Transition from back-arc to foreland basin development in the southernmost Andes: Stratigraphic record from the Ultima Esperanza District, Chile

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ABSTRACT

During the Mesozoic evolution of the southernmost Andes, back-arc basin formation in an extensional setting was followed by foreland basin development in a compressional setting. A stratigraphic record of this tectonic transition is described from Jurassic and Cretaceous rock units exposed in the Patagonian fold-thrust belt at 51°S latitude. Initiation of an extensional deep-marine trough in Late Jurassic time is documented by interstratified rhyolites and marine mudstones in the Tobifera Formation. Development of the Early Cretaceous Rocas Verdes back-arc basin is recorded by Zapata Formation submarine-slope deposits, which draped the west-facing, passively subsiding cratonic margin of the basin. The onset of Andean compressional orogenesis and creation of the Magallanes foreland basin is marked by the influx of coarse-grained sandstone turbidites of the Albian-Cenomanian Punta Barrosa Formation. During Late Cretaceous time, the foreland basin configuration consisted of a narrow foredeep trough bounded by a gently sloping foreland ramp on the craton to the east. This geometry reflects greater flexural subsidence of the stretched and thermally weakened passive back-arc basin margin beneath the load of the obducted Rocas Verdes basin floor. Changes in depositional regime and sediment dispersal patterns in the latest Cretaceous, together with an eastward shift in the Magallanes basin depocenter, record cratonward migration of deformation in the Patagonian fold-thrust belt. Loading of the thicker, cratonic lithosphere caused subsidence of the foreland ramp to form a wide flexural basin.

INTRODUCTION

The positions and geometries of sedimentary basins in southernmost South America changed markedly during the Mesozoic-Cenozoic Andean orogenic cycle. These changes reflect evolving regimes of extension and compression within the overriding plate above the convergent Pacific margin. In the latest Jurassic and Early Cretaceous, coarse clastic sedimentation was localized within a quasi-oceanic back-arc basin that opened near the Pacific margin, called the "Rocas Verdes basin" by Dalziel and others (1974a). The clastic infill and ophiolitic floor of this basin are exposed in the main Cordillera of the southernmost Andes (Fig. 1). During mid-Cretaceous to Neogene Andean compressional orogenesis, coarse clastic sedimentation shifted progressively toward the craton as a retroarc foreland basin developed (Katz, 1964, 1973; Dott and others, 1982; Winslow, 1982; Biddle and others, 1986). The deposits of this depocenter, called the "Magallanes basin" in Chile and the "Austral basin" in Argentina, are exposed in the fold-thrust belt of the Andes and occupy much of the subsurface of South America to the north and east (Fig. 1).

Jurassic to lower Upper Cretaceous rock units exposed in the southern Andes preserve a stratigraphic record of the change from volcanism and sedimentation in an extensional setting to compressional deformation and foreland basin development. I report here the results of the first detailed outcrop investigation of a critical part of this stratigraphic record in southern Chile at latitude 51°S. In this area, extensive exposures of Jurassic and Cretaceous strata occur adjacent to the Sarmiento ophiolite complex, which is the largest remnant of back-arc basin floor exposed in the southern Andes (Fig. 1). Stratigraphic and sedimentologic data from these units provide a more detailed picture of changing basin architectures associated with back-arc basin formation, its subsequent closure and obduction, and development of the Magallanes foreland basin. Equivalent sequences in the subsurface east of the Andes, which are the primary reservoir and source rocks for hydrocarbons in the Magallanes basin, document a linked but significantly different basin history (Gonzalez, 1965; Natland and others, 1974; Biddle and others, 1986). The contrasting tectonostratigraphic evolution recorded by these sequences reflects the influence of prior geologic events on the dynamic behavior of the South American lithosphere during basin development.

Tectonic Framework

The Mesozoic-Cenozoic orogenic cycle in the southernmost Andes began in the Middle to Late Jurassic with a phase of extension characterized by diffuse crustal stretching and widespread silicic volcanism (Bruhn and others, 1978; Gust and others, 1985). This volcanism is inferred to record crustal anatexis during the initial stages of Gondwana breakup (Bruhn and others, 1978; Gust and others, 1985; Dalziel and others, 1987; Kay and others, 1989). The Jurassic silicic volcanic rocks have generally been considered to be "basement" to the Mesozoic-Cenozoic sedimentary depocenters but, as discussed below, extrusion of the volcanics was intimately linked to the early stages of basin development. Subvolcanic crystalline rocks instead constitute mechanical basement, and consist of Gondwanide accretionary assemblages near the Pacific margin and arc and cratonic material farther east (Forsythe, 1982; Dalziel and others, 1987; Ramos, 1988). In the latest Jurassic and Early Cretaceous, the Rocas Verdes back-arc basin opened near the Pacific margin of southern South America, detaching the continental margin arc from South America (Dalziel and others, 1974a; Suarez, 1979; Dalziel, 1981). The back-arc basin filled with uppermost Jurassic and Lower Cretaceous volcaniclastic turbid-

Geological Society of America Bulletin, v. 103, p. 98–111, 15 figs., January 1991.

ites derived mainly from the arc to the west (Suarez and Pettigrew, 1976; Winn, 1978).

The change from extension to compression occurred in the mid-Cretaceous, when arc/craton convergence resulted in partial obduction of the ophiolitic basin floor onto the craton margin (Dalziel and others, 1974a; Bruhn and Dalziel, 1977; Dott and others, 1982; Dalziel, 1986; Dalziel and Brown, 1989). Ongoing compression produced a retroarc fold-thrust belt along the eastern margin of the Cordillera from the Late Cretaceous into the Neogene (Gonzalez, 1965; Katz, 1962, 1973; Winslow, 1982; Wilson, 1983). As defined here, the Magallanes foreland basin was initiated with the onset of compression in the mid-Cretaceous and evolved with the fold-thrust belt into the Neogene.

Geologic Setting of the Study Area

The study area lies in the western fold-thrust belt of the Patagonian Andes at latitude 51°S, within the Ultima Esperanza District of Magallanes Province, Chile (Fig. 1). Upright to steeply inclined folds at the surface are developed above a subsurface thrust system (Wilson, 1983). An east-dipping thrust zone along the western limit of the fold-thrust belt represents the roof thrust of an antiformal duplex that marks the Cordillera/Precordillera boundary (Wilson, 1983). A regionally developed cleavage is present in all of the rock units and, in the Jurassic and Lower Cretaceous rocks, is associated with very low-grade recrystallization.

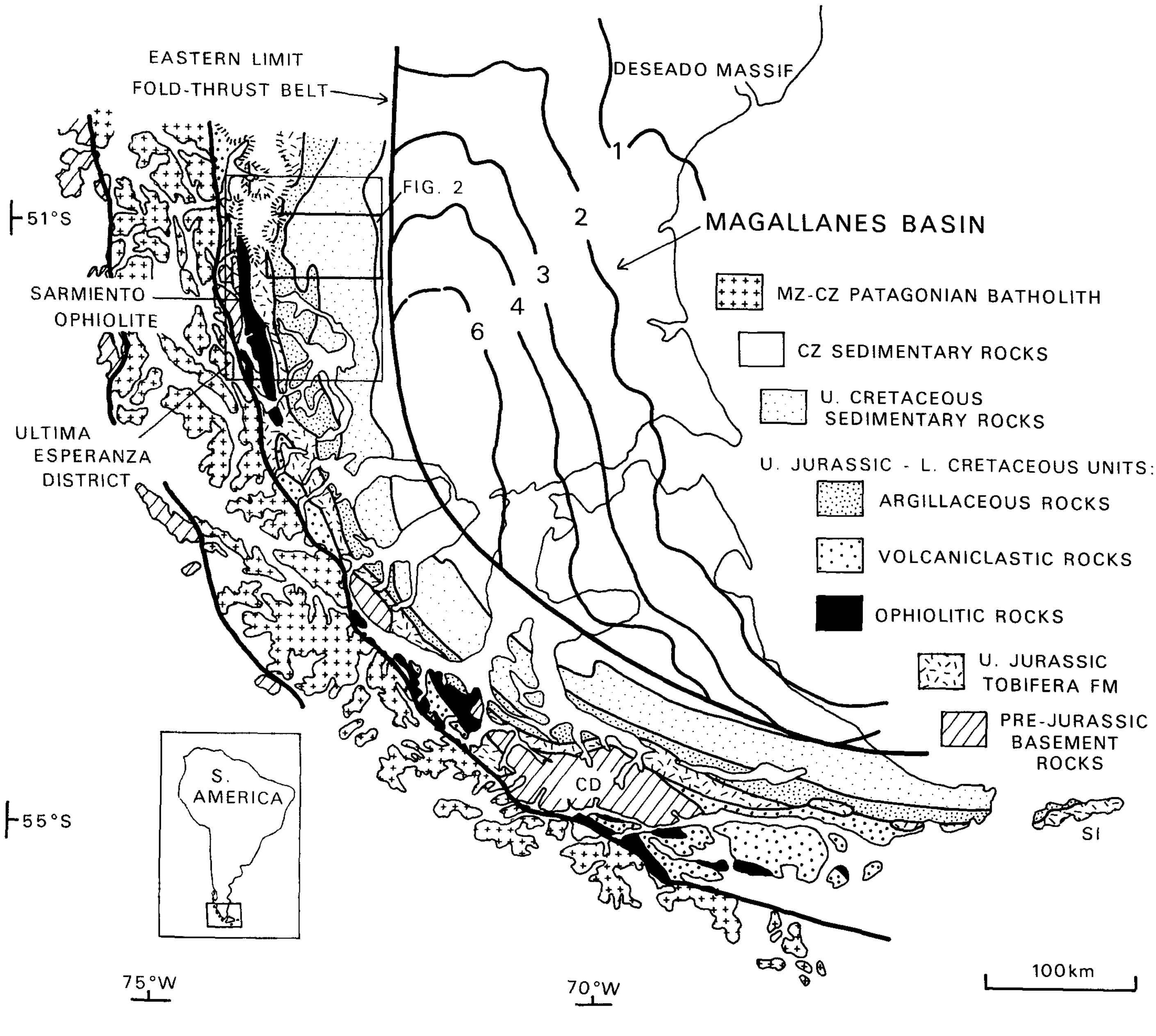


Figure 1. Simplified geologic map of the southern Andes (modified from Dalziel, 1981). Structure contours on top of Jurassic volcanic rocks in kilometers (from Biddle and others, 1986). Hachured pattern: edge of Patagonian ice cap. CD: Cordillera Darwin; SI: Staten Island. Inset shows position of Figure 1 in southernmost South America; boxes show locations of Ultima Esperanza District and Figure 2.

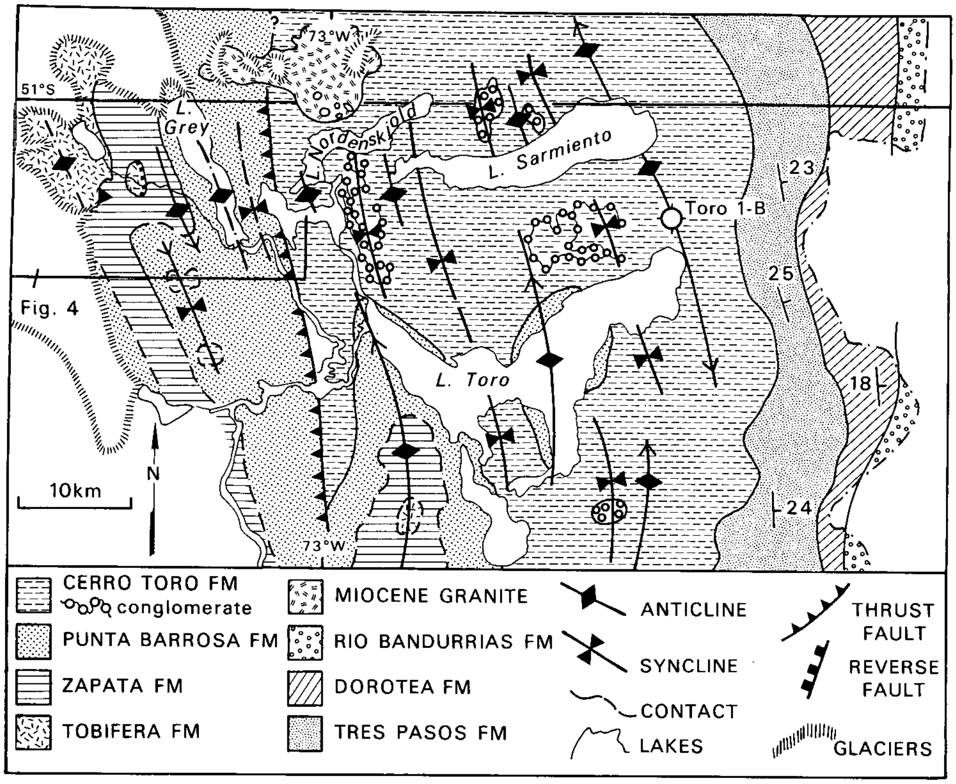


Figure 2. Geologic map of part of the Patagonian fold-thrust belt in Ultima Esperanza (location shown in Fig. 1). Contacts in the east are modified from Cortes (1964). Box shows area of Figure 4. Dot-dash line marks border between Chile and Argentina.

The Mesozoic stratigraphy of the Ultima Esperanza region was first outlined during early scientific expeditions to South America (for example, Nordenskjold, 1899; Hauthal, 1907; Wilckens, 1907; Quensel, 1911; Halle, 1913). Regional studies carried out by ENAP, the Chilean national petroleum company, defined the stratigraphic units shown in Figures 2 and 3 (Cecioni, 1956, 1957; Cortes, 1964; Katz, 1963; Natland and others, 1974; ENAP, 1978). Sedimentologic studies have been carried out on Upper Cretaceous rocks in the area (Scott, 1966; Smith, 1977; Winn and Dott, 1979; Dott and others, 1982; Macellari and others, 1989), but the lower part of the section, which includes the Tobifera, Zapata, and Punta Barrosa Formations (Fig. 3), was examined in detail for the first time in this study. Mapping at 1:50,000 scale was completed for a 400-km^2 area extending eastward from the Patagonian ice cap (Fig. 4). Reconnaissance mapping at 1:100,000 scale in adjacent areas (Fig. 2) was used to modify the 1:250,000 geologic map of the Precordillera region by Cortes (1964).

TOBIFERA FORMATION

The oldest rocks in the Ultima Esperanza Precordillera are part of a Jurassic silicic volcanic unit present throughout southern South America, referred to here as the Tobifera Formation. Tobifera volcanic rocks crop out in local structural inliers and along the eastern margin of the Patagonian ice cap (Figs. 2 and 4). The Tobifera Formation rests with angular unconformity on Gondwanide basement rocks in the Cordillera 20 km to the west of the study area (Forsythe and Allen, 1980; Allen, 1982). The exact thickness of the formation in Ultima Esperanza is unknown because of structural repetition and cover by ice to the west, but it is a minimum of 1 km (Katz, 1963).

Ammonites, belemnites, and inoceramids in mudstones within the upper Tobifera Formation indicate a Late Jurassic (Kimmeridgian-Tithonian) age (Fuenzalida and Covacevich, 1988). Near the Sarmiento ophiolite to the west, Late Jurassic radiolarians are present ~200 m above the base of the formation (Allen, 1982), suggesting that the entire formation may be of Late Jurassic age in the Ultima Esperanza region. Isotopic dates from equivalent silicic volcanics in southern Argentina, however, range from 168–150 Ma (Nullo and others, 1978; Riccardi and Rolleri, 1980; Gust and others, 1985; Biddle and others, 1986), indicating a Middle to Late Jurassic time range for silicic volcanism in southern South America as a whole (Riccardi, 1983; Gust and others, 1985).

Lithology

Mudstones. Black, pyritic, siliceous mudstone horizons from a few centimeters to tens of meters thick are abundant throughout the sequence (Figs. 5 and 6). The mudstones have an even, laterally persistent horizontal lamination. An open-marine origin is clearly indicated by the macrofossil assemblage and the presence of recrystallized radiolarian tests.

Submarine Ash-Fall Tuffs. Tuffs containing abundant devitrified glass shards are locally up to 15 m thick. The tuffs are well sorted, and have even, horizontal bedding and delicate horizontal laminae. These features indicate that vitric ash-fall material was deposited by slow settling into a quiet marine environment.

Tuff Turbidites. Abundant discrete, laterally continuous, tuff turbidite beds from 2 cm to 1 m thick are intercalated with mudstone. They contain rhyolitic lithic debris and quartz and feldspar fragments in a phyllosilicate-rich matrix that most likely represents recrystallized vitric material. Individual beds show T_{a-c} Bouma sequences, and they have sharp bases, load casts, flame structures, and abundant mudstone rip-up clasts (Fig. 5). These sedimentary structures indicate deposition of rhyolitic pyroclastic debris from relatively low-density turbidity currents.

At Cerro Zapata (Fig. 4), the tuff turbidites are amalgamated into composite beds as much as 9 m thick that are lenticular and have channeled bases (Fig. 6). Individual layers within the composite beds are separated by internal scours or by thin, discontinuous mudstone horizons. The thick accumulations of tuff turbidites suggest proximity to source vents.

Rhyolitic Lithic Debris-Flow Deposits. Massive, very poorly sorted lithic debris flows up to 3 m thick are intercalated with the finer-grained rocks (Figs. 5 and 6). The deposits contain angular, nonvesicular rhyolite blocks and less abundant, subrounded, tabular or folded mudstone clasts supported in a fine-grained tuffaceous matrix (Fig. 5). Beds are tabular, with sharp, planar bases (Fig. 5) or occupy broad, shallow channels. Individual beds show coarse-tail grading of rhyolite and mudstone clasts, and reverse grading is present at the base of some beds. Clasts locally project above the tops of beds (Fig. 6). The characteristics of these beds indicate that they formed by en masse deposition from debris flows in which a relatively dense, viscous, fine-grained matrix was capable of supporting coarser clasts during transport (Middleton and Hampton, 1973, 1976; Nardin and others, 1979).

Subaqueous Pyroclastic-Flow Deposits. Tabular, laterally extensive pyroclastic flows up to 5 m thick contain abundant pumice lapilli supported in a tuffaceous matrix, but lack other rhyolitic lithic debris. Coarse-tail normal grading is well developed, and beds typically grade upward into well-laminated tuff. Pumice fragments retain long-tube vesicular texture with no evidence of welding. Both pumice and tabular mudstone rip-ups are oriented subparallel to bedding. The presence of both mudstone and pyroclastic material indicates that tephra derived from eruption columns mixed with water during downslope transport, resulting in turbulent flow. The bedding-parallel clast orientation suggests that a period of laminar flow occurred prior to deposition. The absence of welding and other indicators of high heat content suggest that any original heat contained within the flows was lost during downslope transport.

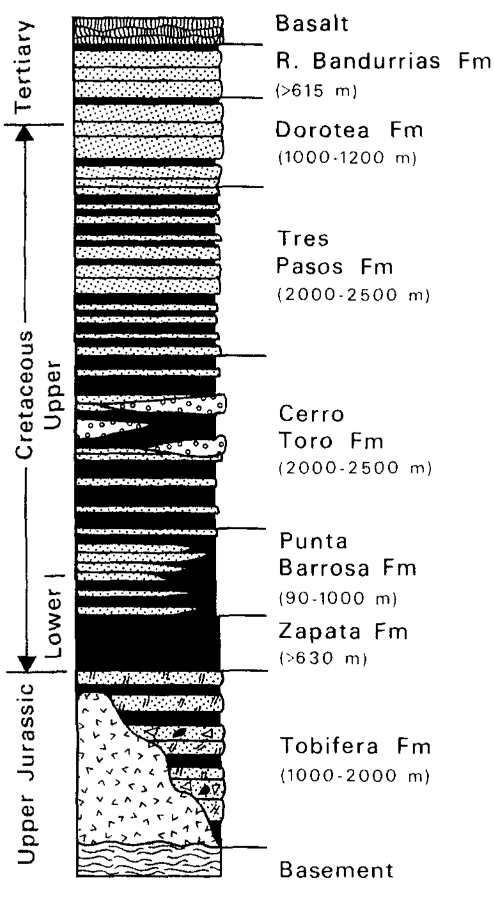


Figure 3. Stratigraphic column in the Ultima Esperanza Precordillera. Patterns indicate relative abundance of mudstone and shale (black), sandstone (stipple), conglomerate (circles), rhyolitic volcaniclastic rocks (stipple with dash) and intrusive rhyolite porphyry (v). In part modified from Katz (1963), Cortes (1964), and Smith (1977).

Penecontemporaneous Hypabyssal Intrusions. Rhyolite and basalt bodies cut the Tobifera section but do not extend into overlying units (Fig. 4). Intrusive rhyolite occurs as coherent bodies or as peperite breccias. The extensive peperite breccias consist of a chaotic mixture of rhyolite fragments and mudstone that formed when rhyolitic magma intruded wet, unconsolidated sediments (Hanson and Wilson, 1986). An intrusive origin for the pillow basalt is indicated by truncation and disruption of bedding in adjacent host rocks, and by the presence of irregular basalt tongues and mudstone xenoliths along the margins.

Depositional Setting

The abundant horizons of black, finely laminated, pyrite-rich carbonaceous mudstone containing marine fossils establish a marine setting for the Tobifera Formation. The mudstones record prolonged periods of hemipelagic sedimentation in a quiet, anaerobic to disaerobic environment punctuated by sporadic influx of rhyolitic debris. Preservation of turbidites and delicate horizontal lamination in the finer-grained rocks, together with the total lack of wave-generated structures, indicate deposition in moderately deep-marine waters below storm-wave base. The presence of inoceramids within the mudstones is consistent with such a setting, as these organisms are believed to have inhabited sublittoral to upper bathyal depths in the Mesozoic (Thiede and Dinkelman, 1977).

Random intercalation of mass-flow and pyroclastic-fall deposits with the marine mudstones documents submarine deposition of the volcaniclastic rocks. The abundance of primary rhyolitic pyroclastic debris requires explosive vesiculation of magma and thus argues for either subaerial or shallow-marine source vents. The absence of accretionary lapilli, however, which are present elsewhere in the Tobifera Formation (Hanson and Wilson, 1990), may indicate that eruption columns did not enter the atmosphere, and that source vents in the Ultima Esperanza region were entirely submarine. Accumulations of coarse-grained, amalgamated, and channelized tuff turbidites, are interpreted to represent submarine aprons formed around volcanic centers. Emplacement of rhyolitic hypabyssal intrusions into unlithified sediments to form extensive peperites documents ongoing magmatic activity during deposition of the sequence. The Tobifera inlier to the southeast of Lago Pingo is onlapped by the highest levels of the overlying Zapata Formation (Fig. 4), indicating that substantial topographic relief was present within the depositional basin.

These new data, together with data from the adjacent Cordillera where Allen (1982) reported black marine mudstones and radiolarian chert near the base of the Tobifera Formation, show that most or all of the sequence in the Ultima Esperanza region was deposited under relatively deep marine conditions upon an erosional surface cut into underlying basement. The Late Jurassic silicic volcanism in Ultima Esperanza was thus associated with development of a deep-marine volcanic trough that preceded opening of the latest Jurassic–Early Cretaceous back-arc basin in the same area. The overlap between mafic and silicic volcanism in the study area, as well as in the lowermost part of the Sarmiento ophiolite complex just to the west (Bruhn and others, 1978; de Wit and Stern, 1981; Allen, 1982), indicates that formation of the ensimatic back-arc basin floor coincided with the waning phases of Tobifera volcanism.

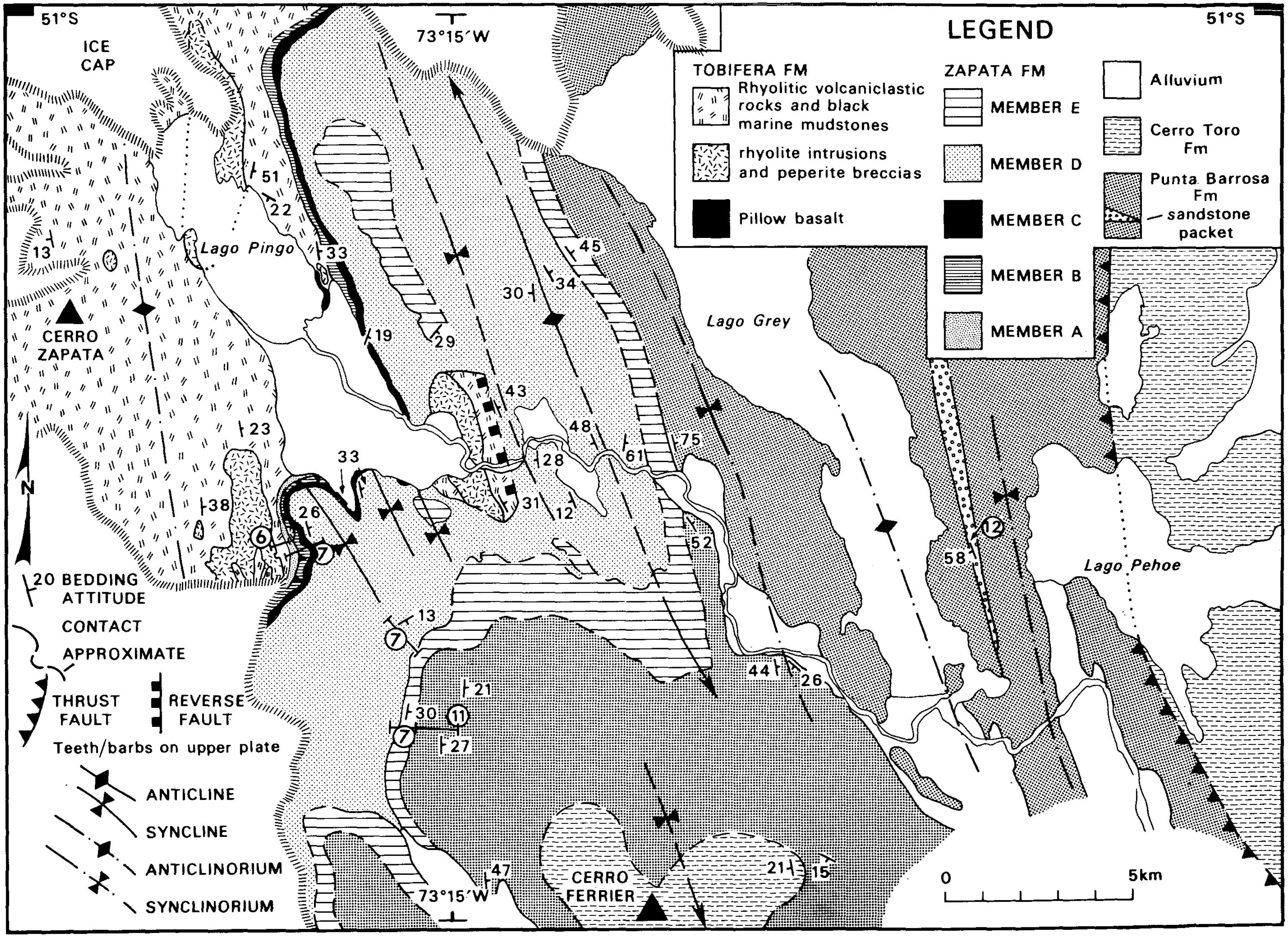


Figure 4. Geologic map of the Tobifera, Zapata, and Punta Barrosa Formations in the westernmost Ultima Esperanza Precordillera (see Fig. 2 for location). Locations of measured sections and photos presented in subsequent figures indicated by circled numbers.

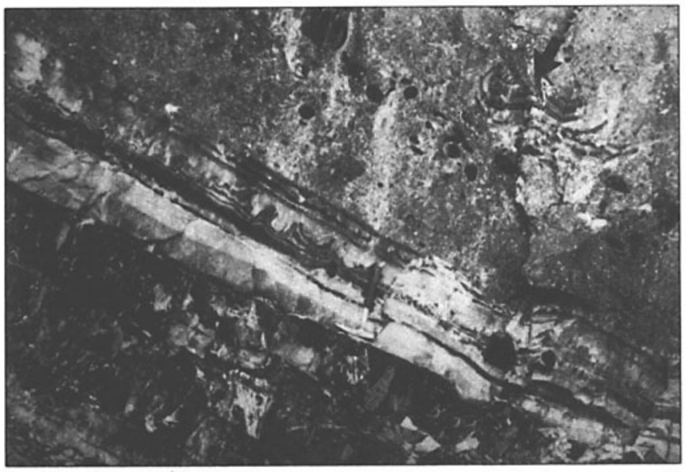


Figure 5. Rhyolitic lithic debris-flow deposit in the Tobifera Formation at Cerro Zapata. Contains light-colored rhyolite clasts and dark-colored mudstone clasts; folded mudstone clast marked by arrow. Hammer handle rests on tuff turbidite with load casts and flame structures. Horizontally bedded, black, marine mudstone makes up base of exposed section.

ZAPATA FORMATION

A Lower Cretaceous unit consisting predominantly of dark mudstones abruptly succeeds the Tobifera Formation in the Ultima Esperanza District. This unit was designated the "Erezcano Formation" by Cecioni (1956) but was renamed the "Zapata Formation" by Katz (1963). The lower contact of the Zapata Formation is defined in this study as the top of the highest volcaniclastic unit within the Tobifera Formation. The upper contact is at the base of the first thick sandstone of the overlying Punta Barrosa Formation. These contacts enclose a distinctive, readily mappable, fine-grained sedimentary package within the Mesozoic succession. *Inoceramus* imprints, belemnites, and ammonites occur throughout the Zapata Formation and indicate a late Tithonian to Albian-Aptian age (Katz, 1963; Cortes, 1964).

The Zapata Formation at the type locality near Cerro Zapata is a minimum of 630 m thick and consists of 5 mappable members, informally designated "A" through "E" (Figs. 4 and 7). Member A is in depositional contact with the underlying Tobifera Formation, but a thrust fault juxtaposes the base of member B against member A. Above the thrust, members B through E form an unbroken section affected only by open folding.

Sedimentology

The Zapata Formation consists of dark mudstones rhythmically interbedded with thin siltstones and very fine-grained sandstones. Bedding is laterally continuous over distances of several kilometers. Except in member C, mudstones in the formation are dark gray to black and contain abundant disseminated pyrite. Black mudstone in member A is identical to mudstone in the underlying Tobifera Formation and shows thin, continuous, undisturbed horizontal lamination. Mudstones throughout the rest of the formation are burrow-mottled, and original lamination has been largely disrupted. Member C forms a prominent interval of white-weathering, intensely bioturbated, highly calcareous mudstone (Fig. 7). Member E is distinguished by a whitish color on weathered surfaces, but is noncalcareous. The trace fossil *Zoophycos* is characteristic of the Zapata Formation above member A. Carbonate concretions are present in all members, reaching up to 2 m in long diameter and 1 m in thickness. Gray, finely laminated calcareous shales from 5 to 25 cm thick occur in member D (Fig. 7).

Thin, laterally continuous beds of quartzose siltstone to very fine-grained sandstone are typically 0.5–3 cm thick, rarely reach 15 cm, and have sharp bases against underlying mudstone. Individual beds typically show T_{a-c} Bouma sequences, indicating deposition from waning, low-density, turbidity currents. The upper surfaces of the siltstones are diffuse and disturbed by bioturbation. Intercalation of these beds with thoroughly bioturbated mudstone intervals records repeated influx of fine-grained turbidites, punctuating periods of slow sedimentation accompanied by intense burrowing.

Member D contains two distinctive types of sandstone turbidite that are not present in the rest of the formation. Tabular and laterally continuous, medium- to coarse-grained micaceous sandstones to 10 cm in thick-

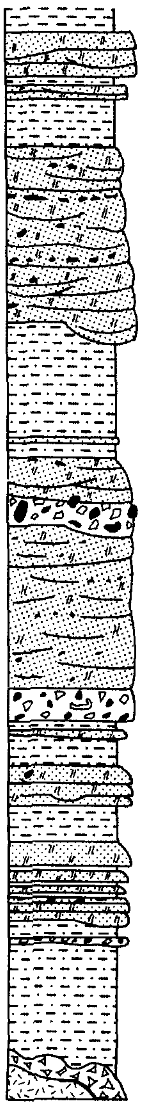
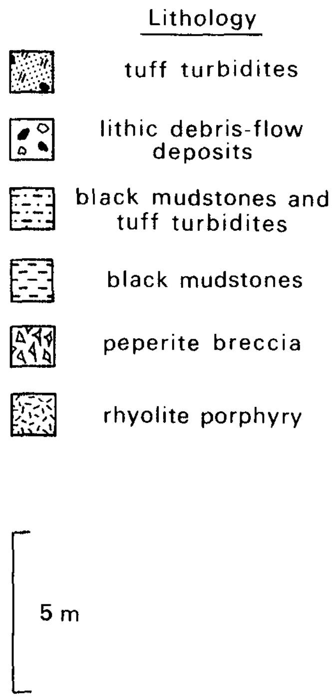


Figure 6. Measured section of part of the Tobifera Formation at Cerro Zapata (Fig. 4). Note thick intervals of amalgamated tuff turbidites and sporadic intercalation of marine mudstone and rhyolitic debris-flow deposits. Peperite breccia and rhyolite porphyry intrude the volcaniclastic section.



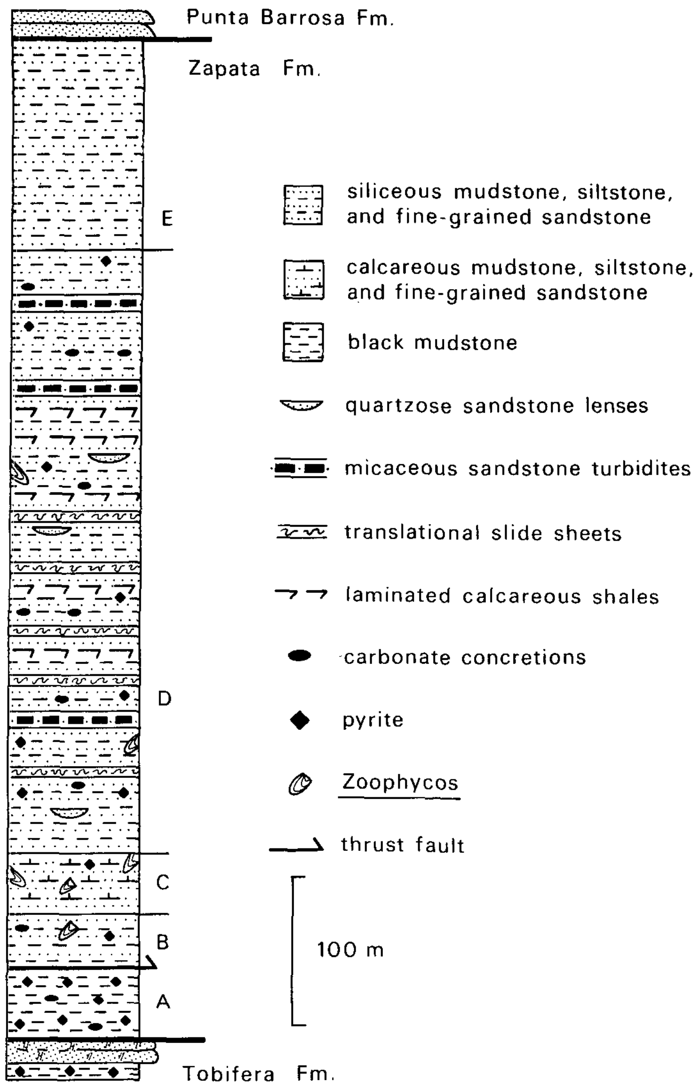


Figure 7. Measured section of the Zapata Formation at the type locality southeast of Cerro Zapata (Fig. 4). Letters designate informal members A through E. Note presence of translational slide sheets and lenticular bodies of medium- to coarse-grained quartzose sandstone turbidites in member D.

ness are T_a turbidites with normal distribution grading and a pronounced bedding-parallel fabric defined by abundant, coarse-grained detrital biotite. Medium-grained quartzose sandstones occur as T_{a-c} turbidites that make up composite beds up to 1.5 m thick. Some of the composite beds are highly lenticular (Fig. 8) and contain silicic detritus up to granule grade, representing the coarsest material in the Zapata Formation. They are interpreted to represent the infill of shallow channels that were incised into the underlying fine-grained deposits. Mudstones and siltstones adjacent to the channels show soft-sediment deformation (Fig. 8), presumably related to channel formation and subsequent loading.

The medium-sand to granule-grade components of the Zapata sandstones consist predominantly of embayed, bipyramidal quartz and plagioclase. Minor amounts of silicic volcanic rock fragments and polycrystalline quartz are also present. This composition shows that the sands were mainly derived from silicic volcanic rocks of the Tobifera Formation. The coarse-grained detrital biotite may also have been derived from the Tobifera volcanics or, alternatively, from pre-Jurassic crystalline basement rocks.

Translational Slide Sheets

The middle section of member D (Fig. 7) contains abundant discrete horizons of disturbed and contorted bedding, here termed "translational slide sheets" following Cook (1979). The translational slide sheets are 0.5 to 5 m thick, with a minimum areal extent of 2 km^2, and have planar bounding surfaces parallel to bedding. They typically have semi-coherent internal layering, and isolated, tight to isoclinal fold closures occur within a disharmonically distorted matrix of mudstone and siltstone. Less intensely disrupted coherent slide sheets with folded but unbroken bedding also are present. Folds within these sheets have a westward vergence and facing direction (Fig. 9), suggesting a westward paleoslope.

The translational slide sheets record repeated gravitational failure on a submarine slope. The hinge lines of soft-sediment folds in all sheets define a north-northwest–south-southeast cluster, with some dispersion within the planes of the sheets (Fig. 10). A southwest paleoslope direction was estimated from the hinge line data using the mean axis method of Woodcock (1979). This analysis assumes the movement direction of a given slide to be perpendicular to the mean hinge line of slump folds within it. Fold closures in the Zapata slide sheets are rootless, which precludes use of fold asymmetry data to constrain the paleoslope direction.

Depositional Setting

The Zapata Formation represents a thick sequence of predominantly fine-grained marine sediments deposited directly after the cessation of Tobifera silicic volcanism. Sedimentary structures throughout the formation indicate deposition below storm wave base. Prolonged accumulation of hemipelagic muds was episodically interrupted by the influx of low-density turbidity currents carrying silt- and sand-sized siliceous detritus.

The characteristics of the Zapata Formation are diagnostic of accumulation in a submarine slope environment. The abundance of organic-rich, intensely bioturbated hemipelagic deposits, the rhythmic intercalation of thin and laterally continuous fine-grained turbidites, and the occurrence of translational slide sheets are all features characteristic of modern and ancient submarine slope deposits (Cook, 1979; Pickering, 1982a; Cook and others, 1982). Coarse-grained, lenticular channel infills

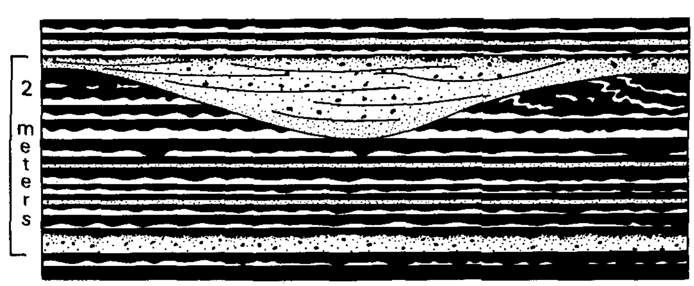


Figure 8. Cross section of lenticular, composite, quartzose sandstone turbidite bed in member D of the Zapata Formation at the type locality. Note channeling, mudstone rip-ups in beds, and soft-sediment disruption of adjacent fine-grained material. Individual, sandstone turbidite occurs beneath the channel fill.

Figure 9. Coherent slide sheet in the Zapata Formation west of Cerro Ferrier, showing westward asymmetry and facing direction (right in photo). Brunton compass for scale (arrowed).

probably represent small-scale examples of the gully systems that dissect modern continental slopes (Cook and others, 1982); channels of similar scale have been described from other ancient slope sequences (Cook and Taylor, 1977; Dott and Bird, 1979). *Zoophycos* occurs in both shallow- and deep-water environments (Bradley, 1973) but is especially common in bathyal settings (Seilacher, 1967; Ekdale and others, 1984); its abundance within the Zapata Formation is thus consistent with a moderately deep marine setting on a slope. The abundance of pyrite and siderite concretions and the presence of laminated calcareous shales are features common to, although not necessarily indicative of, other ancient slope sequences (Cook and Taylor, 1977; Cook and others, 1982).

Slump-fold orientations in the translational slide sheets indicate that the Early Cretaceous Zapata slope faced west. Equivalent rocks near the Sarmiento ophiolite to the west, also designated the Zapata Formation, consist largely of andesitic volcaniclastic turbidites that constitute the Lower Cretaceous sedimentary infill of the Rocas Verdes back-arc basin (Allen, 1982). These relationships indicate that the Zapata slope sequence developed on the passive eastern flank of the back-arc basin. This paleogeography is strongly supported by the fact that the paleoslope direction inferred from slump-fold hinge lines in the Zapata Formation is perpendicular to the axis of the back-arc basin as inferred from the attitudes of sheeted dikes in the Sarmiento ophiolite (Fig. 10).

The change from active Tobifera silicic volcanism to deposition of the Zapata mudstones is interpreted to reflect the mature development of the back-arc basin in the Early Cretaceous. The contact between the two formations records the abrupt quiescence of rhyolitic vents within the submarine Tobifera trough, which ceased to act as local sediment sources. The volcanic trough evolved into a passively subsiding, westward-facing slope that was draped by the Zapata deposits. In the Early Cretaceous, Tobifera silicic volcanic rocks were exposed in the Deseado massif (Fig. 1), and limited exposures of both Tobifera volcanics and crystalline basement rocks were present in isolated fault-block uplifts to the northeast of Ultima Esperanza (Riggi, 1969, 1977; Cecioni and Charrier, 1974; Biddle and others, 1986). Siliceous sandstones within the Zapata Formation thus are inferred to reflect northeast to southwest transport from these source areas onto the west-facing slope.

PUNTA BARROSA FORMATION

The Zapata Formation is conformably overlain by the sand-rich Punta Barrosa Formation. The contact is marked by the abrupt appearance of medium-grained turbidite sandstones that range from 30–100 cm in thickness. The Punta Barrosa Formation grades upward into the shale-dominated Cerro Toro Formation. The contact between the two formations was mapped in this study at the top of the highest thick-bedded sandstone interval. Mapping presented in Figures 2 and 4 differs from earlier regional mapping (Katz, 1963; Cortes, 1964; ENAP, 1978), in which only prominent thick-bedded sandstones in western Ultima Esperanza were assigned to the Punta Barrosa Formation. These prominent sandstone intervals are herein considered to be local lenses within the Punta Barrosa Formation. A complete section of the Punta Barrosa Formation exposed on Cerro Ferrier (Fig. 4) is estimated to be 1,000 m thick. A very sparse ammonite assemblage within the Punta Barrosa Formation indicates a late Albian to Cenomanian age (Cortes, 1964).

Sedimentology

Bed Organization. The Punta Barrosa Formation shows an overall upward increase in grain size and bed thickness. The formation consists of a lower, more shale-rich section and an upper, coarser-grained, sand-rich section containing granule and pebble conglomerates. In the lower section, medium- to coarse-grained sandstone turbidites form individual, tabular, laterally extensive beds from 20 to 150 cm thick within shale-rich sequences (Fig. 11). In contrast, in the upper, sand-rich part of the sequence, thick sandstone beds are segregated into packets from 2 to 40 m thick

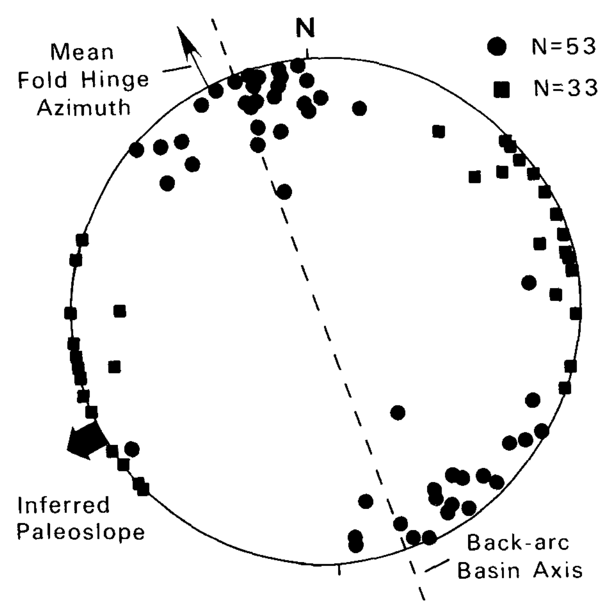


Figure 10. Plot of slump fold hinge lines (solid circles) from translational slide sheets in the Zapata Formation. The inferred paleoslope direction is perpendicular to the mean hinge line azimuth. The dashed line shows the trend of the back-arc basin axis inferred from the orientation of sheeted dikes in the Sarmiento ophiolite complex (solid squares = poles to dikes; data from Allen, 1983). Lower-hemisphere equal-area projection.

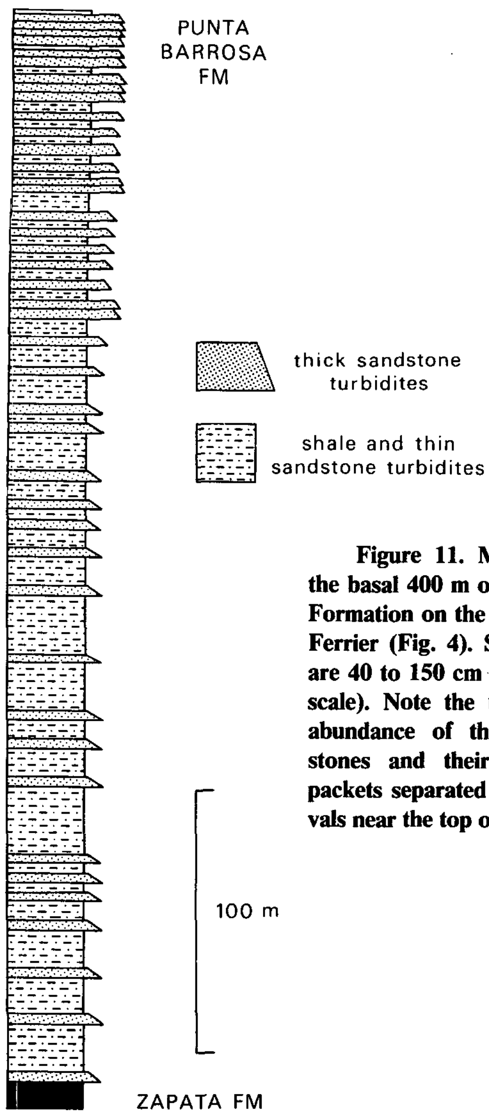


Figure 11. Measured section of the basal 400 m of the Punta Barrosa Formation on the west flank of Cerro Ferrier (Fig. 4). Sandstone turbidites are 40 to 150 cm thick (not shown to scale). Note the upward increase in abundance of thick turbidite sandstones and their arrangement into packets separated by thin shale intervals near the top of the section.

enclosed within sequences of interbedded shale and thin, finer-grained sandstones (Figs. 11 and 12). Vertical trends in bed thickness within packets are either absent or, less commonly, define thickening-upward sequences.

The thicker sandstone packets consist of tabular, composite sandstone beds up to 9 m thick, separated by thin shaly interbeds (Fig. 12). The composite beds contain very coarse-grained sandstone and granule to fine pebble conglomerate in amalgamated layers from 0.5 to 2 m thick separated by shallow scours; distinct channels between beds are rare. The thickness of the packets as a whole, and of the individual composite beds within them, decreases gradually both across and along depositional strike, indicating an overall lenticular geometry. A particularly well-exposed packet was traced for 10 km before it pinched out into finer-grained rocks; it showed a gradual decrease in thickness over this distance (Fig. 4). Abundant channeling was observed in only one packet near the top of the section, which contains a 10-m-deep channel filled with composite beds characterized by abundant internal channeling and local development of basal rip-up conglomerates.

Sedimentary Structures. The thick sandstones within packages are typically massive T_a turbidites with normal distribution grading. Less commonly, the sandstones are massive, lack grading, and contain diffuse horizontal laminae, dish structures, and dispersed shale rip-ups. Very coarse-grained sandstones and granule or pebble conglomerate beds show inverse grading at their bases followed upward by normal coarse-tail or distribution grading. The Punta Barrosa sandstones are inferred to have been deposited by rapid suspension sedimentation from high-energy, high-density turbidity currents (Corbett, 1972; Hiscott and Middleton, 1979; Lowe, 1982).

Outside of the packets, individual, thick, medium-grained sandstones are either T_{a-c} turbidites or, less commonly, T_{a-e} turbidites that were deposited from waning turbulent flows (Walker, 1976). Thin, fine-grained sandstones and siltstones intercalated with shales are typically ripple-form and cross-laminated throughout (T_c turbidites), and represent deposition from traction in relatively dilute, low-velocity turbulent flows (Walker, 1976).

Paleocurrents. Groove and flute casts in Punta Barrosa sandstones document north-northwest to south-southeast paleoflow (Fig. 13). Sediment dispersal was thus parallel to the long axis of the Punta Barrosa depositional basin.

Depositional Setting

An important change in depositional regime is marked by the contact between the Zapata and Punta Barrosa Formations. This boundary records the sudden influx of large volumes of sand transported into the area by turbidity currents. Petrographic studies reported by Dott and others (1982) have shown that the primary source area for the Punta Barrosa sands was a calc-alkalic magmatic-arc terrane, as indicated by the abundance of intermediate-composition volcanic rock fragments. Thus, in the mid-Cretaceous, a major reversal occurred from the eastern source area that supplied silicic volcanic detritus to the Zapata slope, to a western source along the Pacific margin of the continent. This reversal in sediment supply marks rapid construction of the magmatic arc and initial orogenic deformation and uplift in the Main Cordillera of the Andes (Dalziel and others, 1974b; Dott and others, 1982).

Figure 12. Tabular form of thick, subvertical, composite sandstone beds that make up a packet in the upper part of the Punta Barrosa Formation east of Lago Grey (Fig. 4). Top to right. Daypack (circled) rests on fine-grained material that separates the two prominent beds. Internal contacts between layers within the composite beds visible as subtle color changes. Sandstone packet containing the same beds is visible on prominent hillside in the background (arrows).

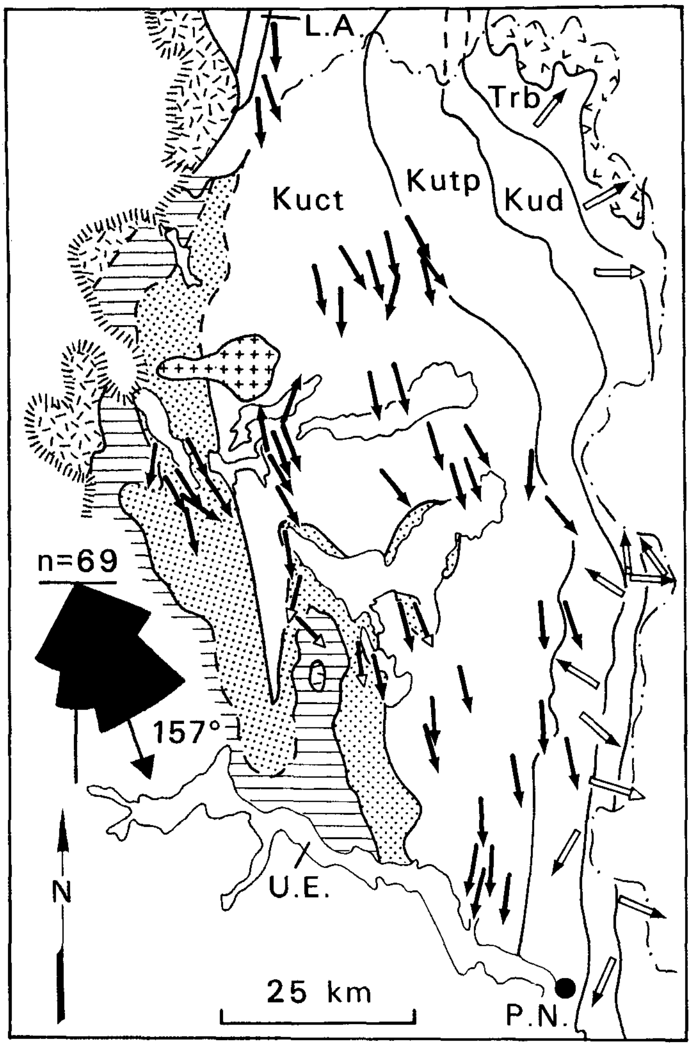


Figure 13. Summary map of paleocurrent trends within the clastic section of the Ultima Esperanza Precordillera. Note consistent southward paleoflow in the Late Cretaceous units and the change to mainly eastward paleoflow in the latest Cretaceous and younger units. Rose diagram shows new data from the Punta Barrosa Formation, including 61 groove and 8 flute orientations. Paleoflow directions from flute orientations shown by arrows with single tails; from cross-bedding shown by arrows with double tails. Solid head: mean of 2 to 95 readings; open head: single readings. Data compiled from Cortes (1964), Scott (1966), Smith (1977), Wilson (1983), and Arbe and Hechem (1984). Random dash: Tobifera Fm; horizontal lines: Zapata Fm; stipple: Punta Barrosa Fm; Kuct: Cerro Toro Fm; Kutp: Tres Pasos Fm; Kud: Dorotea Fm; Trb: Rio Bandurrias Fm (see Fig. 3). U.E.: Seno Ultima Esperanza; P.N.: Puerto Natales; L.A.: Lago Argentino; dot-dash line: border between Chile and Argentina.

The thickness of the Punta Barrosa Formation decreases eastward from 1,000 to 800 m in outcrop. Punta Barrosa sandstones are absent in the Toro 1-B well drilled 50 km (nonpalinspastic) east of Cerro Ferrier (Fig. 2), indicating an eastward pinch-out of the unit (Katz, 1963; Cortes, 1964). The Punta Barrosa sandstones can be traced for only 90 km along strike to the south before they pinch out into a fine-grained sequence (Katz, 1963; Cortes, 1964). The formation thus represents a narrow, elongate, sand-rich unit that pinches out both to the east and south. Together with the uniform axial paleoflow direction (Fig. 13), this indicates that the Punta Barrosa depositional system was confined in a narrow trough that developed on the site of the Early Cretaceous Zapata slope. The Punta Barrosa Formation shows many similarities to ancient, sand-rich, submarine-fan deposits (Mutti and Ricci Lucchi, 1972; Link and Nilsen, 1980; Hiscott, 1980; Howell and Normark, 1982; Shanmugam and Moila, 1988), although the elongate, tongue-shaped geometry of this depositional system suggests that no morphologic "fan" was developed (compare with Pickering, 1982b; Hiscott and others, 1986). The overall coarsening- and thickening-upward trend within the Punta Barrosa Formation records the progradation of this fan-like, sand-rich depositional system southward along the axis of the constricted trough.

DEPOSITIONAL SETTING OF CENOMANIAN-TERTIARY UNITS

The Punta Barrosa Formation represents the lowest part of a thick succession of sand-rich clastic rocks of Late Cretaceous and Tertiary age in Ultima Esperanza (Fig. 3). The Cenomanian-Campanian Cerro Toro Formation, which overlies the Punta Barrosa Formation, consists of hemipelagic shales and sandstone turbidite deposits that enclose the spectacular Lago Sofia conglomerates (Zeil, 1958; Katz, 1963; Cortes, 1964; Scott, 1966) and has been interpreted as levee, levee-flank, and channel-fill facies deposited in the inner part of a submarine-fan system (Winn and Dott, 1979). The lower and middle parts of the overlying, sand-rich Santonian-Maastrichtian Tres Pasos Formation have been interpreted as mid-fan deposits (Smith, 1977; Dott and others, 1982). The Punta Barrosa through middle Tres Pasos units constitute a conformable sequence ~3,000 m thick laid down within a fan-like depositional system characterized by consistent north-to-south paleoflow directions (Fig. 13). The accumulation of this sequence indicates that a stationary, deep-water, trough-like depocenter existed in Ultima Esperanza from Cenomanian to Maastrichtian times.

Beginning in the Maastrichtian, deep-water deposits were succeeded by shallow-marine sediments, and sediment dispersal shifted from the longitudinal system to a more diverse pattern dominated by eastward transport off rising highlands (Fig. 13; Smith, 1977; Dott and others, 1982). Shallow-marine, deltaic, and fluvial environments persisted throughout deposition of the succeeding Maastrichtian-Neogene units (Katz, 1963; Dott and others, 1982; Macellari and others, 1989).

REGIONAL STRATIGRAPHIC RELATIONSHIPS

Middle to Late Jurassic

The linear outcrop belt of the Tobifera Formation in the southern Andes represents one element of the Middle to Late Jurassic rhyolitic volcanic field that extended across southern South America (Gust and others, 1985; Kay and others, 1989). The recognition of marine mudstone intercalations near the top of the Tobifera Formation in these Andean exposures led several workers to suggest that subsidence and marine incursion began in the region during the waning phases of Tobifera volcanism (Feruglio, 1949; Katz, 1963; Dalziel and others, 1974b; Bruhn and others, 1978; Winslow, 1980; Charrier and Covacevich, 1980). More recent work has demonstrated a marine depositional environment for most or all of the unit in Ultima Esperanza (Allen, 1982; this study), southernmost Patagonia and Tierra del Fuego (Fuenzalida, 1984; Johnson, 1987; Fuenzalida and Covacevich, 1988; Hanson, 1987) and Staten Island (Hanson, 1989). In all of these areas, several kilometers of rhyolitic material accumulated in

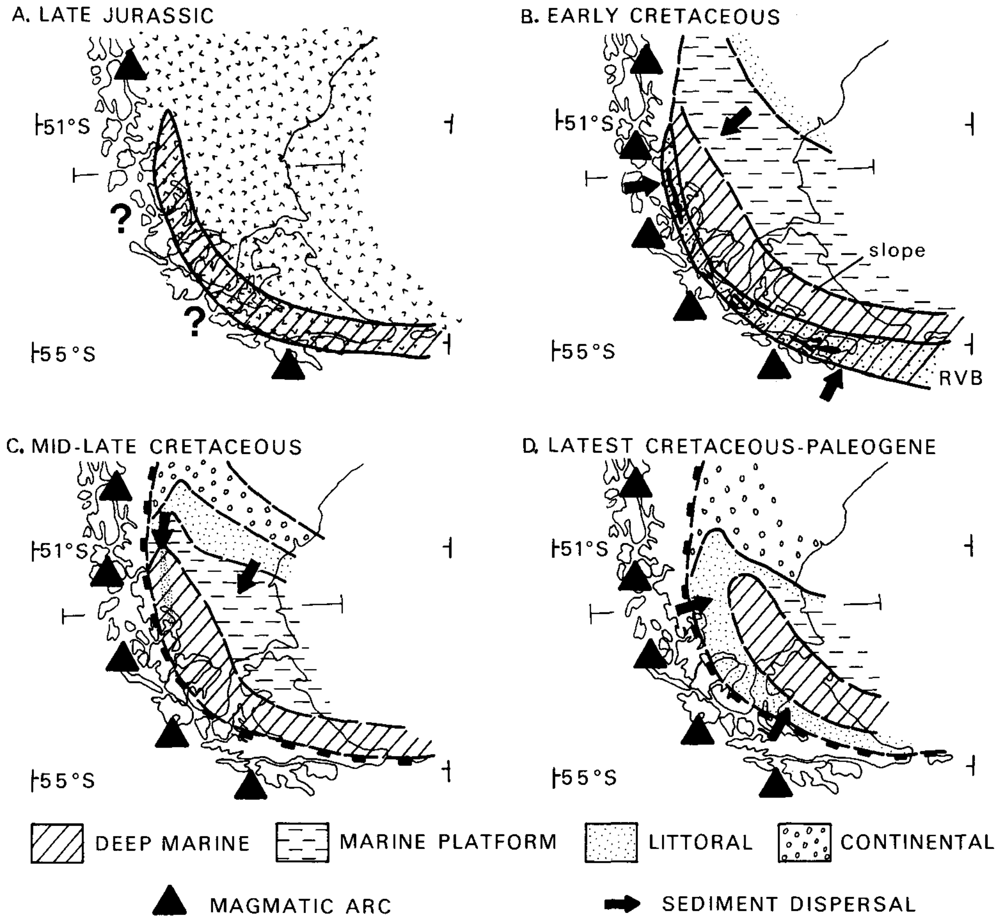


Figure 14. Generalized depositional settings for four stages of basin evolution in southernmost South America. A. Subaerial extension-related rhyolitic volcanism passes westward into deep-marine volcano-tectonic trough. B. Rocas Verdes back-arc basin (RVB; volcaniclastic material stippled; exposed ophiolitic rocks in black) passing cratonward to bathyal submarine slope province and shallow-marine platform. C. Deformation and uplift in Andean protocordillera (barbed line); Punta Barrosa clastics (stipple) deposited in narrow foredeep trough, with shaly foreland ramp to east. D. Deformation and uplift in fold-thrust belt (barbed line); basin shifts eastward. Sources: Riccardi and Rolleri (1980); Russo and others (1980); Aguirre and Ramos (1981); Nullo and others (1981); Malumian and others (1983); Biddle and others (1986); Riccardi (1987, 1988).

marine environments below storm wave base, indicating that a deep-marine volcano-tectonic basin developed for ~1,000 km along the Jurassic continental margin (Hanson and Wilson, 1990). Tobifera-equivalent units to the east of the Andes consist of subaerial silicic volcanic and volcaniclastic rocks, and marine conditions were first established there in the Early Cretaceous after volcanism had ceased (Riggi, 1969, 1977; Natland and others, 1974; Riccardi, 1983; Gust and others, 1985; Biddle and others, 1986). The association of this volcanism with regional extension and normal faulting is well documented from seismic and well data (for example, Gust and others, 1985).

The Middle to Late Jurassic paleogeography thus consisted of an extensive, extension-related subaerial volcanic province that passed into a deep-marine volcanic trough along the Pacific margin (Fig. 14a). South of approximately 50°S latitude, there is little direct evidence for an active continental margin magmatic arc at this time, and thus the inferred "back-arc" setting of this trough remains in question (Hanson and Wilson, 1990).

Early Cretaceous

Lower Cretaceous black mudstones equivalent to the Zapata slope sequence are exposed along the entire length of the southern Andean Precordillera (Fig. 1). North of the study area between latitudes 45°S and 50°S, these have been interpreted as a shallow-marine basin fill (Skarmeta, 1976), or a euxinic marine platform sequence (Nullo and others, 1981). To the south, between latitudes 52°S and 54°S, Winslow (1980) interpreted them as a distal turbidite facies deposited along the cratonward edge of the back-arc basin. Published descriptions of the Lower Cretaceous rocks to the south of Ultima Esperanza (Dalziel and others, 1974b; Nelson and others, 1980; Winslow, 1980) indicate a close similarity to the Zapata Formation as described herein. Recent observations of the Lower Cretaceous rocks in Tierra del Fuego and on Staten Island have confirmed that these units show rock types, sedimentary structures, and trace fossils identical to the Zapata Formation in Ultima Esperanza (R. E. Hanson, 1989, oral commun.). These Lower Cretaceous mudstones are the lateral equivalents of the volcaniclastic turbidite infill of the back-arc basin that opened between the magmatic arc and the craton, and they are interpreted here to represent a regionally developed Early Cretaceous submarine slope that faced westward into the back-arc basin (Fig. 14b).

In the subsurface to the east, the Lower Cretaceous sequence consists of a southwestward-thickening wedge of marine mudstones, shaly limestones, and minor glauconitic sandstones that was deposited under anaerobic to disaerobic conditions (Biddle and others, 1986). Biddle and others (1986) used subsurface data to infer east-to-west shelf, slope, and basin-plain depositional settings for this sequence east of the Andes, but the outcrop data from the Zapata Formation presented here indicate that the bathyal part of the Early Cretaceous slope was well to the west of their study area.

Regional stratigraphic data for the Early Cretaceous document a transition in depositional environment from a marine shelf in the east to a bathyal slope and a back-arc basin along the Pacific margin (Fig. 14b). Spreading occurred within the precursor Jurassic volcanic basin to form the Rocas Verdes back-arc basin, which filled with arc-derived deep-water

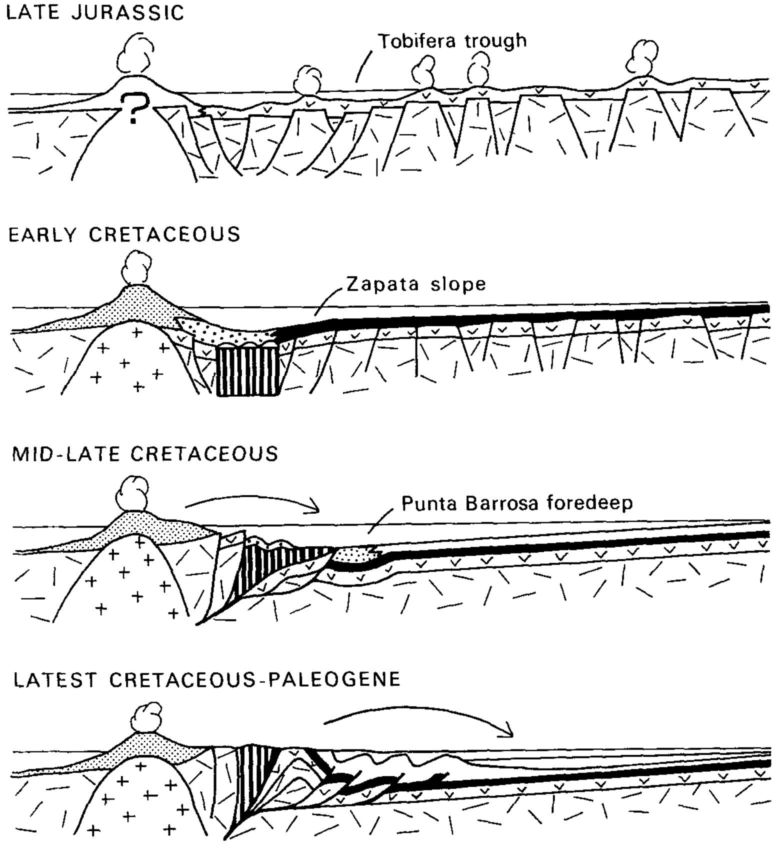


Figure 15. Sequential evolution in basin geometry in southernmost South America. See text for discussion. Arrows indicate principal source areas for clastic material deposited in foredeep. Random dash = crystalline basement; v-pattern = Tobifera Formation; cross-pattern and dense stipple = magmatic arc; heavy vertical line pattern = ophiolite; coarse stipple = Lower Cretaceous arc-derived, volcaniclastic rocks; black = Lower Cretaceous fine-grained, clastic rocks; light stipple = coarse-grained, deep-water, clastic deposits in foredeep; unpatterned = Upper Cretaceous–Tertiary clastic deposits.

submarine fan deposits. A starved, west-facing prism of euxinic, fine-grained clastic material was draped on the subsiding cratonic margin of the basin. The along-strike continuity of the back-arc basin slope province precludes the existence of a physiographic remnant arc along the cratonward margin of the back-arc basin, as had been suggested by previous workers (Dalziel and others, 1974a; de Wit, 1977; Bruhn and others, 1978; Dalziel, 1981). Regional subsidence occurred inboard of the back-arc basin, and oxygen-poor marine waters transgressed eastward over the Jurassic subaerial volcanic province (Cecioni and Charrier, 1974; Gust and others, 1985; Biddle and others, 1986).

Mid-Cretaceous to Late Cretaceous

The paleogeographic setting changed markedly beginning in Albian-Cenomanian time (Fig. 14c), with the onset of the Andean orogeny (Dalziel, 1986). The initial pulse of deformation involved closure and inversion of the Rocas Verdes back-arc basin, including partial obduction of its ophiolitic floor (Dalziel, 1981; Dalziel and Brown, 1989). Concomitantly, deposition of coarse clastic material shifted from the back-arc basin to an elongate deep-water trough developed on the site of the Early Cretaceous Zapata slope. This transformation of the Early Cretaceous back-arc basin/slope/platform setting into a protocordillera/foredeep setting is considered here to mark the inception of the Magallanes foreland basin.

Arc-derived Punta Barrosa sandstones represent the first influx of coarse clastic material into the Magallanes basin. The northern limit of the deep-marine trough is marked by coarse clastic littoral deposits at latitude 49°S (Riccardi and Rolleri, 1980; Nullo and others, 1981; Aguirre and Ramos, 1981). At latitude 50°S, turbidite sandstones with southerly paleocurrent directions (Vilela and Csaky, 1968; Arbe and Hechem, 1984) appear to represent the northern limit of the Punta Barrosa depositional system, which continues for about 90 km to the south (Cortes, 1964), where equivalent sections consist of fine-grained submarine-fan deposits (Winslow, 1980).

A southwestward-thickening wedge of mudstones deposited in an aerobic to disaerobic, open-marine setting in the central and eastern Magallanes basin (Biddle and others, 1986) correlates with the sandy Punta Barrosa section. Seismic reflection patterns indicate a gently southwest-sloping platform in this part of the basin, with sediment dispersal from northeast to southwest (Biddle and others, 1986). Regional relationships thus show that the Punta Barrosa depositional system and equivalent deep-water clastic deposits to the south were confined within a narrow, elongate moat that was topographically isolated from a sloping platform, or foreland ramp, on the craton to the east. Both outcrop and subsurface data indicate that this basin geometry persisted throughout most of the Late Cretaceous, into the Maastrichtian.

Latest Cretaceous through Neogene

Shallow-marine, deltaic and fluvial sedimentation was initiated along the western margin of the basin in the Maastrichtian, and continued there into the Neogene (Macellari and others, 1989). Resedimented Early Cretaceous fossil fragments and palynomorphs (Gonzalez, 1965; Katz, 1973) and changes in the sandstone petrofacies of the units (Manassero and Macellari, 1987; Macellari and others, 1989) reflect structural uplift in the westernmost fold-thrust belt during this interval. The shaly foreland ramp that existed up to this time in the central and eastern basin subsided to become the main basin depocenter (Fig. 14d; Katz, 1964, 1973; Natland and others, 1974; Winslow, 1980, 1982; Biddle and others, 1986). According to Biddle and others (1986), the first sedimentary units in this new, more easterly depocenter with a possible Cordilleran source are of mid-Maastrichtian to mid-Thanetian (Paleocene) age. A major regional erosional unconformity separates these units from mid-Eocene and younger sequences that represent major fan deltas that prograded eastward and northeastward into the central and eastern parts of the foreland basin from the uplifting Cordillera (Biddle and others, 1986).

IMPLICATIONS FOR BASIN EVOLUTION

Four distinct phases of basin development are manifested in the Ultima Esperanza stratigraphic record, and regional stratigraphic relationships show that they represent evolving basin geometries throughout South America to the south of ~50°S latitude. The basin stages must, therefore, be linked to changing tectonic regimes and structural patterns within the southernmost part of the South American plate during Mesozoic and Cenozoic time.

Basin formation in the Jurassic and Early Cretaceous occurred during regional crustal extension. The initial development of a deep-marine trough coincided with the inception of silicic Tobifera volcanism in Ultima Esperanza (Fig. 15a) and along a 1,000-km-long sector of the Pacific margin (Hanson and Wilson, 1990). Differential subsidence of the trough with respect to the rest of the Jurassic silicic volcanic field suggests that more extreme crustal thinning by normal faulting was localized along the margin or, alternatively, that there was a more marked response to stretching along the margin because it is underlain by a thin basement of accreted Gondwanide forearc material (Hanson and Wilson, 1990). The waning phase of Tobifera volcanism is contemporaneous with the onset of the mafic volcanism that generated the ophiolitic floor of the Rocas Verdes back-arc basin. This indicates that extension and subsidence to form the volcano-tectonic trough represents the initial stage of back-arc basin formation. Rocas Verdes basin development in the Early Cretaceous is marked by accumulation of a thick sequence of coarse, arc-derived andesitic clastic material in the back-arc basin, construction of the west-facing Zapata slope sequence on the subsiding basin margin, and marine onlap over the craton to the east (Fig. 15b). Formation of a localized deep-marine trough followed by regional subsidence produced an asymmetric version of the "steers head" configuration typical of extensional basins, which originates from early fault-controlled subsidence followed by regional subsidence due to cooling and contraction of the heated lithosphere (Dewey, 1982; White and McKenzie, 1988). As noted by Biddle and others (1986), the primary hydrocarbon source rocks and main reservoir rocks were laid down during this thermal subsidence phase, before the Magallanes foreland basin was topographically defined.

A fundamental change in basin geometry occurred in late Albian to Cenomanian time when the site of coarse clastic deposition shifted from the back-arc basin to a subsiding trough that developed on the site of the Early Cretaceous back-arc basin slope. Throughout the Late Cretaceous, arc- and cordillera-derived clastic material was funneled axially along this deep-water trough, which was bounded by a gently sloping, sediment-starved foreland ramp on the craton to the east (Fig. 15c). Contemporaneity of this phase of basin development with the onset of Andean compressional deformation and obduction of the back-arc basin floor in the mid-Cretaceous indicates that subsidence resulted from flexure of the craton margin by thrust loading, delineating the Magallanes foreland basin for the first time. The width and depth of a flexural basin formed in response to a given thrust load depends on the flexural rigidity of the lithosphere, with narrow, deep basins forming on lithosphere with a relatively small effective elastic plate thickness (Quinlan and Beaumont, 1984). The segmentation of the Magallanes basin into an extensive foreland ramp on the craton and a narrow "moat" adjacent to the Andean protocordillera therefore reflects a decrease in flexural rigidity from the Atlantic to the Pacific margins, caused by the thinning and thermal weakening of the lithosphere along the passive cratonic margin of the back-arc basin and by the presence of an initially thin accreted crust beneath the margin. The persistence of this basin configuration for ca. 30 Ma suggests that no significant cratonward migration of the thrust load occurred during this interval. This inference is consistent with structural observations that show that crustal thickening was localized along the cratonic margin of the back-arc basin up to the Late Cretaceous in both the Cordillera Darwin area (Bruhn and Dalziel, 1977; Nelson and others, 1980; Dalziel and Brown, 1989) and Ultima Esperanza (Wilson, 1983).

Beginning in the latest Cretaceous, the basin geometry again altered substantially. The western moat was filled to sea level, coarse clastic debris was shed eastward off rising highlands, and the basin depocenter shifted eastward and widened to encompass the former foreland ramp (Fig. 15d). This change in basin configuration is interpreted here to mark the onset of eastward-migrating thrusting in the Patagonian fold-thrust belt. The major late Paleocene to mid-Eocene erosional unconformity throughout the central and eastern Magallanes basin (Gonzalez, 1965; Biddle and others, 1986) may mark development of a flexural peripheral bulge related to this thrust loading. Flexural subsidence in the eastern part of the basin beginning ca. 50 Ma (mid-Eocene) is recorded in subsidence curves calculated from well data by Biddle and others (1986), tracking continued cratonward migration of the thrust belt. This final period of basin evolution reflects tectonic loading of the thicker lithosphere of the South American craton.

ACKNOWLEDGMENTS

I would like to thank Ian W.D. Dalziel for the opportunity to work in southern Chile; National Science Foundation awards to him supported the research. Logistical support in Chile was generously provided by the Empresa Nacional del Petroleo, the Instituto de la Patagonia, and the Parque Nacional Torres del Paine. The help and hospitality of Pepe Alarcon and the late Pochongo Medina in the Parque Nacional Torres del Paine made much of my field work possible. I also thank the many Chilean and American colleagues and friends who assisted me during the field work. Trace fossils were kindly identified by R. H. Dott, Jr. and C. W. Byers of the University of Wisconsin. Discussions with Pitch Allen, Gerard Bond, Raul Cortes, Richard Hanson, and Richard Schweickert clarified the ideas presented here. Special thanks are due to Richard Hanson for unflagging enthusiasm in the field and for reviewing early versions of this manuscript. Constructive reviews by K. T. Biddle, R. H. Dott, Jr., and R. D. Winn, Jr., improved the manuscript.

REFERENCES CITED

Aguirre, M. B., and Ramos, V. A., 1981, Estratigrafia y paleontologia de la Alta Cuenca de Rio Roble, Cordillera Patagonica—Provincia de Santa Cruz: Congreso Geologico Argentino, 8th, San Luis, Actas III, p. 101–138.

Allen, R. B., 1982, Geologia de la Cordillera Sarmiento, Andes Patagonicos, entre los 51° 00′ y 52° 15′ lat. S, Magallanes, Chile: Servicio Nacional de Geologia y Mineria—Chile, Boletin 38, 46 p.

——— 1983, Geologic studies of the Scotia Arc region and Agulhas Plateau [Ph.D. thesis]: New York, Columbia University.

Arbe, H. A., and Hechem, J. J., 1984, Estratigrafia y facies de depositos continentales, litorales y marinos del Cretacico Superior, Lago Argentino, Provincia de Santa Cruz: Noveno Congreso Geologico Argentino, Actas 7, p. 124–158.

Biddle, K. T., Uliana, M. A., Mitchum, R. M., Jr., Fitzgerald, M. G., and Wright, R. C., 1986, The stratigraphic and structural evolution of the central and eastern Magallanes Basin, southern South America, *in* Allen, P. A., and Homewood, P., eds., Foreland basins: International Association of Sedimentologists Special Publication 8, p. 41–61.

Bradley, J., 1973, *Zoophycos* and *Umbellula* (Pennatulacea): Their synthesis and identity: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 13, p. 103–128.

Bruhn, R. L., and Dalziel, I.W.D., 1977, Destruction of the Early Cretaceous marginal basin in the Andes of Tierra del Fuego, *in* Talwani, M., and Pitman, W. C., III, eds., Island arcs, deep sea trenches and back-arc basins: American Geophysical Union, Maurice Ewing Series, Volume 1, p. 395–405.

Bruhn, R. L., Stern, C. R., and de Wit, M. J., 1978, Field and geochemical data bearing on the development of a Mesozoic volcano-tectonic rift zone and back-arc basin in southernmost South America: Earth and Planetary Science Letters, v. 41, p. 32–46.

Cecioni, G., 1956, Leopoldia? payensis Favre: Sua posizione stratigrafica in Patagonia: Societa Italiana della Sciencias Naturales, Alti., v. 95, p. 135–145.

——— 1957, Cretaceous flysch and molasse in Depto. Ultima Esperanza, Magallanes Province, Chile: American Association of Petroleum Geologists Bulletin, v. 41, p. 538–564.

Cecioni, G., and Charrier, R., 1974, Relaciones entre la Cuenca Patagonica, la Cuenca Andina, y el Canal de Mozambique: Ameghiniana, v. 11, p. 1–38.

Charrier, R., and Covacevich, V., 1980, Paleogeografia y bioestratigrafia del Jurasico superior y Neocomiano en el sector Austral de los Andes Meridionales Chilenos (42°-56° latitud Sur): Segundo Congreso Argentino Paleontologico y Bioestratigrafico, Buenos Aires, 1978, Actas V, p. 153-174.

Cook, H. E., 1979, Ancient continental slopes and their value in understanding modern slope development, *in* Doyle, L. J., and Pilkey, O. H., eds., Geology of continental slopes: Society of Economic Paleontologists and Mineralogists Special Publication No. 27, p. 287-307.

Cook, H. E., and Taylor, M. E., 1977, Comparison of continental slope and shelf environments in the Upper Cambrian and lowest Ordovician of Nevada, *in* Cook, H. E., and Enos, P., eds., Deep-water carbonate environments: Society of Economic Paleontologists and Mineralogists Special Publication No. 25, p. 51-83.

Cook, H. E., Field, M. E., and Gardner, J. V., 1982, Characteristics of sediments on modern and ancient continental slopes, *in* Scholle, P. A., ed., Sandstone depositional environments: American Association of Petroleum Geologists Memoir 31, p. 329-364.

Corbett, K. D., 1972, Features of thick-bedded sandstones in a proximal flysch sequence, Upper Cambrian, southwest Tasmania: Sedimentology, v. 19, p. 99-114.

Cortes, R., 1964, Estratigrafia y un estudio de paleocorrientes del flysch Cretaceo del Depto. de Ultima Esperanza, Tesis de grado, Universidad Tecnico del Estado, Santiago, Chile, 117 p.

Dalziel, I.W.D., 1981, Back-arc extension in the southern Andes: A review and critical reappraisal: Royal Society of London Philosophical Transactions, ser. A, v. 300, p. 319-335.

——— 1986, Collision and cordilleran orogenesis: An Andean perspective, *in* Coward, M. P., and Ries, A. C., eds., Collision tectonics: Geological Society of London Special Publication 19, p. 389-404.

Dalziel, I.W.D., and Brown, R. L., 1989, Tectonic denudation of the Darwin metamorphic core complex in the Andes of Tierra del Fuego, southernmost Chile: Implications for Cordilleran orogenesis: Geology, v. 17, p. 699-703.

Dalziel, I.W.D., de Wit, M. J., and Palmer, K. F., 1974a, Fossil marginal basin in the southern Andes: Nature, v. 250, p. 291-294.

Dalziel, I.W.D., Caminos, R., Palmer, K. F., Nullo, F., and Casanova, R., 1974b, South extremity of Andes: Geology of Isla de los Estados, Argentine Tierra del Fuego: American Association of Petroleum Geologists Bulletin, v. 58, p. 2502-2512.

Dalziel, I.W.D., Storey, B. C., Garrett, S. W., Grunow, A. M., Herrod, L.D.B., and Pankhurst, R. J., 1987, Extensional tectonics and the fragmentation of Gondwanaland, *in* Coward, M. P., Dewey, J. F., and Hancock, P. L., eds., Continental extensional tectonics: Geological Society of London Special Publication 28, p. 433-441.

Dewey, J. F., 1982, Plate tectonics and the evolution of the British Isles: Geological Society of London Journal, v. 139, p. 371-412.

de Wit, M. J., 1977, The evolution of the Scotia Arc as a key to the reconstruction of southwestern Gondwanaland: Tectonophysics, v. 37, p. 53-81.

de Wit, M. J., and Stern, C. R., 1981, Variations in the degree of crustal extension during formation of a back-arc basin: Tectonophysics, v. 72, p. 229-260.

Dott, R. H., Jr., and Bird, K. J., 1979, Sand transport through channels across an Eocene shelf and slope in southwestern Oregon, *in* Doyle, L. J., and Pilkey, O. H., eds., Geology of continental slopes: Society of Economic Paleontologists and Mineralogists Special Publication No. 27, p. 327-342.

Dott, R. H., Jr., Winn, R. D., Jr., and Smith, C.H.L., 1982, Relationship of late Mesozoic and early Cenozoic sedimentation to the tectonic evolution of the southernmost Andes and Scotia Arc, *in* Craddock, C., ed., Antarctic geoscience: Madison, Wisconsin, University of Wisconsin Press, p. 193-203.

Ekdale, A. A., Bromley, R. G., and Pemberton, S. G., 1984, Ichnology: The use of trace fossils in sedimentology and stratigraphy: Society of Economic Paleontologists and Mineralogists, Short Course Notes No. 15, 317 p.

ENAP, 1978, Mapa geologico de la XII Region, Magallanes y Antarctica Chilena, Chile, 1:500,000: Empresa Nacional del Petroleo, Depto. de Exploraciones, Santiago, Chile.

Feruglio, E., 1949, Descripcion geologica de la Patagonia: Dirreccion Generale Yacimientos Petroliferas Fiscales, Buenos Aires, Volumes I-III, 334 p.

Forsythe, R., 1982, The late Paleozoic to early Mesozoic evolution of southern South America: A plate tectonic interpretation: Geological Society of London Journal, v. 139, p. 671-682.

Forsythe, R. D., and Allen, R. B., 1980, The basement rocks of Peninsula Staines, Region XII, Province of Ultima Esperanza, Chile: Revista Geologica de Chile, No. 10, p. 3-15.

Fuenzalida, R., 1984, Nuevas ideas sobre la geologia de los Andes Magallanicos: Servicio Nacional de Geologia y Mineria—Chile, Miscelanea No. 4, p. Q1-9.

Fuenzalida, R., and Covacevich, V. C., 1988, Volcanismo y bioestratigrafia del Jurasico Superior y Cretacico Inferior en la Cordillera Patagonica, Region de Magallanes, Chile: Congreso Geologico Chileno, 5th, Tomo III, p. 159-183.

Gonzalez, E., 1965, La cuenca petrolifera de Magallanes: Revista Minerales, v. 91, p. 1-20.

Gust, D. A., Biddle, K. T., Phelps, D. W., and Uliana, M. A., 1985, Associated Middle to Late Jurassic volcanism and extension in southern South America: Tectonophysics, v. 116, p. 223-253.

Halle, T. G., 1913, Some Mesozoic plant-bearing deposits in Patagonia and Tierra del Fuego and their floras: Stockholm, Sweden, Svenska vetenskap-akademien, Handlingar 51, p. 3-58.

Hanson, R. E., 1987, Facies of a Jurassic submarine rhyolitic volcano-tectonic rift zone, southern Andes, Chile: Geological Society of America Abstracts with Programs, v. 19, no. 7, p. 693.

——— 1989, Paleovolcanology of Jurassic extension-related submarine rhyolites on Staten Island, southernmost Argentina: Geological Society of America Abstracts with Programs, v. 21, p. A14.

Hanson, R. E., and Wilson, T. J., 1986, Large-scale rhyolite peperites in a Jurassic submarine volcano-tectonic rift setting, southern Andes, Chile: Geological Society of America Abstracts with Programs, v. 18, no. 6, p. 791.

——— 1990, Submarine rhyolitic volcanism in a Jurassic proto-marginal basin, southern Andes, Chile and Argentina, *in* Harmon, R. S., and Rapela, C., eds., Andean magmatism and its tectonic setting: Geological Society of America Special Paper (in press).

Hauthal, R., 1907, Croquis geologico de la region entre el Lago Argentino y el Seno Ultima Esperanza, 1:500,000. Naturforschenden Gesellschaft Freiburg, v. 15, Pl. 1.

Hiscott, R. N., 1980, Depositional framework of sandy mid-fan complexes of Tourelle Formation, Ordovician, Quebec: American Association of Petroleum Geologists Bulletin, v. 64, p. 1052-1077.

Hiscott, R. N., and Middleton, G. V., 1979, Depositional mechanics of thick-bedded sandstones at the base of a submarine slope, Tourelle Formation (Lower Ordovician), Quebec, Canada, *in* Doyle L. J., and Pilkey, O. H., eds., Geology of continental slopes: Society of Economic Paleontologists and Mineralogists Special Publication No. 27, p. 307-326.

Hiscott, R. N., Pickering, K. T., and Beeden, D. R., 1986, Progressive filling of a confined Middle Ordovician foreland basin associated with the Taconic Orogeny, Quebec, Canada, *in* Allen, P. A., and Homewood, P., eds., Foreland basins: International Association of Sedimentologists Special Publication No. 8, p. 309-326.

Howell, D. G., and Normark, W. R., 1982, Sedimentology of submarine fans, *in* Scholle, P. A., ed., Sandstone depositional environments: American Association of Petroleum Geologists Memoir 31, p. 365-380.

Johnson, C. N., 1987, Geologia de la precordillera al sur del Seno Almirantazgo, Tierra del Fuego [Taller de Titulo II]: Santiago, Chile, Universidad de Chile, 74 p.

Katz, H. R., 1962, Fracture patterns and structural history in the sub-Andean belt of southernmost Chile: American Association of Petroleum Geologists Bulletin, v. 47, p. 595-603.

——— 1963, Revision of Cretaceous stratigraphy in Patagonian Cordillera of Ultima Esperanza, Magallanes Province, Chile: American Association of Petroleum Geologists Bulletin, v. 47, p. 506-524.

——— 1964, Some new concepts on geosynclinal development and mountain building at the southern end of South America: International Geological Congress, 22nd, Proceedings, Part IV, p. 241-255.

——— 1973, Contrasts in tectonic evolution of orogenic belts in the southeast Pacific: Royal Society of New Zealand Journal, v. 3, p. 333-362.

Kay, S. M., Ramos, V. A., Mpodozis, C., and Sruoga, P., 1989, Late Paleozoic to Jurassic silicic magmatism at the Gondwana margin: Analogy to the Middle Proterozoic in North America?: Geology, v. 17, p. 324-328.

Link, M. H., and Nilsen, T. H., 1980, The Rocks Sandstone, an Eocene sand-rich deep-sea fan deposit, northern Santa Lucia Range, California: Journal of Sedimentary Petrology, v. 50, p. 583-601.

Lowe, D. R., 1982, Sediment gravity flows: II. Depositional models with special reference to the deposits of high-density turbidity currents: Journal of Sedimentary Petrology, v. 52, p. 279-297.

Macellari, C. E., Barrio, C. A., and Manassero, M. J., 1989, Upper Cretaceous to Paleocene depositional sequences and sandstone petrography of southwestern Patagonia (Argentina and Chile): Journal of South American Earth Sciences, v. 2(3), p. 223-239.

Malumian, N., Nullo, F. E., and Ramos, V. A., 1983, The Cretaceous of Argentina, Chile, Paraguay and Uruguay, *in* Moullade, M., and Nairn, A.E.M., eds., The Phanerozoic geology of the world, Volume II. The Mesozoic, B: Amsterdam, The Netherlands, Elsevier, p. 265-304.

Manassero, M. J., and Macellari, C. E., 1987, Late Cretaceous sandstone petrofacies of the Patagonian Cordillera between Lago Argentino and Rio Turbio (Argentina): Geological Society of America Abstracts with Programs, v. 19, no. 7, p. 757-758.

Middleton, G. V., and Hampton, M. A., 1973, Sediment gravity flows: Mechanics of flow and deposition, *in* Middleton, G. V., and Bouma, A. H., eds., Turbidites and deep sea sedimentation: Society of Economic Paleontologists and Mineralogists, Pacific Section, Short Course, p. 1-38.

——— 1976, Subaqueous sediment transport and deposition by sediment gravity flows, *in* Stanley, D. J., and Swift, D.J.P., eds., Marine sediment transport and environmental management: New York, Wiley, p. 197-218.

Mutti, E., and Ricci Lucchi, R., 1972, Le torbiditi dell'Apennino settentrionale: Introduzione all'analysi di facies: Societa Geologica Italiano Memoir No. 11, p. 161-199. English translation by T. H. Nilsen, 1978, International Geology Review, v. 20, p. 125-166, American Geological Institute Reprint Series No. 3.

Nardin, T. R., Hein, F. J., Gorsline, D. S., and Edwards, B. D., 1979, A review of mass movement processes, sediment and acoustic characteristics, and contrasts in slope and base-of-slope systems versus canyon-fan-basin floor systems, *in* Doyle, L. J., and Pilkey, O. H., eds., Geology of continental slopes: Society of Economic Paleontologists and Mineralogists Special Publication No. 27, p. 61-73.

Natland, M. L., Gonzalez, E., Canon, A., and Ernst, M., 1974, A system of stages for correlation of Magallanes basin sediments: Geological Society of America Memoir 139, 126 p.

Nelson, E., Dalziel, I.W.D., and Milnes, A. G., 1980, Structural geology of the Cordillera Darwin—Collisional style orogenesis in the southernmost Chilean Andes: Eclogae Geologicae Helvetiae, v. 73, p. 727-751.

Nordenskjold, O., 1899, Geological map of the Magellan Territories, with explanatory notes: Wissenschaftliche Ergebnisse der Schwedischen Expedition nach den Magellanslander 1895-1897, Volume I, Stockholm, p. 81-85.

Nullo, F. E., Proserpio, C., and Ramos, V. A., 1978, Estratigrafia y tectonica de la vertiente este del Hielo Continental Patagonico, Argentina-Chile: VII Congreso Geologico Argentino, Actas I, p. 455-470.

Nullo, F. E., Proserpio, C. A., and Blasco, G., 1981, El Cretacico de la Cuenca Austral entre el Lago San Martin y Rio Turbio, *in* Volkheimer, G., ed., Cuencas Sedimentarias del Jurasico y Cretacico de America del Sur, Volume 1: Buenos Aires, Argentina, Comité Sudamericano del Jurásico y Cretácico, p. 181-220.

Pickering, K. T., 1982a, A Precambrian upper basin-slope and prodelta in northeast Finnmark, north Norway—A possible ancient upper continental slope: Journal of Sedimentary Petrology, v. 52, p. 171-186.

——— 1982b, The shape of deep-water siliciclastic systems: A discussion: Geo-Marine Letters, v. 2, p. 41-46.

Quensel, P. D., 1911, Geologisch-petrographische Studien in der patagonischen Cordillera: Geological Institutionen University of Upsala, Bulletin 9, p. 1-113.

Quinlan, G. M., and Beaumont, C., 1984, Appalachian thrusting, lithospheric flexure and the Paleozoic stratigraphy of the eastern interior of North America: Canadian Journal of Earth Sciences, v. 21, p. 973-996.

Ramos, V. A., 1988, Late Proterozoic-early Paleozoic of South America—A collisional history: Episodes, v. 11, p. 168-174.

Riccardi, A. C., 1983, The Jurassic of Chile and Argentina, *in* Moullade, M., and Nairn, A.E.M., eds., The Phanerozoic geology of the world, Volume II. The Mesozoic, B: Amsterdam, The Netherlands, Elsevier, p. 201-263.

——— 1987, Cretaceous paleogeography of southern South America: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 59, p. 169-195.

——— 1988, The Cretaceous System of southern South America: Geological Society of America Memoir 168, 161 p.

Riccardi, A. C., and Rolleri, E. O., 1980, Cordillera Patagonica austral: Segundo Simposio Geologico Regional Argentino, Academia Nacional de Ciencias, Cordoba, Volume 2, p. 1163-1306.

Riggi, J. C., 1969, Rasgos sedimentologicos, paleogeograficos y paleoclimaticos de la seccion inferior de la Cuenca Austral en la zona comprendida entre los Pozos SC.1-SC.EM.1-SC.CD.1 (Santa Cruz): Actas Cuartas Jornadas Geologicas Argentinas, v. 1, p. 509-517.

——— 1977, La Formacion Springhill en el subsuelo de Tierra del Fuego, Asociacion Geologica Argentina Revista, v. 32, p. 176-189.

Russo, A., Flores, M. A., and Di Benedetto, H., 1980, Patagonia austral extrandina: Segundo Simposio Geologico Regional Argentino, Academia Nacional de Ciencias, Cordoba, Volume 2, p. 1430-1462.

Scott, K. M., 1966, Sedimentology and dispersal pattern of a Cretaceous flysch sequence, Patagonian Andes, southern Chile: American Association of Petroleum Geologists Bulletin, v. 50, p. 72-107.

Seilacher, A., 1967, Bathymetry of trace fossils: Marine Geology, v. 5, p. 413-428.

Shanmugam, G., and Moila, R., 1988, Submarine fans: Earth Science Reviews, v. 24, p. 383-428.

Skarmeta, J., 1976, Evolucion tectonica y paleogeografica de los Andes Patagonicos de Aisen (Chile) durante el Neocomiano: Primero Congreso Geologico Chileno, Actas I, p. B1-B15.

Smith, C.H.L., 1977, Sedimentology of the Late Cretaceous (Santonian-Maestrichtian) Tres Pasos Formation, Ultima Esperanza District, southern Chile [M.S. thesis]: Madison, Wisconsin, University of Wisconsin, 129 p.

Suarez, M., 1979, A late Mesozoic island arc in the southern Andes, Chile: Geological Magazine, v. 116, p. 181-190.

Suarez, M., and Pettigrew, T. H., 1976, An upper Mesozoic island-arc-back-arc system in the southern Andes and South Georgia: Geological Magazine, v. 113, p. 305-328.

Theide, J., and Dinkelman, M. G., 1977, The occurrence of *Inoceramus* remains in late Mesozoic pelagic and hemipelagic sediments, *in* Supko, P. R., and Perch-Nielsen, K., and others, Initial reports of the Deep-Sea Drilling Project, Volume 39: Washington, D.C., U.S. Government Printing Office, p. 899-910.

Vilela, C. R., and Csaky, A., 1968, Las turbiditas en los sedimentos Cretacicos de la region de Lago Argentino (Provincia de Santa Cruz): Tercera Jornada Geologica Argentina, v. I, p. 209-225.

Walker, R. G., 1976, Facies models. 2. Turbidites and associated coarse clastic deposits: Geoscience Canada, v. 3, p. 25-36.

White, N., and McKenzie, D., 1988, Formation of the "steer's head" geometry of sedimentary basins by differential stretching of the crust and mantle: Geology, v. 16, p. 250-253.

Wilckens, O., 1907, Erlauterungen zu R. Hauthals Geologischer Skizze des Gebietes zwischen dem Lago Argentino und dem Seno de la Ultima Esperanza (Sud-patagonien): Naturforschenden Gesellschaft Freiburg, v. 15, p. 75-96.

Wilson, T. J., 1983, Stratigraphic and structural evolution of the Ultima Esperanza foreland fold-thrust belt, Patagonian Andes, southern Chile [Ph.D. thesis]: New York, Columbia University, 360 p.

Winn, R. D., Jr., 1978, Upper Mesozoic flysch of Tierra del Fuego and South Georgia Island: A sedimentologic approach to lithosphere plate restoration: Geological Society of America Bulletin, v. 89, p. 533-547.

Winn, R. D., Jr., and Dott, R. H., Jr., 1979, Deep-water fan-channel conglomerates of Late Cretaceous age, southern Chile: Sedimentology, v. 26, p. 203-228.

Winslow, M. A., 1980, Mesozoic and Cenozoic tectonics of the fold and thrust belt in southernmost South America and stratigraphic history of the cordilleran margin of the Magallanes basin [Ph.D. thesis]: New York, Columbia University.

——— 1982, The structural evolution of the Magallanes Basin and neotectonics in the southernmost Andes, *in* Craddock, C., ed., Antarctic Geoscience: Madison, Wisconsin, University of Wisconsin, p. 143-154.

Woodcock, N. H., 1979, The use of slump structures as paleoslope orientation estimators: Sedimentology, v. 26, p. 83-101.

Zeil, W., 1958, Sedimentation in der Magallanes Geosynklinale mit besonderer Berucksichtigung des Flysch: Geologisches Rundschau, v. 47, p. 425-433.

Manuscript Received by the Society July 17, 1989
Revised Manuscript Received April 26, 1990
Manuscript Accepted May 3, 1990