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Deformation produced by the subduction of a Palaeozoic turbidite sequence in northern Chile

C. M. Bell

SUMMARY: A deep-sea basin-plain turbidite succession of Ordovician to Devonian age is exposed between 25°30′ and 27°S in the coastal region of northern Chile. In the S these turbidites form a large-scale tectonic mélange. Alkali basalt pillow lavas and tuffs are interstratified with the sediments to the S of 26°20′S. The rocks were metamorphosed to the greenschist facies and subjected to two phases of deformation subsequent to the formation of the mélange. The first phase is characterized by large-scale, tight chevron folds and an axial planar cleavage. The folds of the second phase are smaller and more upright and accompanied by a crenulation cleavage. The fold axes of both phases trend between W–E and NW–SE, oblique to the N–S trending Mesozoic and Cenozoic Andean structures. Folds are predominantly asymmetric and overturned towards the S and SW, suggesting an origin by underthrusting towards the N and NE. The deformed succession is interpreted as a subduction complex produced by oblique underthrusting of an oceanic plate towards the N and NE during late Devonian to Carboniferous times.

Most rocks in northern Chile are of Mesozoic or Cenozoic age. They consist of volcanic, sedimentary and plutonic rocks, belonging to the Andean magmatic complex, which were produced by the eastward subduction of an oceanic plate beneath the South American continental margin (James 1971). The result of this long-lived subduction process is that the evidence for pre-Mesozoic geological events in northern Chile is sparse and fragmentary. A notable exception, however, is seen in the virtually continuous strip of Palaeozoic strata exposed between 25°30′ and 27°S which extends for some 150 km from N to S and between 10 and 30 km inland (Fig. 1) and forms the subject of the present study.

The rocks in this area comprise deformed turbidites and basic volcanic rocks and are known as the 'metasedimentary basement' by Chilean geologists. Most of the previous work in this area has been by members of the Servicio Nacional de Geologiá y Mineriá (previously the Instituto de Investigaciones Geologicas). An unpublished map at a scale of 1:100,000 covering the N of the area was produced by Ulriksen (1979). Mercado (1978, 1980) and Naranjo (1978) have produced maps of the southern section at a similar scale.

The extremely dry climate of the Atacama desert results in excellent rock exposures, particularly along the rocky coast and in the deeply incised and steep-sided dry river valleys or quebradas.

Regional geology

Scattered occurrences of Palaeozoic rocks are found throughout Chile but detailed correlations have yet to be made, primarily due to the shortage of reliable age determinations. Most of the strata are unfossiliferous marine sediments which have been deformed and metamorphosed, some to a high grade (Gonzalez-Bonorino & Aguirre 1970; Aguirre *et al.* 1972; Hervé 1977). These rocks are most abundant in southern Chile and become progressively less common towards the N. No definitely pre-Palaeozoic strata have yet been identified in Chile.

The sedimentary and volcanic rocks described here have been metamorphosed to the greenschist facies (Miller 1970; Aguirre *et al.* 1972). They were intruded by Lower Permian monzogranites (267 ± 8 Ma: Zentilli 1974) and have been correlated on lithological grounds with the El Toco Formation of possible Devonian or Carboniferous age (Harrington 1961; Garcia 1967). The identification of nine genera (*Gordia* spp., *Laevicyclus* sp., *Lophoctenium* spp., *Nereites* sp., *Paleodictyon* sp., *Planolites* spp., *Rusophycus* sp., *Scalarituba* sp., *Scolicia* spp.) of trace fossils collected during the present investigation strongly suggests an Ordovician to Devonian age (fossil identification by V. Covacevich).

Relatively undeformed andesitic volcanic and volcaniclastic sedimentary rocks of Mesozoic age overlie the Palaeozoic basement with a major unconformity in many parts of Chile. Intruded into these sedimentary and volcanic rocks are abundant plutons ranging in age from late Palaeozoic to Upper Cretaceous. The most abundant intrusive rocks in the Palaeozoic strata in the Chañaral area (Fig. 1) are Palaeozoic tonalites and granites (Naranjo 1978; Mercado 1978, 1980). There is evidence of an eastward migration of magmatic foci during the Mesozoic (Naranjo 1978).

Palaeozoic stratigraphy

The Palaeozoic strata are readily subdivided into two mappable units (Fig. 1). In the N are turbidites of the

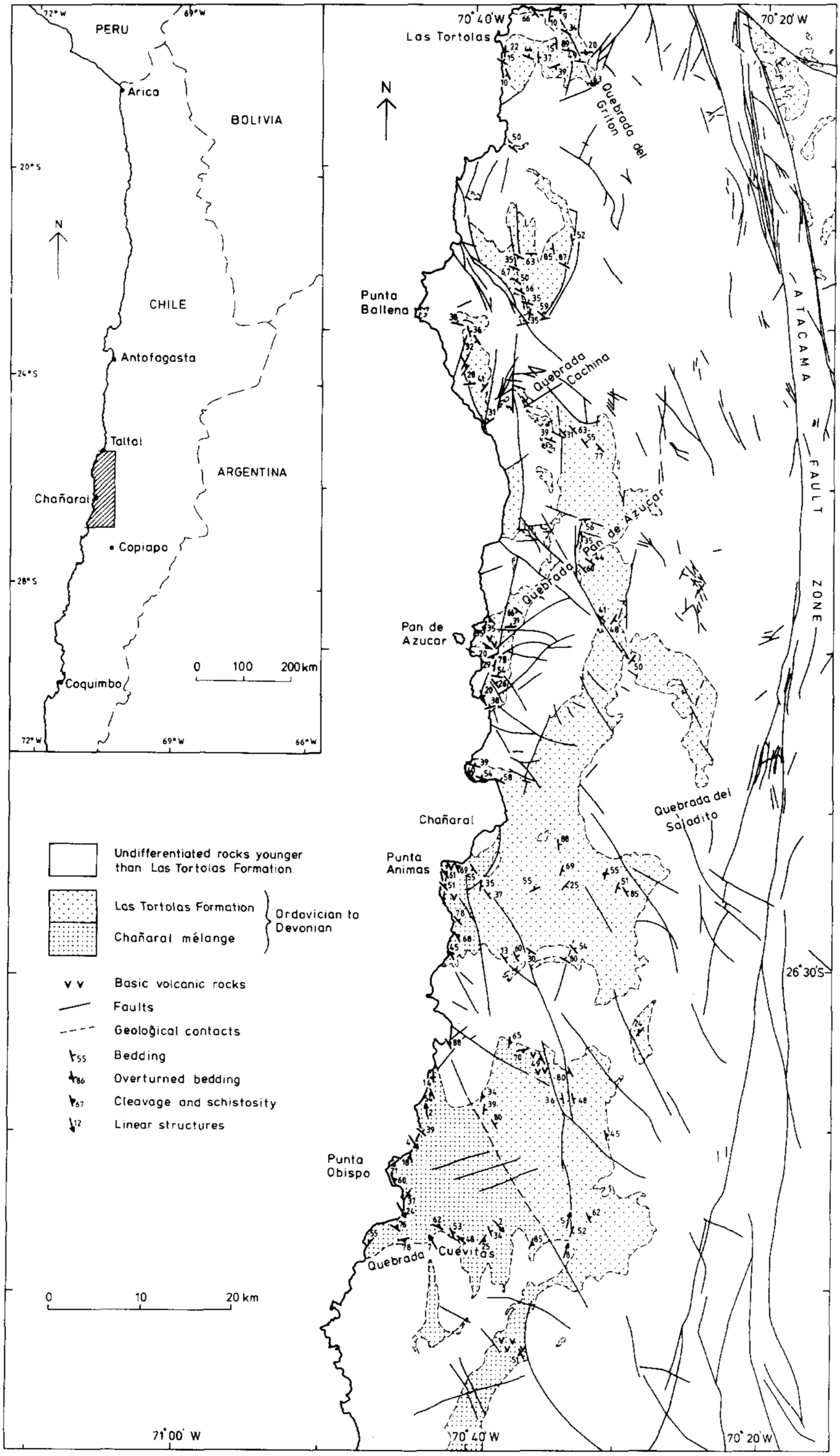

FIG. 1. Locality and geological map showing the distribution of Palaeozoic sedimentary and volcanic rocks in the Chañaral area between 25°30′ and 27°S.

Las Tortolas Formation (informal name given by Ulriksen 1979) and in the S, the Chañaral mélange (informal name). Metamorphosed alkali basalt pillow lavas and tuffs are interstratified with the sediments at Pan de Azucar and to the S of Chañaral. The mélange is the deformed equivalent of the Las Tortolas Formation.

Las Tortolas Formation

This formation is a thick and monotonous clastic sequence of interbedded sandstones and mudstones with rare limestones and cherts. Sedimentary structures and rock types indicate a deep-sea basin-plain turbidite association (Mutti & Ricci-Lucchi 1978; Bell 1982). The thickness of the formation is not known due to the intense deformation.

Sandstones of the Las Tortolas Formation are generally fine- to very fine-grained. Rare medium-grained sands with scattered granules and pebbles occur only at the base of large-scale graded units. The immature sandstones can be classified as feldspathic greywackes, but as the matrix is predominantly secondary in origin (Moore 1979) the rocks are more correctly classified as lithic and feldspathic arenites. The lithic clasts include andesite, volcanic glass and low-grade metasediment together with rare diorite and granodiorite. The tectonically corrected palaeocurrent direction indicated by flute marks (28 observations) is predominantly towards the SE (Bell 1982).

Rare limestone horizons up to 4 m thick are interstratified with the clastic turbidites and appear to be more common towards the S. Most of the limestone is massive and recrystallized but in places it consists of tightly-packed, broken or bent calcitic plates about 1 mm thick and up to 10 mm long. A fibrous structure of long filaments divided into cells trends perpendicular to the length of the plates. These fossils probably represent calcareous red algae such as the crustose Corallinaceae (R. Riding, pers. comm.).

Within the limestones are very small spines resembling the spicules of glass sponges of the class Hexactinellida (Bell 1982). The limestone horizons were probably deposited by turbidity currents carrying calcareous debris possibly derived from Devonian-Carboniferous algal beds on fossil tidal flats some 150 km to the E (Sepúlveda & Naranjo 1982). Associated with the limestones are massive and apparently unfossiliferous cherts (up to 0.6 m thick) of possible pelagic origin.

Chañaral mélange

The Chañaral mélange is a sequence of chaotically deformed distal turbidites enclosing isolated blocks of pillow lava, tuff and limestone. Exposures in the coastal area S of Chañaral extend at least 70 km from N to S and up to 30 km from W to E. No mélange has previously been described from Chile, a fact which led Hsü (1974) to suggest a landward migration of the Mesozoic subduction zone in the central Andes.

Rock types of the Chañaral mélange are identical to those of the Las Tortolas Formation, the only difference being the extreme disassociation and mixing of the strata. Most contacts between the mélange and the Las Tortolas Formation are apparently faulted. However, E of Punta Animas the sediments of the Las Tortolas Formation become progressively more deformed over a distance of about 2 km to form the Chañaral mélange. In this area strips of less deformed sediment similar to those found in the Franciscan complex (Hsü 1974; Jones *et al.* 1978) and other mélange terrains (G. F. Moore & Karig 1980) are interlayered with the intensely deformed mélange.

The mélange is typically a breccia of sandstone blocks varying in diameter from millimetres up to many metres, in an essentially pelitic matrix. The largest single block observed was of limestone and approximately 150 m in diameter. The blocks are extremely irregular in shape but some are well-rounded and commonly pillow- or spindle-shaped. Most blocks have well-defined boundaries against the matrix but others have irregular boundaries which grade into the matrix. Few blocks retain any traces of the original bedding and thin-section studies indicate that many are themselves composed of a sandy mélange. This evidence of a complex history of formation is similar to that described from the Dunnage mélange (Hibbard & Williams 1979). The mélange ranges in composition from interlocking sandstone blocks with little matrix to massive mudstones enclosing a few whisps of sand. No truly exotic clasts or high-grade metamorphic rock types such as those of the Franciscan mélange (Hsü 1968) were recorded.

Certain structures within the mélange are indicative of soft-sediment deformation. For example, mud whisps are commonly observed in sandstone blocks and some sandstone blocks are more pelitic towards their rims. Angular fractures are rare, suggesting that much of the deformation was essentially ductile. The mélange was probably produced by the mechanical breakdown and mixing of both unconsolidated and partly lithified (coherent but plastic) sediments. There is, however, no evidence of subaqueous sorting or sedimentary contacts within the mélange to suggest an olistostromal origin. Similarly the great extent and uniformity of the mélange, together with the deep-sea basin-plain depositional environment are strongly suggestive of a tectonic rather than an olistostromal origin. However, as with other similar mélanges (Jones *et al.* 1978), the distinction is impossible to make with certainty. The presence of soft sediment deformation structures does not preclude a tectonic origin, as the subduction of incompletely lithified sediment would produce structures indistinguishable from those of olistostromes.

In most places the matrix of the mélange has an irregular tectonic fabric similar to that seen in the pervasively sheared matrix of the Franciscan and other mélanges (Hsü 1968). This fabric is generally confined to the matrix, but in some zones of intense deformation the blocks are themselves strongly elongated (up to 10 times). Thin-section study of both blocks and matrix indicates a complex deformational history. Early cataclastic deformation was followed by the development of a schistosity, itself deformed by later crenulation cleavage. The fabric of the matrix is highly irregular: cleavage planes intersect in a diamond pattern and are commonly deflected around more resistant blocks. Much of the fabric was clearly produced after the development of the mélange but it has proved impossible to determine whether the formation of the mélange was accompanied by the development of a pervasively sheared matrix. Similar doubts as to the time of development and the significance of the fabric in the Franciscan mélange have been expressed by Page (1978) and Cowan (1978).

The dominance of extensional structures such as boudins indicates that the mélange was produced by lengthening or shearing along bedding. Folds are very rare and no regular pattern of folding was determined. A similar paucity of small-scale folds is a characteristic feature of many mélanges (G. F. Moore & Karig 1980; Cowan 1974; Hsü 1974) whereas others exhibit a distinctive pattern of early folds (Moore & Wheeler 1978).

Volcanic rocks

Basic volcanic rocks are associated with the Las Tortolas Formation at Pan de Azucar and they become more common towards the S in the Chañaral mélange (Fig. 1). Contacts are faulted at Pan de Azucar but in the mélange the volcanic rocks and the sediments are apparently interstratified. Hyaloclastite breccias and amygdaloidal pillow lavas are indicative of submarine eruption. In thin section the lavas are seen to be of greenschist facies and consist of radiating masses of actinolite together with albite, zoisite, epidote, calcite and iron oxide. Low Zr/Nb ratios and high Sr, Ni and Cr (Bell 1982) indicate that these rocks resemble alkali basalts rather than normally depleted mid-ocean ridge basalts (chemical analyses by Dr A. D. Saunders). These volcanic rocks are the product of submarine eruptions during the accumulation of the sediments. Their geochemical characteristics suggest an origin related to a mid-plate oceanic island.

Environment of deposition and deformation

The sedimentary structures and lithology of the distal turbidites indicate a deep-sea basin-plain depositional environment. The sandstone provenance of intermediate igneous rocks and metasediments suggests a source area similar to the present-day Andean cordillera. These factors, together with the presence of a large-scale tectonic mélange, indicate that the rocks are part of a subduction complex produced by underthrusting of an oceanic plate.

Tectonic deformation

No detailed account of the deformation of the Palaeozoic metasedimentary and volcanic rocks of this area has been published previously. In the southern part of the area Naranjo (1978) and Mercado (1978, 1980) have both described N–S trending folds with axial planes dipping predominantly towards the E. Miller (1970) recorded NNW–SSE trending minor folds at Punta Animas, S of Chañaral. Farther N, Ulriksen (1979) and Miller (1970) also identified NW–SE trending folds. Naranjo (1978) described isoclinal folding and multiple deformation to the E of Pan de Azucar.

The recognition of large-scale tectonic structures in the Palaeozoic strata has been hindered by the paucity of way-up indicators and marker horizons. Bedding planes are recognizable in most exposures but most sedimentary structures have been destroyed by the metamorphism and tectonic deformation.

Two phases of deformation affected the Palaeozoic strata prior to the intrusion of Lower Permian monzogranites. These phases are readily distinguished in the field by their distinctive styles. The fold hinges of both phases are near parallel and vary from an E–W trend in the N to a NW–SE trend in the S (Fig. 2). Folds are generally overturned towards the S and SW.

First phase of deformation

The first phase of deformation produced large-scale, tight asymmetrical chevron folds accompanied by a planar fabric. Few folds were observed in the field due to their relatively large-scale, near-parallel limbs, and the masking effects of later deformation. The interlimb angle varies from 3° to 35° with an average of approximately 20°. Wavelengths range from 10 to 200 m and average about 100 m. The folds are straight-limbed, often with barely apparent hinge zones. Fold axes are horizontal to gently plunging (Fig. 2) and the axial surfaces are moderately inclined with dips ranging from 25° to 75° and averaging 52°. In the N the fold axes trend predominantly E–W with axial surfaces dipping N (Fig. 2) but farther S the fold axes swing towards the NW–SE with axial surfaces dipping NE. The folds are thus predominantly overturned towards the S and SW over the whole area. This distinctive style of folding probably originated by bedding-parallel shear or slip (Ramsay 1974).

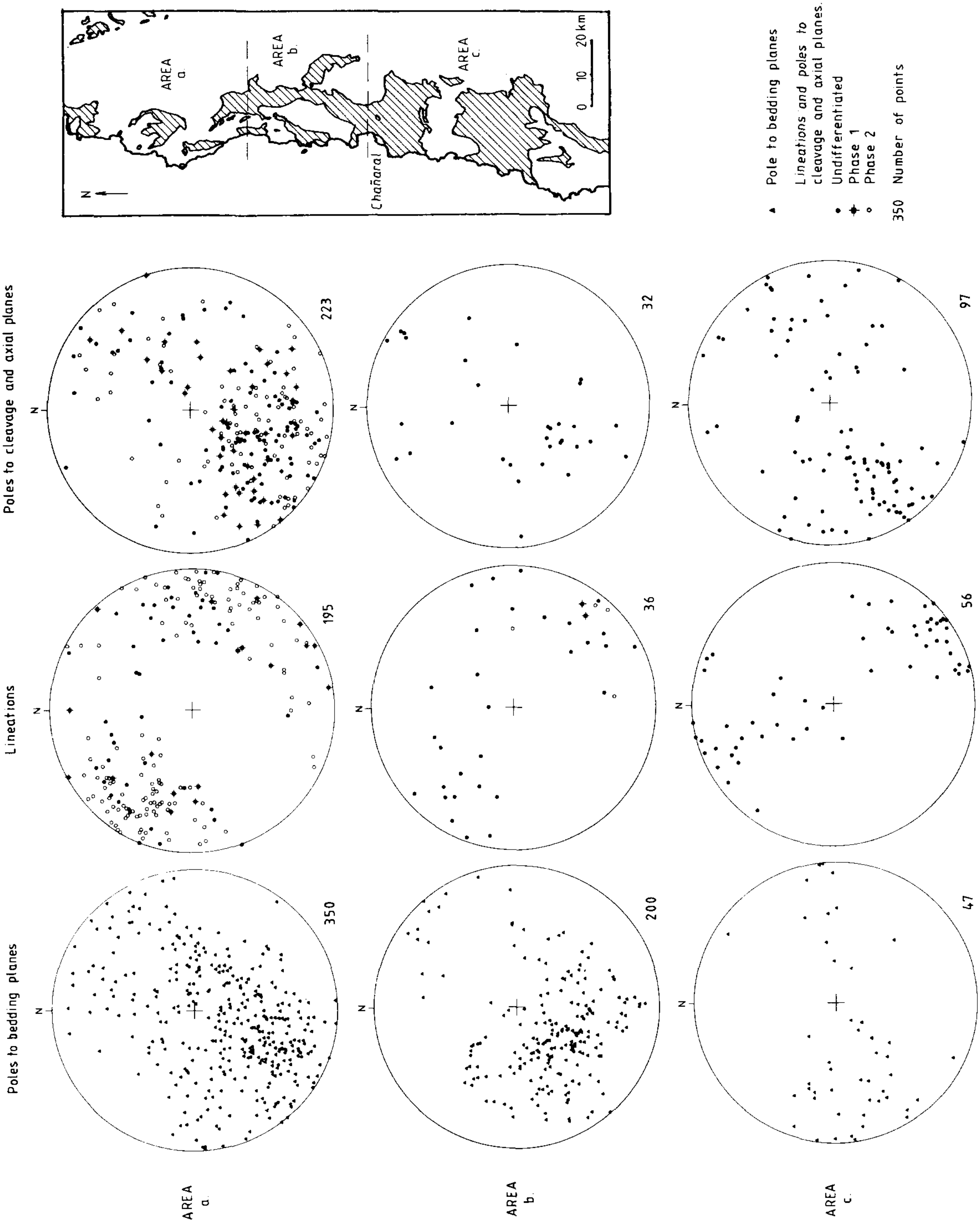

FIG. 2. Attitudes of poles to bedding planes, lineations (fold axes and bedding/cleavage intersections), poles to cleavage/schistosity and axial planes of minor folds in three sub-areas of the Chañaral area.

A prominent and pervasive axial planar foliation is particularly obvious in the field. This foliation can be readily recognized by its low angle to bedding planes, but the orientation varies due to the effects of subsequent deformation. Stereographic plots (Fig. 2) indicate that in the N the foliation dips moderately towards the NNE whereas farther S it dips ENE. The foliation is well developed in pelitic horizons where it has produced a schistosity parallel or sub-parallel to the bedding. In sandstones the foliation occurs either as a spaced cleavage with planes 5 to 15 mm apart, or as a schistosity with persistent regular laminations between 1 and 5 mm thick. In places this schistosity resembles fine bedding laminations but in thin section it can be seen to be a spaced cleavage. The recognition in thin section of lenticular, strained quartz and feldspar porphyroclasts surrounded by unstrained polycrystalline aggregates suggest that the foliation was produced by cataclasis accompanied by dynamic recrystallization. The relationship between the early fabric and the metamorphic minerals suggests that this deformation occurred during low-grade regional metamorphism.

Second phase of deformation

This phase of deformation is locally very intense but elsewhere virtually absent. Like the first phase no consistent regional variations in intensity were observed. The folds are easily recognized in the field by their relatively small size, open crenulation style, and the development of a distinctive crenulation cleavage. The second phase structures invariably deform the earlier foliation but few re-folded first phase folds were observed in the field.

The folds are asymmetric to symmetric, commonly form zig-zag or box patterns and are disharmonic. They vary in wavelength from millimetres up to 20 m and have an interlimb angle of 35° to 120° with an average of about 80°. Hinge zones are rounded but the fold limbs tend to be straight.

A crenulation cleavage invariably accompanies the second folding and locally gives rise to a marked linear fabric. The orientation of the cleavage and of the axial planes of the folds is very variable (Fig. 2). The cleavage approximates to, but does not always parallel, the axial surfaces of the folds. Cleavage planes and axial surfaces commonly form an irregularly cross-cutting conjugate pattern which suggests that the cleavage planes were not formed perpendicular to the minimum principal stretch but instead are the product of conjugate shearing (Borradaile 1978; Boulter 1979; Gray 1981). Second stage cleavages and axial planes are generally more steeply dipping than those of the first phase, but both sets dip persistently towards the N and NE.

Inverted fold envelopes

Of the 113 way-up indicators observed and recorded in this area approximately two-thirds indicated overturned bedding. Virtually all strata which young towards the S or SW are overturned. This suggests the presence of locally inverted fold envelopes which could have originated only by progressive rotation of structures. The work of Ghosh (1966) and Sanderson (1980) has explained similar rotation by an underthrusting mechanism.

Faulting

The whole area is bounded to the E by the giant N–S trending Atacama fault zone which extends for at least 600 km and was active mainly during mid-Cretaceous times (Naranjo 1978). Mercado (1978) has suggested that movement continued up to recent times. The Atacama fault was described by Arabasz (1971) as an initial strike-slip fault with subsequent normal block faulting. The discovery of a 1 km wide mylonite zone N of Quebrada del Saladito by Naranjo (1978) suggested that the fault had a major strike-slip component of movement. The sense of movement has yet to be determined. There is a possibility that the rocks of the Las Tortolas Formation and the Chañaral mélange, to the W of the fault zone, have undergone considerable lateral transport since the time of their accretion.

In the coastal areas to the W of the Atacama fault zone (Fig. 1) the majority of large faults trend in a NW–SE direction. Lateral displacement of strata along faults, particularly to the N of Quebrada Pan de Azucar and E of Chañaral, indicates that many of these faults had a sinistral strike-slip component of up to 1.5 km (Naranjo 1978). Naranjo also observed smaller faults with dextral strike-slip movement to the E of Pan de Azucar. Few minor and no major thrust faults were identified within the Palaeozoic sediments during the present investigation.

Origin and significance of the deformation

The stratigraphical and structural evidence presented above suggests that the rocks of the Las Tortolas Formation and the Chañaral mélange form part of a subduction complex produced by underthrusting of an oceanic plate towards the NE probably during Upper Palaeozoic times. The mélange represents the product of deformation of semi-consolidated sediment possibly at the base of the inner trench slope (G. F. Moore & Karig 1980). The sediments were subsequently subjected to greenschist facies metamorphism and polyphase deformation within the subduction complex.

Accretionary wedges are commonly described as comprising imbricated slices of sediment each a

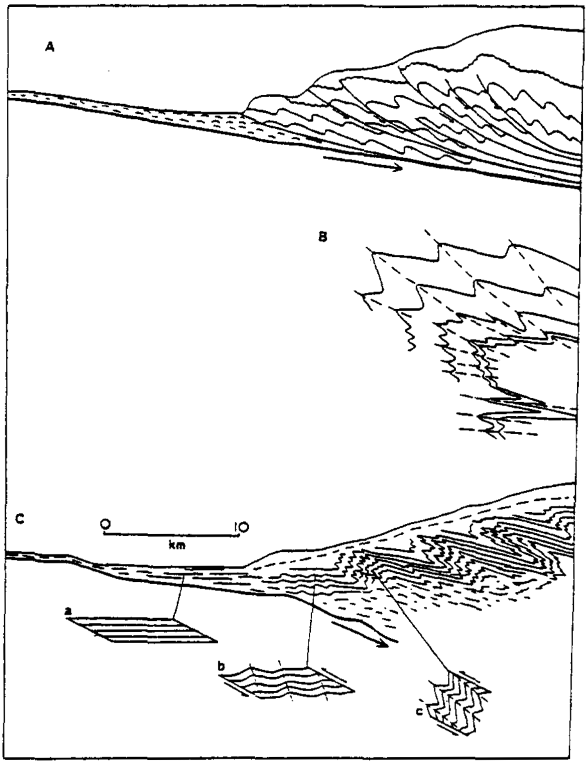

FIG. 3. Schematic diagrams of deformation produced by underthrusting. A. Trench margin model (Seely *et al.* 1974) showing deformation associated with accretion of imbricate thrust wedges. B. Diagrammatic cross-section (Sanderson 1980) showing rotation and tightening of folds as a result of underthrusting in N Cornwall. C. Conceptual model of the formation of chevron folds in the Las Tortolas Formation. (a) Undeformed basin plain sediments. (b and c) Development of chevron folds and packets of overturned strata by simple shear (diagrams after Ghosh 1966). Structures not to scale.

kilometre or less in width (J. C. Moore & Karig 1976) interleaved with belts of mélange (Karig 1974; Seely *et al.* 1974) (Fig. 3A). The wedge is produced by the accretion of sediment at the base of the trench slope, and whilst the wedge is forming, younger sedimentary sequences accumulate in slope basins (J. C. Moore & Karig 1976; G. F. Moore & Karig 1980) and in the fore-arc basin (Scholl & Marlow 1974). Current models of subduction complexes are therefore dominated by the presence of alternating slices of mélange and less-deformed sediment separated by major thrust faults, and unconformably overlain by pockets of less-deformed sediment. The pattern of folding within the accretionary wedge is of asymmetric chevron-style folds, predominantly overturned seawards and with variable interlimb angles, 9° in the Shikoku subduction zone (J. C. Moore & Karig 1976) and between 50° and 80° on Nias Island (G. F. Moore & Karig 1980). Axial surfaces are initially inclined gently landward (J. C. Moore & Karig 1976) but become progressively steepened due to rotation resulting from the wedging effect of the underthrust slices (Seely *et al.* 1974; G. F. Moore & Karig 1980).

By contrast with this well-established explanation for the pattern of deformation seen in subduction complexes, Sanderson (1980) has shown (following the experimental work of Ghosh 1966) that the shear couple resulting from underthrusting initially produces folds with steeply-dipping axial planes which become progressively tighter, with rotation of axial surfaces towards the horizontal, as underthrusting progresses (Fig. 3B). The ultimate stage in this deformation may produce an overturned fold envelope. Tanner & Macdonald (1982) have applied this model to the turbidites of South Georgia Island and have concluded that the pattern of deformation indicates partial subduction or underthrusting of the floor of this Mesozoic back-arc basin beneath the contemporary island arc.

The rocks of the Las Tortolas Formation were deposited in a continuous sequence on a deep-sea basin-plain. There is no evidence of unconformities, stratigraphic discordances or facies changes to suggest deposition in a slope or fore-arc basin. Similarly there are no facies variations of the type suggested by J. C. Moore & Karig (1976) to be indicative of trench deposition (Grow 1973). In addition, there is no evidence of the abundant thrust faults found in the typical accretionary wedge. Most accounts of subduction zone deformation emphasize the importance of these thrust faults (Hsü 1968, 1971, 1974; Seely *et al.* 1974; Jones *et al.* 1978; Moore & Wheeler 1978; G. F. Moore & Karig 1976). However, J. C. Moore & Karig (1976) found no positive evidence for large-scale thrust faulting and no stratigraphic repetitions in the Shikoku subduction zone; similarly Tanner & Macdonald (1982) noted the apparent scarcity of thrust faults on South Georgia. A possible explanation for the absence of thrust faults in the Las Tortolas Formation is that the turbidites were deposited as an undisturbed sequence and subsequently deformed by initiation of subduction beneath the sediment pile. This model (Fig. 3C), based on that developed for South Georgia (Tanner & Macdonald 1982), is contrasted with the more usual model of piecemeal accretion of trench deposits.

The first phase of deformation in the Las Tortolas Formation is dominated by large-scale chevron-style folds very similar in style and scale to those of Nias Island (G. F. Moore & Karig 1980), SW Alaska (Moore 1973), South Georgia (Tanner & Macdonald 1982) and Cornwall (Sanderson 1980). In these two latter areas the deformation has been ascribed to thrust-type shear, with initially upright chevron folds being progressively tightened and rotated towards the horizontal, as a result of continued underthrusting. In the absence of evidence for underthrust slices this model is preferred as an explanation for the deformation in the Las Tortolas Formation. It also explains the probable presence of overturned fold envelopes (Sanderson 1980).

The second phase of deformation, characterized by open upright folds and associated crenulation cleavage is also closely matched by that in NE South Georgia (Tanner & Macdonald 1982), where it was ascribed to horizontal shortening related to basin closure.

González-Bonorino (1971) and Hervé (1977) have described a paired metamorphic belt in S Chile in the region of latitude 40°S where accretion of turbidites and lavas was associated with magmatic activity resulting from NE-directed subduction during Devonian to Carboniferous times. The younger basement complex to the S of 49°S has been described by Nelson *et al.* (1980) as an accretionary prism developed along a convergent segment of the late Palaeozoic to early Mesozoic Pacific margin of Gondwanaland. The trends of the late Palaeozoic folds in Chile commonly vary between E–W and NNW–SSE (Katz 1972; Zeil 1979), oblique to the N–S trending Mesozoic and Cenozoic Andean structures. This may possibly be explained by NE-directed subduction beneath the N–S trending Gondwana margin.

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