, 1964; Cortés 1965; Scott 1966). According to Riccardi & Rolleri (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 Sofía Conglomerates or Lago Sofía Member) (Cecioni 1956, 1957*a*, 1958; Katz 1963, 1964; Cortés 1965; Scott 1966; Riccardi & Rolleri 1980; Dott *et al.* 1976, 1982; Winn & Dott 1979; Shultz 2002; Beaubouef 2004). This formation corresponds to the 'Chondrites Flysch' of Cecioni (1957*a*) and extends eastwards into western Argentinian territory (Riccardi & Rolleri 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 Sofía Conglomerates' consist of clast- and matrix-supported conglomerates, and sandstones and pebbly sandstones, corresponding to turbidity currents and debris flows (Winn & Dott 1979) forming stacked levee–channel complexes (Beaubouef 2004).

The Cerro Toro Formation is overlain by the upward-shallowing clastic succession of the late Campanian–Maastrichtian Tres Pasos Formation (Katz 1963, 1964; Scott 1960; Cortés 1965; Natland *et al.* 1974) comprising 1000–2000 m of mudstone and turbiditic sandstones, and interpreted as the deposits of a delta-fed submarine slope depositional apron (Shultz 2002). These deposits have been correlated into eastern Argentina with the Cerro Cazador Formation (Riccardi & Rolleri 1980; Malumián & Caramés 1997).

The Tres Pasos Formation is conformably overlain by the Maastrichtian–Palaeocene Cerro Dorotea Formation (Hoffstetter *et al.* 1957; Katz 1963, 1964; González *et al.* 1965), defined in the Sierra Dorotea close to the international borderline (see Hünicken 1955; Cecioni 1957*b*; Riccardi & Rolleri 1980). This 300–500-m-thick homoclinal, east-dipping and upward-coarsening clastic unit contains the first continental carbonaceous deposits since Late Jurassic times. It is conformably overlain by the continental Río Turbio Formation (Hünicken 1955), which consists of a 600–650-m-thick succession of bluish-green coarse sandstones with cross-bedding, conglomerates and thin shale intercalations deposited in marine and continental environments, as well as with tuff layers and coal seams in its upper part. Coal intercalations are intensively exploited in Argentina (Vivallo *et al.* 1999). The Río Turbio Formation, which is fully exposed in Argentinian territory, is covered by younger formations also deposited in the eastward-migrating foreland basin. According to Malumián & Caramés (1997) the contact between the Cerro Dorotea and the Río Turbio formations coincides with the Palaeocene–Eocene boundary.

In the **Riesco Island–Seno Skyring** section the evolution is slightly different, although it is possible to recognize a similar basin evolution (Fig. 3.65). The Aptian–Albian (probably somewhat younger) upper member of the Erezcano Formation (Prieto 1993; Mella 2001) reflects little change from the lower member beneath. It consists of a 300–500-m-thick succession of bioturbated siltstones and white tuffaceous sandstones of probable turbiditic character. These tuffs and volcanic components in the sandstones suggest some volcanic activity at that time (incipient volcanic arc). Subsurface equivalents of the upper member of the Erezcano Formation in the Springhill Platform are the 'Lutitas con Ftanitas' (Phthanite Shales). This unit is conformably overlain by the Aptian–Albian eastward-wedging Canal Bertrand Formation (subsurface equivalent is the Margas Fm; Mella 2001) which interfingers upward with the La Pera Complex (see Mella 2001) and consists of an alternation of shales and turbidites in which the sandstone/shale ratio gradually increases upwards. These deposits have been interpreted as turbiditic lobes that record the inception of tectonism in the western regions. Thus the Canal Bertrand Formation is considered to be broadly correlative with the Punta Barrosa Formation, although the turbiditic sedimentation occurred earlier in this region than in Última Esperanza. The La Pera Complex is a 400–800-m-thick volcanic and volcaniclastic wedge that interfingers with shallow marine deposits of Turonian age.

Gradual shallowing of the basin and abundant sediment supply from the uplifted regions is recorded by overlying Late Cretaceous deposits (Barcarcel, Rosa, Fuentes and Rocallosa formations; Thomas 1949; Cecioni 1955*c*, 1956; Hoffstetter *et al.* 1957; Charrier & Lahsen 1968, 1969; Lahsen & Charrier 1972; Hünicken *et al.* 1975, 1977; Mella 2001). Shallow platform sediments of the Barcarcel (500 m thick) and the Fuentes (1220 m thick) formations, consisting of shales, siltstones and glauconitic sandstone intercalations with frequent ripple marks and cross-bedding, alternate with thick, hard conglomeratic 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 'Sandy 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 Glauconitica' in the Springhill Platform.

The middle Eocene Tres Brazos and Leña Dura formations conformably overlie the Agua Fresca Formation (see Mella 2001). These deposits probably correspond to adjacent eastward prograding deltaic lobes consisting of fine to medium glauconitic sandstones, calcareous sandstones, siltstones, conglomeratic lenses, and thin coal seams. The upper part of the overlying Late Eocene?–Oligocene coal-bearing Loreto Formation (see Vivallo *et al.* 1999) and the next overlying units (El Salto and Palomares formations) correspond to continental deposits. This infilling of the foreland basin culminated with deposition of the Miocene El Salto Formation. The subsurface equivalent of the Tres Brazos and Leña Dura formations is the Bahia Inútil Group, and the equivalent of the Loreto and El Salto formations are the Brush Lake and Filaret formations (Gonzalez *et al.* 1965; Mella 2001).

In latest Oligocene and Late Miocene times plutonic intrusions dated at 12 ± 2 Ma, 13.1 ± 1 Ma (Halpern 1973) and 28 Ma (K–Ar) (Skarmeta & Castelli 1997) were emplaced into the basin deposits (Cerro Toro Formation) to form the Torres del Paine laccolith. Recent magmatic arc activity related to the subduction of the Pacific–Antarctic Plate underneath the southern part of the South American continent in this region is almost non-existent. The few volcanic edifices that are developed are named, from north to south: Lautaro, at 49ºS; Viedma, at 49º50′S, Aguilera, at 50ºS; Reclus–ex Mano del Diablo at 51ºS (Harambour 1988); Monte Burney, at 52º15′S.

The triple-junction in the Chilean Patagonian Andes
The Chilean triple-junction at 47ºS deserves special mention because it is linked to an asthenospheric window underneath southern Patagonia, the Liquiñe–Ofqui Fault Zone (LOFZ), the distinctive geological evolution of the Tres Montes Peninsula and apparently also to a gap in the volcanic arc. The triple-junction is located at the western extremity of the Taitao Peninsula and corresponds to the intersection of the continental margin (Patagonian Andes) with the active spreading Chile Ridge that separates the Nazca Plate from the Pacific–Antactic Plate (Herron *et al.* 1981; Cande *et al.* 1987) (Fig. 3.70). At this point sediments are channelled north and south of the junction along the trench axis. Regional plate reconstructions indicate that the Chile Ridge collided with the southern tip of South America *c.* 14 million years ago and that since then the triple-junction has shifted northward to its present position (Cande & Leslie 1986; Mpodozis *et al.* 1985; Cande *et al.* 1987). This situation implies the existence below southern Patagonia of subducted segments of the Chile Ridge and consequently of a segmented asthenospheric window (Herron *et al.* 1981; Forsythe *et al.* 1986; Cande *et al.* 1987; Nelson *et al.* 1994). This window would have allowed uprise of great volumes of primitive alkali basalts to the SE of the present triple-junction in late Miocene times (Skewes & Stern 1979; Forsythe *et al.* 1986; Espinoza 2003; Espinoza *et al.* 2003; 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.* 1979*a*; 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 Penas, a possible pull-apart basin linked to the movement of the LOFZ (Forsythe & Nelson 1985; Forsythe *et al.* 1985, 1986; Mpodozis *et al.* 1985; Bourgois *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
Apatite 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; Harambour & Soffia 1988; Soffia & Harambour 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 42°S; this suggests that the general tectonic environment here was

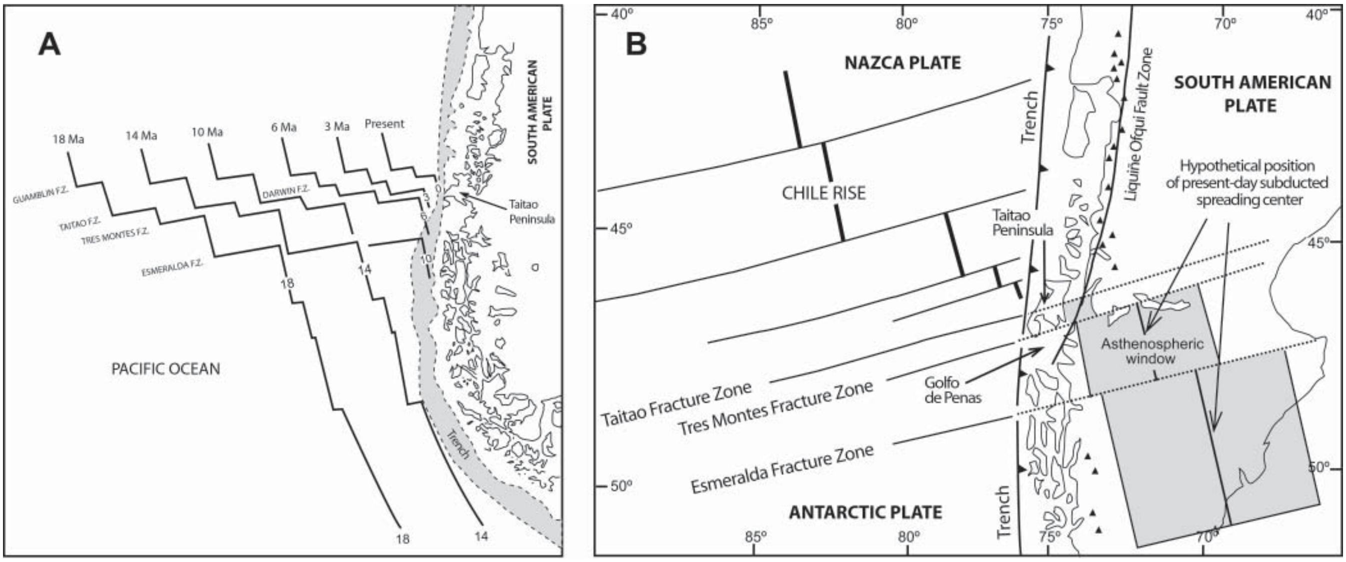


Fig. 3.70. The triple-junction in southern Chile (see Figs 3.2, 3.57 & 3.64). (**A**) Location of the Chile Ridge relative to South America in the last 18 Ma. Eastward migration of the active Chile Ridge during the last 14 Ma (since collision with the continent) caused a gradual northward shift of the triple-junction, which is presently located at *c.* 47ºS at the Taitao Peninsula. Collision of the ridge with the continent occurred 13–14 million years ago (modified from Cande *et al.* 1987). (**B**) Present-day location of the Chile Ridge, the Antarctic and South American plate boundary and the triple-junction between these plates. Interpretative location of the ridge segments below South America and asthenospheric window is indicated (modified from Murdie *et al.* 1993).

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 42ºS (Subhercynian or Peruvian phase), although the palaeogeography south and north of 42ºS 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 *c.* 50ºS 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 deposition 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 Última 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 (latest 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.

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