The palaeogeography of the Lower Cretaceous Aysén Basin of southern Chile

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Abstract

The Aysén Basin of southern Chile is an elongate sedimentary basin situated to the east of the Patagonian batholith in an ensialic back-arc environment. It contains marine sediments of the Coyhaique Group which provide an important stratigraphic marker between similar volcanic successions. The Coyhaique Group comprises, limestones (Toqui Formation), black shales (Katterfeld Formation) and sandstones (Apeleg Formation), which outcrop over an area of approximately 150km N/S and 75km E/W with an estimated thickness of more than 1200m.

The Toqui Formation limestones are discontinuous across the basin and were laid down in a marine transgression which occurred during the Tithonian to Berriasian. In the NE, oyster banks indicate a NW/SE aligned shoreline which probably marked the margin of the basin. By the beginning of the Hauterivian, black shales of the Katterfeld Formation had become regionally extensive. They represent a restricted basin with fluctuating levels of oxygen which controlled the faunal communities. Fossil assemblages indicate a connection to the more southerly basins and a clear link to South African fauna. The influx of the higher energy sandstones of the Apeleg Formation during the late Hauterivian marked the opening of the basin from the south and east to normal marine conditions. Trace fossils indicate a shallow marine setting. This formation probably represents tidal sandstones deposited in a shelf environment.

The sediments of the Aysén Basin reflect gentle subsidence of the back-arc basin, probably due to thermal relaxation related to a period of volcanic quiescence, from the lower Berriasian until at least the late Hauterivian, possibly early Barremian. Unlike the basins to the north and south, there is no evidence of subduction related rifting during the formation of the basin.

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Through this project I have met many wonderful people from this country and abroad who became friends as well as colleagues. This work was made possible with funding from a PhD research studentship awarded by Cheltenham and Gloucester College of Higher Education.

DECLARATION

This thesis is a product of my own work and
the views expressed in the dissertation
are my own and not those of the University

Marisia Jean Townsend

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CHAPTER 1

INTRODUCTION

1.1 Aims of the thesis

The aim of this thesis is to produce a palaeogeographical reconstruction of the central part of the Aysén Basin of southern Chile during the Lower Cretaceous using palaeoecological and palaeoenvironmental data.

1.2 Background to the thesis

1.2.1 Location

The Aysén Basin is part of a composite backarc basin which extends throughout Region XI of Chile, from 43°S to 47°S, and east of the Patagonian batholith, into Argentina. It is poorly defined in previous literature compared to adjacent basins. Riccardi (1988) showed the area as a northwesterly extension of the Magallanes Basin which linked into the San Jorge Basin (Figure 1.1). At the same location, Ramos (1989), identified a distinct sedimentary basin which he named the Rio Mayo Embayment. The basin opened towards the Pacific Ocean in the west and was isolated from the Magallanes Basin and the San Jorge Basin (Figure 1.2). The uncertainty of the boundaries of the Aysén Basin is due to the lack of investigation of this area and the sporadic and discontinuous nature of the exposures. However, the Aysén Basin forms a distinct unit of volcanic and sedimentary strata bounded by mountainous areas. The western border is formed by the granitic rocks of the Patagonian batholith, whilst to the north and east is the Somuncurá Massif. The southeastern boundary is marked by the Deseado Massif (Figure 1.1).

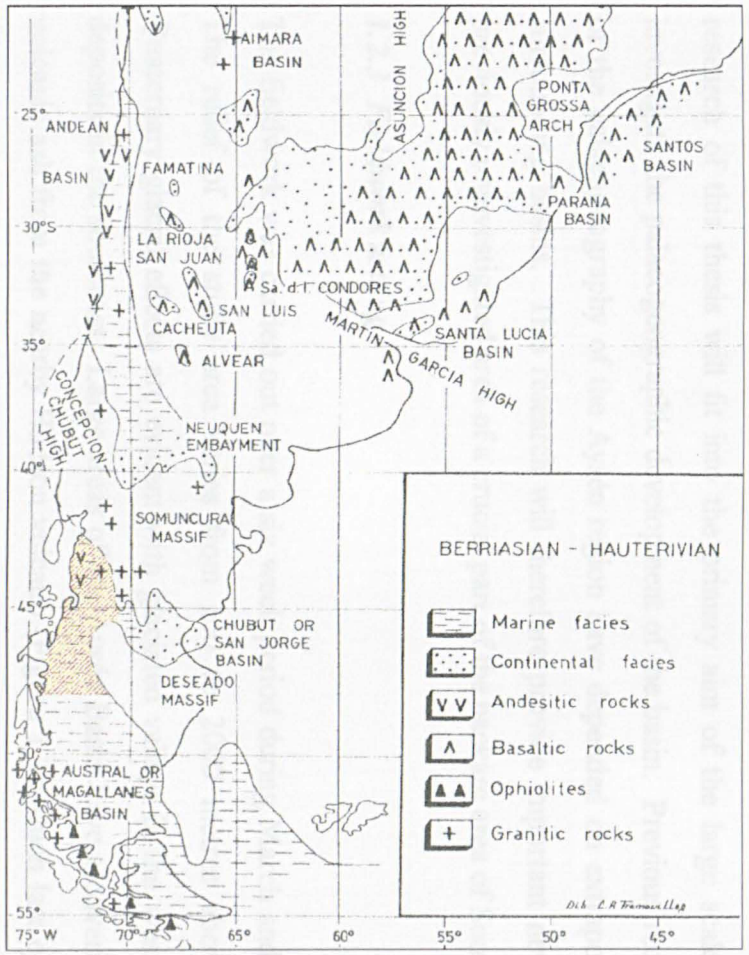


Figure 1.1: Riccardi's (1988) palaeogeographic map of southern South America during Berrriasian-Hauterivian time, indicating the major basins. The Aysén Basin (orange) is tentatively depicted as an extension of the Magallanes Basin which linked into the San Jorge Basin.

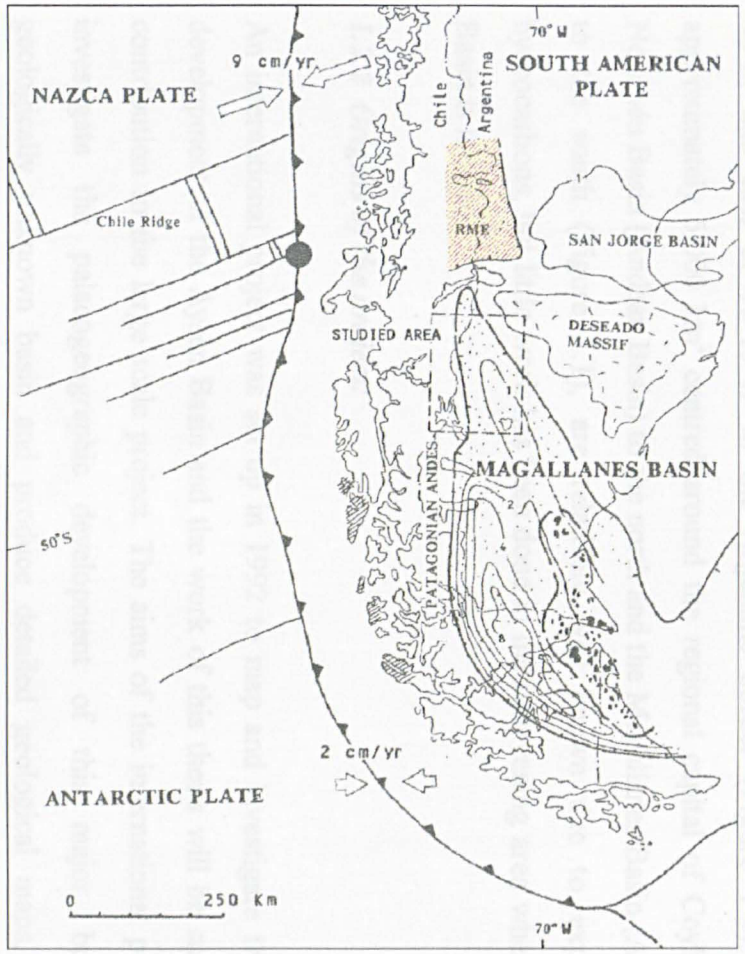


Figure 1.2: Ramos' (1989) map of southern South America showing the Aysén Basin (orange), which he named the Rio Mayo Embayment (RME), as an independent basin opening towards the west. The solid circle is the Aysén Triple Junction. Dashed lines enclose the Springhill platform, the main oil producing area of southern South America.

The area of the present study covers the central part of the Aysén Basin from 45°S to 46°S, and east of 73°15'W to the Argentine border (Figure 1.3), an area of approximately 5,000 km^2 centred around the regional capital of Coyhaique. The Neuquén Basin (Andean Basin) to the north and the Magallanes Basin (Austral Basin) to the south (Figure 1.1), are relatively well known due to exploration for hydrocarbons, but little work has been done in the intervening area where the Aysén Basin is located.

1.2.2 Origins of the project:

An international project was set up in 1992 to map and investigate the geological development of the Aysén Basin and the work of this thesis will be submitted as a contribution to the large scale project. The aims of the international project are to investigate the palaeogeographic development of this major, but relatively geologically unknown basin and produce detailed geological maps, whilst also assessing the economic geology of the area. Geologists from Britain, France, New Zealand, Argentina and the USA are involved. A preliminary inspection of the thesis research area by the principal investigators of the project during the 1993 field season, identified a theme for further research which has resulted in the present study. The research of this thesis will fit into the primary aim of the large scale project, to investigate the palaeogeographic development of the basin. Previous reconstructions of the palaeogeography of the Aysén region have depended on extrapolations from surrounding basins. This research will therefore provide important new data on a previously uninvestigated area of a crucial part of the backarc area of South America.

1.2.3 Fieldwork setting

The fieldwork was carried out over a six week period during March and April 1994. The relief of the study area varies from 100 to 2000 metres above sea-level. Quaternary glacial effects are evident with glaciated valleys in the west and glacial deposits in the southeast. Large areas of the Aysén Basin have a covering of recent volcanic ash from the nearby Hudson volcano (Figure 1.3) which last ejected ash in August 1991.

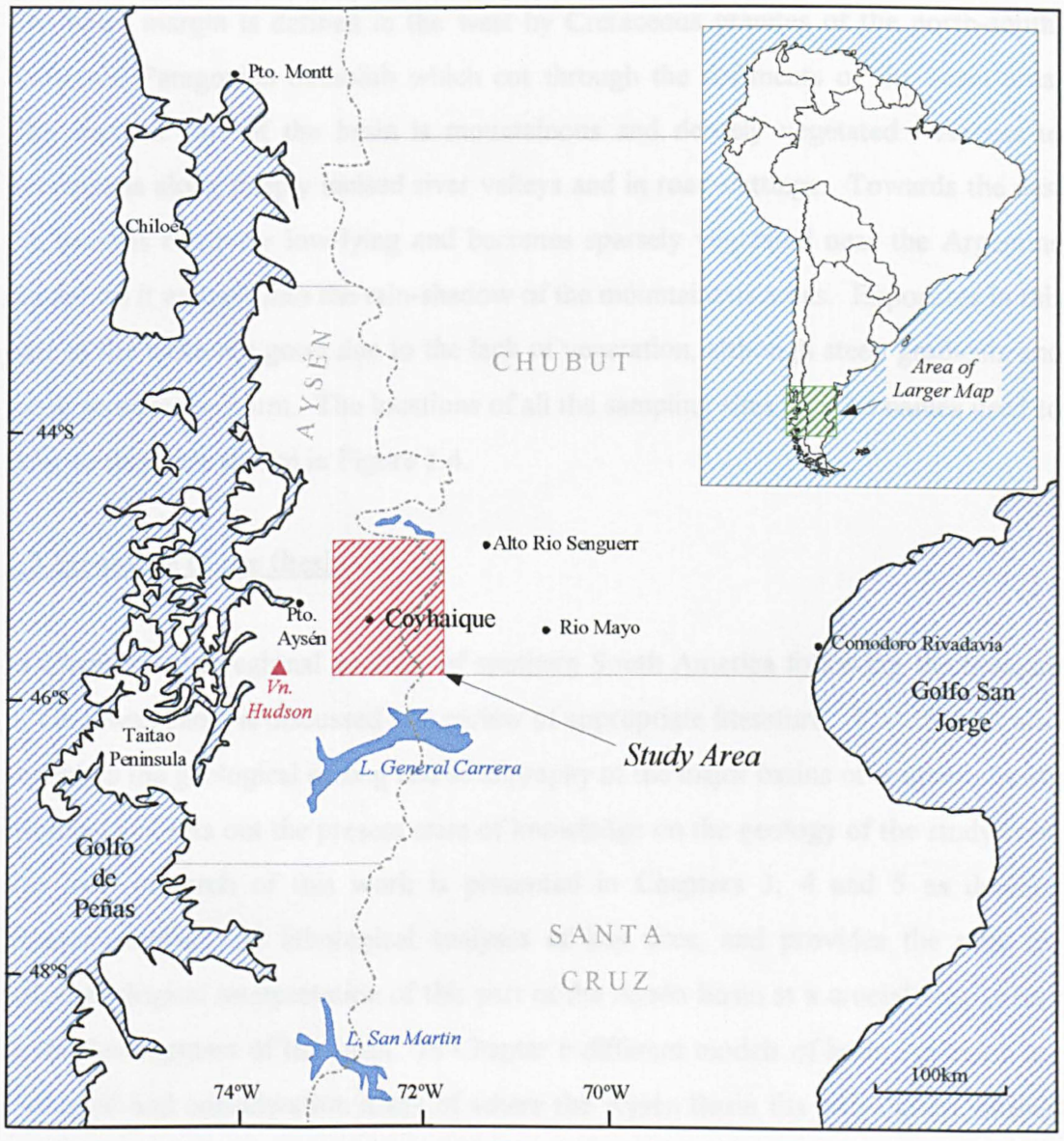


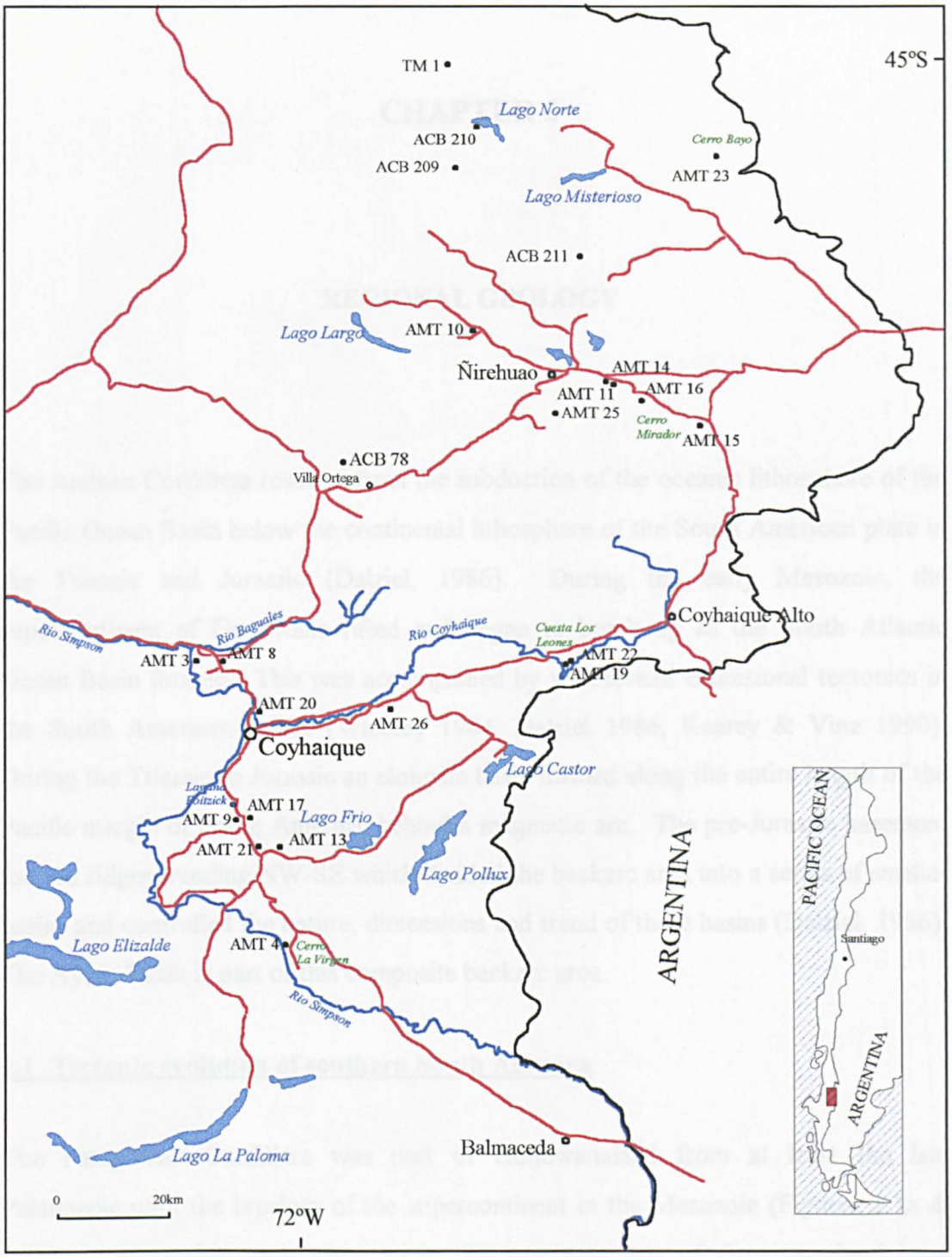
Figure 1.3: Map of northern Patagonia showing the study area in relation to the Aysén region and areas of Argentina named in the thesis.

The basin margin is defined in the west by Cretaceous granites of the north-south, elongated Patagonian batholith which cut through the sediments of the basin area. The western part of the basin is mountainous and densely vegetated. Access to outcrops is along deeply incised river valleys and in road-cuttings. Towards the east the basin is relatively low-lying and becomes sparsely vegetated near the Argentine border, as it extends into the rain-shadow of the mountainous areas. Exposures in this part of the basin are good due to the lack of vegetation, although steep gradients and loose scree is the norm. The locations of all the sampling sites and the names used to refer to them are shown in Figure 1.4.

1.3 Structure of the thesis

In Chapter 2 the regional geology of southern South America following the breakup of Gondwanaland, is discussed in a review of appropriate literature. This chapter also considers the geological setting and stratigraphy of the major basins of southern South America and sets out the present state of knowledge on the geology of the study area. The new research of this work is presented in Chapters 3, 4 and 5 as detailed palaeontological and lithological analyses of key sites, and provides the data for palaeoecological interpretation of this part of the Aysén Basin at a crucial time period in the development of the basin. In Chapter 6 different models of basin evolution are evaluated and consideration made of where the Aysén Basin fits into the established models. The palaeoenvironmental evidence of the preceding work is used in Chapter 7 to present an interpretation of the palaeogeography of the Aysén Basin during the Lower Cretaceous.

The project was made possible by funding and technical support from the Department of Geography and Geology, Cheltenham and Gloucester College of Higher Education, whilst the Chilean Geological Survey financed the fieldwork and all logistical support in Chile as part of the Aysén Geology Project.



AMT 3	Los Baguales	AMT 15	Cerro Mirador	AMT 25	El Tranquilo
AMT 4	Río Simpson	AMT 16	Coichel	AMT 26	Galpos
AMT 8	Baguales Vista	AMT 17	La Rosita	ACB 78	Villa Ortega
AMT 9	Foitzick	AMT 19	Cuesta Los Leones	ACB 209	El Gato
AMT 10	Estero Pedregoso	AMT 20	El Rincon	ACB 210	Lago Norte
AMT 11	Baño Nuevo	AMT 21	El Salto	ACB 211	Los Mallines
AMT 13	La Fraille	AMT 22	Coyhaique Alto	TM 1	Toqui Mine
AMT 14	Baño Nuevo (W)	AMT 23	Cerro Bayo		

Figure 1.4: Location of sampling sites in the Aysén Basin

CHAPTER 2

REGIONAL GEOLOGY

The Andean Cordillera resulted from the subduction of the oceanic lithosphere of the Pacific Ocean Basin below the continental lithosphere of the South American plate in the Triassic and Jurassic (Dalziel, 1986). During the early Mesozoic, the supercontinent of Gondwana rifted and began to break up as the South Atlantic Ocean Basin formed. This was accompanied by widespread extensional tectonics in the South American sector (Windley 1984, Dalziel 1986, Kearey & Vine 1990). During the Triassic to Jurassic an elongate basin formed along the entire length of the Pacific margin of South America, behind a magmatic arc. The pre-Jurassic basement formed ridges trending NW-SE which divided the backarc area into a series of smaller basins and controlled the nature, dimensions and trend of these basins (Dalziel, 1986). The Aysén Basin is part of this composite backarc area.

2.1 Tectonic evolution of southern South America

The Patagonian Cordillera was part of Gondwanaland from at least the late Palaeozoic until the breakup of the supercontinent in the Mesozoic (Figures 2.1a & 2.1b). Events that took place while the area was part of Gondwanaland have influenced the evolution of northern Patagonia (Uliana & Biddle, 1987).

2.1.1 Precambrian to late Palaeozoic basement

The possibility of a Precambrian basement to southern South America was noted by Dalziel & Elliot (1971). Ramos (1988) proposed that the major part of South America was a complex collage of allochthonous, cratonic blocks brought together by

Figure 2.1a: Palaeocoastline map showing South America as part of Gondwanaland during the Triassic when material was accreting to the continental landmass and before the initiation of the supercontinent breakup. (Smith *et al.* 1994).

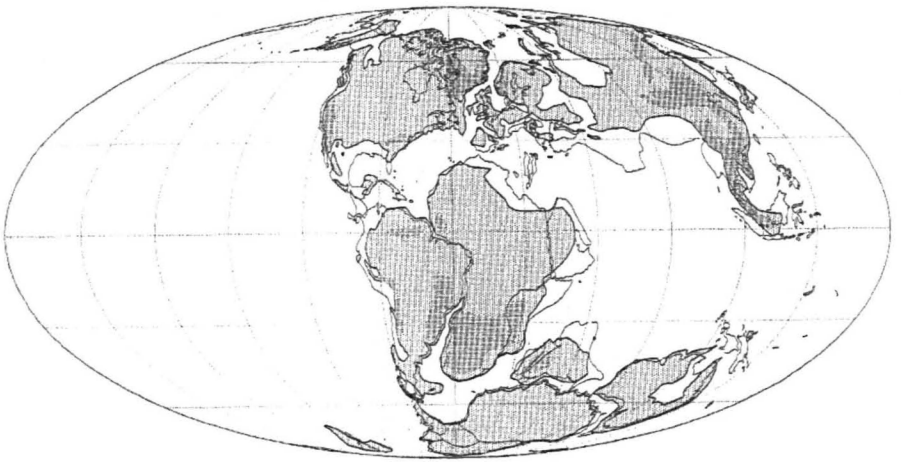


Figure 2.1b: Palaeocoastline map for Barremian - Hauterivian. Extension has resulted in the opening of the South Atlantic Ocean and a number of NW trending grabens which have become marine near the west coast of South America (Smith *et al.* 1994).

collisions and amalgamations from late Proterozoic to early Palaeozoic times. This suggestion has recently been re-established by Tankard *et al.* (1995) who state that the continental lithosphere of southwest Gondwanaland was largely formed before the end of the Proterozoic and is composed of a series of fault-bounded terranes which accumulated around the Archaean cratons. The evidence for this theory refers mainly to blocks and cratons in central and eastern Argentina however, Ramos (1988) reports that ages of gneiss complexes south of Lago General Carrera and northwest of the Somuncurá Massif (Figures 1.3 & 2.2) indicate that this area is also composed of amalgamated terranes. Basement terranes have been documented by Forsythe & Mpodozis (1983) on the archipelago at 50°S, near the southwest tip of South America, and it has been suggested that the docking of these allochthonous terranes contributed to the Gondwanidian orogenic deformational event (Ramos, 1989). Forsythe (1982) and Dalziel (1986) consider this to be unlikely and suggest that the entire Palaeozoic basement represents an unusually wide convergent margin. A recent palaeomagnetic study by Rapalini (1998), on Silurian - Devonian clastic sediments exposed in northeast Patagonia, disproves Ramos' (1988) theory involving collision of allochthonous terranes during the late Precambrian and shows that the main tectonic events that affected northern Patagonia agree with those of Gondwanaland. The magnetization results indicate an Early to Late Permian age for the main tectonic events which folded the sediments of northern Patagonia. Palaeozoic magnetic poles from Patagonia agree with those from Gondwanaland since Devonian times, suggesting that Patagonia did not undergo important displacements relative to South America since those times and was part of Gondwanaland since the early Palaeozoic (Rapalini, 1998).

In the late Palaeozoic, subduction of the Pacific Ocean Basin below the Gondwana supercontinent created a broad accretionary wedge, as the sediments scraped off the downgoing plate and piled up against the continental margin (Table 2.1).

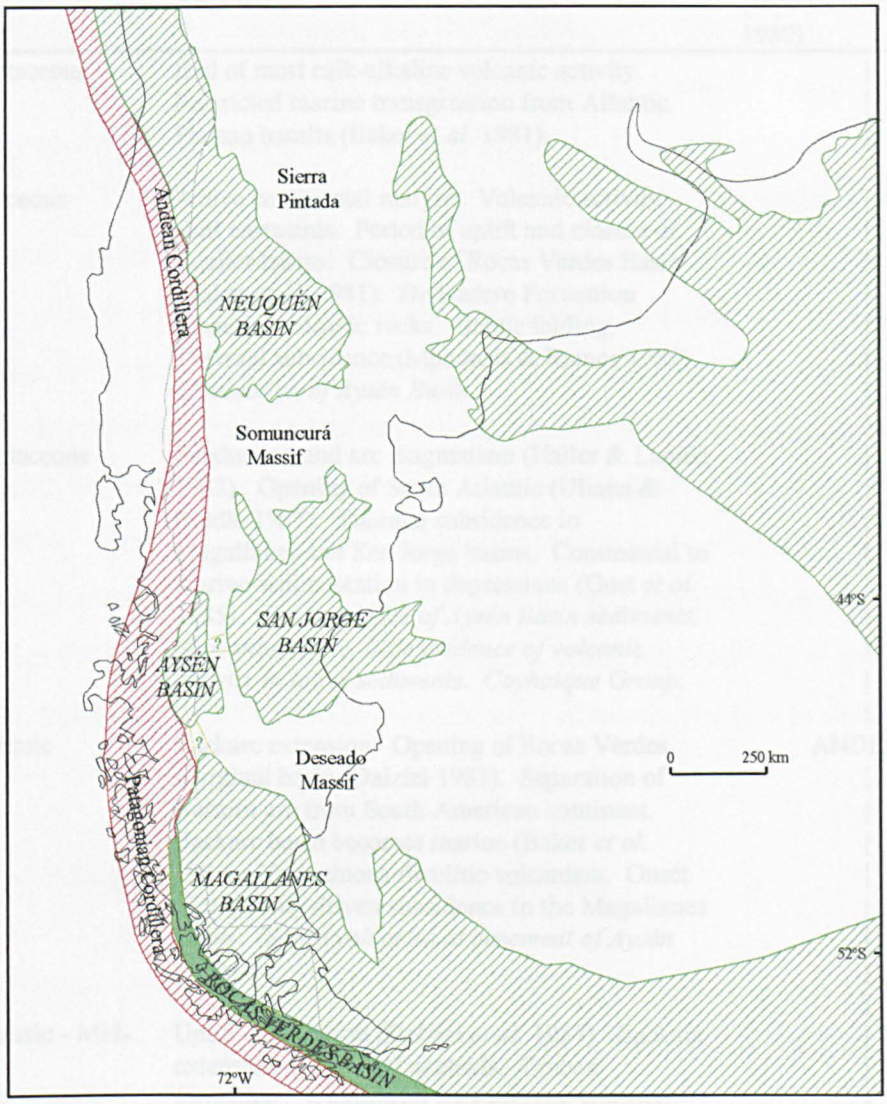


Figure 2.2: Early Mesozoic major sedimentary basins of southern South America and other areas named in the study. (based on Malumián & Ramos, 1984).

Time	Major tectonic/magmatic events in South America	Tectonic cycle (Mpodozis & Ramos 1989)
Upper Cretaceous	End of most calk-alkaline volcanic activity. Restricted marine transgression from Atlantic. Plateau basalts (Baker *et al.* 1981).	
Mid-Cretaceous	Active continental margin. Volcanic activity shift eastwards. Period of uplift and closure of smaller basins. Closure of Rocas Verdes Basin (Baker *et al.* 1981). Divisadero Formation subaerial volcanic rocks. Gentle folding. Thermal subsidence (Mpodozis & Ramos 1989). *Elimination of Aysén Basin.*	
Lower Cretaceous	Subduction and arc magmatism (Haller & Lapido 1982). Opening of South Atlantic (Uliana & Biddle 1987). Thermal subsidence in Magallanes and San Jorge basins. Continental to marine sedimentation in depressions (Gust *et al.* 1985). *First evidence of Aysén Basin sediments. Slow subsidence, little evidence of volcanic activity in lower sediments. Coyhaique Group.*	
Upper Jurassic	Backarc extension. Opening of Rocas Verdes marginal basin (Dalziel 1981). Separation of western arc from South American continent. Backarc basin becomes marine (Baker *et al.* 1981). Voluminous rhyolitic volcanism. Onset of thermally driven subsidence in the Magallanes Basin. *Ibañez volcanics at basement of Aysén Basin.*	ANDEAN
Upper Triassic - Mid-Jurassic	Uplift and erosion (Dalziel *et al.* 1987). Backarc extension and crustal anatexis. Graben formation. Widespread mid-Jurassic volcanic event (Gust *et al.* 1985). Initiation of Gondwanaland breakup (Dalziel *et al.* 1987).	
Permian to Mid-Triassic	Subduction, accretion and arc magmatism. Convergent margin system - accretionary prism, magmatic arc, forearc basin (Forsythe 1982). Deformation and low-grade metamorphism Uliana & Biddle 1988).	----------
Carboniferous to Permian	Subduction, accretion and arc magmatism along the western edge of Gondwanaland. Samfrau geosyncline (du Toit 1937).	GONDWANIDIAN

Table 2.1: Summary of major tectonic and magmatic events in southern South America during the Upper Palaeozoic to Mesozoic. Aysén Basin events shown in italics.

Semi-continuous subduction from the late Palaeozoic to the present day resulted in a convergent margin system in which three zones can be recognised east to west; i) a fold and thrust belt which corresponds to the Samfrau geosyncline of du Toit (1937), (Figure 2.3) ii) a magmatic arc iii) a forearc accretionary wedge along the Pacific margin (Forsythe 1982, Dalziel 1986, Ramos 1989).

Deformation associated with the subduction and known as the Gondwanidian Orogeny, is seen in both the forearc and backarc areas (Table 2.1). High-grade deformation accompanied by low-grade metamorphism occurred throughout the thick sediments and was the first sign of tectonic instability after a tectonically quiescent period from the Late Permian to the Mid-Triassic (Dalziel *et al.* 1987, Uliana & Biddle 1988). The Gondwanidian Orogeny was responsible for imposing a structural grain on the basement rocks which influenced the subsequent evolution of the basinal areas. The Palaeozoic basement has an arcuate structural pattern, convex to the southwest, but in the west the trend swings more northwards, parallel to the magmatic arc (Dalziel, 1986).

2.1.2 Mid to Late Jurassic extension and volcanism

Crustal extension began in the Late Triassic and persisted throughout most of the Jurassic (Gust *et al.* 1985, Dalziel *et al.* 1987, Uliana & Biddle 1987). The extension formed a number of NNW trending, fault-bounded, grabens which followed the basement grain. The troughs are mostly half-grabens with the faults on the eastern margins showing the greatest displacement (Gust *et al.*, 1985).

Several major, regional, erosional unconformities separate variably deformed Upper Triassic, Jurassic and Lower Cretaceous strata, from less deformed Upper Cretaceous sequences in southern South America and are well documented (Gust *et al.*, 1985, Dalziel *et al.* 1987, Uliana & Biddle 1988). Although hard to date precisely, the fill of the grabens indicate that uplift and erosion immediately preceded the deposition of the overlying strata. The erosional event occurred in a number of tectonic environments which pre-dated the graben formation and igneous activity associated with the fragmentation of Gondwanaland.

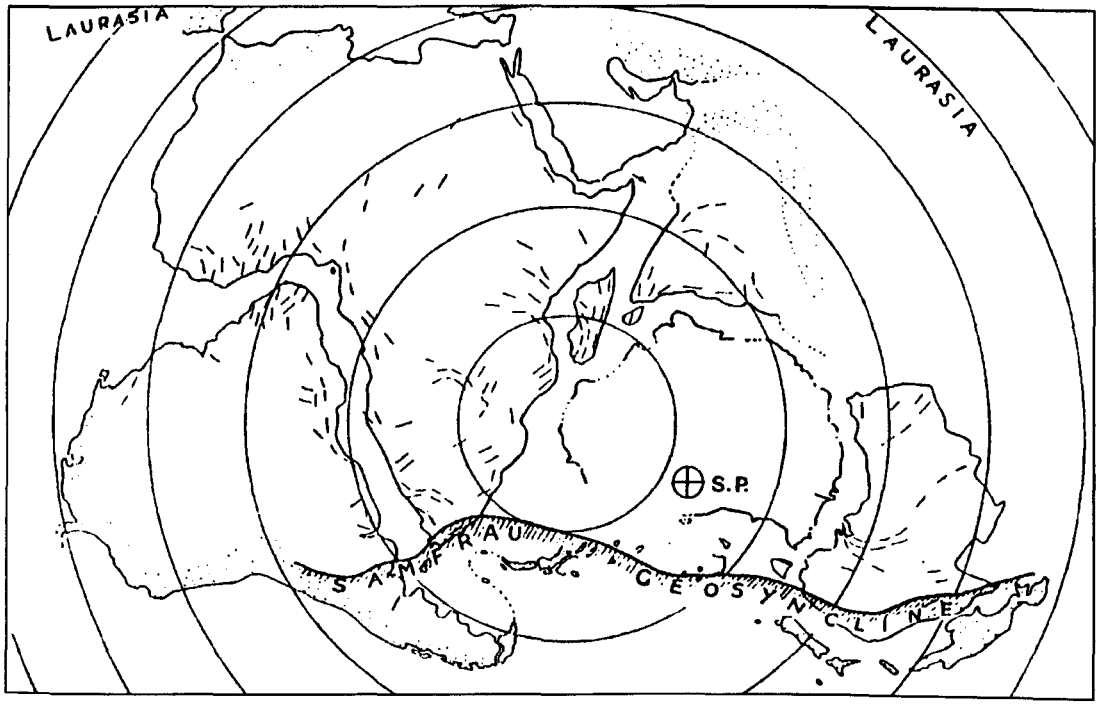


Figure 2.3: The Gondwana supercontinent during the Palaeozoic as depicted by du Toit (1937) showing the Samfrau Geosyncline which du Toit interpreted as the ancestral Pacific margin of Gondwanaland. In the late Palaeozoic, the Samfrau Geosyncline was an epicratonic backarc basin, separated from the Pacific realm by a magmatic arc. The short lines indicate the Precambrian basement grain (du Toit, after Lawver & Scotese 1987).

During the early stages of extension the basins were small and isolated but the active extension and subsidence enlarged the graben system and created a composite, linear backarc system along the entire Pacific margin of South America, approximately parallel to the continental landmass (Uliana & Biddle, 1988). In Patagonia, the influence of the Gondwanidian structural grain, and the fault induced subsidence of the NNW trending grabens, allowed a marine incursion from the west (Figure 2.4a). The amount of extension and subsidence varied, with segments of the backarc basin being separated by basement ridges trending northwest - southeast (Dalziel, 1986). In the far south, the crustal stretching culminated in the opening of the Rocas Verdes marginal basin, part of a two stage development of the Magallanes Basin (also known as the Austral Basin in Argentina) (Ramos, 1989). On the western margin of South America, in northern Patagonia, the Aysén Basin formed, and in central, west South America, the Neuquén Basin (Andean Basin). Inland, the San Jorge Basin (Chubut) formed as one of a series of small continental basins (Macellari, 1988). These are the major basins of southern South America (Figure 2.2) and are discussed individually in section 2.2.

Extension, normal faulting and volcanism commenced some considerable time before the major Mid-Jurassic volcanic event and continued throughout the Jurassic period (Gust *et al.*, 1985). In the Mid-Jurassic, extension was accompanied by intense, silicic to bimodal volcanism, which is evident as far north as 40°S (Figure 2.4b). Marked variations in thickness of the deposits show that the area was undergoing active subsidence and proves that the volcanism and extension occurred contemporaneously (Bartholomew & Tarney, 1984) (Table 2.1). The volcanics are mainly andesitic in the north and rhyolitic in the south. The rhyolites are thought to be products of crustal anatexis of sedimentary source material formed through the input of hot basaltic magma into the base of the crust (Haller & Lapido 1982, Gust *et al.* 1985), although there is also evidence of subduction derived calk-alkaline magmatic products (Uliana & Biddle, 1988). Igneous activity was at a maximum during the Middle Jurassic but had almost ceased by the end of the Jurassic (Uliana & Biddle, 1987).

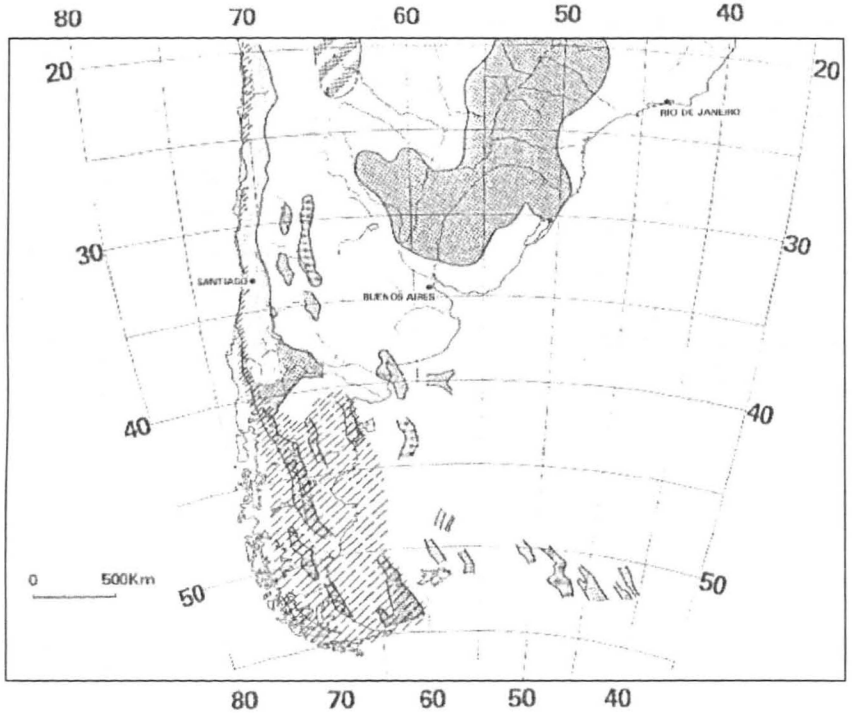


Figure 2.4a: Middle Jurassic (160Ma) palaeogeographic reconstruction of southern South America illustrating Jurassic extension and the creation of NNW trending grabens. *Dark grey areas represent non-marine sedimentation, light grey areas are marine sedimentation and lined areas indicate regions affected by igneous activity* (Uliana & Biddle 1988).

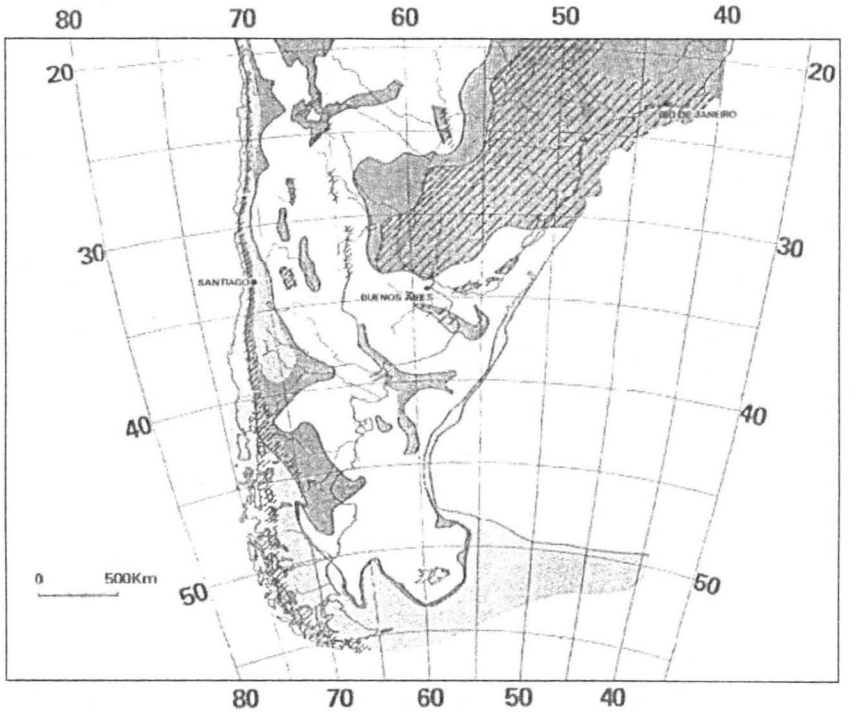


Figure 2.4b: Neocomian palaeogeographic reconstruction of southern South America illustrating Early Cretaceous rifts, sags and magmatic arc. The increase in marine sedimentation due to thermal subsidence and a marine transgression from the west is marked. *(Key as in Figure 2.4a).* (Uliana & Biddle 1988).

2.1.3 Lower Cretaceous subsidence

The extension that began in the Late Triassic and culminated in the opening of the Rocas Verdes Basin in the Late Jurassic, led to the opening of the South Atlantic and the rifting apart of South America and Africa during the Early Cretaceous (Table 2.1). Once sea-floor spreading began in the South Atlantic Ocean, faulting declined in Patagonia and thermally driven subsidence began in the backarc areas. Basins close to the western margin of South America accumulated marine deposits while large areas of the southern South America interior became depocentres for continental clastic materials (Uliana & Biddle, 1987, 1988).

2.1.4 Mid-Cretaceous uplift and closure of the basins

Uplift and deformation of varying intensity occurred in the deposits of the composite backarc trough along the whole length of the Cordillera in the mid-Cretaceous. An overall increase in clastic input was related to an increase in the magnitude of arc-related activity. The trend is thought to reflect a change in the dynamics of subduction beneath southern South America caused by a shift eastwards in volcanic activity (Bartholomew & Tarney 1984, Dalziel 1986, Uliana & Biddle 1988). Synchronous, massive, batholith emplacement together with uplift and a marine retreat resulted in crustal shortening to varying degrees and induced the closure of the Rocas Verdes Basin. The Aysén region was less affected by the shortening event and closure of the basin area was more likely to be due to vertical uplift of the basin or the sediment source area creating an increase in the amount of clastic input to the basin. The Divisadero Formation volcanic rocks (Table 2.1) overlie the marine sediments and mark a change to fluviatile - lacustrine conditions. The Divisadero Formation is almost entirely subaerial and is taken to have developed in the tectonic setting of an active continental margin (Baker *et al.*, 1981).

2.1.5 Tertiary to Recent plateau basalts and triple junction

In the Tertiary, massive outpourings of basalt created the Patagonian plateau basalts found south of 41°S. They were erupted in regular pulses of activity and presently

cover approximately 10% of the area as isolated mesetas (Baker *et al.* 1981, Ramos & Kay 1992). Ramos & Kay (1992) deduce that the plateau basalts are closely related to ridge and trench interactions along the coastal margin in the west, and that the volume of basalt erupted is directly related to the length of ridge segment which collides with the trench.

The area around Coyhaique is the present day site of the Chile Triple Junction, the point of intersection of the Nazca, Antarctic and South American plates. The intersection has gradually migrated northwards at varying rates of between 10mm per year and 160mm per year until its present location at 46°S. However, the current location of the Chile Triple Junction has no effect on the Early Cretaceous deposits of the study.

2.2 Major basins of southern South America; their geological setting and stratigraphy

2.2.1 The Rocas Verdes Basin

It is well established that the Rocas Verdes Basin developed along the Pacific margin of southern South America in the Late Jurassic and Early Cretaceous (Dalziel *et al.* 1974, Bruhn *et al.* 1978, Dalziel 1981, Wilson 1991). It occurs as a narrow, discontinuous belt of mafic igneous rocks which extend from 50°S to Cape Horn, a distance of more than 350km (Figure 2.2). The present width of the basin ranges from 5km in the northern part, to 50km in the south, at Tierra del Fuego. The original width of the basin is difficult to estimate but Dalziel (1981) concludes that the present width must be considerably less than the width of the original backarc basin which may have been twice as wide. This is due firstly to the sedimentary fill extending onto the margin of the South American mainland, and secondly because the mid-Cretaceous deformation caused tectonic shortening and likely underthrusting of part of the basin floor. The southeast limit of the Rocas Verdes Basin is unclear but it is known to extend to the northern Scotia Ridge and may have been related to the opening of the Weddell Sea Basin and the backarc basins in the Atlantic Peninsula (Barker & Griffiths 1972, Farquharson 1982). South Georgia island, at the eastern end of the north Scotia Ridge is interpreted as an off-faulted part of the floor and infill of the Rocas Verdes Basin (Bruhn & Dalziel, 1977).

During the Late Jurassic, extension along the Pacific continental margin of southern South America led to the development of a submarine trough and basaltic magmas were intruded into the continental crust behind a volcanic complex to the west (Bruhn *et al.*, 1978). The trough formed a narrow marginal basin as the volcanic arc became detached from the continental landmass (Figure 2.5). Submarine volcanism produced oceanic type crust of pillow lavas and sheeted dykes at small spreading centres, and these formed part of the floor of the basin. The outcrops of the upper part of the ophiolite suite are known as the Tortuga and Sarmiento Complexes (Suárez, 1979). Although the complex is commonly referred to as ophiolitic, only the upper part of a complete ophiolite complex is present, i.e. there are no ultramafic rocks.

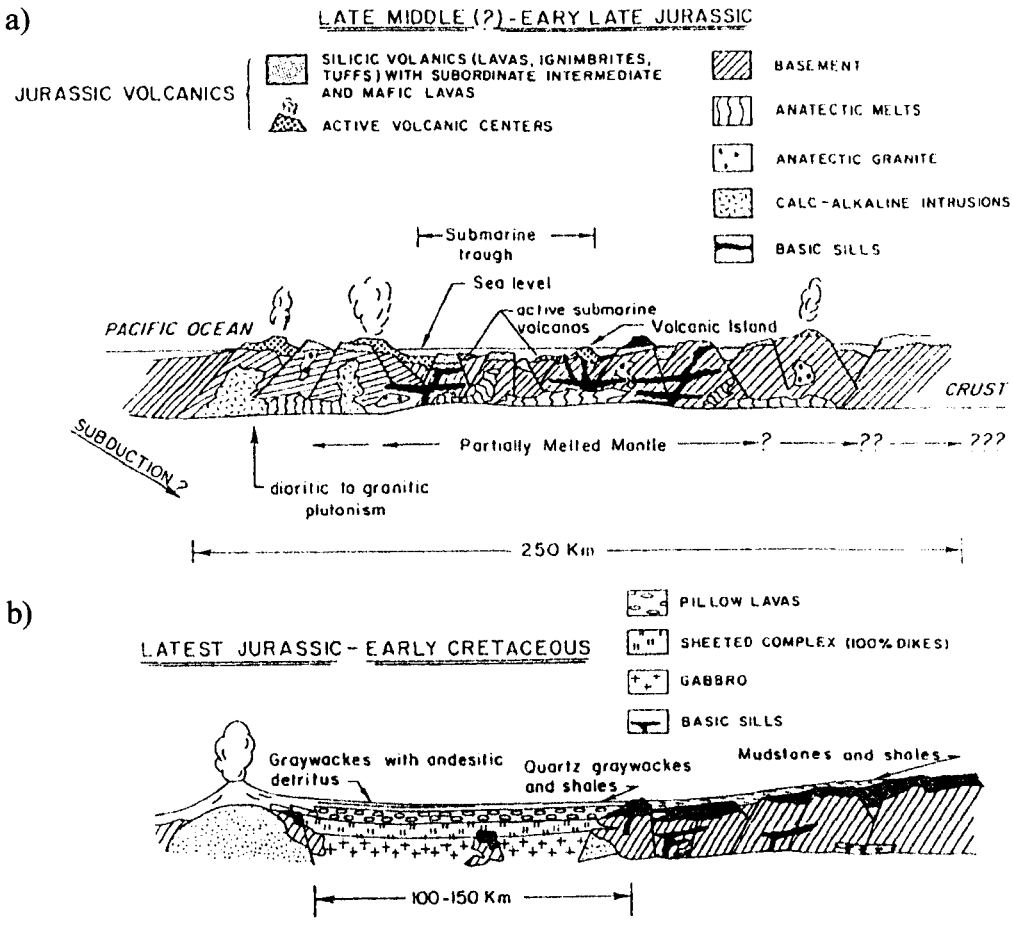


Figure 2.5: a) Tectonic setting in southernmost South America during the Middle Jurassic, after crustal extension and the opening of the submarine trough behind the island arc.

b) Tectonic setting of the Rocas Verdes Basin during the Early Cretaceous when oceanic crust had been extruded at small spreading centres and incorporated into the basin floor. The basin was also receiving andesitic detritus from the island arc to the west, whilst mudstones and shales accumulated on the continental margin to the east.

(after Bruhn *et al.* 1978)

Bruhn and Dalziel (1977) state that this is probably due to the level of exposure, and that the ultramafics exist at depth. However, there is the possibility that the part of the ophiolite suite that is present is allochthonous and could represent old oceanic crust that became trapped between South America and the island arc. Dalziel (1981) notes that the main mafic bodies have tectonized contacts but asserts the mafic rocks were emplaced during the formation of the basin. Bruhn and Dalziel (1977) emphasize that the floor of the basin is autochthonous and is close to its original level with respect to the frontal arc, and uplifted with respect to the remnant arc. This uplift and movement towards the continent is interpreted as an aborted obduction episode.

In the Early Cretaceous, the marginal basin was the site of volcaniclastic, andesitic turbidites derived mainly from the active arc on the Pacific side but also from the inactive silicic remnant arc on the Atlantic side (Yahgan Formation). Contemporaneously, the rifted continental shelf of the marginal basin subsided gently and more than 1.5km of black shales and marls (Zapata Formation) of comparatively shallow water facies (<500m) accumulated on the shelf (Bruhn & Dalziel, 1977). The basement that was intruded by the mafic bodies contained rocks as young as Permian which were unconformably overlain by a layer of volcanic rocks interbedded with shallow marine sedimentary layers containing Upper Jurassic fossils (Dalziel, 1981). As the turbidites do not show any evidence of intrusions or mafic volcanics, it can be deduced that the emplacement of the ophiolitic rocks and hence the opening of the basin, occurred in the Late Jurassic.

The basin closed as a result of mid-Cretaceous compression, as the volcanic arc moved back towards the continental margin. The basin floor assemblages were variably deformed with zones of intense deformation alternating with virtually undeformed rocks across the width of the basin. The sedimentary fill was more uniformly deformed with folding and thrusting towards the Atlantic side. It is possible the basin closure was transitional from the narrow northern segment to the wider southern segment on South Georgia (Dalziel *et al.* 1974, Dalziel 1981).

The arc, backarc and remnant arc terranes were all uplifted relative to the stable South American continent. Between the uplifted Cordillera and the stable interior, the foreland Magallanes Basin developed (Dalziel, 1981).

2.2.2 The Magallanes Basin

The Magallanes Basin is also known as the Austral Basin in Argentina. It stretches from 54°S to approximately 47°S and covers roughly 160,000 km^2 (Figure 2.2). It narrows northwards where the basin boundary becomes indistinct, but in the south, at its widest part it is about 370 km across (Biddle *et al.*, 1986). Some authors (Riccardi 1988, Ramos & Aguirre-Urreta 1994) include the Aysén Basin as a northerly extension to the Magallanes Basin, as shown in Figure 1.1, but this study shows that the Aysén Basin has distinct structural and lithological attributes and should be regarded as a discrete basin.

The Magallanes Basin is the main oil producing area of southern Argentina and Chile. The production is derived mainly from the Springhill platform and formation located in the eastern sector of the basin (enclosed by dashed lines in Figure 1.2).

The Magallanes Basin is bounded by the Patagonian batholith to the west and the Deseado Massif to the east and northeast. The Deseado Massif is a Precambrian block of low grade metamorphic rocks covered by mainly Middle to Upper Jurassic volcanics and pyroclastics. The block was stable and rigid in Mesozoic and Cenozoic times, acting as a craton (Riccardi 1988, Aguirre-Urreta 1990, Riccardi *et al.* 1992).

The basement rocks found along the western and southern edges of the basin are representative of material accreted to the South American continent during the late Palaeozoic and early Mesozoic (Forsythe 1982, Biddle *et al.* 1986). The origin of the rocks in the eastern part of the basin is less certain. Forsythe (1982) suggests that they may have formed in the late Palaeozoic forearc setting that was established along the Gondwana supercontinent, but Ramos (1989) thought it was more likely that they represent the continental crust of Gondwanaland. Both theories are possible as the basin covers such a large area and as Biddle *et al.* (1986) state, they could result from

a combination of events and processes. The forearc assemblage basement rocks are sequences of strongly folded greywackes and shales, up to 2,000 metres thick and, further north, metasandstones and slates (Ramos, 1989).

In the Magallanes Basin, Triassic to Jurassic extension and normal faulting created NNW trending grabens and half-grabens, accompanied by widespread silicic volcanism (Bruhn *et al.* 1978, Gust *et al.* 1985). The Jurassic volcaniclastic sequence unconformably overlies the Palaeozoic deposits and comprises acidic tuffs and ignimbrite flows which are andesitic to rhyolitic in composition. These rocks are termed the El Quemado Complex, which is equivalent to the Tobifera Formation further south (Table 2.2) and has generally been considered to be the basement to the sedimentary depocentres (Wilson 1991, Ramos & Aguirre-Urreta 1994). The volcanic rocks show large changes in thickness, from being absent on some basement highs, to more than 2,000 metres thick in graben areas (Alvarez-Marrón *et al.*, 1993). This volcaniclastic sequence covers most of the Deseado Massif.

In the Late Jurassic and Early Cretaceous, the Rocas Verdes Basin (section 2.2.1) opened near the Pacific margin, creating a marginal basin behind a magmatic arc. This is regarded by many authors (Biddle *et al.* 1986, Mpodozis & Ramos 1989) to be the first stage in the opening of the Magallanes Basin in the south.

The Magallanes Basin is filled with more than 8 km of sediments that range in age from Late Jurassic to Cenozoic (Biddle *et al.* 1986, Aguirre-Urreta 1990). The Mesozoic clastic sequence begins with a transgressive sandstone sheet known as the Springhill Formation. They are coarse-grained, quartzose sandstones which are interpreted to be of both fluvio-deltaic and marine, shallow-water origin (Ramos 1989, Alvarez-Marrón *et al.*, 1993). The Springhill Formation sandstones are the major hydrocarbon reservoirs in the eastern and northern parts of the basin. They are generally thin, with an average thickness of 30 to 40 metres and a maximum thickness of 130 to 150 metres (Riccardi, 1988), but are sometimes absent over some basement highs. The age of the Springhill Formation varies from Tithonian to Valanginian, dated by macrofauna and microfauna fossil remains. The wide age range is attributed to diachronism (Riccardi, 1988).

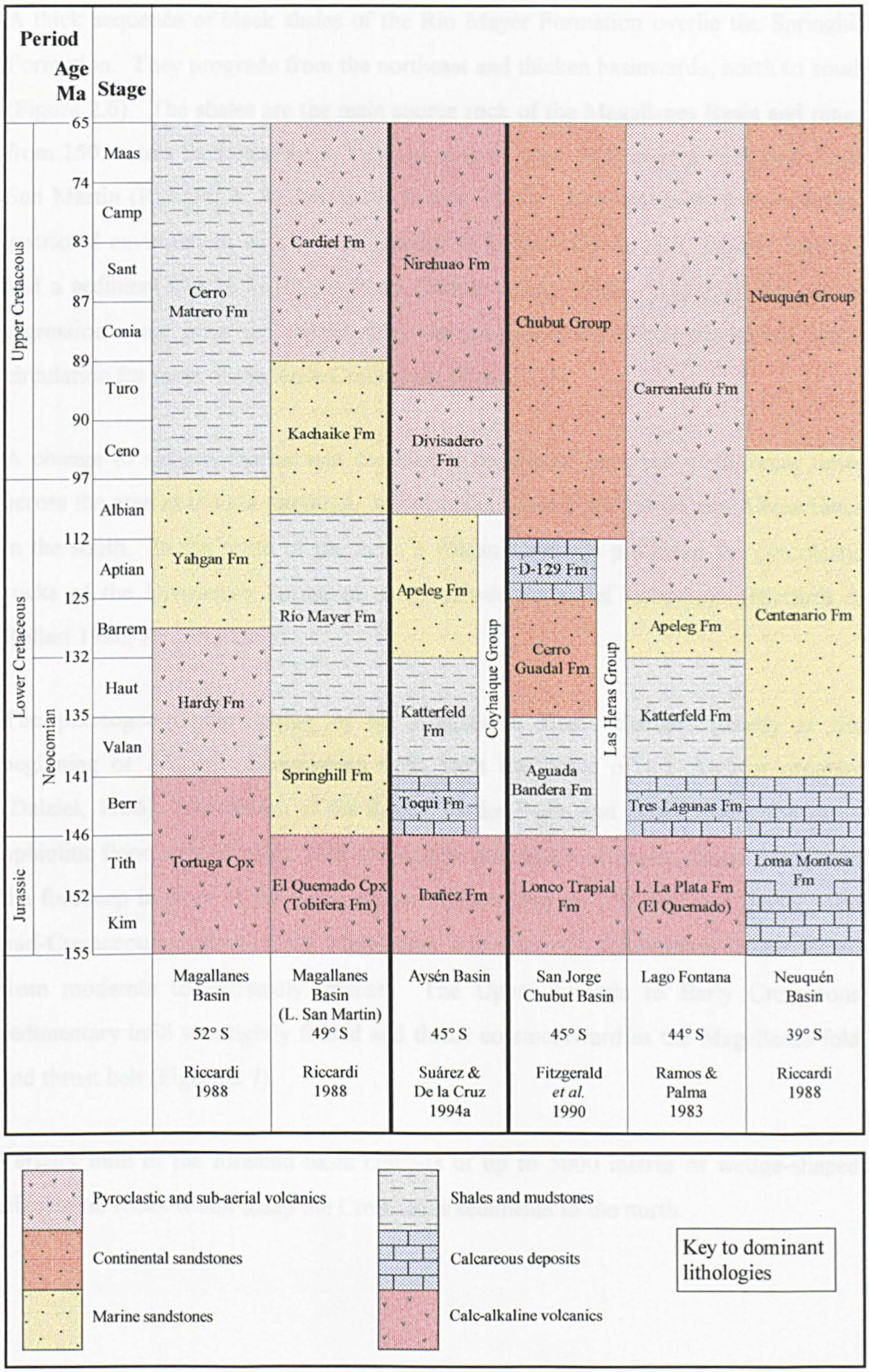


Table 2.2: Formations and generalised lithologies of major basins in southern South America. Compilation showing the relationship of the Aysén Basin to other basins. (Geochronology based on Harland *et al.* 1990).

A thick sequence of black shales of the Rio Mayer Formation overlie the Springhill Formation. They prograde from the northeast and thicken basinwards, north to south (Figure 2.6). The shales are the main source rock of the Magallanes Basin and range from 250 metres thick near Lago Fontana to more than 1000 metres thick near Lago San Martin (Riccardi & Rolleri 1980, Ramos 1989). They represent a low energy, restricted environment with a large amount of terrestrially derived, organic material and a sediment source to the northeast (Biddle *et al.*, 1986). Although the marine ingression came from the Pacific side, the magmatic arc effectively barred water circulation for most of the Early Cretaceous (Figure 2.7).

A change to shallow marine and continental conditions occurred at different times across the area as the sea retreated, in Barremian times in the north, and Albian times in the south. In the north of the basin a volcanic episode produced the pyroclastic rocks of the Divisadero Formation and equivalent Cardiel Formation (Riccardi & Rolleri 1980, Riccardi 1988).

The palaeogeographic setting of the Magallanes Basin changed greatly at the beginning of Albian - Cenomanian time, with the onset of the Andean orogeny (Dalziel, 1986). The closure of the Rocas Verdes Basin and partial obduction of its ophiolitic floor created uplift, with consequent deposition of coarse clastic material in the foredeep in front of the rising protocordillera (Figure 2.7) (Wilson, 1991). The mid-Cretaceous closure of the Magallanes Basin caused deformation which varied from moderate to extremely intense. The Upper Jurassic to Early Cretaceous sedimentary infill was tightly folded and thrust continentward as the Magallanes fold and thrust belt (Figure 2.7).

Tertiary infill of the foreland basin consists of up to 5000 metres of wedge-shaped siliciclastic rocks which onlap the Cretaceous sediments to the north.

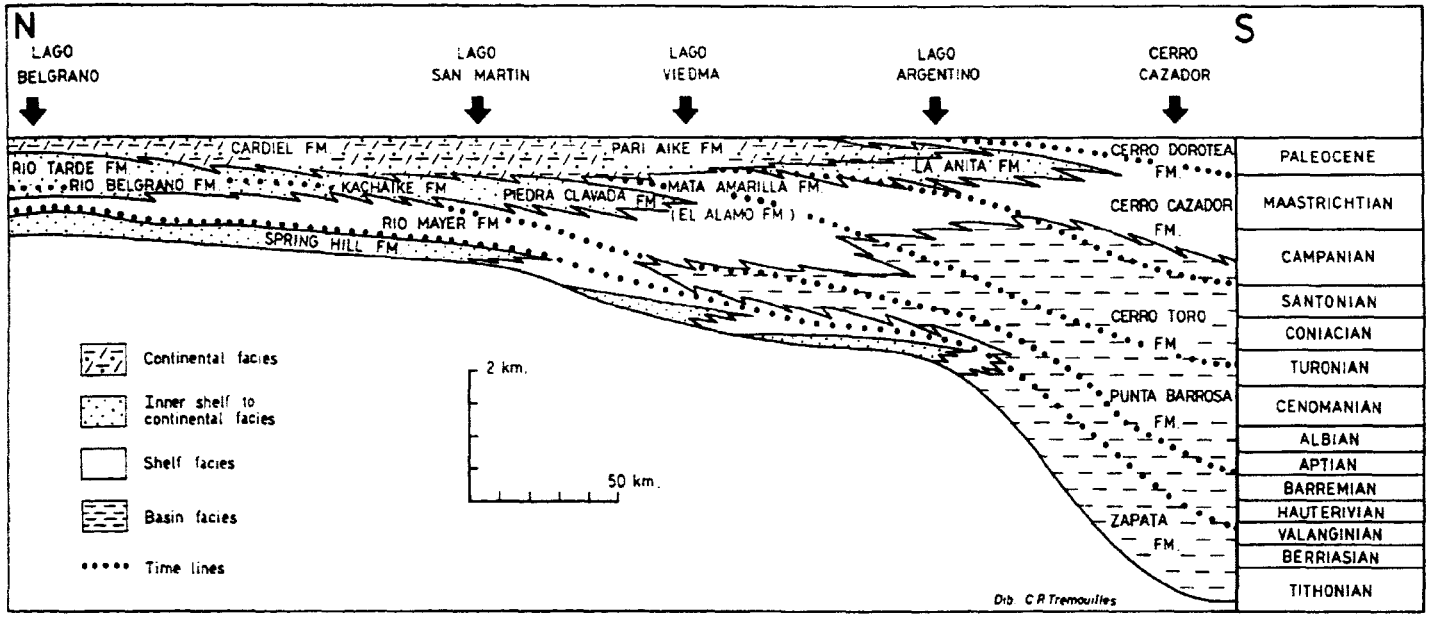


Figure 2.6: Generalised scheme from north to south across the Magallanes Basin showing relationships between formations. The Springhill Formation appears as three, discontinuous, prograding intervals. The black shales of the Rio Mayer Formation thin northwards to about 250 metres thick, from approximately 1000 metres thick near Lago San Martin. The variation in thickness is likely to be due to thermal relaxation of the main basin partly controlled by the basement structure. The closure of the marginal basin in Albian times and the uplift of the cordillera is reflected in the foredeep deposits of the Magallanes Basin (Riccardi 1988).

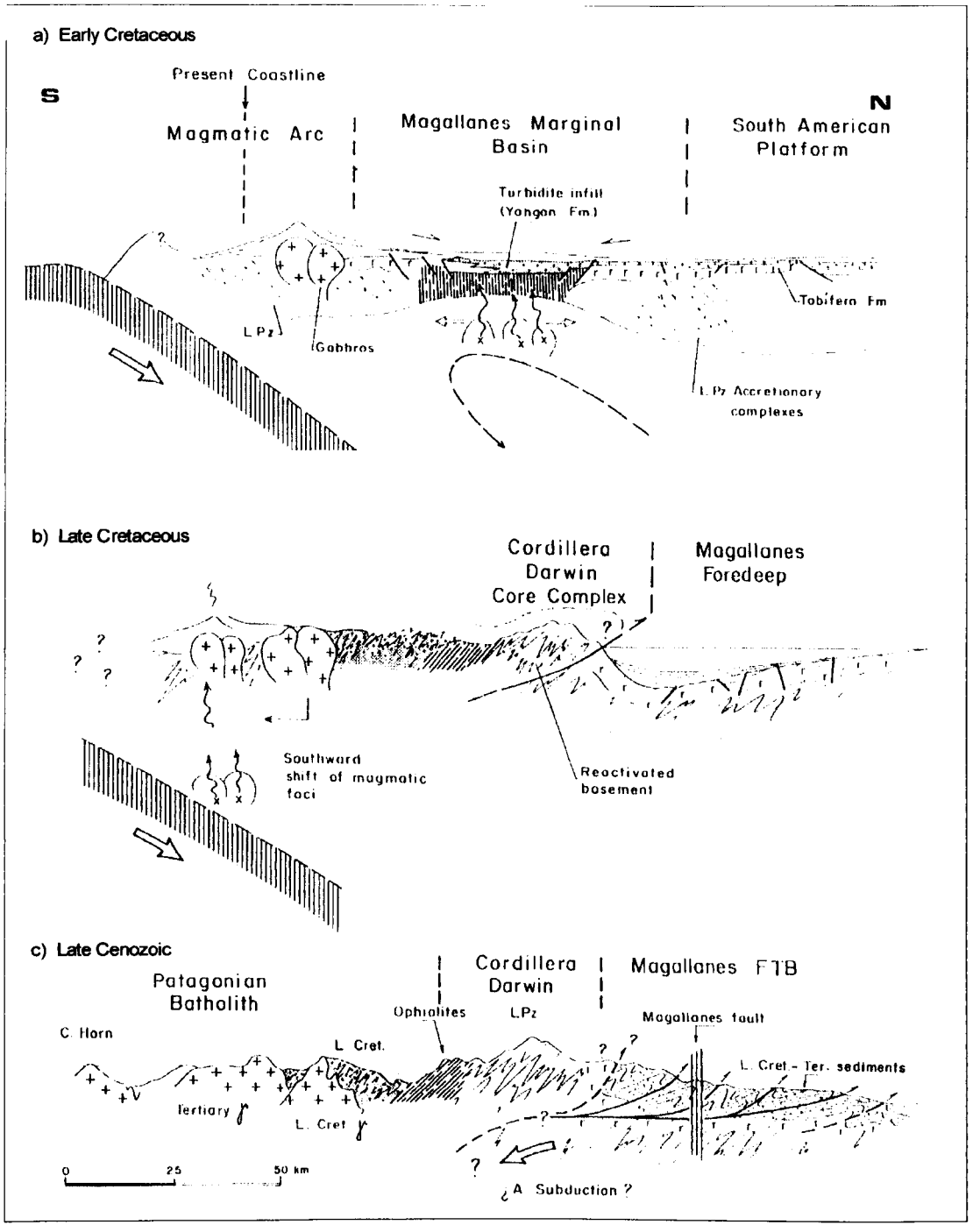


Figure 2.7: Structural and palaeogeographical sections across the Magallanes Basin.

a) Early Cretaceous formation of a marginal basin behind a magmatic arc. Turbiditic input originates in the magmatic arc and marine to continental deposits are laid down on the continental margin.

b) Late Cretaceous closure of the marginal basin and uplift of the cordillera resulting in initiation of the Magallanes Foredeep.

c) Cenozoic migration of the orogenic front eastwards producing a fold and thrust belt. (Mpodozis & Ramos 1989).

2.2.3 The Neuquén Basin

The Neuquén Basin is known as the Andean Basin in Argentina with an embayment in the southeast termed the Neuquén Embayment (Hallam, 1991). It lies within the southern part of the central Andes, between 36°S and 40°S and stretches eastwards into Argentina (Figure 1.1). It is flanked by the Somuncurá Massif to the south, the Sierra Pintada to the north and the Andean Cordillera to the west (Figure 2.2). It is a triangular shaped basin covering more than 160,000 km^2 (Vergani *et al.*, 1995).

The tectonic evolution of the Neuquén Basin began during the Late Triassic as northeasterly extension, related to a pre-Gondwanaland-breakup tectonothermal-event and resulted in a series of normal faulted half grabens in the Palaeozoic basement rocks (Figure 2.8). The resulting basins received thick accumulations of volcaniclastic materials deposited in both marine and non-marine environments and with subordinate carbonates and evaporites. These faulted basins, trending principally north-south and northwest-southeast, formed the initial Neuquén Basin (Legarreta & Uliana 1991, Vergani *et al.* 1995).

During the Late Jurassic to Early Cretaceous, the basin downwarped in an early sag phase and a marine embayment formed (Mitchum & Uliana 1985, Riccardi 1987). The marine transgression spread from northwest to southeast and calcareous sediments, the Loma Montosa Formation (Table 2.2), devoid of volcanics, were deposited in the eastern areas. During the early stages of this transgression, water remained shallow and slightly hypersaline and an area of tens of thousands of square kilometres was covered by a stromatolite zone. As the transgression advanced, a well diversified fauna developed and a wide basinal area emerged (Legarreta & Uliana, 1991). However, at the southeast margins, continental deposits were laid down (Riccardi, 1987). The subsidence was thermally controlled and Ramos (1990) interprets this to reflect an important thinning of the continental crust. An extensional tectonic regime persisted into the Early Cretaceous and caused the depocentres to move eastwards.

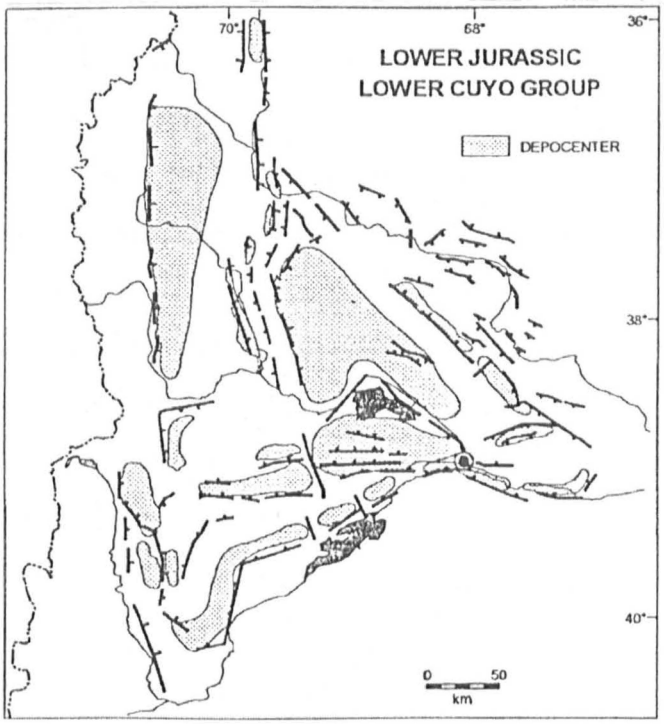


Figure 2.8: A series of superimposed structural styles affected the Neuquen Basin throughout its evolution. The principal fault trends are north/south and northwest/southeast. The system of extensional faults formed the initial Neuquen Basin. The major depocentres are shown light grey. Dark grey areas are lakes.
(after Vergani *et al.*1995)

At the end of the Berriasian, an influx of siliciclastic detritus inhibited carbonate deposition and the basinal area became reduced by progradational sedimentation. The reduction in extent of the depositional site continued into the Valanginian, when a short marine regression took place. The regression was short lived and succeeded by a transgression in late Valanginian - Hauterivian times (Centenario Formation, Table 2.2 and Figure 2.9). During the Hauterivian, several episodes of marine shallowing caused an increase in clastic deposition with evidence of aeolian/fluvial sands, evaporites and extensive mudflats. The area was rapidly flooded and a euxinic basin which accumulated organic rich, ammonite shales, was resumed. Towards the end of the Hauterivian a shallow water facies of thin, unfossiliferous shales and ooidal packstones-grainstones accumulated in large areas. The basin continued to shallow and by the end of the Albian, broad alluvial aprons of clastic sediments dominated (Riccardi 1987, Legarreta & Uliana 1991, Vergani *et al.* 1995). The basin fill does not show any evidence of deformation until the mid-Cretaceous when the Neuquén Group was initiated in the foreland deep in front of a fold and thrust belt. During the Upper Cretaceous the basin was dominated by the terrestrial depositional systems of the Neuquén Group (Table 2.2).

2.2.4 The San Jorge Basin

The San Jorge basin of Argentina is also known as the Chubut Basin (Figure 2.2). In Jurassic to Cretaceous times it extended over most of Argentine Patagonia from the Somuncurá Massif in the north, to the Deseado Massif in the south however, the present day San Jorge Basin is much smaller and trends east - west, between 45°S and 47°S. Like the other major southern South American basins, it is predominantly an extensional basin imposed on Palaeozoic basement. Northwest to southeast trending grabens formed and filled during a Triassic to Early Jurassic rift phase. The sedimentary history is however, different to the other age equivalent basins such as the Neuquén and Magallanes basins.

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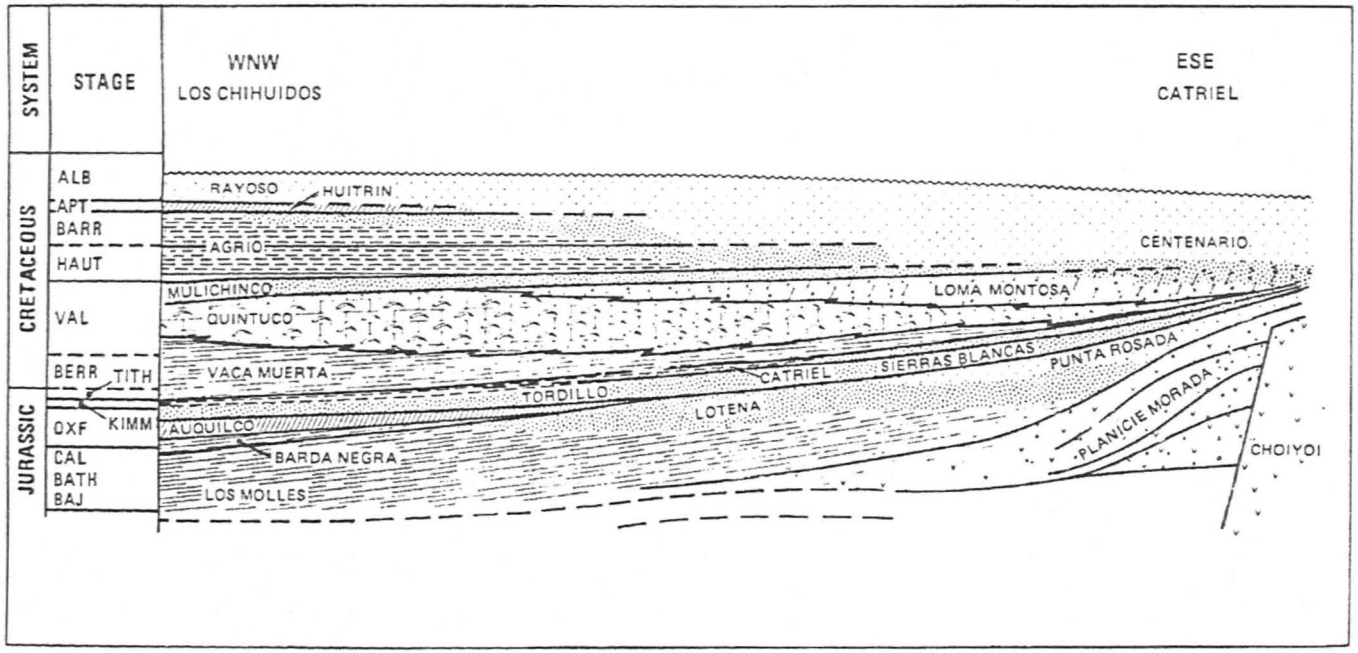

Figure 2.9: Regional stratigraphic diagram of the Neuquen Basin showing the variation of facies across the basin and the main formations and groups during the Upper Jurassic and Early Cretaceous.
(Mitchum & Uliana 1985).

The widespread volcanic event which produced the El Quemado Complex in southern basins, was also present in the San Jorge Basin and produced the Lonco Trapial volcanics (Table 2.2), a succession of acidic and intermediate volcanic rocks which cover most of central and southern Patagonia and fill the northwest trending graben. Normal faulting continued at a decreased rate into the Early Cretaceous and thick, lacustrine, black shale sequences, with oolitic limestones bordering them, as well as fluvial-deltaic clastic rocks, were deposited until Aptian times. The sediments, known as the Las Heras Group, represent an early sag environment (Riccardi, 1988). Although the black shales are generally interpreted as lacustrine, Fitzgerald *et al.* (1990) report several occurrences of marine levels within them. The likely setting for the deposition of the black shales is an extensive lake complex bordered by fluvial systems with minor but repeated marine incursions. It suggests there was an intermittent opening to the west, through the Aysén Basin.

Most normal faulting ceased as oceanic crust began to form in the South Atlantic and a late sag phase commenced in the Albian when fluvial sediments were laid down. This sequence of sandstones, siltstones, mudstones and pyroclastics contain hydrocarbons but are devoid of reliable stratigraphic markers. The depositional setting was dominantly continental, subaerial and shallow sub-aqueous (Lesta *et al.*, 1980). Deposition was controlled by alluvial processes associated with swamps and shallow, ephemeral alkaline lakes (Fitzgerald *et al.*, 1990).

Uplift and compressional deformation occurred during the Tertiary and created a north trending fold belt and imposed an easterly tilt on the sedimentary successions. The basin stopped receiving sediments during the early Tertiary.

2.2.5 The Aysén Basin

The Aysén Basin stretches from 43°S to 47°S and lies directly to the east of the present Patagonian batholith. It is bordered by the Somuncurá Massif to the north and the Deseado Massif to the east and southeast. The Aysén Basin is approximately 500km long, in a north-south direction and is less than 150 km at its widest point where it extends into Argentina, near the entrance to the San Jorge Basin. The boundaries of the basin are unclear and, as already discussed in section 2.2.2, several authors include the Aysén Basin as a northerly extension to the Magallanes Basin and a link in to the San Jorge Basin of Argentina. The uncertainty of the boundaries of the basins in this area has led to overlap in their naming and identification. Ramos (1989) named this area the Rio Mayo Embayment and included the Lago Fontana and Lago La Plata areas of Ramos (1976, 1981) and Ramos & Palma (1983) within the Rio Mayo Embayment.

2.2.5a The Aysén Basin in Argentina

The Rio Mayo Embayment was first described by Aguirre-Urreta & Ramos (1981) and formally named by Ramos in 1989. The area described by Aguirre-Urreta & Ramos (1981) as the Rio Mayo Embayment straddles the international boundary between Chile and Argentine Patagonia (Figure 2.10). It occurs to the west of the San Jorge Basin and includes the Lago Fontana and Lago La Plata areas of western Argentine Patagonia (Figure 2.11). It also encompasses the eastern part of the Aysén Basin study area. Ramos (1989) described the Rio Mayo Embayment as a small marine basin which developed to the east of the Patagonian batholith where it received sediments which were interpreted as those of a pro-grading delta. Mpodozis & Ramos (1989) further described a small intra-arc basin to the west of the Rio Mayo Embayment which they named the Katterfeld Basin (Figure 2.12). Ramos and Palma (1983) suggested that the northern segment of the basin was bordered by a western volcanic arc and an eastern, external volcanic arc in the areas of Lago Fontana and Lago La Plata, and that to the south the external volcanism faded out, while to the north, the intra-arc basin disappeared and the volcanism formed a unified arc (Figure 2.10).

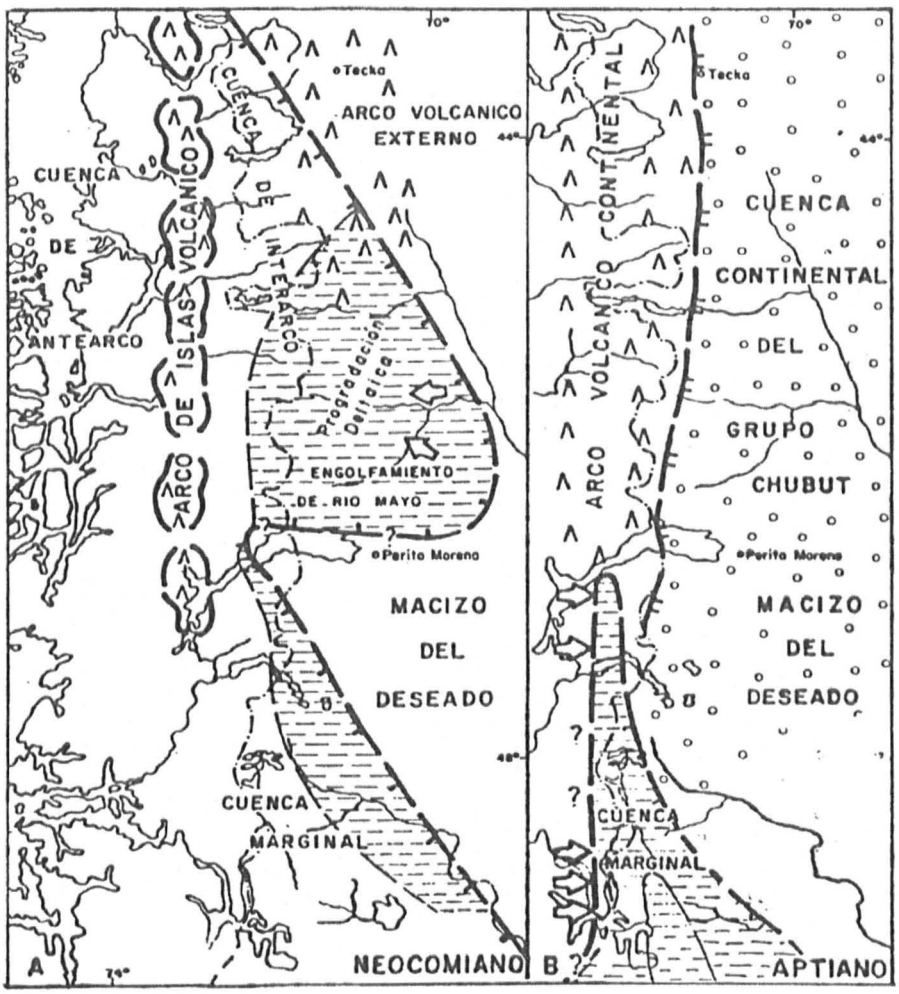

Figure 2.10: The Rio Mayo Embayment of Argentina and Chile as first proposed by Aguirre-Urreta & Ramos (1981). They interpreted the sediments as those of a pro-grading delta in an intra-arc basin setting.

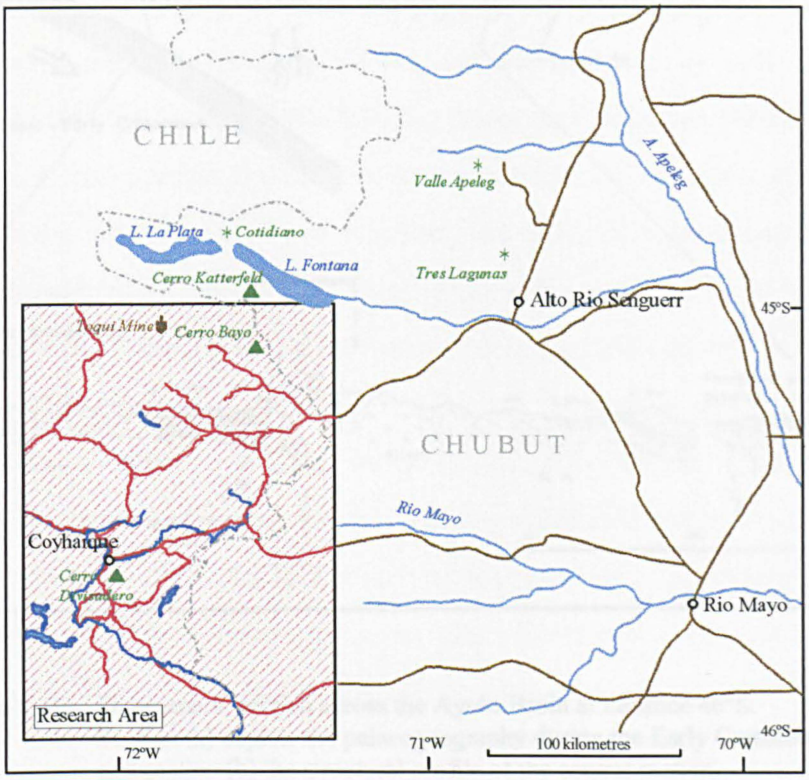

Figure 2.11: Location map showing the research area and the type localities of the Coyhaique Group. Much Argentinian work has focused on the Lago Fontana and Lago La Plata areas.

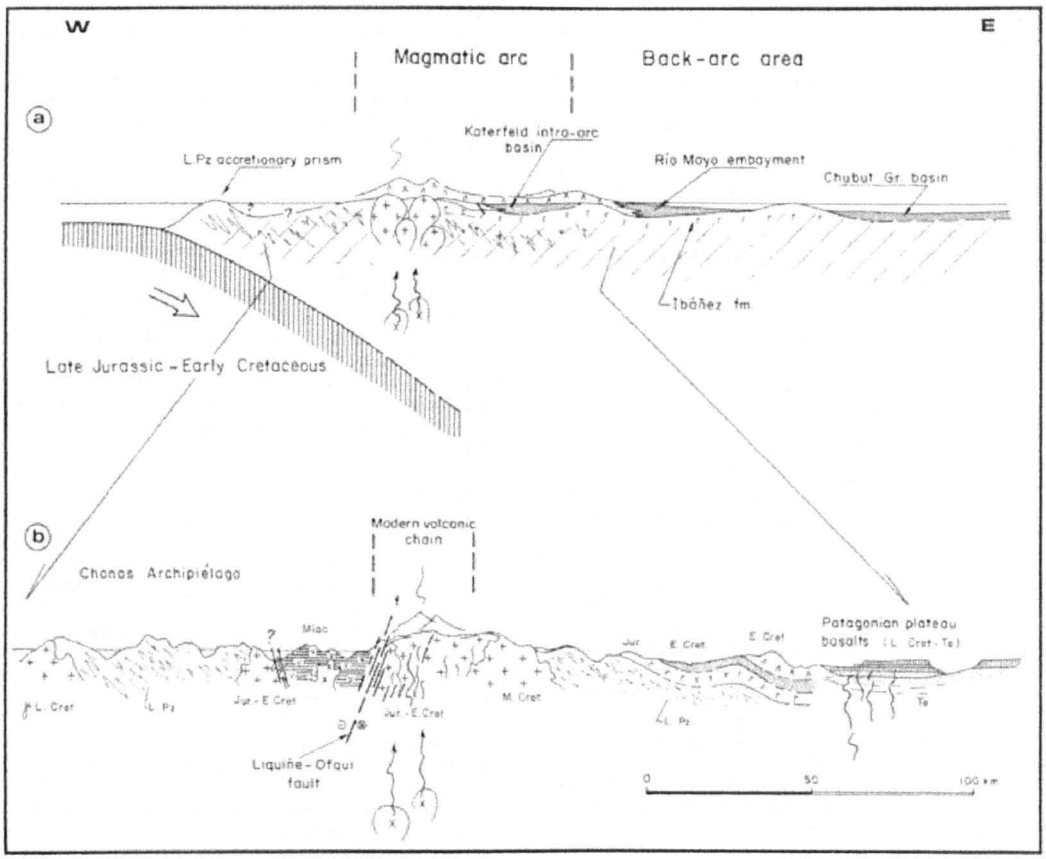

Figure 2.12: Diagramatic section across the Aysén Basin at Latitude 46°S. Section (a) depicts the palaeogeography during the Early Cretaceous and section (b) the structural profile of the central section. (Mpodozis & Ramos 1989).
The diagram shows the Rio Mayo Embayment and Katterfeld Basin as separate depocentres. The Rio Mayo Embayment is described as a marine basin and the Katterfeld Basin as an intra-arc basin.

The basement of the area is a volcanic sequence which Ramos and Palma (1983) named the Lago La Plata Formation (Table 2.2). In the west of the region, between Lago Fontana and Lago La Plata, the volcanics are overlain and intercalated with a calcareous sequence which was first described by Ramos (1976) and which he named the Cotidiano Formation (Figure 2.11). He described isolated banks of coral reef less than five metres thick, containing crinoids and echinoderms as well as bivalves. Further south, near the border with Chile, the calcareous outcrops are thicker, up to 30 metres thick, and contain abundant oysters identified as *Gryphaea* sp. by Ramos (1976). Near the top of the southern exposure, the shelly content disappears and the sediments become more pelitic. In later studies Ramos (1981) noted that the calcareous deposits of the Cotidiano Formation interfinger with the upper levels of the Lago La Plata Formation which comprise banks of calcareous sands and andesitic rocks and tuffs which he also included in the Cotidiano Formation. He deduced that the deposits correspond to the end of the Jurassic volcanism which at this time was aerial and intermittent. From the faunal content Ramos (1981) tentatively assigned an age of Oxfordian to Kimmeridgian to the Cotidiano Formation and the contemporaneous part of the Lago La Plata Formation.

According to Ramos & Aguirre-Urreta (1994) the main basin fill began when a turbiditic series of sediments covered the volcanic relief and the discontinuous calcareous beds. The sediments are principally calcareous sandstones and conglomerates which were defined near Alto Rio Senguerre in Argentina, by Ploszkiewicz & Ramos (1977) as the Tres Lagunas Formation (Figure 2.11 and Table 2.2). The Tres Lagunas Formation only outcrops in the sector south of Lago Fontana (Ramos 1981, Suárez & De La Cruz 1992) and the only complete succession showing the top and base is near the Chile border, west of Cerro Katterfeld (Figure 2.11). The stratigraphic relationship varies from east to west across the area. In the east, at the type locality, the base of the sequence is represented by a thick conglomerate which overlies an erosive unconformity, but further west, near the Chile border, the contact is apparently concordant (Ramos, 1981). The Tres Lagunas Formation is more than 100 metres thick at the type locality even though the top of the sequence is missing. Here the thick conglomerate and calcareous sands are interbedded with thin horizons

of black shales and fining upwards conglomerates containing fragments of *Trigonia* sp. and other bivalves. Near the Chile border the complete succession of the Tres Lagunas Formation is less than 35 metres thick. At this location Ramos (1981) discovered isolated examples of the oyster *Exogyra* sp. in a micritic limestone, as well as fragments of bivalves and a species of belemnite which he identified as *Belemnopsis patagoniensis.* He suggested that the Tres Lagunas Formation represents a period of time from the Tithonian to the Berriasian and perhaps even into the lower Hauterivian (Table 2.2).

Overlying the Tres Lagunas Formation is more than 350 metres of black shales of the Katterfeld Formation defined by Ramos (1976) from Cerro Katterfeld in Argentina (Figure 2.11). Various authors have recorded measurements of up to 1,000 metres thick for the Katterfeld Formation in the type locality but Ramos (1981) pointed out that the area is noticeably faulted and the lower figure is likely to be more accurate. Ramos (1976, 1981) recorded the lithology to be of monotonous black or dark-grey shales, fissile, highly indurated and containing abundant concretions, especially in the top section of the formation. Ramos (1981) also found the top third of the Katterfeld Formation to have the most abundant faunal content of belemnites, and ammonites which he identified as *Favrella americana* and *F. wilckensi.* In the west of the Rio Mayo area, Ramos (1981) found the Katterfeld Formation to show different characteristics. He recorded numerous intrusions of basic dykes and granitic veins which recrystalised the shales and formed hornfels and sometimes slates. Ramos & Palma (1983) also reported similar metamorphic characteristics of the Katterfeld Formation north of Lago La Plata, near the international border with Chile, from where they recorded poorly preserved examples of the belemnite *Belemnopsis patagoniensis* and the ammonite *Favrella* sp. At the type locality, towards the top of the formation, interbedded sandstones appear within the shales. In the upper part of the sequence the shales have a lighter, grey-green appearance and contain lenticular laminations of fine sand. These horizons pass transitionally into the Apeleg Formation (Ramos 1981).

The Apeleg Formation is a sequence of sandstones, sandy conglomerates and mudstones first defined in Argentina by Ploszkiewicz & Ramos (1977) (Figure 2.11) where it covers a large area of northwest Chubut Province. North of the Rio Apeleg it is overlain by Lower Cretaceous volcanics, whilst near Alto Rio Senguerr it overlies the Tres Lagunas Formation. At the type locality the Apeleg Formation is described as being composed of principally medium to coarse-grained sandstones, light brown to yellow, or rarely, green in colour. The sandstones are well stratified in sets of variable thickness from centimetres to metres thick and frequently showing cross-lamination with occasional ripple surfaces (Ploszkiewicz & Ramos 1977, Ploszkiewicz 1987). The interbedded conglomerates are imbricated and often contain abundant silicified woody material. The conglomerates are more frequent in the eastern exposures near Alto Rio Senguerr where the Apeleg Formation is up to 1,200 metres thick. Near the top of the sequence black mudstones become dominant and some thin horizons of carbonaceous material are found (Ploszkiewicz & Ramos, 1977). Near Lago Fontana, Ramos (1981) found there were no exposures of the Apeleg Formation in the western locations, possibly due to non-deposition. However, on the flanks of Cerro Katterfeld (Figure 2.11) he identified a thick sequence of the Apeleg Formation where it passes transitionally and abruptly upwards from the black shales of the Katterfeld Formation as the percentage of sand increases. Ramos (1981) classified the Apeleg Formation into three members at this location. He found the lower member to be composed of grey-green sandstones and shales and less than 100 metres thick. The middle sandstone member varies between 700 and 1,200 metres in thickness and displays distinctive cross-stratification and ripple cross-bedding, whilst the uppermost member is a coarse-grained sandstone, approximately 200 metres thick, containing lenses of conglomerate. Carbonaceous material is common throughout the Apeleg Formation and in thin section the lower part of the sequence contains some glauconite pellets.

Ramos (1981) interpreted the Apeleg Formation to be deposits of a pro-grading delta. The lower member is typical of a delta front and found exclusively in the western sector of the Lago Fontana area, near Cerro Katterfeld. He thought the middle member represented subaqueous deposition on a delta platform and the upper member

to be characteristic of a flood plain and restricted to the eastern sector near Tres Lagunas. He suggested that the three members integrated laterally and passed from principally marine deposits in the west to continental deposits in the east.

2.2.5b: The Aysén Basin in Chile

The Aysén region in Chile is relatively unknown geologically. The first descriptive work on the granites around Puerto Aysén (Figure 1.3) was published by Quensel, 1911 (*in* Niemeyer *et al.*, 1984). Later, Heim, 1940 (*in* Niemeyer *et al.* 1984) described the same volcanic series but proposed new names for the distinct units of the Divisadero Formation and, around Lago General Carrera, the Ibañez series volcanics. Niemeyer *et al.* (1984), reported that the first systematic study of the Aysén region was carried out by Fuenzalida & Latcham (1935). They described a section from Puerto Aysén to the Argentine border and identified a series of lithological units including "black clays of Coyhaique". They correlated the units with descriptions of similar outcrops further south near Lago San Martin in Argentina (Figure 1.3). In later work, Fuenzalida 1954 (*in* Niemeyer *et al.*, 1984), referred to the black clays as "black shales of Coyhaique" and assigned them an age of Tithonian to Neocomian.

In the 1960s and 1970s exploration for hydrocarbons and other minerals resulted in the first detailed geological maps of the area and work on the stratigraphy. Important mineral resources were identified around Lago General Carrera (Figure 1.3) in the south of the Aysén Basin and at El Toqui and Ñirehuao towards the north (Niemeyer *et al.*, 1984). A palaeogeographic study of the area between 45^0 & 46^0 LS by Skarmeta (1976) proposed the area could be divided into three tectonic regions; a marginal basin, island-arc and continental shelf. In this work, Skarmeta identified a number of key sites where the Neocomian sediments are well exposed and have continued to provide a source of new data for more recent studies, including Location AMT 19 of this study.

The Coyhaique Formation was named by Lahsen (1966) after the sequence of shales and fossiliferous sandstones found near Coyhaique, and previously described by Fuenzalida & Latcham (1935). The name became well established in the literature of the 1970s until further work on the sediments by Haller & Lapido (1980). This work extended the stratigraphic range and established clear facies changes between the sediments. Haller & Lapido (1980) linked all the sediments and re-named them the Coyhaique Group (Table 2.2). The type locality of the Coyhaique Group was described in some detail by Niemeyer *et al.* (1984) who proposed an age of Berriasian to Aptian for the sediments from their faunal content. This work identified a correlation between similar sediments in Argentina especially the black shales of the Katterfeld Formation and the Rio Mayer Formation to the south of the study area (Table 2.2).

More recent work by Suárez & De La Cruz (1992) extended the nomenclature already established in Argentina by Ramos (1981), into Chile. The sediments of the Coyhaique Group were therefore represented as; calcareous limestones of the Cotidiano Formation, sandstones and conglomerates of the Tres Lagunas Formation, shales of the Katterfeld Formation and sandstones of the Apeleg Formation. In 1994 Suárez & De La Cruz further refined their earlier work and were able to make distinctions between the Argentinian sediments and the Chilean deposits. They re-named the lower units of the Cotidiano Formation and the Tres Lagunas Formation in Chile as the Toqui Formation (Suárez & De La Cruz, 1994a &b) (Table 2.3). Suárez & De La Cruz (1994a) recognised three facies associations within the sedimentary and pyroclastic rocks of the Toqui Formation which they recorded as calcareous, sandstone and pyroclastic members. All three units are found in the surroundings of the Toqui mine (Figure 2.11) but only sporadically across the area of the study basin.

The lithologies found in the Aysén Basin are therefore; basement rocks of the Ibañez volcanics, sediments of the Coyhaique Group which include the Toqui Formation, Katterfeld Formation and Apeleg Formation and volcanics of the Divisadero Formation (Table 2.3). The sediments of the Coyhaique Group therefore provide an important stratigraphic marker between the similar volcanic successions of the Ibañez and Divisadero Formations and will be discussed in the following three chapters.

Period	Age Ma	Stage	Formation/ Group	Lithology/Event	Interpretation
	90	Ceno	Divisadero Fm	Subaerial calc-alkaline volcanism	Magmatic activity in western arc
	97	Albian		Local unconformity. Deformation and uplift	Minor folding and thrusting
Lower Cretaceous	112	Aptian	Apeleg Fm (Coyhaique Group)	Rippled sandstones and shales in thick sequences. Uniform appearance. Some cross-bedded sequences and mudflake breccias. Sandstones are well-sorted and medium to fine grained. Abundant trace fossils.	Tidal sandbars in a narrow seaway. Slow rates of sedimentation with constant energy levels over long periods of time. Deposition in shallow water below storm wave base.
	125	Barrem			
	132	Haut	Katterfeld Fm (Coyhaique Group)	Calcareous black shales and mudstones. Homogeneous appearance. High organic content. Regionally extensive, laterally continuous and generally parallel bedded. Restricted fossil fauna.	Embayment with restricted circulation. Deposition from suspension onto a flat sea-bed under tectonically quiet conditions. Slow rates of subsidence.
Neocomian	135	Valan			
	141	Berr	Toqui Fm (Coyhaique Group)	Limestones, sandstones and pyroclastics	Marine transgression across active volcanoes and isolated bioherm formation
	146	Tith	Ibañez Fm	Subaerial silicic volcanism. Rhyolitic/ dacitic and andesitic ignimbrites and breccias. Regional unconformity	Western magmatic arc. Volcanism related to subduction processes. Uplift and erosion
Jurassic	152	Kim			
	155				
Palaeozoic/Triassic			Basement	Metasedimentary basement rocks. Meta-sandstones and slates.	Terrane accretion. Sediments piled up against the continental margin due to subduction of downgoing plate.

Table 2.3: Stratigraphic chart of the Aysén Basin (Chronology based on Harland *et al.* 1990).

(For the regional geology of the Aysén Basin see Appendix A)

CHAPTER 3

THE TOQUI FORMATION

The Toqui Formation is the basal unit of the Coyhaique Group (Tables 2.2 & 2.3). It occurs as discontinuous outcrops throughout the study area (Figure 3.1) and is composed of sedimentary and pyroclastic rocks, defined by Suárez and De La Cruz (1994a) as calcareous, arenaceous and pyroclastic members (Chapter 2, Section 2.2.5b). In this study the sporadic outcrops of the Toqui Formation were generally not distinctive enough to be clearly identified to a specific member and so the lithology at each site is described. The type locality is in the Río Toqui district where good outcrops of the differing lithologies of the formation occur near the Toqui mine (Figure 3.1, Location TM 1).

3.1 General characteristics of the Toqui Formation

In the study area, four identifiable lithologies could be attributed to the Toqui Formation: a shelly limestone, a bedded limestone, a volcanic sandstone and some infrequent pyroclastic beds. The lithologies do not always occur together. Sometimes they overlie each other and occasionally they are interbedded which makes it difficult to define them into separate members.

The Toqui Formation is most commonly found as a shelly limestone represented by isolated oyster banks or bioherms (also referred to as coquinas in Chilean literature). At the type locality, Suárez and De La Cruz (1994a) recorded a maximum thickness of 40 metres for the calcareous member of the Toqui Formation, but in the study area exposures are less than 15 metres thick.

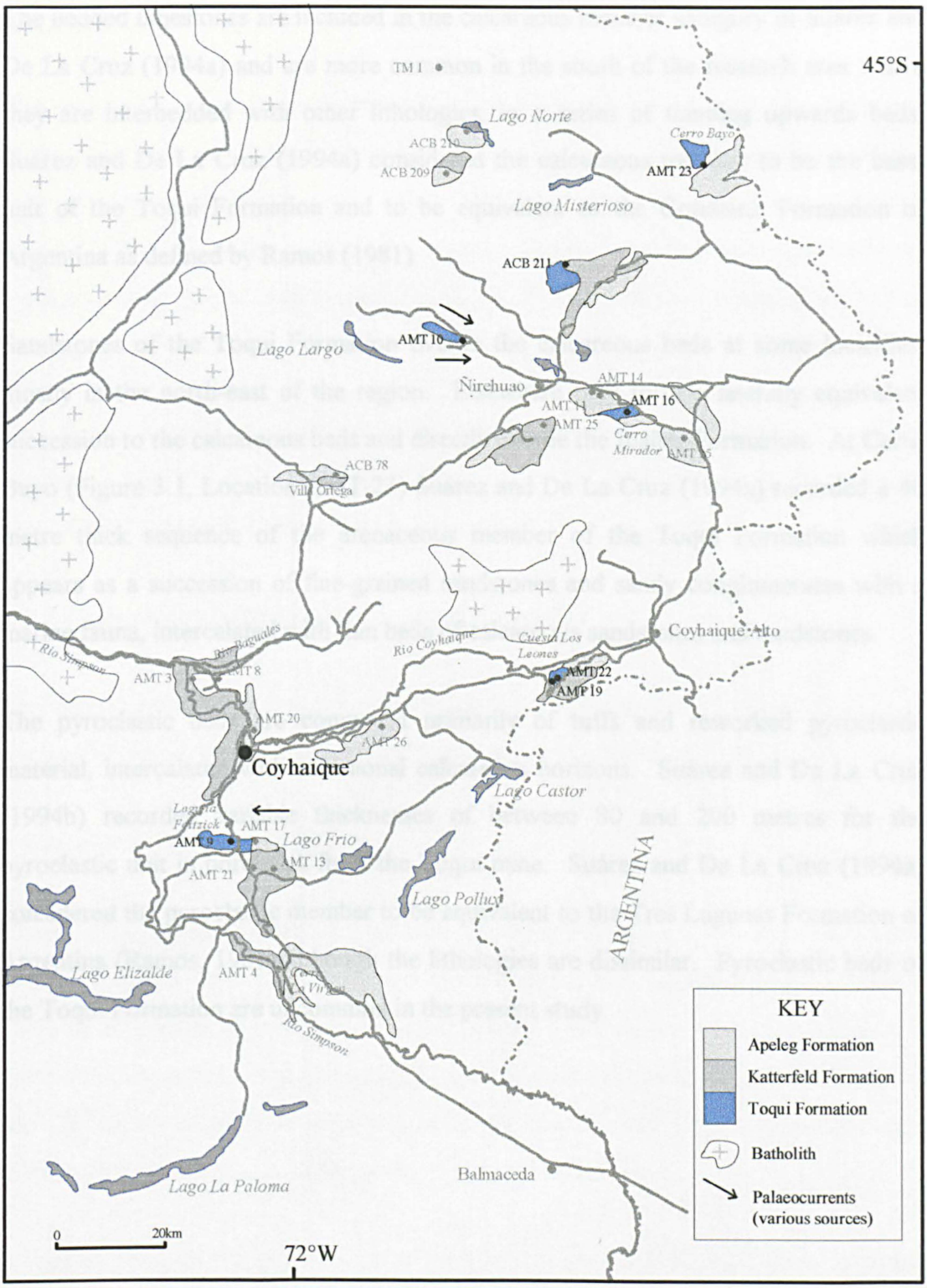

Figure 3.1: The Aysén Basin study area highlighting discontinuous outcrops of the Toqui Formation and the main sampling locations.

Basin Margins: The Toqui Formation deposits represent a gentle marine transgression over the subaerial volcanics of the Ibañez Formation during the Berriasian. The basin may have extended to the Somuncurá Massif in the north but the NE margin occurred near the Argentine border east of Ñirehuao. The basin extended eastwards into Argentina through a restricted opening to the San Jorge Basin and the Formation is equivalent to the Cotidiano Formation near Lago Fontana. The basin is likely to have been separated from the Magallanes Basin in the south by a topographic high near Lago General Carrera (Figure 1.3).

The bedded limestones are included in the calcareous member category of Suárez and De La Cruz (1994a) and are more common in the south of the research area where they are interbedded with other lithologies, in a series of thinning upwards beds. Suárez and De La Cruz (1994a) considered the calcareous member to be the basal unit of the Toqui Formation and to be equivalent to the Cotidiano Formation of Argentina as defined by Ramos (1981).

Sandstones of the Toqui Formation overlie the calcareous beds at some localities, mostly in the north-east of the region. Elsewhere they form a laterally equivalent succession to the calcareous beds and directly overlie the Ibañez Formation. At Cerro Bayo (Figure 3.1, Location AMT 23) Suárez and De La Cruz (1994a) recorded a 40 metre thick sequence of the arenaceous member of the Toqui Formation which appears as a succession of fine-grained sandstones and sandy conglomerates with a marine fauna, intercalated with thin beds of calcareous sandstones and mudstones.

The pyroclastic beds are composed primarily of tuffs and reworked pyroclastic material, intercalated with occasional calcareous horizons. Suárez and De La Cruz (1994b) recorded variable thicknesses of between 80 and 200 metres for the pyroclastic unit in boreholes from the Toqui mine. Suárez and De La Cruz (1994a) considered the pyroclastic member to be equivalent to the Tres Lagunas Formation of Argentina (Ramos, 1981) although the lithologies are dissimilar. Pyroclastic beds of the Toqui Formation are uncommon in the present study.

3.2 Sampling locations of the Toqui Formation

The shell banks of the Toqui Formation are most common in the north and north-east of the study area and bedded limestones are more frequent south of Coyhaique where they are intercalated with some thin pyroclastic beds. The sandstone generally appears in the east of the region. The outcrops of the Toqui Formation are relatively small and occur sporadically throughout the basin, however they comprise distinctive lithologies.

The Toqui Formation outcrops of this study are shown in Figure 3.1. Good exposures were found at: AMT 9, AMT 10, AMT 16, AMT 19, AMT 22, AMT 23 and ACB 211 and will be described in some detail.

3.2.1 Calcareous shell banks of the Toqui Formation

The calcareous unit is found as shell banks in the north-east of the study area at Estero Pedregoso and Los Mallines (Figure 3.1, AMT 10 and ACB 211).

3.2.1a Estero Pedregoso (AMT 10)

Eight kilometres to the NW of Ñirehuao, at Estero Pedregoso (Figure 3.1, Location AMT 10), the Toqui Formation occurs as a shelly bioherm, the major part of which is made up of a series of banks of oysters (Figure 3.2 and Figure 3.3). The bioherm is more than 250 metres long, stretching in a NE/SW direction and dipping gently to the SE. It reaches a height of 14.5 metres in the central section but is less than eight metres thick on the flanks, which gives it a lensoid appearance.

A section through the central part of the bioherm shows a subtle change in faunal composition from the base upwards (Figure 3.4). At the base of the section is a pebble size conglomerate with a coarse-sand size matrix (bed 1). The clasts are sub-rounded rhyolite, poorly sorted and with a maximum size of 30mm. This bed shows some slight cross-bedding with the direction of palaeocurrents to the east. The maximum measured thickness of bed 1 is 3.4 metres but the base is not exposed.

Figure 3.2: The oyster bioherm of the Toqui Formation at Estero Pedregoso (AMT 10). The maximum thickness is 14.5m in the central section, thinning to 8m on the flanks, resulting in a lensoid shape. The bank is orientated in a NE/SW direction.

Figure 3.3: The flanks of the bioherm showing the sandy, mudstone bed (2) which contains a varied fauna including, bivalves, vertebrate teeth and ostracods. It is overlain by the oyster banks of beds 3 & 4 (Figures 3.4 and 3.5) which are thinner on the flanks of the bioherm than in the central section. (Rule is 0.5m).

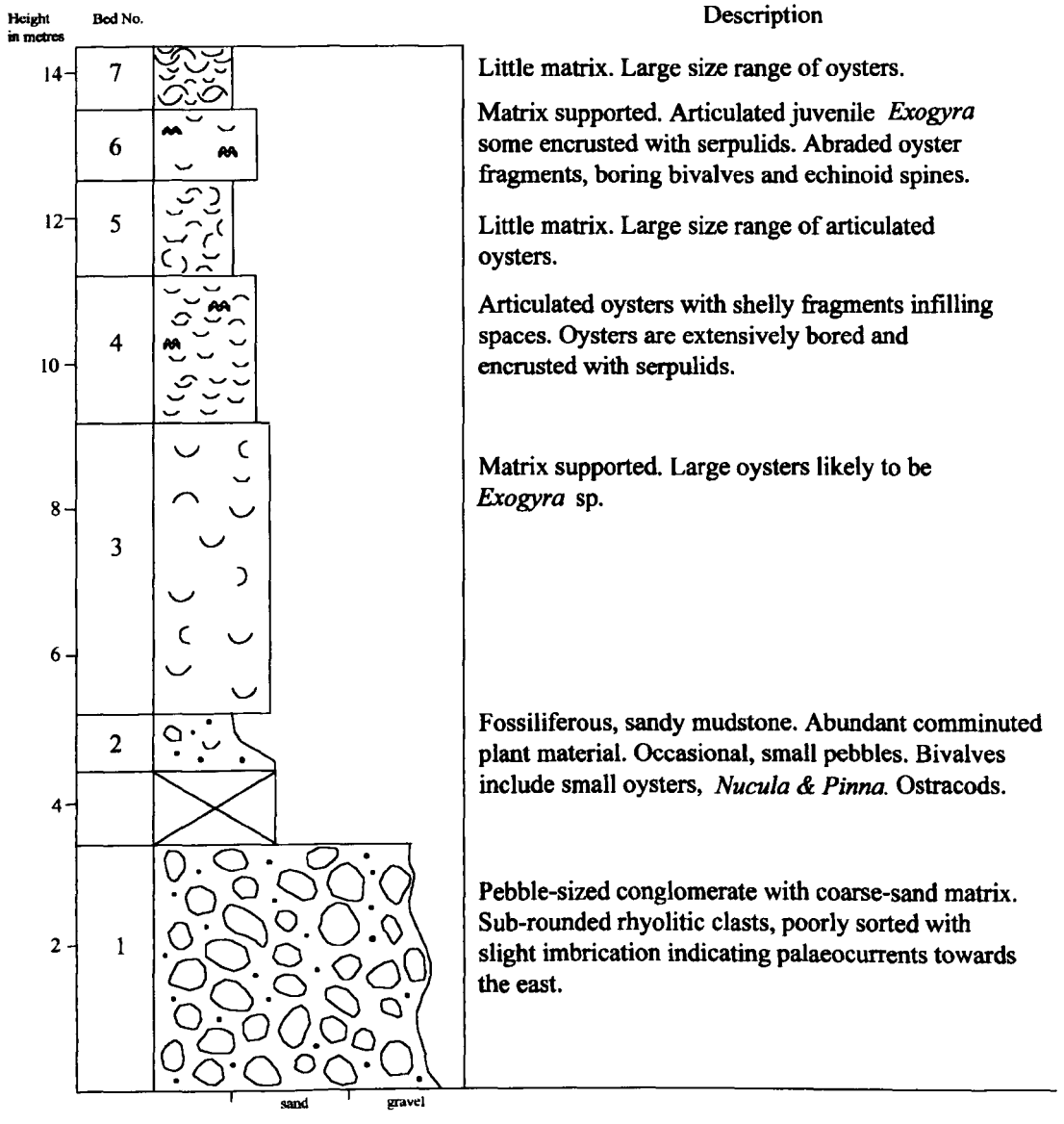

Figure 3.4: Measured section at Estero Pedregoso (AMT 10).

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Overlying this bed is a thin sandy/mudstone unit (Figure 3.3 and Figure 3.4, bed 2), varying in thickness from a few centimetres to a maximum of one metre. This horizon contains a varied fossil assemblage. It is dark-grey to brown in colour, lithified but still quite soft, calcareous and of a homogeneous appearance with no ripples or laminations. The most common fossil found is a small bivalve, averaging 4mm in length and likely to be *Nucula* sp. It is sub-triangular in shape and displays very fine, radial ribs. These bivalves are articulated and show no preferred orientation. Impressions of the bivalve *Pinna* sp. are found, and scattered, small, gryph-shaped oysters and shelly fragments are common. Several, small, unidentifiable vertebrate teeth were found in this bed. The horizon also contains abundant plant material and some small, 10mm, sub-rounded volcanic pebbles. The silty material was processed for microfossils and two species of ostracods, both averaging 1mm in length, have been tentatively identified. The first is a smooth, ovate species with both ends having a rounded carapace but being wider at the posterior end. The ostracod is likely to be *Cytherella* sp. The second species is elongate with rounded ends. The carapace has an ornamentation of fairly strong ridges. This species has been identified as *Cytherelloidea* sp. A further unidentified species was also found. It is ovoid with inflated valves and slightly elongate in lateral view. The left valve is slightly larger than the right with a protruding selvage at the edges of the carapace. The carapace shows a concentric pattern of punctuations which probably define the site of fine spines which have since been lost.

Above the mudstone bed rise a series of five oyster banks (Figure 3.3). The banks decrease in thickness upwards from four metres at the base to under one metre at the top of the sequence. The oysters have inequivalve shells, the right valve is flat to concave whilst the left valve is much larger than the right valve and convex. The left valve has an oval outline, is spirally curved and has a rounded keel. Some examples appear to be spiralled in the early stages of growth and then become straight. The thick shells show foliaceous growth but have no ribs evident. The oysters in all the units are the same, flat lying, gryph-type oyster, likely to be *Exogyra* sp. They are mostly articulated and found in life position, i.e. concave up. All the beds contain a varying amount of a creamy white, micritic cement. A one metre square quadrant was

used to measure the oyster concentrations in each unit, and measurements of shell size and percentage of micrite content were also taken. The results can be seen in Figure 3.5.

Bed 3 (Figure 3.4) is four metres thick with a high percentage (approximately 50%) of a micritic matrix. The oysters show a relatively small size range of between 60mm and 100mm in length, with few juveniles. No other fauna was seen in this horizon. Bed 4 is two metres thick and contains much less matrix, approximately 10% (Figure 3.6a). From Figure 3.5 it can be seen that the oysters range in size between 40mm and 90mm, but the majority are in the 60mm - 70mm size range. The oysters are unbroken and extensively bored. Some shelly material and calcareous tubes of serpulids are found between the whole specimens. Bed 5 is 1.3 metres thick and contains a small amount of matrix, less than 5% (Figures 3.5 and 3.6b). It has the largest size range of oysters from 30mm to 120mm with the average size at 80mm. Bed 6 is one metre thick with 40% matrix and a smaller size range of oysters, averaging 40mm (Figure 3.6c). Some of the oysters in this bed are found concave down which is unusual in this sequence, although 70% are found in life position and articulated. Most show evidence of boring action by other organisms. Between the articulated *Exogyra* are abraded fragments of other oysters. A few specimens of the boring bivalve *Mytiloida lithophaga* were found in this unit as well as some echinoid spines. Some oysters are encrusted with serpulid tubes. Many calcareous tubes of boring bivalves were found within the micritic matrix. The tubes are up to 100mm long and 10mm in diameter and are found intertwined in clumps of six or more (Figure 3.7). Unit 7, the topmost bed, is 0.9 metres thick at its highest point and contains very little matrix. The oysters show a wide range of sizes from 40mm to 150mm with an average size of 75mm. The oysters grow into and on top of each other using their neighbour as a base to fix upon.

In the north and northeast of the basin it was common to find large boulders of the coquina limestone washed down the rivers. These looked identical to the exposure at Estero Pedregoso and indicated that the oyster bank build-up was more extensive than the few outcrops that have been observed.

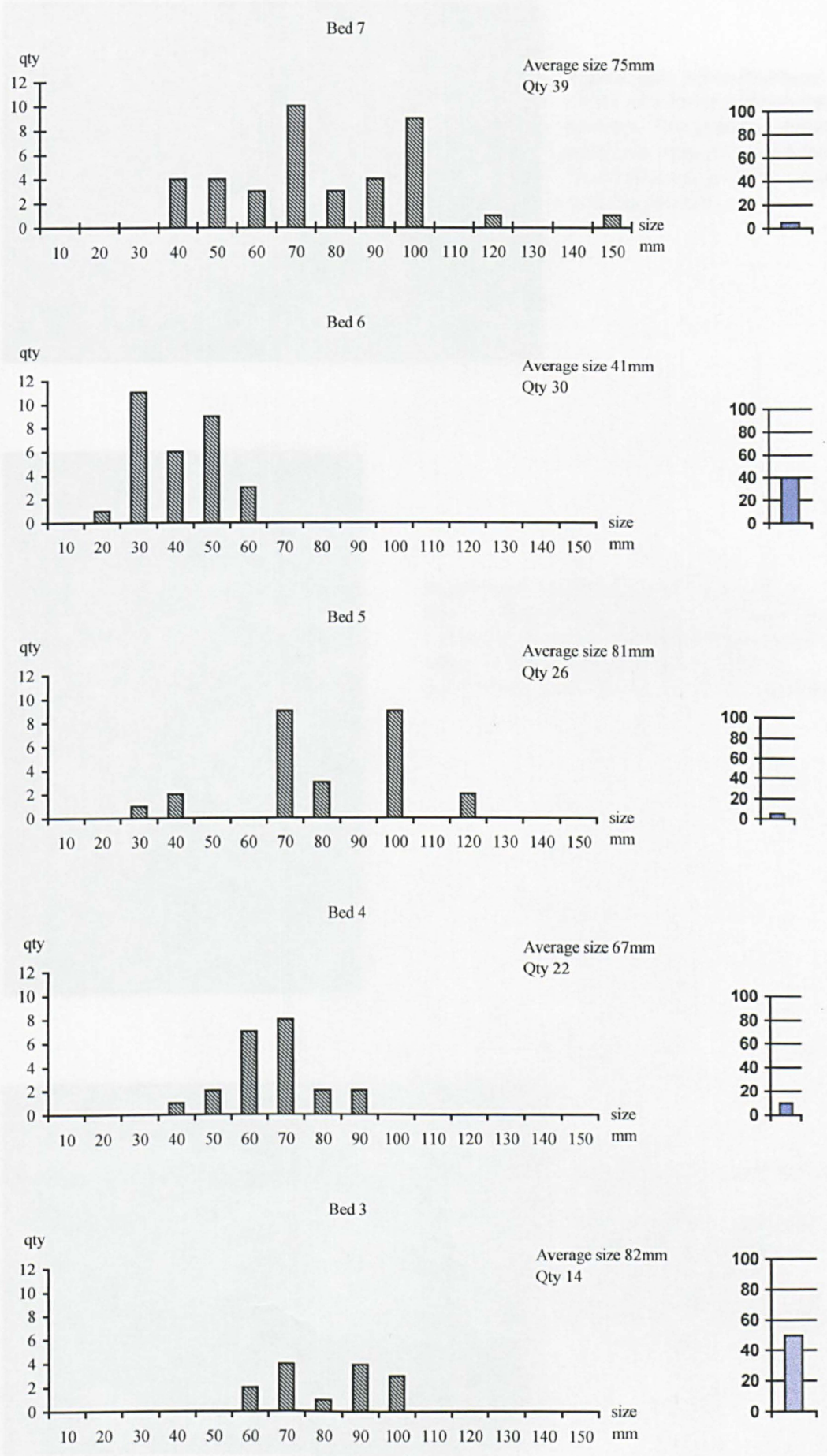


Figure 3.5: Estero Pedregoso (AMT 10). Composition of oyster banks.

Figure 3.6a: Estero Pedregoso (AMT 10). Bed 4 of the oyster bioherm. This bed has a relatively small size range of large oysters which indicates a mature colony with few juveniles.

Figure 3.6b: Estero Pedregoso (AMT 10). Bed 5 contains very little matrix and has a wide size range of oysters, compatible with a thriving colony of juvenile to adult oysters. This is indicative of favourable environmental conditions.

Figure 3.6c: Estero Pedregoso (AMT 10). Bed 6 is matrix supported with a predominantly juvenile oyster population. The boring bivalve *Mytiloida lithophaga* and fragments of echinoid spines were found in association with the oysters.

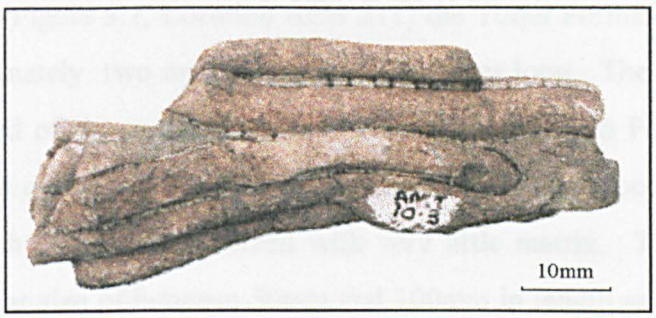


Figure 3.7: Calcareous, infilled tubes of bivalves which bored into the soft sediments and lined their burrows to prevent collapse.

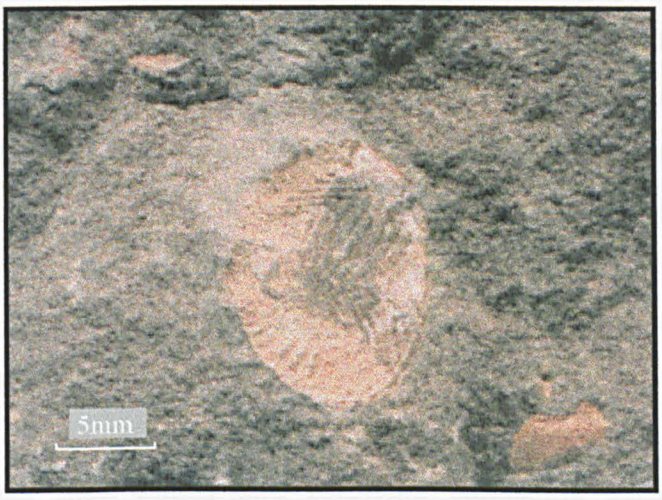


Figure 3.8: Shallowly convex, oval shaped, valve of *Ptychomya koenini* found in a fossiliferous sandstone of the Toqui Formation at ACB 211. Also found in the same horizon were the bivalves *Trigonia* and *Nucula* as well as unbroken gastropods and abundant plant material.

3.2.1b Los Mallines (ACB 211)

At Los Mallines (Figure 3.1, Location ACB 211) the Toqui Formation is present as a low hill, approximately two metres high and 20 metres long. The outcrop is almost entirely composed of the same flat lying oysters seen at Estero Pedregoso, possibly *Exogyra* sp. The oysters are mostly concave up and in growth position, growing on top and into each other and cemented with very little matrix. The shells are of a moderately regular size of between 50mm and 100mm in length and have undergone recrystallisation.

To the west of the oyster bioherm is a two metre thick exposure of parallel-bedded mudstones and shales interbedded with oyster-rich horizons. This outcrop is fossiliferous and contains much organic material. The fossils found include a variety of gastropods and bivalves of various sizes, including some fragments of *Pinna* sp. and *Trigonia* sp. which indicate that they grew to a size of more than 150mm in length. The *Trigonia* belong to the species *Steinmannella* possibly *herzogi* sp. Also found here were some oval shaped, shallowly-convex, bivalves average size 10mm with a distinctive "chevron" pattern. These are identified as the bivalve *Ptychomya koeneni* (Figure 3.8). The mudstone and shale sequence is also found on the eastern side of the oyster bioherm where it contains the same faunal assemblage. Here the beds are overlain by calcareous sands of the Toqui Formation, and contain abundant shelly material and plant fragments, mostly stem fragments up to 300mm long and randomly orientated.

3.2.2 Calcareous shell banks and overlying sandstone beds of the Toqui Formation

The oyster banks of the Toqui Formation are found in association with fossiliferous sandstones in the north-east of the study area at Coichel (Figure 3.1, AMT 16) and Cerro Bayo (Figure 3.1, AMT 23).

3.2.2a Coichel (AMT 16)

At Coichel (Figure 3.1, Location AMT 16) the Toqui Formation occurs in a faulted area and has been interpreted as a horst feature by other researchers (Suárez and De La Cruz, 1994a). The shelly limestone of the Toqui Formation is in contact with volcanics, thought to be of the Ibañez Formation, both above and below. The shelly limestone has a much more varied fauna than at Estero Pedregoso. The large, bored oysters are abundant but interspersed with broken fragments of bivalves, gastropods and echinoids and encrusted by serpulids. The micrite between the shelly fragments contain bundles of intertwined, tubes with a calcareous lining most likely made by boring bivalves. A coarse, volcanic conglomerate with cobble-size clasts up to 200mm in diameter overlies the oyster rich horizon and is interbedded with a coarse sandstone. The conglomerate contains the same shelly debris infilling gaps between clasts, as well as a coral species binding the structure together.

3.2.2b Cerro Bayo (AMT 23)

The small, isolated exposure at Cerro Bayo (Figure 3.1, Location AMT 23) is approximately 10 metres long and five metres high. It overlies, and is in contact with, an ignimbrite of the Ibañez Formation. The outcrop appears as a low hill mainly made up of flat lying oysters, possibly *Exogyra* sp. These oyster shells are found in life position and articulated. The largest measured shell is 150mm long. The shells are very bored, most likely by small bivalves. The matrix is a very fine micrite at the base of the exposure comprising of about 10% of the bed. Some calcareous tubes, likely to be the borings of bivalves such as *Mytiloida lithophaga* sp. are present within the matrix rich horizon. The amount of matrix decreases up the sequence until a negligible amount is present, but broken shell fragments and small shells act as a matrix fill. The top two metres of the outcrop is colonised by the colonial, reef building, branching coral, *Scleractinia* possibly *porites* sp. The oyster concentration decreases as the corals became more dominant.

The sandstone unit is well exposed near Cerro Bayo (Figure 3.1, Location AMT 23), where it is more than 40 metres thick and, at this location, conformably overlies the

calcareous member of the Toqui Formation and underlies the Katterfeld Formation (Suárez & De La Cruz, 1994a). It is principally composed of volcaniclastic material and contains Upper Berriasian marine fossils (Suárez and De La Cruz, 1992). The outcrop appears as a fine to medium grained, creamy-grey volcanic sandstone containing abundant fossil bivalves. Some muddy ripples appear near the base of the exposure and plant fragments and woody material with attached serpulids are common. Fossils include abundant *Entolium* sp. *Nucula* sp. *Astarte* sp. *Chlamys* sp. and *Trigonia* sp.

3.2.3 Bedded limestones alternating with sandstones and pyroclastic beds of the Toqui Formation

To the south of Coyhaique the Toqui Formation appears sporadically as bedded limestones integrated with arenaceous and pyroclastic beds. There is no evidence of the oyster banks found in the northeast of the study area. A major exposure was studied south of Laguna Foitzick (Figure 3.1, AMT 9).

3.2.3a Foitzick (AMT 9)

A distinctive outcrop of the Toqui Formation occurs eight kilometres south of Coyhaique, at Foitzick (Figure 3.1, Location AMT 9). The outcrop extends for approximately one kilometre in a NW/SE direction and reaches a maximum height of about 200 metres. It dips at an angle of approximately 15 degrees to the SW (Figure 3.9). The bedded limestones interfinger with the arenaceous and pyroclastic units. The Ibañez Formation appears at the base of the outcrop as an angular volcanic breccia more than 30 metres thick. This is overlain by a 20 metre thick succession of stratified, flow banded rhyolite. An erosional unconformity separates the Ibañez Formation volcanics from the overlying Toqui Formation. The base of the Toqui Formation is a 30 metre thick, poorly sorted, conglomerate containing large rhyolitic boulders up to five metres in width. Some of the blocks are very angular whilst others are very rounded. This bed is overlain by a series of horizontal, parallel bedded, thinning upwards beds of alternating volcanic sandstones and limestones (Figure 3.10).

Figure 3.9: Foitzick (AMT 9). The beds dip at an angle of approximately 10° SW, which exposes the NE face of the outcrop. The Ibañez Formation occurs as an angular, volcanic breccia at the base of the sequence on the right of the photo. It is overlain by a 20m thick succession of the Toqui Formation which is well exposed on the small hill to the left of the picture.

Figure 3.10: The Toqui Formation at Foitzick (AMT 9). Thinning upwards, parallel beds of limestone and volcanic sandstones, overlying a coarse conglomerate with rotated rhyolitic blocks (R) more than 4m wide. An erosional unconformity (u) separates the Toqui Formation from the underlying flow banded rhyolite (rh) at the top of the Ibañez Formation. (Height of outcrop approximately 30m).

The angle of dip exposes a complete section of the Toqui Formation which is easily accessible and was used to log the succession. The measured section is just over 20 metres high (Figures 3.11, 3.12 and 3.13). The basal bed is a poorly sorted, oligomictic, rhyolitic, conglomerate (Figure 3.11, bed 1). Some of the boulders are up to four metres wide, whilst the smallest clasts are 100mm in diameter. The large boulders are very angular whereas the smaller clasts are very rounded.

The large, angular boulders do not appear in the top metre of the conglomerate bed (Figure 3.11, bed 2). The boulders are encrusted with oysters in life position and, in one case, the oysters extend out into the sediment at either side of the boulder, indicating a palaeo-horizon. There is a calcareous matrix and serpulids and corals are found growing in the gaps between the boulders. Bed 2 (Figure 3.11, Sample AMT 9.1), contains large oyster fragments which have been bored by endolithic algae and other larger boring organisms, the borings have then become infilled with micrite, as in Figure 3.14a. This sample also shows rounded fragments of coralline red algae, which are commonly found in this bed. Echinoid plates are present throughout the lower beds. Figure 3.14b, shows some cellular plant material in the centre of the photograph and at the bottom left, two sutured grains, one of which has been replaced by a drusy sparite cement, but the micritic envelope which formed around the outside of the grain is still visible. There are two generations of cement present in bed 2 and they are both seen in Figure 3.14c. An acicular, fringing cement has formed on the outer edge of the oyster fragment in the centre of the picture. In-between the shelly fragments, a blue-stained, coarse, ferroan, sparite infills the pore spaces.

The overlying five metres (Figure 3.11, beds 3, 4 and 5), belong to the arenaceous unit of the Toqui Formation. These are volcanic sandstones with a calcareous matrix. They contain sub-rounded pebbles up to 50mm in diameter. Throughout the volcaniclastic sandstones, large articulated oysters of *Exogyra* sp. are found. Bed 3 contains many shell fragments, calcareous red-algae and some echinoid spines. Bed 5 (Figure 3.11, Sample AMT 9.6), is part of the arenaceous member of the Toqui Formation and the sub-rounded volcanic clasts are clearly seen. Figure 3.15a, shows sub-rounded, rhyolitic clasts on the left and right of the picture. A calcareous, syntaxial, cement has overgrown some of the bivalve fragments.

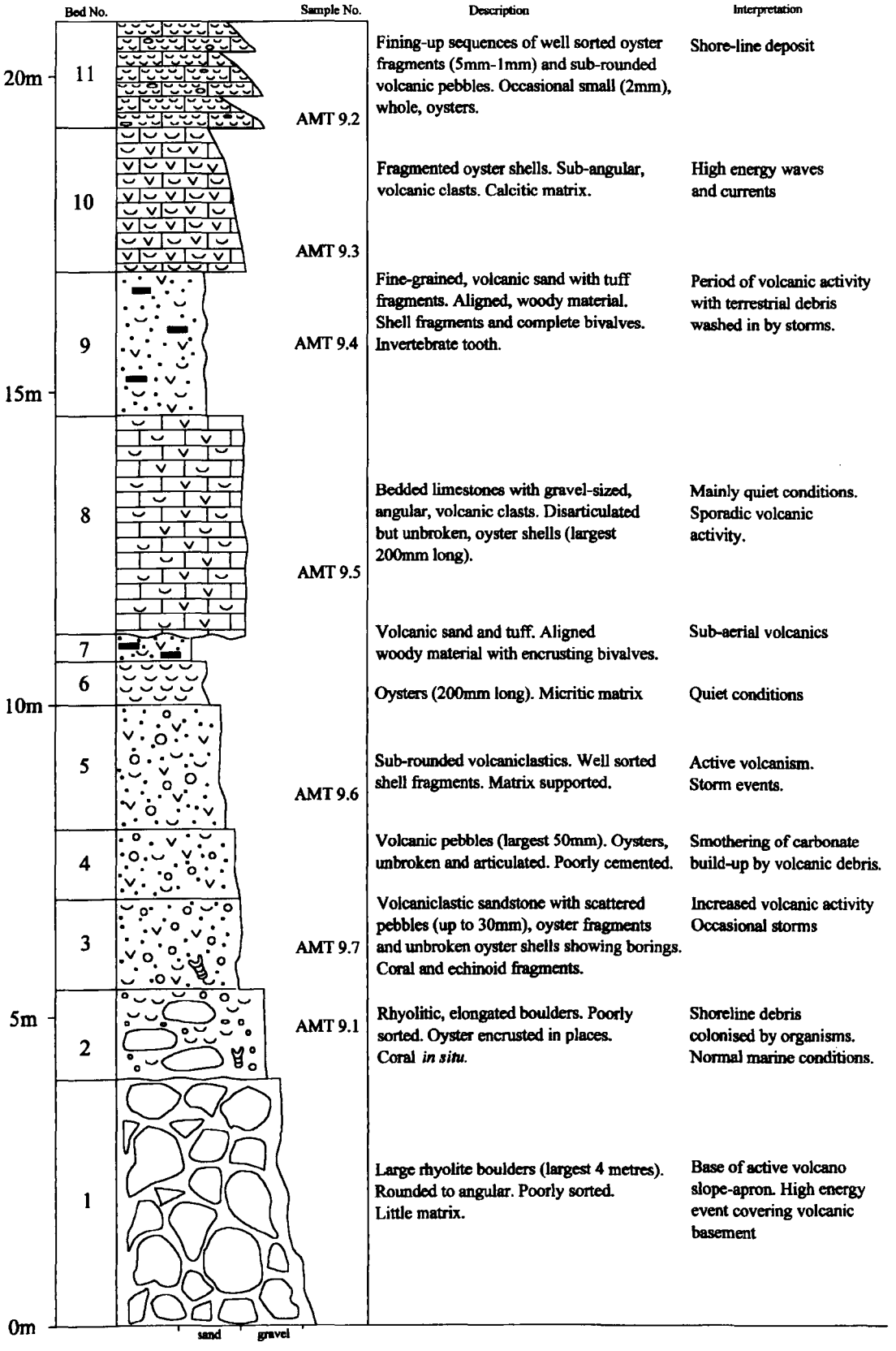


Figure 3.11 Measured section at Foitzick (AMT 9)

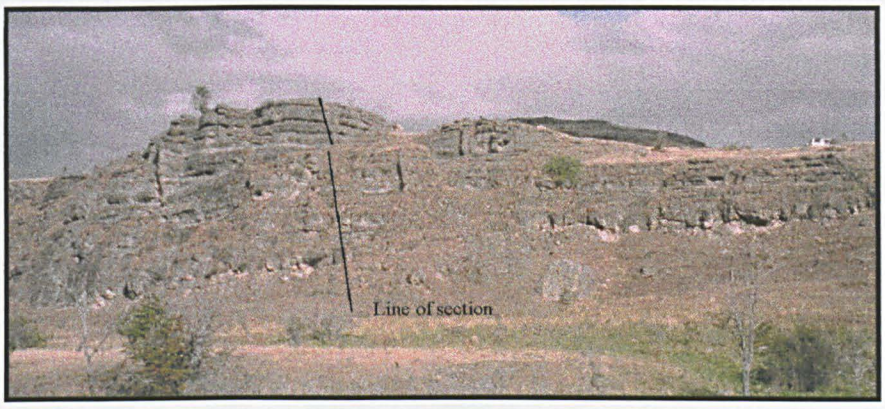


Figure 3.12: Foitzick (AMT 9). Site of the measured section (Figure 3.11) of the Toqui Formation. Height of outcrop approximately 20m.

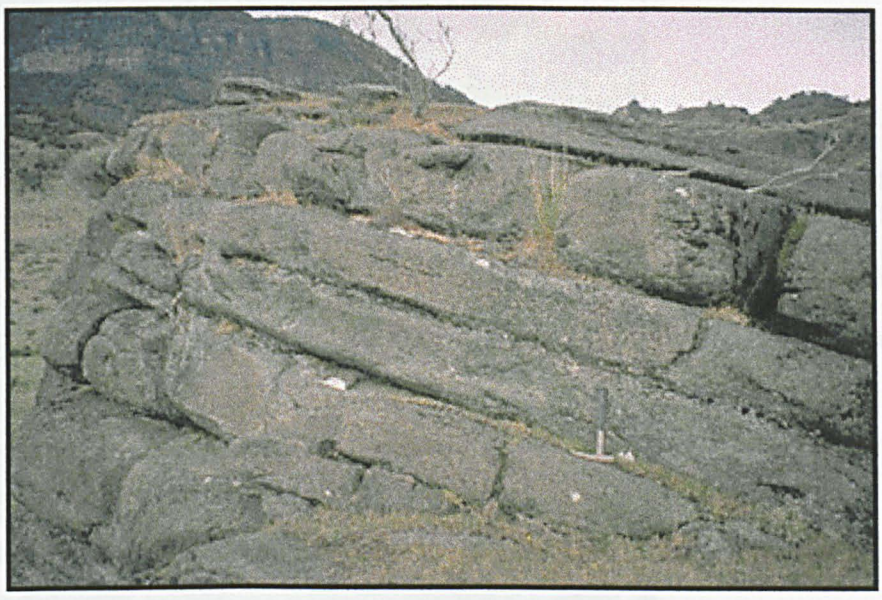


Figure 3.13: The upper beds of the outcrop at Foitzick (AMT 9). Hammer 300mm. A series of thinning upwards sequences of shelly debris (Beds 10 &11 of Figure 3.11). The top 2m of the exposure is composed almost entirely of fragmented oyster shells (5mm-1mm) in fining upwards sequences. The lower 2m also contains fragmented oysters as well as sub-angular volcanic clasts within a calcitic matrix.

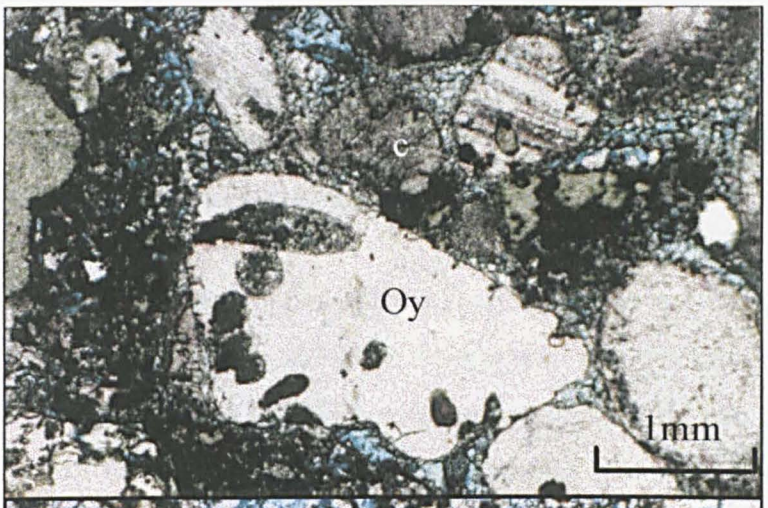


Figure 3.14a (AMT 9.1)

Large oyster fragment (Oy) showing borings of endolithic algae and other larger organisms. The borings have been filled with micrite. Rounded fragments of coralline algae (c) occur throughout the horizon.

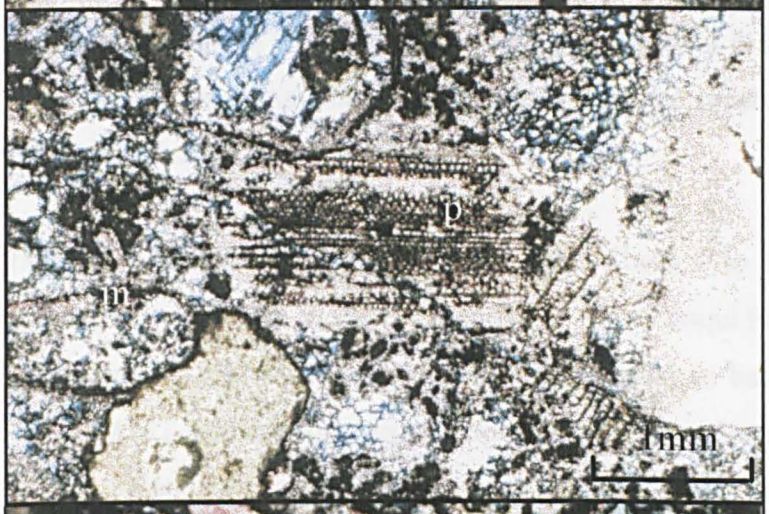


Figure 3.14b (AMT 9.1)

Cellular plant material (p) found throughout this horizon. Clasts showing sutured contacts and micritic envelopes (m) which developed before suturing.

Figure 3.14c (AMT 9.1)

Two generations of cement. A coarse grained, ferroan, sparite cement (Sp) infilling pore spaces between oyster fragments, and an acicular, fringing cement (Ac) around the bioclasts. The fringing cement appears as a rim of fine crystals of equal thickness around the grains.

(All peels stained with Alizarin Red S and potassium ferricyanide)

A dark rim of micrite is seen on many of the grains and shelly fragments. An interlocking, pink stained, coarse calcite cement is seen clearly in Figure 3.15b. Other grains show an asymmetric cement fringe with a radial, fibrous fabric, growing on only one side of some of the volcanic clasts.

Some large oysters, up to 200mm long, appear in bed 6 (Figure 3.11). These oysters have encrusted each other and there is very little micritic matrix between them. The oysters have been bored by other organisms. At the top of the oyster bed is a thin unit of fine grained volcanic sand with small (5mm), angular clasts and 100mm long (average) fragments of woody material which is aligned in an E/W direction. Several pieces of the woody material have bivalves attached (probably Pteriid bivalves), and some show borings of *Teredo* worms.

The top 10 metres of the measured sequence (beds 8 to 11) consists of a series of thinning upwards beds of the calcareous unit of the Toqui Formation interbedded with a two metre thick unit (bed 9) of sandstones similar to bed 7. Bed 8, (Figure 3.11, Sample AMT 9.5) contains large oysters (up to 200mm long) which show evidence of borings by endolithic algae as well as fragmented oyster shells incorporated into the matrix. In this bed there are many examples of angular, volcanic clasts with attached fronds of algae. The pore spaces between the allochems are filled with a microsparite but many of the clasts have an isopachous, acicular, fringing cement as in Figure 3.16a. The volcanic sand of bed 9 contains small, rounded oyster fragments which show micritic envelopes, some aligned, fibrous, woody material, echinoid spines and several vertebrate teeth. The four metre thickness of beds 10 and 11 (Figure 3.11, AMT 9.3 and AMT 9.2) consist of four, fining and thinning upwards sequences (Figure 3.13). These topmost units are composed almost entirely of oyster shell fragments and contain little matrix. The fragments fine upwards from 5mm to 1mm in each sequence. There are occasional very small (2mm) unbroken oysters and some sub-angular to sub-rounded volcanic pebbles. The shell fragments show the same, slight lineation in an E/W direction. In AMT 9.3 (Figure 3.16b) the calcitic cement infills the pore spaces between grains. There is also evidence of dissolution cracks. Figure 3.16c shows an example of the reticulate coralline algae commonly found in beds 10 and 11.

Figure 3.15a: AMT 9.6 (Thin section, plain polarised light). Sub-rounded, rhyolitic clasts (rh), calcareous algae (al) and oyster fragments (Oy) showing micritic rims. A calcareous, syntaxial cement overgrowth (c) can be seen on some of the bioclasts.

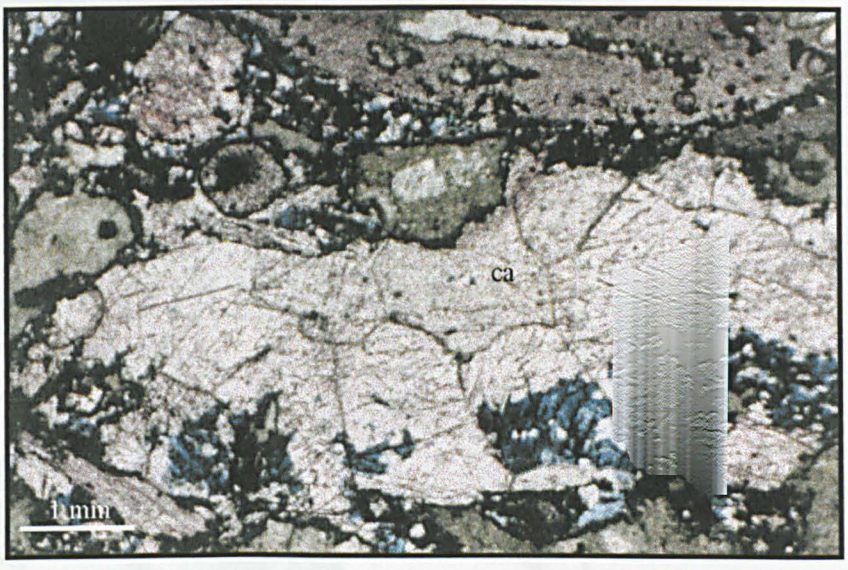


Figure 3.15b: AMT 9.6 (Stained peel)
Interlocking, pink-stained, coarse calcitic cement (ca) infilling pore spaces between allochems.

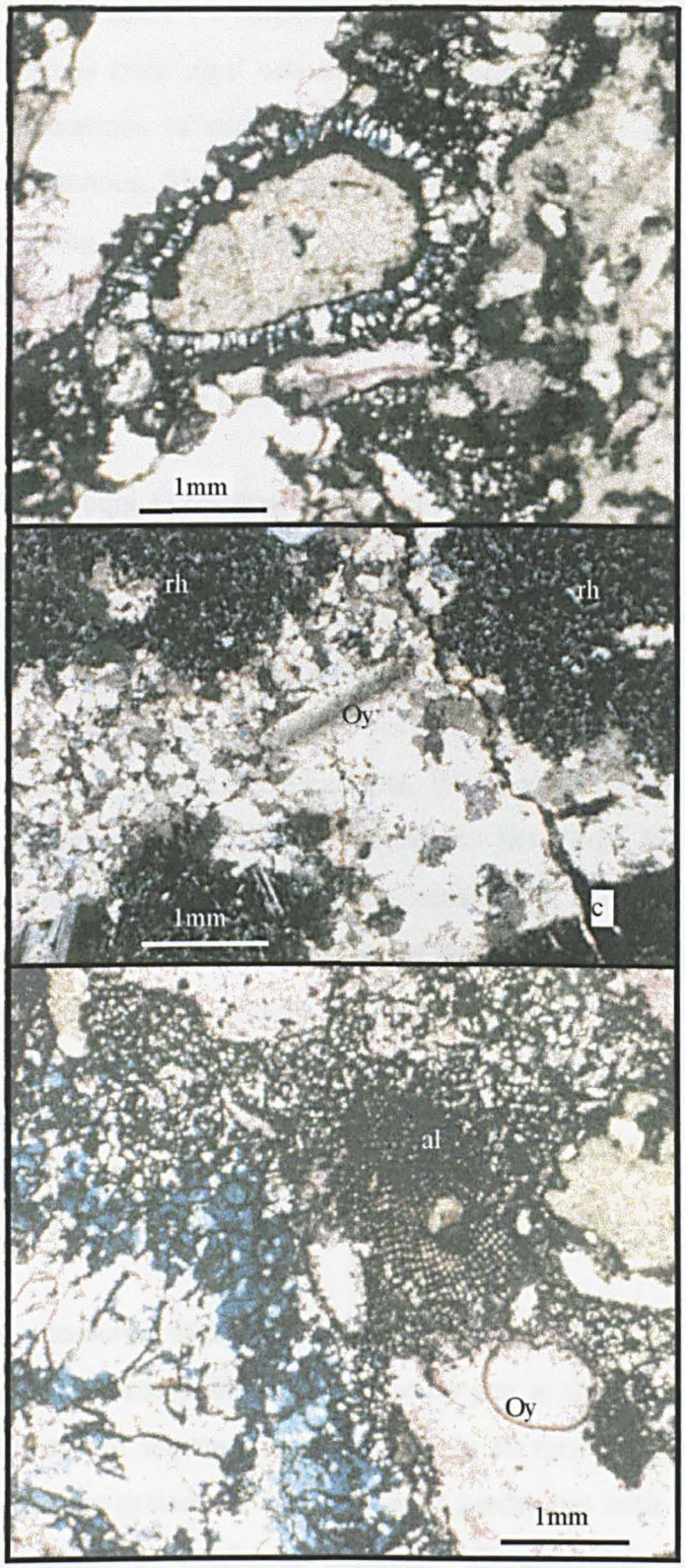


Figure 3.16a: AMT 9.5 (Stained peel) Sub-rounded, micritised clast with acicular, fringing cement.

Figure 3.16b: AMT 9.3 (Thin-section, crossed polars). Rounded, rhyolitic clasts (rh) and oyster fragments (Oy) in a calcitic cement. Removal of the calcite cement by percolating pore-waters has resulted in dissolution cracks filled with clay (c) forming between grains.

Figure 3.16c: AMT 9.3 (Stained peel) Distinctive, reticulate appearance of coralline algae (al). Oyster fragments (Oy) bored at the margins by endolithic algae and then infilled with micrite.

Sample AMT 9.2 contains well sorted and rounded fragments of oysters which show borings from algal activity as in Figure 3.17a. This photograph also shows two generations of cement on the oyster fragment near the top of the picture. An isopachous, fringing cement around the grain and a coarse, ferroan, sparite, cement infilling the gaps between allochems. Figure 3.17b shows the same coarse ferroan cement surrounding coralline algae and an echinoid spine.

3.2.4 Sandstone beds of the Toqui Formation

The Toqui Formation is represented by calcareous sandstones at two sites in close proximity in the Río Coyhaique canyon, to the southeast of Cuesta Los Leones (Figure 3.1, Location AMT 19 and AMT 22). This is the type locality for the Coyhaique Group (Niemeyer *et al.*, 1984) and the three formations of the group; Toqui Formation, Katterfeld Formation and Apeleg Formation, are represented at this outcrop. Suárez & De La Cruz, 1994a, point out that this area is faulted and so the assignment of the basal sequence to the Toqui Formation is tentative but highly probable and accounts for the locations of the two sites. Figure 3.18 shows the deeply incised Río Coyhaique and the faulted area at the type locality. The sandstone lithology is also present at Cerro Bayo (AMT 23) as already described in section 3.2.2b and at Foitzick (AMT 9) where it is interbedded with the other lithologies of the Toqui Formation.

3.2.4a Cuesta Los Leones (AMT 19)

A measured sequence of the calcareous sandstones was recorded from the most southerly accessible point of the canyon at this location (Figure 3.19). The Toqui Formation sequence is approximately 30 metres thick and comprises of a series of very-fine-grained to coarse-grained sandstones with varying mud content which are separated by mudstone partings. Cross-bedding on a 200mm scale and some laminated and carbonaceous horizons occur within the sandier beds. Some sandy-conglomerate beds containing ripped-up mudflakes appear within the sequence. The conglomerates show normal grading from gravel-size to very-fine-sand size. Some of the units show erosive bases whilst others display parallel laminations.

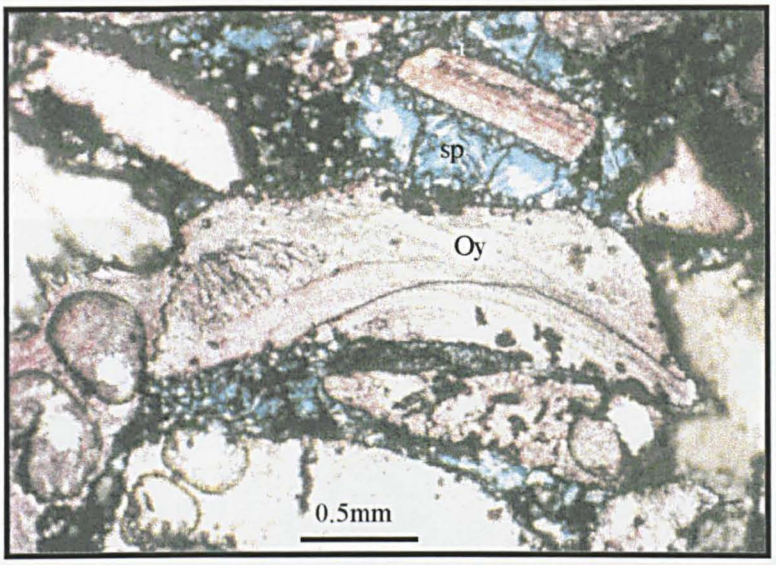


Figure 3.17a: AMT 9.2 (Stained peel).
Oyster fragments bored by endolithic algae (Oy). Two generations of cement can be seen, an isopachous fringing cement (i) around some of the oyster fragments and a coarse ferroan sparite (sp) infilling gaps between bioclasts.

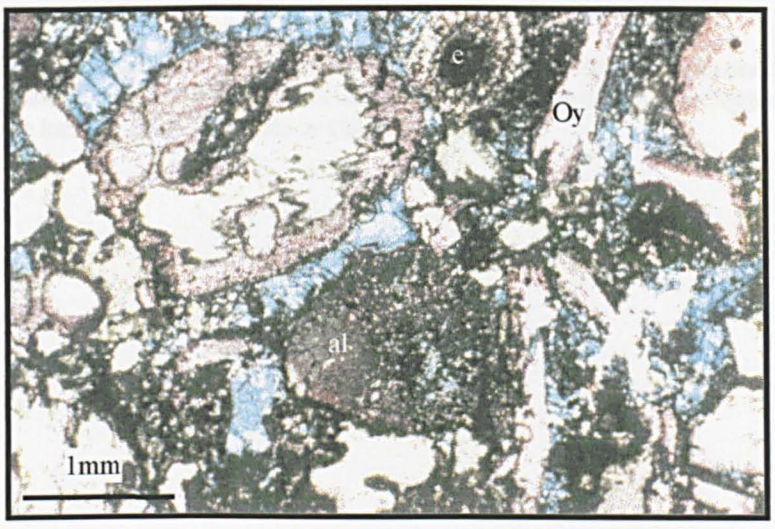


Figure 3.17b: AMT 9.2 (Stained peel).
Oyster fragments (Oy), coralline algae (al) and echinoid spine (e) in a coarse, ferroan cement.

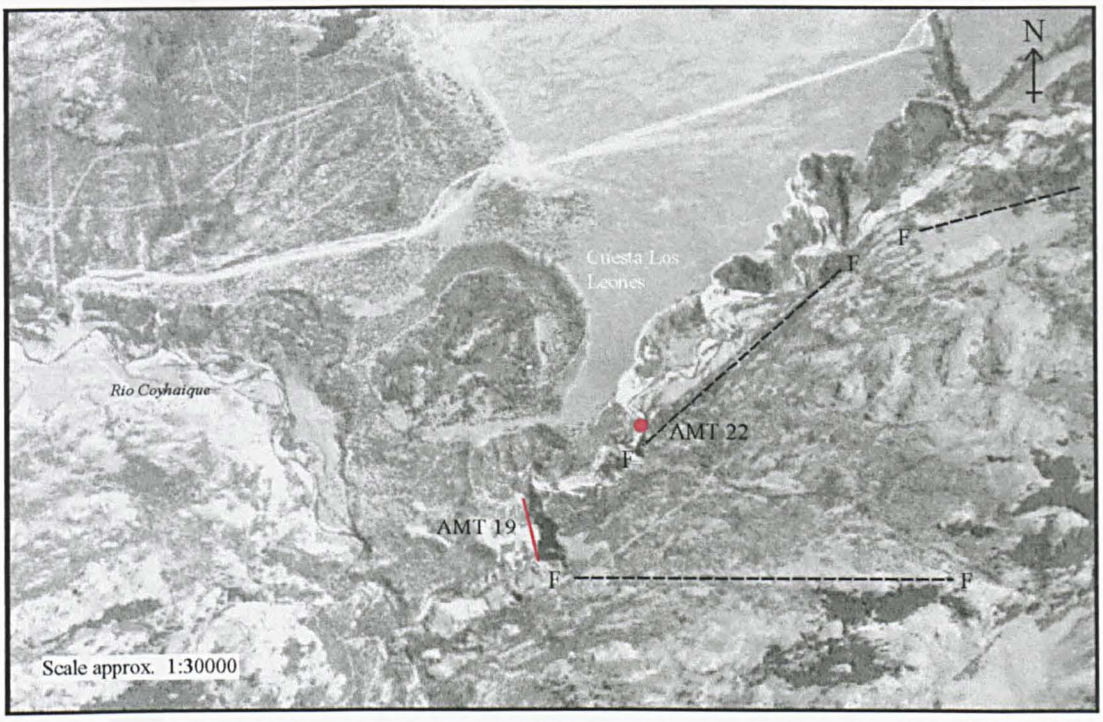


Figure 3.18: Aerial photograph showing the locations of Cuesta Los Leones (AMT 19) and Coyhaique Alto (AMT 22) in the deeply incised valley of the Río Coyhaique. This is the type locality of the Coyhaique Group.

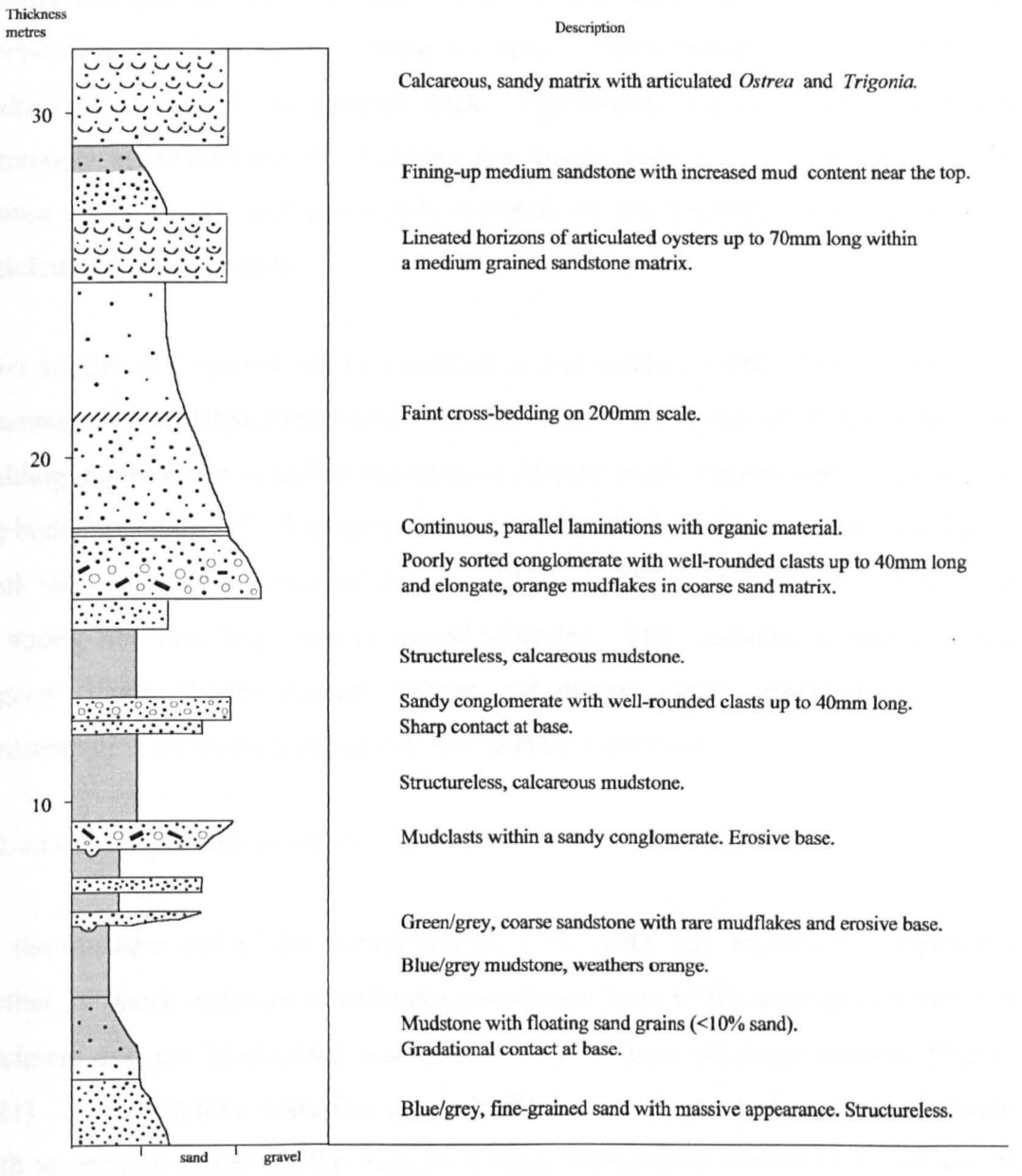


Figure 3.19: Measured section of the sandstone lithology of the Toqui Formation at Cuesta Los Leones (AMT 19)

Twenty-five metres from the base of the exposed sequence, a cycle of bands of recrystallised bivalves occur. There are three bivalve horizons within a medium grained sandstone unit, two metres thick. The bivalves are up to 70mm long and commonly articulated but not showing any strong orientation. The shells are of *Ostrea* and *Trigonia* and are clearly exposed on the bedding surface of a large anticlinal fold found nearby.

Two species of *Trigonia* can be identified at this location (AMT 19). Some loose examples of articulated *Steinmannella herzogi* were found in the scree slopes near the bedding plane, whilst complete examples of *Megatrigonia rogersi* were abundant on the bedding plane itself. *Steinmannella herzogi* (Figure 3.20a) has a distinctive, large, thick shell which is convexly inflated. The flanks of the shell have an ornamentation of strong ribs with large, evenly spaced tubercles. The examples of *Megatrigonia rogersi* (Figure 3.20b) average 100mm and display strong, rounded ribs. The ornamentation produces a distinctive 'fan' shaped expression.

3.2.4b Coyhaique Alto (AMT 22)

At the northern end of the section (Figure 3.18, AMT 22), faulting has exposed a further 10 metre sequence of calcareous sandstone beds which average one metre in thickness and are interbedded with thin (0.3m) horizons of shelly material (Figure 3.21). The sandstone units fine upwards from a coarse grained sandstone, always with an erosive contact at the base, to a hard, fine-grained sandstone at the top of each unit. Mudflakes occur throughout the sandstones and, in the finer grained sands, flasers of the coarser sands are found. The shelly horizons are of *Trigonia* and *Ostrea* and occur in thin layers. The shells are disarticulated but unbroken and show no preferred orientation (Figure 3.22). They are embedded in a coarse-grained sandstone matrix but in-between the shelly horizons there is a very fine-grained area which contains no shelly material.

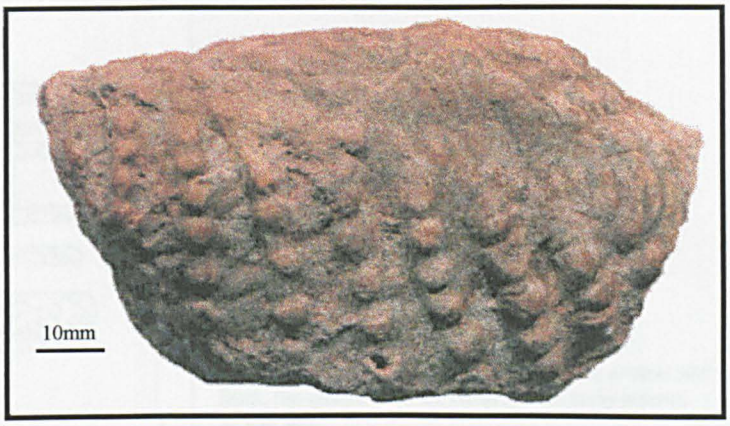


Figure 3.20a: Thick shelled *Steinmannella herzogi* displaying highly ornamented, strong, tuberculate ribs. Collected from the sandstone beds of the Toqui Formation at Cuesta Los Leones (AMT 19).

Figure 3.20b: *Megatrigonia rogersi* on the bedding plane at Cuesta Los Leones (AMT 19). Strong rounded ribs gives a fan shaped outline. (Pen 120mm)

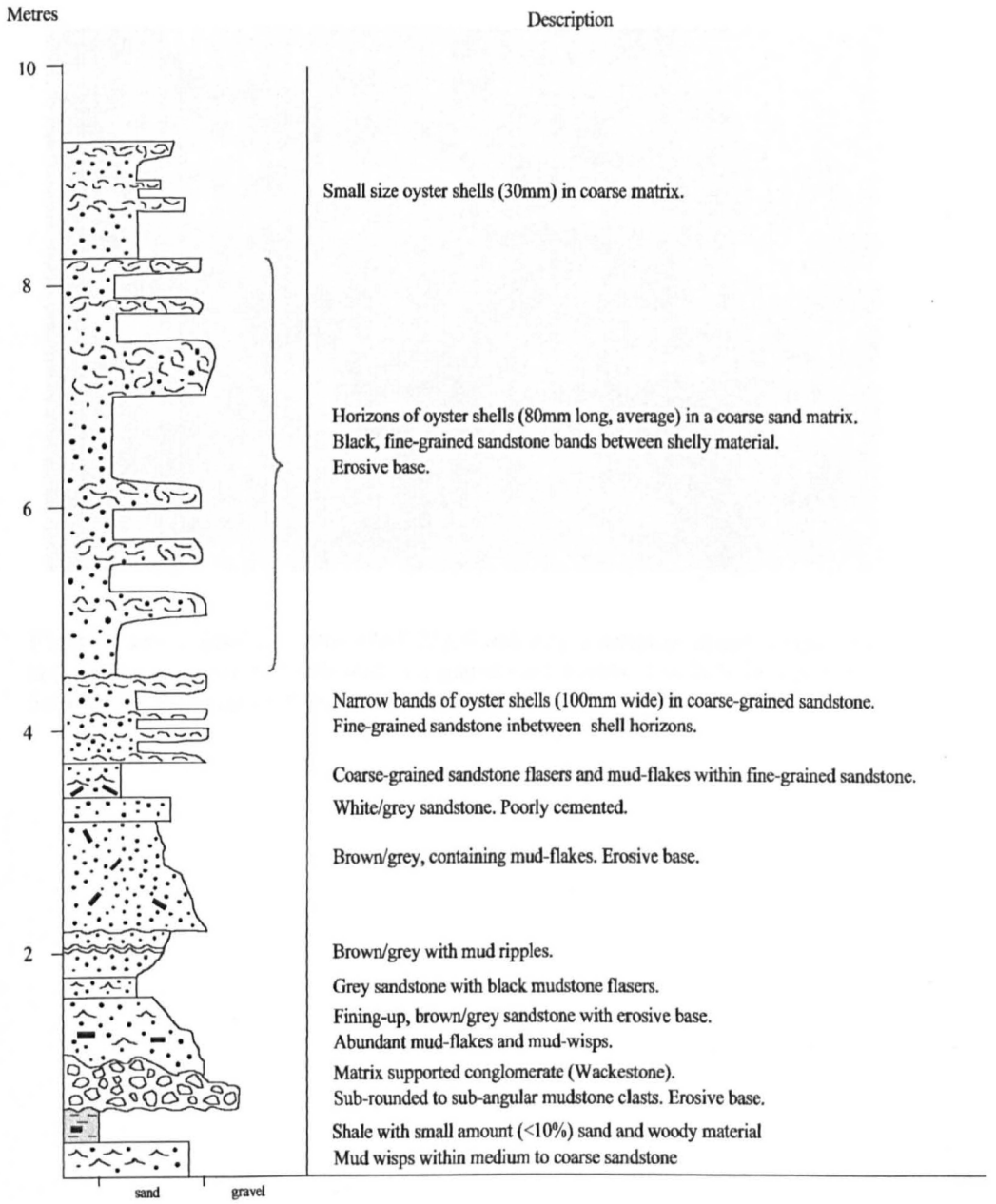


Figure 3.21: Measured section at Coyhaique Alto (AMT 22).

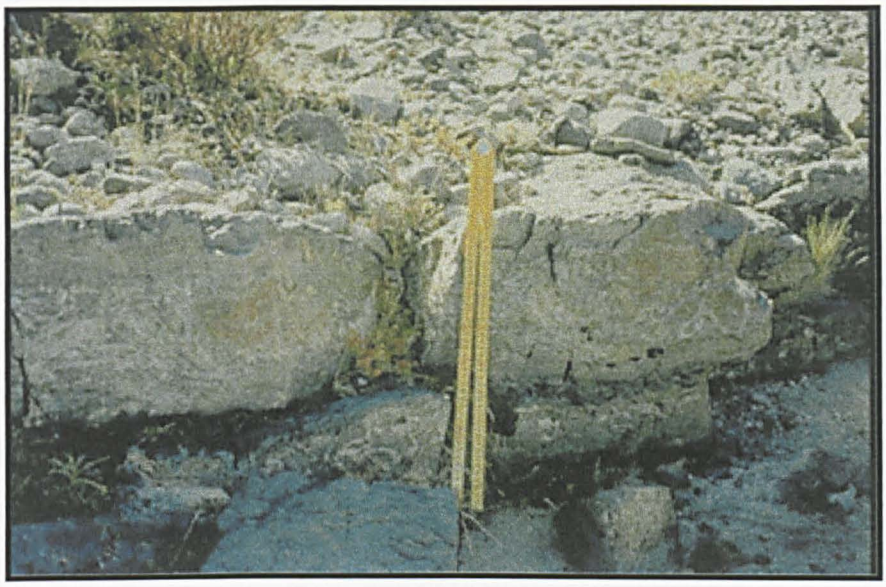


Figure 3.22: Coyhaique Alto (AMT 22). Randomly orientated, disarticulated *Ostrea* and *Trigonia* in narrow bands within a coarse-sand matrix. Likely to be storm beds. Sand beds inbetween shell horizons are fine-grained.

3.3 Interpretation of the Toqui Formation

3.3.1 General overview of the Toqui Formation in the Aysén Basin study area

The Toqui Formation shows a distinct variation in its lithological expression from NE to SW across the study area. Although the exposures are sporadic and discontinuous, it is clear that the extensive oyster banks are only found in the north and NE of the region whilst along the eastern margins of the study area the Toqui Formation is represented by fossiliferous, calcareous sandstones. South of Coyhaique the Toqui Formation appears as a bedded limestone, interbedded with sandstones and volcaniclastic strata and, although containing usually isolated oysters, they do not form the distinctive shell banks seen in the north of the region.

A variety of terms have been used to describe biogenic accumulations. In this work the substantial buildups of oyster banks are termed bioherms. Tucker and Wright (1990), define a bioherm as a small lensoid, biogenic buildup, whilst the term biostrome is related to a more tabular and bedded biogenic deposit. Carbonate buildups are laterally restricted reef mounds which possess topographic relief but lack *in situ* skeletal framework. They are usually rich in bioclastic material or mud.

3.3.2 Interpretation of the calcareous shell banks of the Toqui Formation

The shell banks of the Toqui Formation found in the north and NE of the study area are dominated by a flat lying oyster, possibly *Exogyra* sp. Coarse sandstones and conglomerates are found in association with the oysters at all the study sites.

At Estero Pedregoso (AMT 10) the base of the measured sequence is a conglomerate with a volcanic provenance. Slight imbrication of the sub-rounded clasts of the conglomerate suggests a moderate current in an easterly direction. The overlying, variably thick, soft, sandy mudstone of bed 2 (Figure 3.4) is likely to be a channelised deposit. It contains an assemblage of small bivalves, no larger than 5mm which have been washed in by gentle currents that did not disarticulate them but buried them rapidly. The bivalves were mainly *Nucula* sp. which are shallow burrowers in soft

sediments. The few *Pinna* sp. that were found confirm that the sediment had some sand content to support their semi-infaunal mode of life. Like *Nucula*, they prefer a sheltered, stable environment.

The sandy mudstone also provided some ostracod fauna. Ostracods are useful as indicators of salinity tolerances and sea-floor depths, as well as an important tool in biozonation. The family Cytherellidae, to which both sub-genera *Cytherella* and *Cytherelloidea* belong, is a euhaline form (normal salinity), although known to be able to tolerate a brackish environment (van Morkhoven, 1963). *Cytherella* is found at all depths although *Cytherelloidea* is common in shallower environments. The thick-shelled carapace and ornamentation of strong ribs on *Cytherelloidea* also indicates a shallow, moderately high energy environment. Cytherellidae are benthonic organisms and therefore dependent on the nature of the bottom sediment, which influences the size, shape and ornamentation of the ostracod. The thick shell and fairly strong ribs of *Cytherelloidea* confirms that the substrate had some sand content within the mud and the smooth form of *Cytherella* indicates that it was a burrower in soft sediments. The third, unidentified ostracod was of a spinose form which also suggests a soft sediment where the spines give support to the mass of the organism. Almost all the carapaces found were whole and articulated which indicates a high sedimentation rate (Brasier, 1980), where the organisms were covered quickly before the shells had time to separate or become disarticulated by other scavengers. Both *Cytherella* and *Cytherelloidea* cover a long time range, from Jurassic to Recent (Moore, 1961) and so are not very useful for helping to date the sediments at Estero Pedregoso. They are also a geographically widespread genus. A species of *Cytherella* has been reported by Malumian and Nañez (1983) from the Rio Mayer Formation, south of the study area, near Lago San Martin (Figures 1.3 & 2.6) and both forms are found in the Neuquén Basin (Figure 2.2) to the north of the Aysén Basin (Musacchio and Abrahamovich, 1984). The *Cytherelloidea* species found at Estero Pedregoso has similarities in its shape and ornamentation with *Cytherelloidea agrioensis* which is found in the Neuquén Basin, and this has been dated as Hauterivian.

The undulating nature of the deposit and the varied content of the bed, such as the abundant organic material, and unidentifiable vertebrate teeth, indicate that this was a

channel which cut through the coarser conglomeratic bed below and became infilled quickly with washed-in fauna and detritus as the flow waned. Sedimentation was rapid and covered the bivalves before they had time to disarticulate.

The major part of the Toqui Formation at Estero Pedregoso is the succession of oyster banks. All oysters explicitly like shallow water, no deeper than 80 metres (Clarkson, 1979) and except for a few genera, require warm waters. The shallow depths therefore make them sensitive to fluctuations in temperature (Turek *et al.*, 1988). Flat lying oysters such as *Exogyra* require a firm and stable substratum to anchor onto. If the floor is too soft or muddy, oyster larvae cannot find suitable attachment areas. Oysters are therefore confined to a specific environment where waves and currents are too weak to shift sand and too strong to allow mud to settle (Stenzel, 1971). The conglomerate and sandy mudstone underlying the oyster banks at Estero Pedregoso provided ideal conditions for initial colonisation by the oysters. Once the oyster community had colonised the sediments, succeeding generations used their predecessors as anchor points.

Oyster bioherms have been documented off the coast of North Carolina, U.S.A. (Grave, 1901). In this work it was noted that three to seven generations of oysters grew ontop of each other and died from overcrowding or smothering by sediments. The elongate oyster build-ups formed at right angles to the shoreline and currents, which flowed parallel to the coast. Strong currents rushed past the outer ends and sluggish currents ran along the axis. The oysters at the current end were kept clean of sediment and survived better, obtaining nutrients from the food in the currents. This therefore encouraged growth of the bioherm in a linear direction. The bioherm at Estero Pedregoso most probably evolved in a similar manner. It is elongate in a NE/SW direction which would indicate a proximal shoreline running NW/SE.

The distinctive banks of the Estero Pedregoso bioherm reflect how environmental conditions fluctuated. Size and distribution of the oysters and other organisms within the beds are a key factor in interpreting environmental conditions. Almost all the oysters are in life position i.e. more than 75% are concave up. This applies throughout the bioherm. The beds thin upwards with bed 3 being the thickest at more

than four metres and the top bed, bed 7, less than one metre thick. Apart from serpulids which encrust many of the oysters, the only other fauna that were found were in bed 6 where the boring bivalve *Mytiloida lithophaga* was collected along with fragments of echinoid spines. Many of the oysters show evidence of borings which have become infilled with micrite. Bivalves could be responsible for some of the borings but it is also likely that many were made by sponges, although no evidence of these organisms remain. Sellwood (1978) notes that oysters often comprise more than 90% of the fauna in a bioherm and a similar high percentage would apply to the oyster banks at Estero Pedregoso. Serpulids are frequently found encrusting tops of shells where they take advantage of food-carrying currents. Other annelids burrow into soft sediments and mine the organic detritus within the sediments, leaving networks of calcareous burrows and tubes. Some evidence of these types of annelids are present at Estero Pedregoso but more common are the intertwined, calcareous tubes of boring bivalves (pers. comm. V. Covacevich). It is recognised that generally, burrows are associated with shallow, turbulent water with a high concentration of suspended food (Sellwood, 1978). The crevices in-between shells in many of the oyster beds are filled with debris derived from fragmented oysters which further indicates high energy environmental conditions. The micrite, also found in the crevices and in many of the cavities within the oysters, is composed of lime mud which is expelled after feeding by the oysters themselves. Beds 3 and 6 have a higher percentage of micrite than the other beds and this could account for the slightly higher faunal diversity as the environment became less adverse to other organisms.

Size and distribution of the oysters are key indicators to the environment of these beds. The colonising bed, bed 3, has a large average size of mature oysters with very few juveniles but a high percentage of micritic matrix. It is likely that after the successful establishment of the oyster colony and favourable conditions prevailing for a considerable time, a drop in current velocity resulted in sediments accumulating on the bioherm. The oyster bank quickly became covered in fine muds and this prevented young oysters finding suitable attachment areas. Conditions once more became favourable and the oyster colony re-established itself. Bed 4 shows a similar growth cycle to the previous bed except for smaller oysters and less micrite content.

The bioherm was now well established and bed 5 comprises of a complete range of juvenile to mature oysters in a stable environment with ample nutrients supplied by nearshore currents. Once again conditions changed and perhaps overcrowding caused the mature oysters to die and be replaced by juveniles and the more varied fauna of bed 6. The uppermost bed, bed 7, reflects favourable conditions once again as there is a spread of oysters from juvenile to adult before the complete smothering of the bioherm.

The cause of these cycles of oyster growth on the bioherm could be related to changes in salinity. It is likely that large inputs of fresh water from coastal rivers could have caused reduced salinity in these nearshore areas in which it was impossible for other euhaline organisms to survive. Salinity is one of the main controls on the distribution of aquatic biota (Hudson, 1990). Combinations of high fertility and high stress factors results in brackish water faunas containing few species but many individuals. Fürsich (1981) also found that low values of species diversity result when salinity is considerably below norm. Some species of ostracods, bivalves and gastropods are tolerant of brackish water but are usually of low taxonomic diversity (Hudson, 1990). The Cytherellidae ostracods which were found in bed 2 are known to be tolerant of brackish water (van Morkhoven, 1963) and strengthen the case for lower salinities in the bioherm. Oysters can be divided into two families; Ostreidae and Gryphaeidae. All oysters originated in euhaline waters and to this day Gryphaeidae are restricted to them (Stenzel, 1971). However, among the Ostreidae, groups evolved that were able to penetrate into brackish estuaries and lagoons (Stenzel, 1971). Brackish waters of marginal marine environments are often productive because of nutrients introduced from the land areas. The oyster bioherms with their restricted fauna at Estero Pedregoso comply with a reduced salinity environment although this is not conclusive. Echinoid spines were present in bed 6 and, although echinoids indicate normal salinity, the low density of their skeletal elements means they are transported very easily and may have been deposited on the bioherm by currents. It is likely the bioherm endured a range of salinities and echinoids could have colonised the beds and grazed on algae, which covered the oysters, when salinity was at a normal level (Fürsich, 1981). It is also likely that if the

cause of salinity changes was input from fresh water rivers, that this fluctuation prevented marine organisms colonising the bioherm at times when low values of diversity resulted and the only species that could consistently tolerate the changes were the oysters.

At Estero Pedregoso, the basal conglomerate with a sandy matrix had a provenance in the volcanic terrane to the west of the area. Moderate currents flowing towards the east, deposited the sub-rounded clasts which formed the base of the bioherm. A fairly strong current scoured out a channel in the conglomerate and deposited the sandy mudstone as current velocities dropped. It is possible that this was a distributary channel of a fresh water river which transported organic debris and vertebrate remains. The oyster banks colonised the conglomerate and sandy substrate and exploited the favourable conditions, building a substantial series of oyster banks. Probable salinity fluctuations prevented other organisms colonising the build-up and resulted in low species diversity and large individuals as the oysters exploited their niche in the environment. Each generation of oysters used the previous one as a fixation point to establish themselves and eventually the mature oysters could not obtain enough nutrients and became smothered. Regular cycles occurred which stopped development but as conditions improved the oyster community again became established. The bioherm eventually died probably due to inundation by sediments from a terrestrial source as currents waned.

The smaller outcrop of oyster banks at Cerro Bayo (AMT 23) follows a similar pattern to Estero Pedregoso but individual banks are less well defined. However, at the top of the sequence the oysters are suddenly colonised by a branching coral. The coral is well established and eventually excludes the oysters. This shows that at this location and in this part of the sequence normal marine conditions prevailed. At Los Mallines (ACB 211) the oyster banks are associated with fossiliferous muds and shales containing the bivalve *Ptychomya koenini.* This bivalve has previously been recorded in the San Jorge Basin (Figure 2.2 & Section 2.2.4) by Olivero (1983) and Riccardi (1988) and has been given an age range of Valanginian to lower Hauterivian by the ammonite fauna that was found in association with it.

3.3.3 Interpretation of the inter-bedded limestones, sandstones and pyroclastic beds of the Toqui Formation

At Foitzick (AMT 9), the base of the Toqui Formation comprises a poorly sorted conglomerate containing large, angular, rhyolitic boulders. The palaeosurface of the original topography can be seen in Figure 3.10 and the chaotic formation of this unit is suggestive of a sub-aqueous volcanic apron which received debris from the steep flanks of an intermittently active volcano. The larger, angular blocks have not been transported great distances but the infill between the boulders is composed of smaller, rounded cobbles and pebbles indicative of water-born deposition. The lack of any finer-grained matrix suggests fairly strong currents which would have removed the smaller particles. Tectonic activity and deposition of the angular boulders ceased although strong currents continued to transport and deposit material along the volcanic shoreline. The boulder bed acted as a base for a reefal system and organisms such as corals, coralline-algae, echinoids and bivalves including oysters colonised the topography. The fauna present indicates a shallow sea of normal salinity and a high energy environment. Regular influxes of volcanic material smothered the organisms and prevented the reef from growing, but as conditions stabilised the volcanic debris once again became colonised. Volcanic activity became less regular and enabled carbonate sediments to accumulate in an open platform environment. Periodic storms washed in terrigenous material and sporadic volcanic events produced air-fall tuffs. The fining-up sequences of well-sorted shell beds at the top of the section reflect a high energy environment with strong currents which winnowed the shell fragments. Currents ran along shore in an E/W direction depositing the shelly material as a transgressive beach deposit.

The skeletal grains that are found in the calcareous beds are almost all derived from oyster shells. Many fragments are rounded or abraded, show boring by endolithic algae and are rimmed by micrite envelopes. Endolithic algae, or blue/green algae, are typically found in the shallow marine environment where, after the micro-organisms die, their borings become infilled with micrite (Adams *et al.*, 1984). The disintegration of blue/green, calcareous algae is widely regarded as the main process of micrite production in shallow marine environments (Tucker, 1981). Repeated

boring and infilling results in a dense micrite envelope replacing the outer part of the skeletal grain. Micritic envelopes due to endolithic algae can be used as a depth criterion, indicating deposition in the photic zone, at depths of less than 100 metres (Tucker, 1981). The evidence shows that skeletal debris at Foitzick was transported to the sea-floor by currents where it was attacked by boring micro-organisms. The resulting cavities then became infilled with micrite cement which eventually formed a rim around the shell fragment.

Thin sections and peels taken throughout the measured section show diagenetic processes such as cementation. The carbonate staining process of the peels gives an indication of iron content and distinguishes ferroan from non-ferroan cements which reflect oxidising conditions of pore-waters. A ferroan calcite cement appears blue when stained and indicates reducing conditions which are more likely to occur at depth than near surface (Adams *et al.*, 1984). It is not uncommon to find that cement crystals are zoned, implying changes in the chemistry of the pore-waters. In many limestones early cement generation is composed of non-ferroan calcite, while the main cement phase, sparite, is ferroan (Tucker, 1981). In many of the samples two generations of cement can be distinguished, a thin fringe of fibrous calcite around the grains or bioclasts and a coarse sparite infilling pore spaces. Sparry cements therefore post-date the acicular fringing cements.

The fibrous calcite cements are usually isopachous and these are recognised as precipitating in a marine environment below the water table (phreatic zone) where the pores are constantly filled with water (Tucker, 1981), possibly in a sub-tidal setting. Occasional grains did show an asymmetric fringe of the fibrous calcite and this is likely to reflect precipitation in the vadose zone where the cement formed on the underside of the grains as water percolated through. The non-ferroan character of fibrous calcites is consistent with a syn-sedimentary, submarine or inter-tidal origin, while the ferroan nature of many sparry calcites is explained by precipitation in the meteoric, phreatic zone where reducing conditions prevail (Evamy, 1969). Sparite is the typical cement of near-surface, meteoric diagenesis but it can also precipitate in a deep burial environment, perhaps beneath 100s metres of sediment, then it would be

associated with fractured fibrous cement fringes which are not seen in the Foitzick location.

There is little evidence of grain or bioclast fracture in the thin-sections but Figure 3.14b shows grain to grain compaction which occurred after micritisation. Stress is concentrated at the points where the grains meet. Some evidence of compaction is seen in the selective dissolution of the calcium carbonate material leading to the inter-penetration of one grain by another resulting in a concavo-convex contact. This is the first stage of pressure solution. Compaction is only slight and the original porosity of the beds is still present. Where pressure solution is more intense dissolution veins occur as in Figure 3.16b, usually as a result of overburden or tectonic pressure (Boggs, 1987). A residue of insoluble material is left behind, usually parallel to the bedding, as the calcium carbonate is reprecipitated as a cement elsewhere. Reprecipitation is considered to be one of the main sources for limestone cementation specifically of ferroan sparite cements (Tucker, 1981). Cementation is most widespread in areas of high current activity such as along shorelines where sea-water is pumped through the sediments (Tucker & Wright, 1990). It is unlikely there was a large overburden in the Foitzick sediments as there is little evidence of grain fracture. It is more likely that the sparry calcite precipitated in a near-surface meteoric environment, perhaps after a slight, relative sea-level fall, so that rainwater affected the carbonates.

The Toqui Formation at Foitzick is representative of a reef and high energy shoreline of an active volcano.

3.3.4 Interpretation of the sandstone beds of the Toqui Formation

The sandstone lithology occurs mainly in the northeast of the study area. It is composed principally of muddy volcanic sandstones with occasional sandy conglomerate beds and contains marine fossils.

At Cuesta Los Leones (AMT 19), the sandy conglomerates with mudflakes and erosive bases are likely to have been deposited by turbidity currents. The graded bedding, cross-bedded units and parallel laminations, recorded in Figure 3.19, are also characteristic of turbidite horizons (Tucker 1981, Boggs 1987). The turbidite sequences alternate with calcareous mudrocks which were deposited from suspension as flow waned. The well-graded sandstone beds and thin, fine-grained nature of the units suggest a low-density, distal turbidite (Tucker, 1981). The distinctive shelly horizons of trigonias and oysters seen at Coyhaique Alto (AMT 22), are likely to be event beds laid down as storms re-worked local shell banks similar to those seen at Estero Pedregoso (AMT 10). Previous authors (Reyes 1970, Suárez & De La Cruz 1994a) interpreted them as *in situ* bioherms but upon closer examination it is clear that the thin bands of shells are randomly orientated and many are disarticulated. The shelly horizons have a coarse-sand matrix but the intervening areas are very fine-grained carbonate sands. This characteristic is typical of settling out of material from suspension after storms subside. Similar pavements of oysters in a carbonate mud have been recorded from Cretaceous deposits in Kansas U.S.A. by Kidwell (1991). The Kansas beds are interpreted as an event concentration of storm re-working of local biostromes in an open, epicratonic sea.

At Cerro Bayo (AMT 23) Covacevich (personal communication, 1994) identified *Megacucullaea* and the ammonite *Neocosmoceras* sp. from which he deduced the Upper Berriasian age of these deposits. In earlier work Charrier & Covacevich (1980) described the fossiliferous sandstone of the Toqui Formation at Cerro Bayo. They recorded cross-bedded sandstones with a fauna of bivalves including *Entolium, Chlamys, Astarte* and *Trigonia* as well as the ammonite *Neocosmoceras.* A similar fauna which included *Neocosmoceras* was collected by Olivero (1983) from outcrops of the Tres Lagunas Formation in Chubut (Figure 2.11) and also recorded in the Springhill Formation at Lago San Martin (Figure 1.3) by Riccardi (1977). Although the same bivalve fauna was collected at Cerro Bayo, no examples of *Neocosmoceras* were found and so the age of the Toqui Formation sandstones cannot be confirmed at this location.

Trigonias are well documented in southern Chile and in particular around the Coyhaique location (Reyes 1970, Reyes & Perez 1979). A systematic collection of examples of trigonia from Coyhaique Canyon at Cuesta Los Leones (AMT19) was made but not documented by Fuenzalida during the 1950s and 1960s. The collection was re-examined and recorded by Reyes (1970) when he discussed the stratigraphical significance of the species found. Fuenzalida had assigned the trigonia he found near Coyhaique to mid-Valanginian due to their affinity with African trigonia. Reyes (1970) re-evaluated the fauna which included *Megatrigonia rogersi, Iotrigonia stowi* and *Steinmannella herzogi*. All these forms are found in the marine strata of South Africa and range from upper Valanginian - lower Hauterivian. Reyes (1970) also noted their association with the ammonite *Favrella* which he considered Hauterivian and which was found in the overlying sediments and so he assigned an age of late Valanginian - early Hauterivian to the Toqui sandstones. However, Riccardi (1970) attributed *Favrella* an age of Hauterivian - Barremian whilst Leanza & Leanza (1973) concluded it was Aptian. A further study was made at the same location by Reyes & Perez (1979) who documented *Megatrigonia, Iotrigonia* and *Steinmannella* which they interpreted to be Hauterivian - Barremian, in line with Riccardi's evaluation of the age range of *Favrella*. Leanza & Leanza (1973) had made their observations on *Favrella* from the Neuquén Basin which were of a different species to the ones found in the Aysén region and were possibly of a different stratigraphical range. Reyes (1970) also pointed out that trigonias in northern Chile and Peru have a different form to those from southern Chile whilst Crame (1984) notes that trigonias as well as other fauna found in Antarctica are very different to those found in southern South America.

All southern South American genera of Lower Cretaceous trigonia have an Indo-African affinity (Reyes & Perez, 1979). The link is further strengthened by the common occurrence of gryphae oysters which are completely absent in Antarctica but common in South African faunal lists (Crame, 1984).

3.3.5 Summary of the Toqui Formation

The Toqui Formation occurs as small, discontinuous outcrops throughout the study area but it is clear that the formation varies from NE/SW across the region.

To the north the Toqui Formation most commonly occurs as oyster banks or bioherms. The overall picture of the bioherm environment is one of persistent moderate to high energy, in warm seas that could support a suspension feeding community. Influxes of fresh water from nearby land areas brought abundant nutrients in the form of organic material but also caused fluctuations in salinity levels which restricted the faunal community that was able to colonise the bioherm. At Estero Pedregoso (AMT 10), the orientation of the oyster bank indicates the margin of a shallow basin, less than 80 metres deep, where fairly strong currents ran parallel to a NW/SE shoreline.

To the NE of the study area, near the Argentine border, the Toqui Formation appears more commonly as calcareous sandstones. In Coyhaique Canyon (AMT 19 & AMT 22), the alternating, thin, normally-graded sandstones and inter-bedded calcareous mudrocks are characteristic of low-density distal turbidites and the frequent occurrence of storm beds of oysters and trigonia indicate high energy events re-working local shell banks. This location was deeper and further into the basin than the oyster banks and consequently salinity was more likely to have been at normal levels to support the trigonia and other fauna.

In the SW of the Aysén Basin study area at Foitzick (AMT 9), three lithologies of the Toqui Formation are present. Bedded limestones overlie the conglomerates and sandstones and are inter-bedded with some volcanic sand horizons. The fauna indicates a warm, shallow sea of normal salinity which allowed a reef to form. Depths were never more than 100 metres and always in the photic zone. Several episodes of cementation occurred but it is unlikely that the deposits experienced a deep burial environment. The Toqui Formation at Foitzick formed in predominantly high energy reef and shore-line environments, near a sporadically active volcanic source which marked the final stages of the Ibañez Formation volcanism.

The lack of any correlatable stratigraphic fossils in the Toqui Formation makes it difficult to be precise about the age of the deposits. Nowhere in the Toqui Formation of the study area was the Hauterivian ammonite *Favrella* collected, although other authors have reported it near the top of the sequence and Covacevich described an occurrence of the upper Berriasian ammonite *Neocosmoceras* from near the Argentine border. The trigonias identified by Reyes & Perez (1979) were abundant at Cuesta Los Leones and similar South African trigonid give an age range of upper Valanginian - Hauterivian. This correlates with the tentatively proposed age distribution of the bivalve *Ptychomya koenini* and the ostracod fauna from Estero Pedregoso (AMT 10).

The Toqui Formation possibly ranges from upper Berriasian to lower Hauterivian but is more likely to have an upper Berriasian date.

CHAPTER 4

THE KATTERFELD FORMATION

The Katterfeld Formation is the middle unit of the Coyhaique Group (Table 2.2). It is rarely seen to be in contact with the underlying Toqui Formation in the study area except at Los Baguales (Figure 4.1, AMT 3) where it overlies a volcanic tuff of the Toqui Formation. Suárez & De La Cruz (1994a) confirm that the Katterfeld Formation has a conformable contact with the Toqui Formation in the Aysén Basin, but that it directly overlies the Ibañez Formation at some locations, although the contact is poorly exposed.

The Katterfeld Formation was first described from Cerro Katterfeld, Argentina by Ramos (1976). He identified a thick succession of marine black shales very near the border of Chile and Argentina and to the southwest of Lago Fontana (Figure 2.11). The shales had previously been included in the Coyhaique Formation of Niemeyer *et al.* (1984). Equivalent, marine, black shales also occur in the far south of Argentina, in the Magallanes Basin, where they are known as the Río Mayer Formation. They outcrop extensively throughout the Magallanes Basin and decrease in thickness towards the north of the basin where they have a measured thickness of 700 metres (Riccardi, 1988). The fossils that are found within the Río Mayer Formation and the underlying Springhill Formation correspond closely with the fauna found within the Katterfeld Formation black shales.

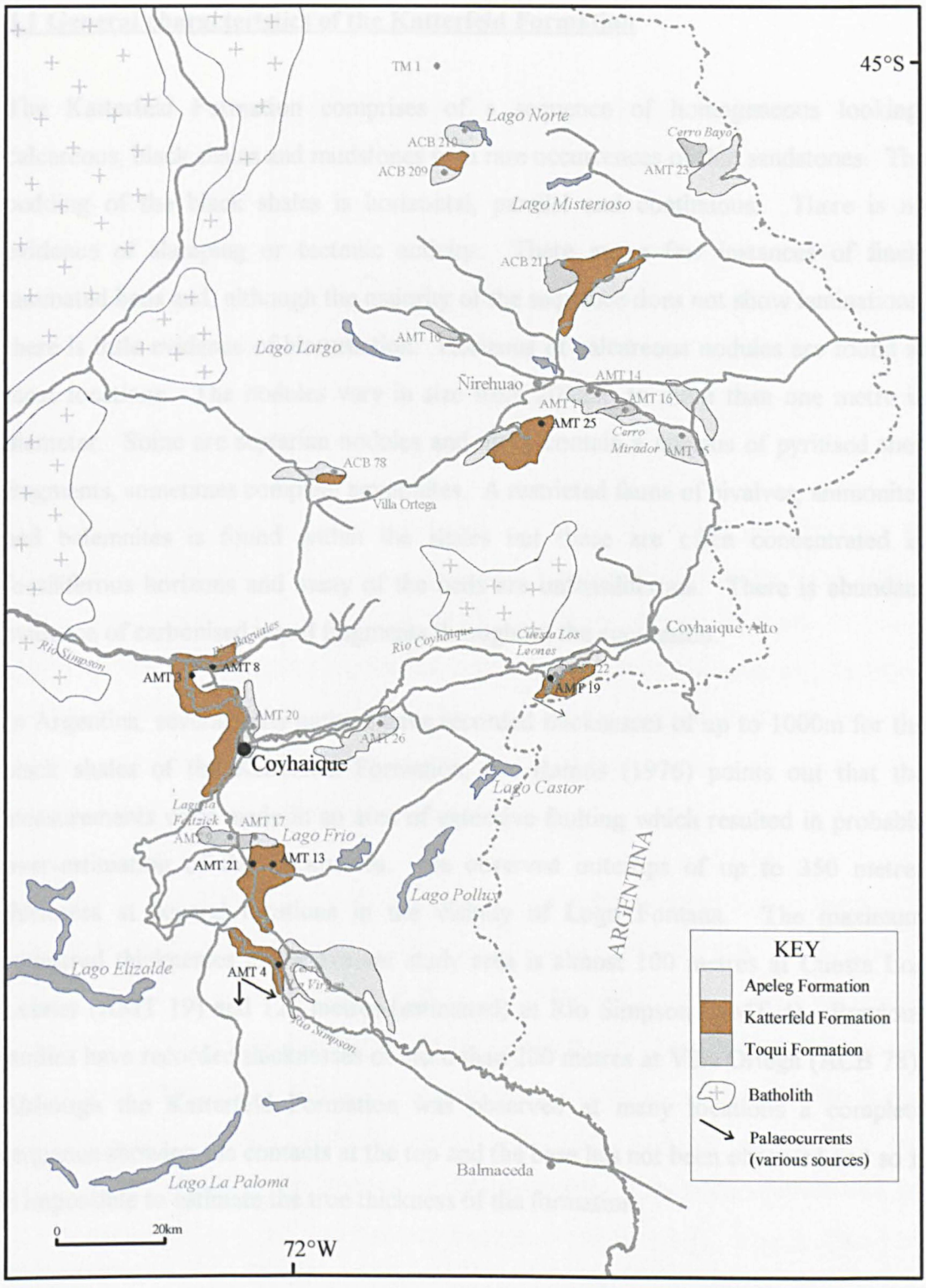


Figure 4.1: The study area highlighting major outcrops and sampling sites of the Katterfeld Formation.

Basin Margins: The Katterfeld Formation was deposited over a regionally extensive area during the Late Berriasian to Upper Hauterivian. The Somuncurá Massif formed the northern margin of the basin and the western margin was defined by a volcanic chain which restricted the opening to the Pacific Ocean. The basin extended well into Argentina to the east where a thick sequence of the Katterfeld Formation was identified near Lago Fontana (Figure 2.11) as pro-delta deposits (Ramos & Palma, 1983). The basin had a restricted opening to the south and the Magallanes Basin where equivalent strata is known as the Rio Mayer Formation.

4.1 General characteristics of the Katterfeld Formation

The Katterfeld Formation comprises of a sequence of homogeneous looking, calcareous, black shales and mudstones with rare occurrences of thin sandstones. The bedding of the black shales is horizontal, parallel and continuous. There is no evidence of slumping or tectonic activity. There are a few instances of finely laminated beds and, although the majority of the sequence does not show laminations, there is little evidence of bioturbation. Horizons of calcareous nodules are found at most locations. The nodules vary in size from 200mm to more than one metre in diameter. Some are septarian nodules and many contain a nucleus of pyritised shell fragments, sometimes complete ammonites. A restricted fauna of bivalves, ammonites and belemnites is found within the shales but these are often concentrated in fossiliferous horizons and many of the beds are unfossiliferous. There is abundant evidence of carbonised wood fragments throughout the succession.

In Argentina, several early authors have recorded thicknesses of up to 1000m for the black shales of the Katterfeld Formation, but Ramos (1976) points out that the measurements were made in an area of extensive faulting which resulted in probable over-estimation of the thicknesses. He observed outcrops of up to 350 metres thickness at several locations in the vicinity of Lago Fontana. The maximum measured thicknesses in the present study area is almost 100 metres at Cuesta Los Leones (AMT 19) and 125 metres (estimated) at Río Simpson (AMT 4). Previous studies have recorded thicknesses of more than 200 metres at Villa Ortega (ACB 78). Although the Katterfeld Formation was observed at many locations a complete sequence showing the contacts at the top and the base has not been observed and so it is impossible to estimate the true thickness of the formation.

4.2 Sampling locations of the Katterfeld Formation

In the Aysén Basin the shales outcrop extensively throughout the study area, mainly in road-cuttings and river valleys. The most important localities within the study area are; Los Baguales (AMT 3), Baguales Vista (AMT 8), Río Simpson (AMT 4) and Cuesta Los Leones (AMT 19), the type locality of the Coyhaique Group in Coyhaique Canyon. Smaller outcrops at La Fraille (AMT 13), El Salto (AMT 21) and El Tranquilo (AMT 25) also offered some fossil evidence.

4.2.1 Los Baguales (AMT 3)

The Katterfeld Formation is well exposed at two locations within one kilometre of each other, Los Baguales (AMT 3) and Baguales Vista (AMT 8) (Figure 4.1 & Figure 4.2), approximately 10 kilometres north of Coyhaique. At Los Baguales approximately 70 metres of interbedded shales and mudstones outcrop in a steep sided gully (Figure 4.3 & Figure 4.4). The loose, scree slope and dense vegetation in the bottom of the gully made access difficult but it was possible to log all of the section in some detail from various access points.

The base of the measured section (Figure 4.5) is a hard grey/cream volcanic tuff containing small pyrite nodules. This bed is more resistant than the overlying mudstones and shales and creates a platform at the base of the gully which the stream cuts down to. Above the tuff are 15 metres of interbedded blue/grey mudstones and brown/grey shales which gradually become more fissile up sequence. The lower mudstone unit contains non-calcareous, iron-coated nodules but neither the mudstones nor shales of the lower units contain fossils. Approximately half-way up the measured sequence, the beds become fossiliferous (Figure 4.5). Slightly distorted, turreted and globose forms of gastropods are found in abundance, with a few burrowing bivalves, in a grey limestone. The laminated mudstone and fissile shale above the limestone unit also contain abundant gastropods including trochiform (turreted) and globose naticiform species, possibly *Bathrotomaria,* which has a flat base and side exhalent opening. A few burrowing bivalves are also present as well as much organic matter.

At 36 metres above the base of the measured sequence, two highly fossiliferous units occur. The lower unit contains a varied fauna of burrowing bivalves up to 60mm long in a reddish/grey, calcareous mudstone matrix (Figure 4.6). The bivalves include *Pleuromya, Protocardia, Tancredia, Modiolus, Astarte* and *Trigonia.* The overlying unit (Figure 4.7) is even more fossiliferous but with much smaller specimens of infaunal, burrowing bivalves between 2mm and 5mm long. The bivalves include *Astarte* and *Protocardia* and are of a uniform size with very little matrix between them. Above the two fossiliferous units the outcrop is a homogeneous, orange/black, fissile shale with fewer bivalves than the previous beds but these include *Astarte, Nucula* and *Grammatodon,* and near the top, a distinctive horizon containing *Entolium.*

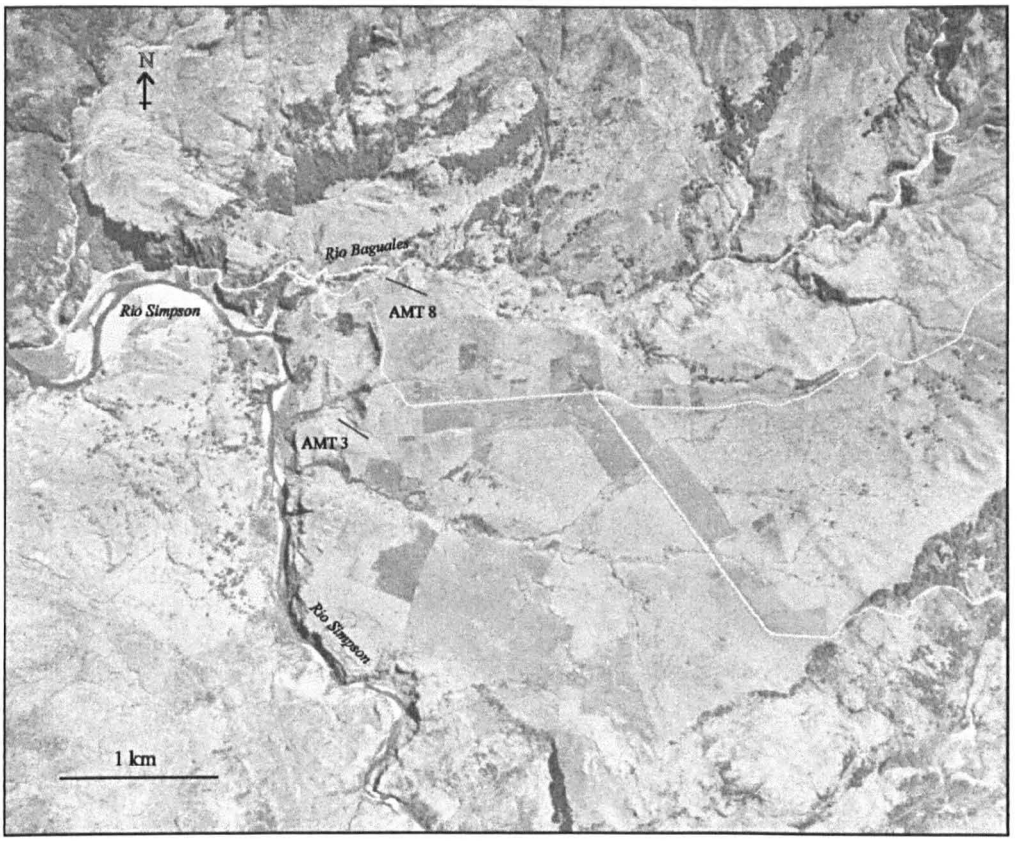


Figure 4.2: Aerial photograph of Los Baguales (AMT3) and Baguales Vista (AMT 8). Their locations are less than 1 km apart.

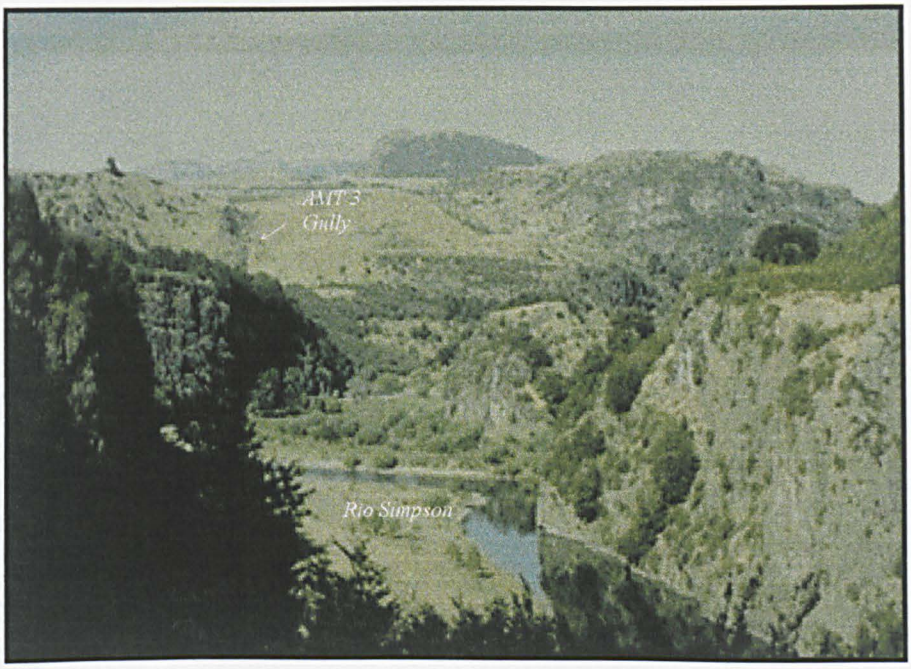


Figure 4.3: The gully at Los Baguales (AMT 3) looking towards the SE across the Río Simpson valley.

Figure 4.4: The exposure of the Katterfeld Formation shales and mudstones at Los Baguales (AMT 3). Outcrop approximately 70 metres high.

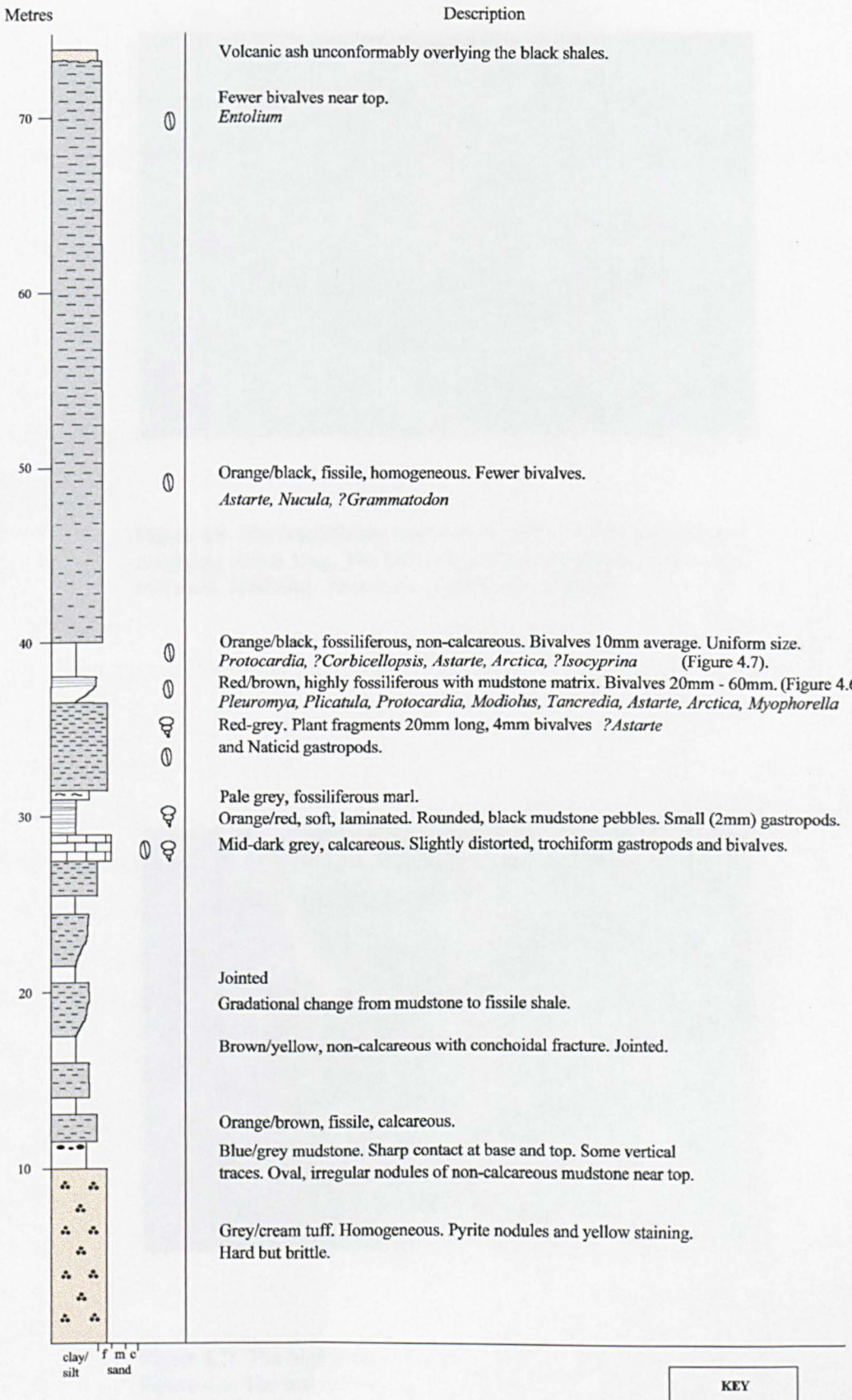


Figure 4.5: Measured section at Los Baguales (AMT 3).

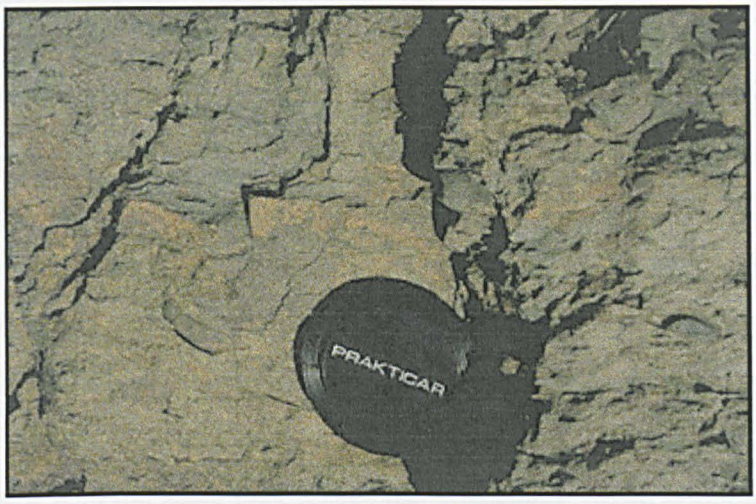


Figure 4.6: The fossiliferous mudstone at AMT 3 containing bivalves averaging 40mm long. The burrowing bivalves include; *Pleuromya, Plicatula, Modiolus, Tancredia, Astarte* and *Trigonia.*

Figure 4.7: The highly fossiliferous unit overlying the bed shown in Figure 4.6. The bed consists of small, infaunal bivalves of a uniform size (average 10mm). The bivalves include *Protocardia* and *Astarte.*

4.2.2 Baguales Vista (AMT 8)

The Baguales Vista outcrop is less than 1km northeast of Los Baguales gully (Figure 4.1 & Figure 4.2). Approximately 50 metres of parallel bedded, black shales and mudstones outcrop in a road cutting near the Río Baguales bridge (Figures 4.8, 4.9 & 4.10). The road dips through the beds over a distance of about 1km and so it is possible to sample throughout the sequence. At the most easterly end the shales are cut through by vertical andesitic dykes (Figure 4.9) which have been dated to the end of the Cretaceous period. The westerly end of the measured sequence (Figure 4.11) starts near the bridge with the same cream/grey tuff containing pyrite nodules, seen at Los Baguales. This area is highly faulted and the beds are distorted. Some septarian nodules almost one metre long are found in the calcareous shales. They are ellipsoidal and aligned parallel to the bedding planes but contain no evidence of fossils. The black shales vary in fissility, sometimes becoming less fissile but hard and blocky. Some small (60mm) nodules with an iron-oxide coating and containing a fine-grained, orange sand are found in the lower beds (Figure 4.11). Large pieces of woody, carbonaceous material are found throughout the section often associated with iron pyrites. The less fissile beds contain occasional limid and pecten type bivalves including *Oxytoma* and *Meleagrinella* and abundant *Entolium.* Many impressions of large ammonites up to one metre in diameter are present in the upper beds of the measured section (Figure 4.11). Some show bifurcating ribs but all can be attributed to the *Favrella* genus. Two different species of belemnites were collected in the same beds as the ammonites, *Belemnopsis aucklandica* and *B. extenuatus.* The black mudstones do not show any evidence of laminations nor much presence of bioturbation. The only trace fossils seen are small areas of a fine, branching system, *Chondrites* and these are quite common in the homogeneous, hard, black mudstones. At 20 metres from the base of the measured sequence is a one metre thick, fining-up sandstone unit with sharp contacts at the base and top. Higher up sequence, at 35 metres, a thinner, conglomeratic bed with an erosive base occurs. It is composed of poorly sorted and poorly cemented, rounded, volcanic clasts (10 - 100mm diameter) in a fine-grained, calcitic matrix.

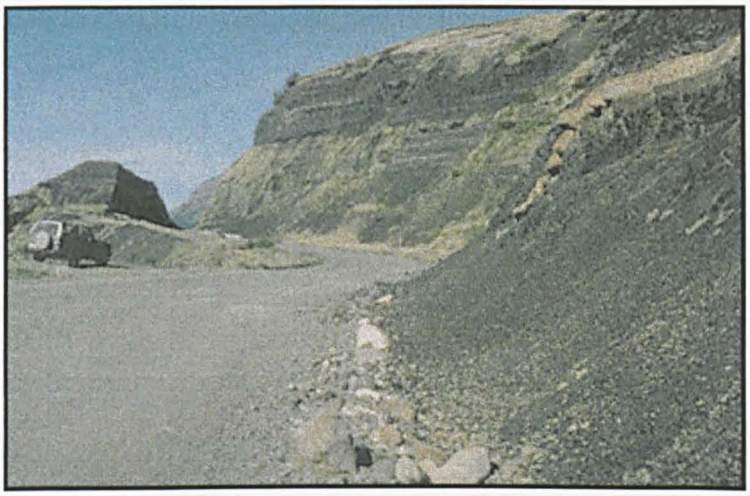


Figure 4.8: Outcrop of the Katterfeld Formation at Baguales Vista (AMT 8). Parallel bedded, black shales and mudstones approximately 50 metres high. Volcanic sill in foreground.

Figure 4.9: Vertical, andesitic dykes dated to the end of the Cretaceous period cutting through the interbedded shales and mudstones of the Katterfeld Formation.

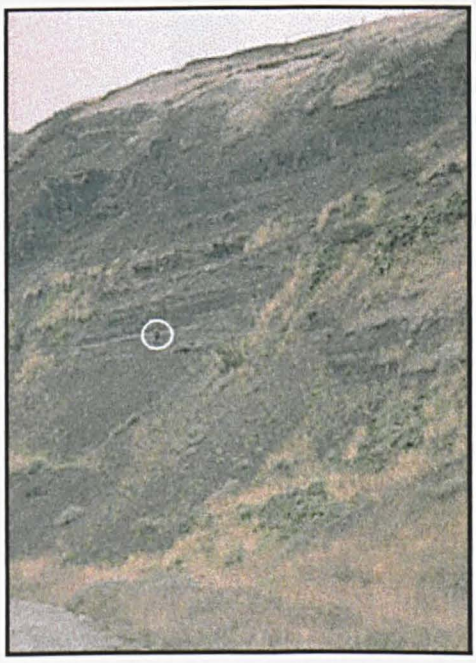


Figure 4.10: Parallel bedded, fissile shales and hard, black mudstones of the Katterfeld Formation. (Figure circled for scale). The ammonites and belemnites were collected from the hard, prominent band approximately 10 metres above the figure.

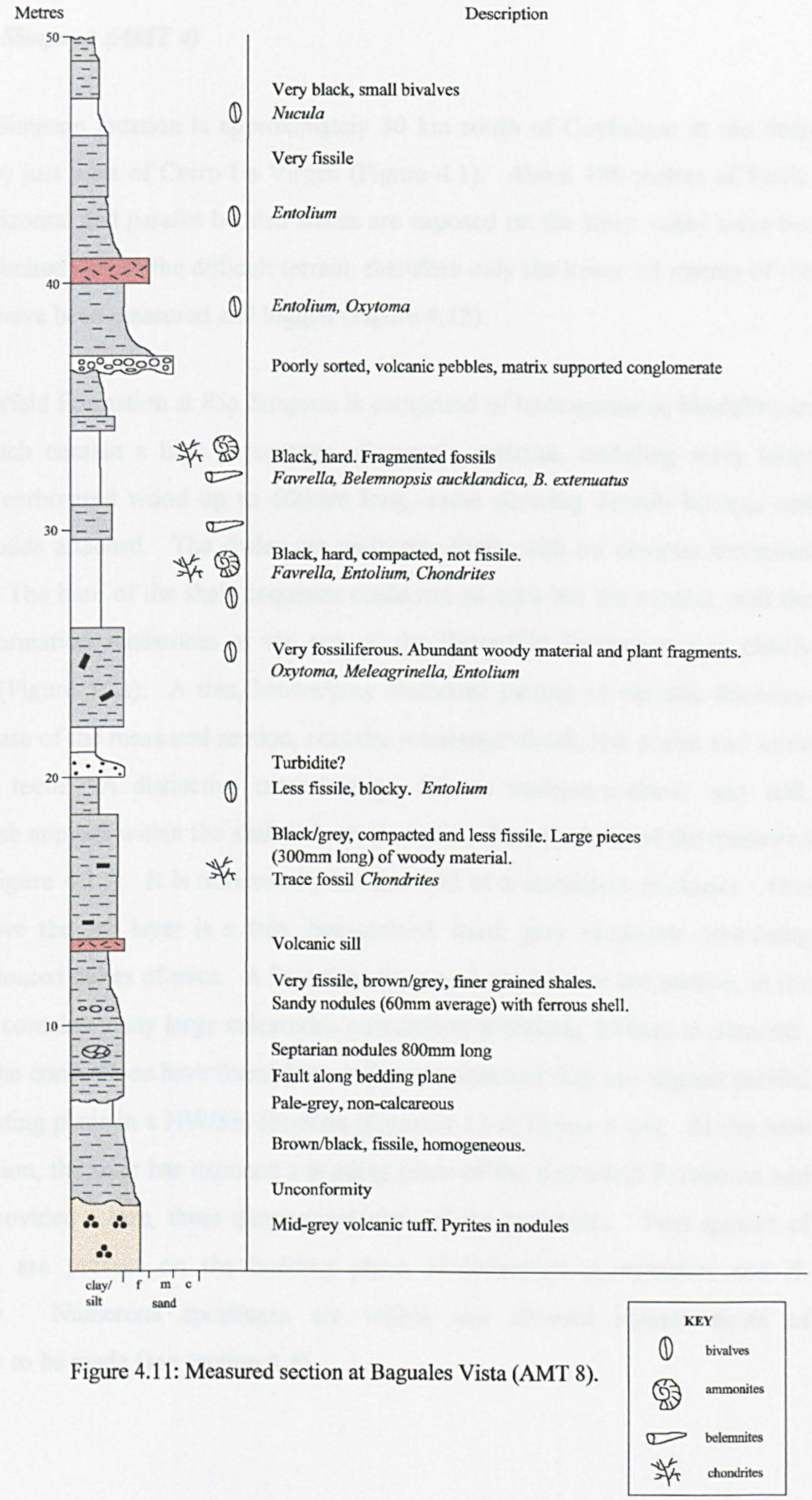


Figure 4.11: Measured section at Baguales Vista (AMT 8).

4.2.3 Río Simpson (AMT 4)

The Río Simpson location is approximately 30 km south of Coyhaique in the deep river valley just west of Cerro La Virgen (Figure 4.1). About 100 metres of fissile, black, horizontal and parallel bedded shales are exposed on the steep valley sides but access is limited due to the difficult terrain, therefore only the lower 18 metres of the exposure have been measured and logged (Figure 4.12).

The Katterfeld Formation at Río Simpson is composed of homogeneous, black/brown shales which contain a high percentage of organic material, including many large pieces of carbonised wood up to 500mm long, some showing *Teredo* borings and with serpulids attached. The shales are uniformly fissile with no obvious laminated horizons. The base of the shale sequence could not be seen but the contact with the Apeleg Formation sandstones at the top of the Katterfeld Formation was clearly observed (Figure 5.2a). A thin, brown/grey mudstone parting of variable thickness near the base of the measured section, contains numerous bluish, fish scales and some vertebrate teeth. A distinctive cream/orange, fine to medium-grained, very soft, volcanic ash appears within the shales about six metres from the base of the measured section (Figure 4.12). It is horizontally bedded and of a consistent thickness. One metre above the ash layer is a thin, fine-grained, hard, grey sandstone containing bronze coloured flakes of mica. A limestone band near the base of the section, in the river bed, contains many large calcareous concretions averaging 600mm in diameter. Many of the concretions have formed around ammonites and they are aligned parallel to the bedding plane in a NW/SE direction (Figure 4.13 & Figure 4.14). At the base of the section, the river has exposed a bedding plane of the Katterfeld Formation and this has provided a rare, three dimensional view of the sediments. Two species of belemnites are present on the bedding plane, *Belemnopsis aucklandica* and *B. extenuatus*. Numerous specimens are visible and allowed measurements of orientation to be made (see section 4.4).

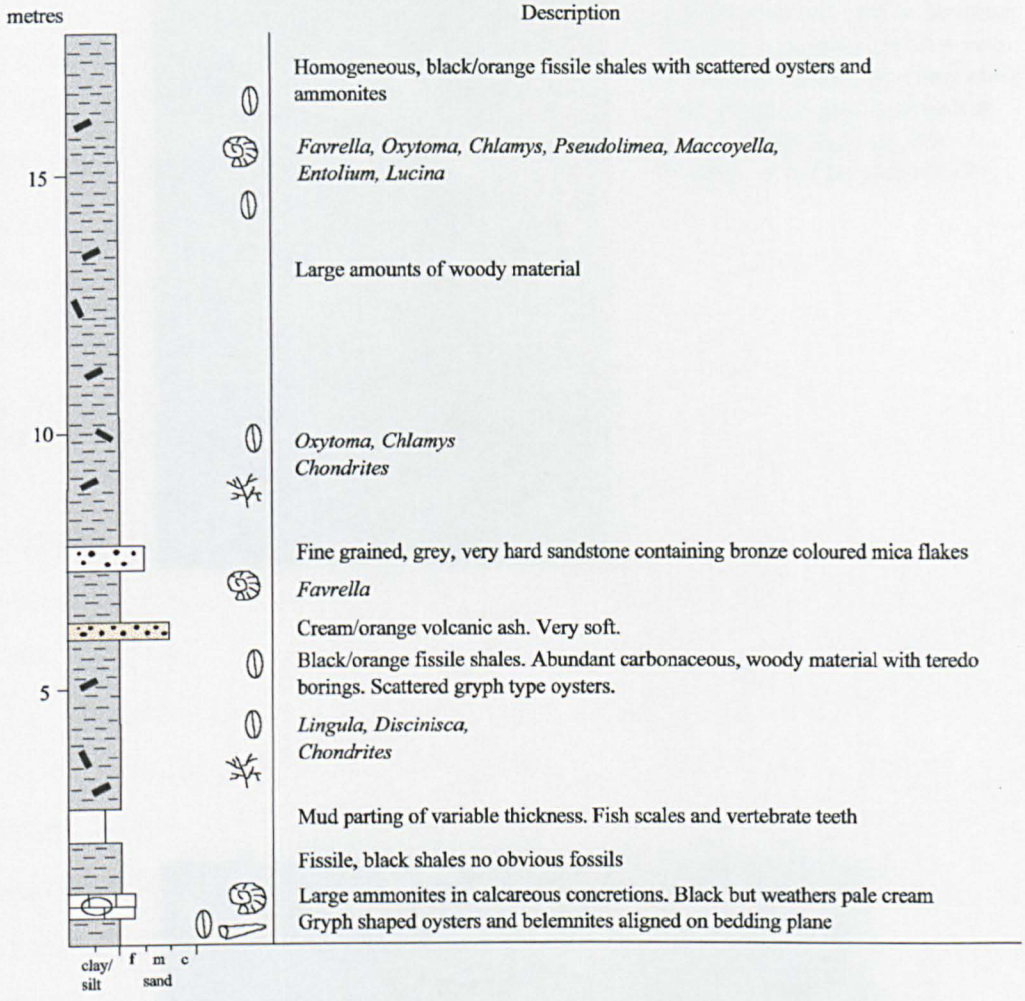


Figure 4.12: Measured section at Río Simpson (AMT4)

Figure 4.13: Concretions in the limestone horizon at Río Simpson (AMT 4). The concretions are up to one metre in diameter and many have formed around ammonites. View looking along the Río Simpson valley towards the SE.

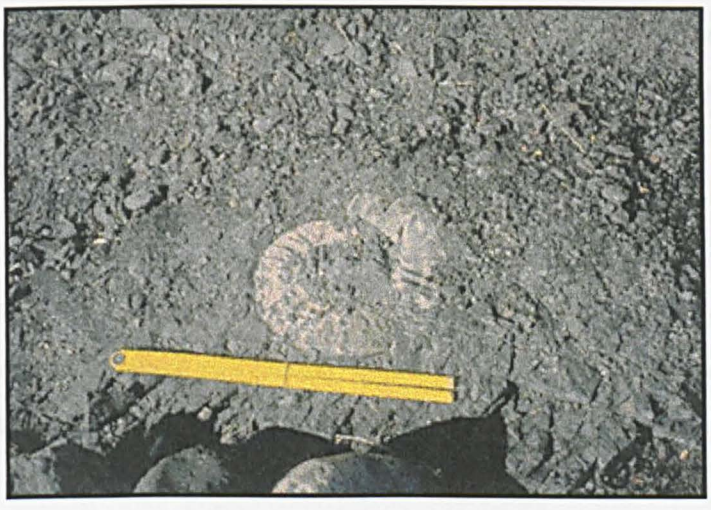


Figure 4.14: The ammonite *Favrella* embedded in the black shales of the Katterfeld Formation at Río Simpson (AMT 4)

Gryph shaped oysters are also common on the same bedding surface. They have a strongly curved left valve which is composed of thick, foliated calcite and hence well preserved (Figure 4.15 & Figure 4.17). Most specimens were found concave down or sideways, i.e. not in life position. The attachment area of many of the oysters show distinct ridges suggesting that they were originally attached to something with high relief, most likely an ammonite. Two species of brachiopods were found at this location, *Lingula* and *Discinisca*. This is the only site that brachiopods were found. Trace fossils were not common. The only ichnospecies that was found was *Chondrites*.

4.2.4 Cuesta Los Leones (AMT 19)

The Katterfeld Formation at Cuesta Los Leones (AMT 19) forms the major part of the lithology of the Coyhaique Group which is found in the deeply incised valley of the Río Coyhaique. The area is faulted to the north and east of the outcrop (Figure 4.19 & Figure 3.18) and the beds of the measured section show a consistent dip of 44^0 NW, which allows access to the complete sequence of the Katterfeld Formation along the river bed (Figure 4.20 & Figure 4.21). The Katterfeld Formation grades conformably from the Toqui Formation sandstones and storm deposited shell beds at the base of the sequence (section 3.2.4a & Figure 3.19) and grades into the Apeleg Formation sandstones at the top of the sequence.

At the base of the measured section (Figure 4.22) an orange/brown mudstone overlies the shelly, storm beds of the Toqui Formation. Within this bed are very small (2mm) articulated Veneridae bivalves and Naticid gastropods. Near the top of the mudstone unit some fish teeth and scales are present together with traces of iron pyrites. The mudstone also contains lenses of fine-grained, grey sandstone containing small, rounded clasts of the surrounding mudstone. The mudstone gradually becomes fissile and darker in colour. Fossils are rare but some faint, small, burrow structures are present and abundant carbonised woody material.

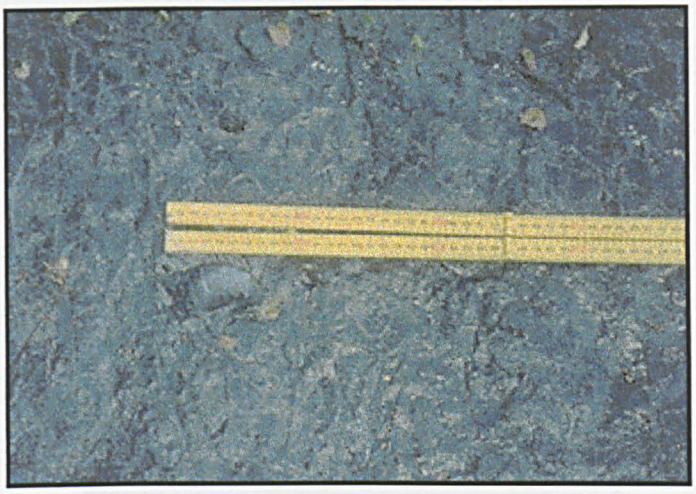


Figure 4.15: Gryph shaped oyster likely to be *Aetostreon.* Most were found concave down or on their side on the bedding plane at the base of the Río Simpson valley.

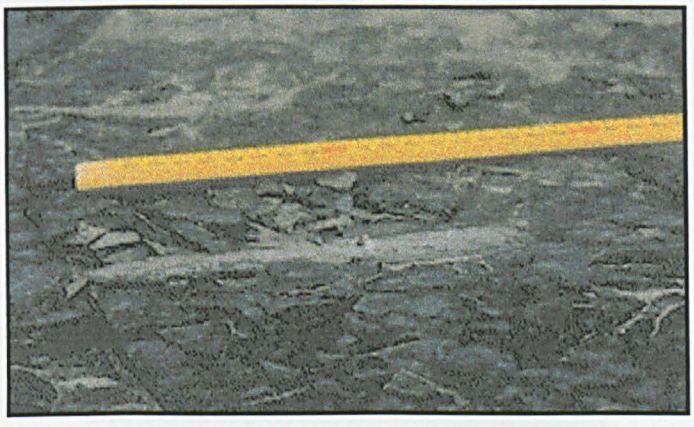


Figure 4.16: *Belemnopsis aucklandica* embedded in the black shales of the Katterfeld Formation at Río Simpson. Specimen 150mm long.

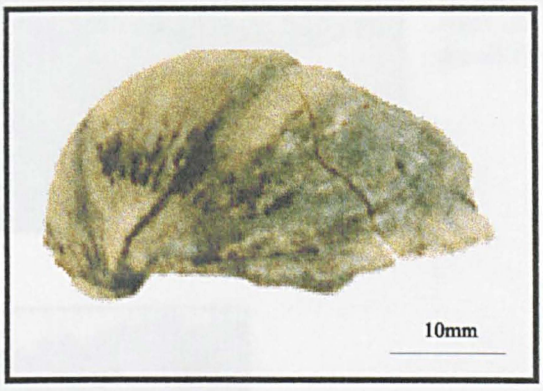


Figure 4.17: A gryph shaped oyster collected from Río Simpson likely to be *Aetostreon.* The left valve is strongly curved with a thick calcite shell. Some specimens showed circular borings possibly made by gastropods although none were collected at the Río Simpson locality.

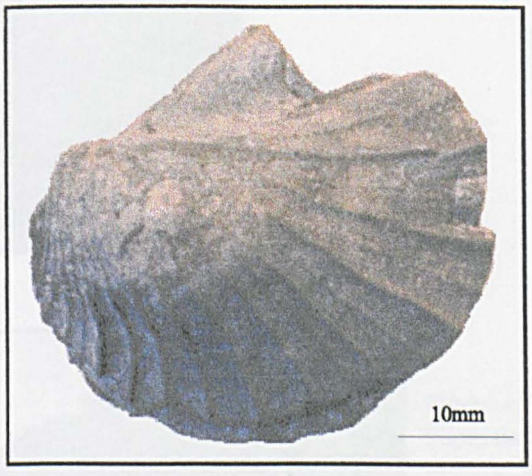


Figure 4.18: The byssally attached bivalve *Oxytoma oxytoma* found in the black shales of the Katterfeld Formation at Río Simpson.

Figure 4.19: Río Coyhaique Canyon looking southwest towards AMT 19 which is around the first bend. AMT 22 is at the base of the cliff in the foreground.

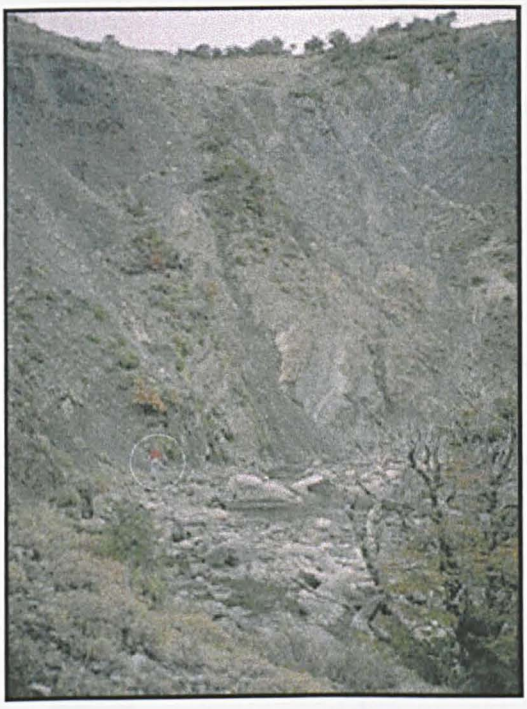


Figure 4.20: The faulted area near the base of the Katterfeld Formation sequence. Figure circled for scale.

Figure 4.21: The Katterfeld Formation looking up sequence towards the NW. The homogeneous black shales contain rare mudstone partings and occasional sandstone units. The beds dip towards the NW and so the sequence can be followed upstream.

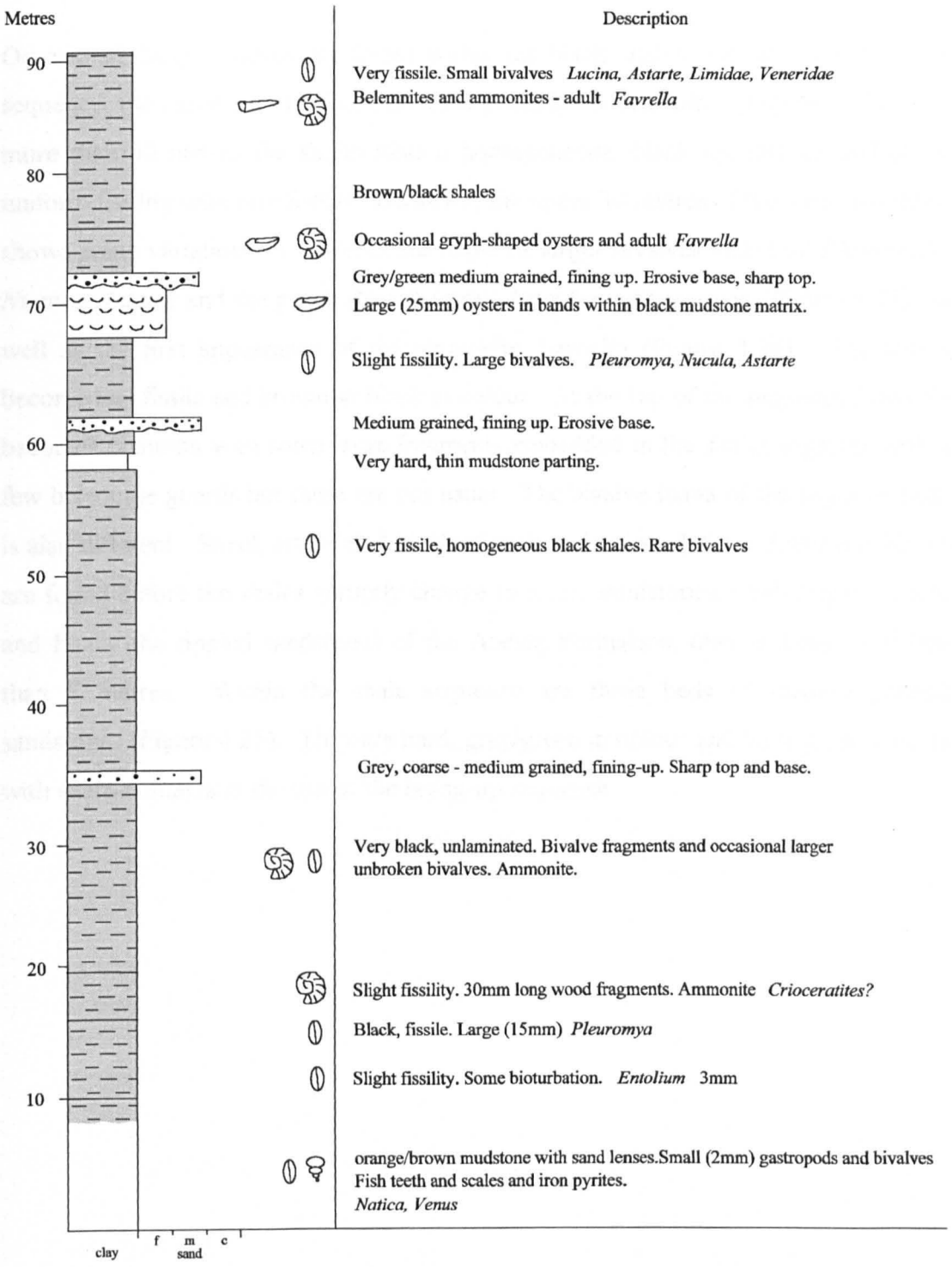
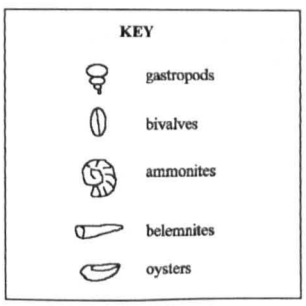


Figure 4.22: Measured section of the Katterfeld Formation at Cuesta Los Leones (AMT 19)



Occasional, large bivalves are found within the black shales and, at 20 metres up-sequence, the outer whorl of an ammonite possibly *Crioceratites* (Figure 4.23). For more than 50 metres the shales have a homogeneous, black appearance and show uniform fissility with rare fossils. However, the upper 30 metres of the shale sequence shows some variation. A more varied fauna of larger bivalves including *Pleuromya, Nucula, Astarte* and the gryph shaped oyster *Aetostreon* are present (Figure 4.22), as well as the first appearance of the ammonite *Favrella* (Figure 4.24). The shales become less fissile and brownish/black in colour. At the top of the sequence *Favrella* becomes common with some large fragments embedded in the shales together with a few belemnite guards but these are not usual. The bivalve fauna of the topmost beds is also different. Small, articulated bivalves such as *Lucina, Astarte, Lima* and *Venus* are found before the shales abruptly change to sandy mudstones, mud-draped ripples and finally the rippled sandstones of the Apeleg Formation, over a distance of less than 5 metres. Within the shale sequence are three beds of medium-grained sandstones (Figure 4.25). They are hard, grey/green in colour and have erosive bases with sharp contacts at the top of the fining-up sequence.

Figure 4.23: A fragment of the outer whorl of a loosely coiled species of ammonite likely to be *Crioceratites*. The fragment displays strong, slightly flexuous ribs with spiky tubercles on the shoulders, occasional tubercles in a ventro-lateral position and irregular tubercles at the dorsal rim.

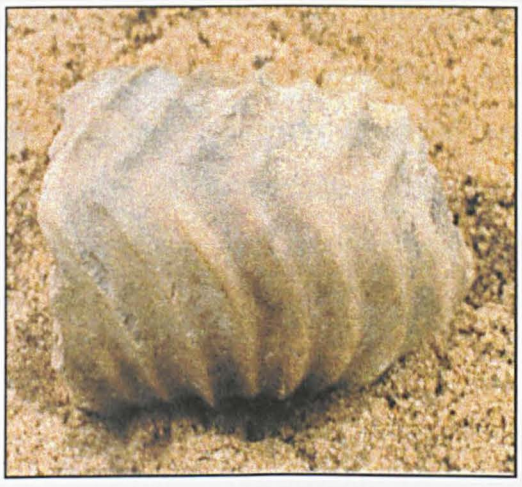


Figure 4.24: A fragment of the outer whorl of the ammonite *Favrella* collected from Cuesta Los Leones (AMT 19).

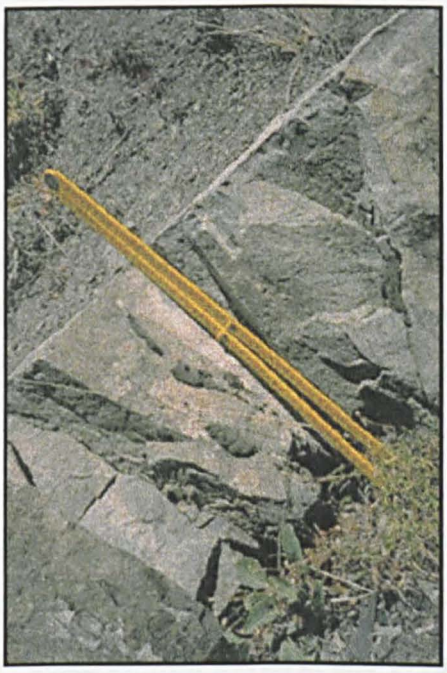


Figure 4.25: One of the fining upward sandstone beds found within the black shales of the Katterfeld Formation at Cuesta Los Leones (AMT 19). The unit has an erosive base and sharp top and is likely to be a turbidite.

4.3 Fossils in the Katterfeld Formation

A restricted fossil fauna was found throughout the Katterfeld Formation. Some sites produced a more varied fauna than others but ammonites, belemnites, bivalves, gastropods and some trace fossils were documented at various locations.

4.3.1 Ammonites

Ammonites are commonly found wherever the upper sequences of the Katterfeld shales occur. They are mostly found as impressions in the indurated sediments, with no remnant of the original shell. Sometimes they are present in calcareous nodules, where the actual specimen is preserved. Although fairly common, only one genus, *Favrella,* has been positively identified with two species *F. americana* and *F. wilckensi.* One other fragmented specimen was collected from Cuesta Los Leones (AMT 19) which has not been positively identified but is possibly a heteromorph of *Crioceratites.*

4.3.1a Favrella

Family	Berriasellidae (Spath 1922)
Subfamily	Neocomitinae (Spath 1924)
Genus	*Favrella* (Douville 1909)

Favrella was first described by Favre (1908) from western Santa Cruz Province, southern Argentina. Favre collected specimens from three different localities and this has led to some uncertainty as to the precise age of the genus. Ages proposed range from late Jurassic to Aptian (Riccardi, 1970). Riccardi re-assessed the type material and concluded that Favre's age range of Hauterivian to Barremian was correct. Thomson and Farquharson (1984) point out that if *Favrella* is a member of the Neocomitinae sub-family, it would be unlikely to range into the Barremian since the youngest representative of that sub-family is mid-Hauterivian. Leanza and Leanza (1973) disagree with Riccardi and propose an early Aptian age for *Favrella.* They argue that some of the specimens Riccardi worked on belong to a separate genus

which they named *Pseudofavrella* and they accepted a Hauterivian age for this genus. Furthermore, they proposed that *Favrella* is restricted to Patagonia and the Magallanes Basin whereas *Pseudofavrella* occurs further north in the Neuquén Basin and northern South America. Thomson and Farquharson (1984) investigated the proposed early Aptian date for *Favrella* and concluded that it is unlikely to be correct as the fauna it has been correlated with is of doubtful Aptian age. They proposed that *Favrella* is most probably Hauterivian in age and this has now been accepted by most authors.

***Favrella americana* (Favre 1908):** *Favrella americana* has been described as having a variable form by several authors (Favre 1908, Arkell *et al.*, 1957, Riccardi 1970), this makes it difficult to identify positively. The main feature which distinguishes it from *Favrella wilckensi* is strong, well-spaced ribs (Figure 4.26). It is evolute in form with a rounded whorl section. The ribs are strong, simple and strongly projected (curved forwards) on the shoulder and venter giving the distinctive 'chevron' pattern associated with *Favrella*. The juvenile form sometimes has irregular, bifurcating ribs, sometimes umbilical tubercles and spines on the shoulders. In the adult form, *F. americana* loses the tubercles and spines. Riccardi (1988) refined his previous work and dated *F. americana* as lower Hauterivian.

***Favrella wilckensi* (Favre 1908):** *Favrella wilckensi* has an early stage of thin, bifurcate ribs with periodic tubercles and a late stage with spaced, simple and intercalated ribs. Like *F. americana* it is evolute in form with similar, strongly projected ribs giving the 'chevron' pattern. It is distinguished from *F. americana* by the more delicate, sometimes intercalated nature of its ribs. It is considered to be the younger species with a proposed age of upper Hauterivian (Riccardi, 1988).

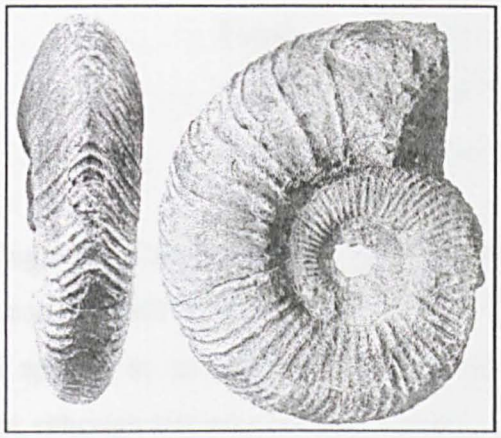


Figure 4.26: Ventral and lateral view of *Favrella wilckensi* illustrated by Favre, 1908 (pl XXXII fig.2) (from Thomson & Farquharson, 1984).

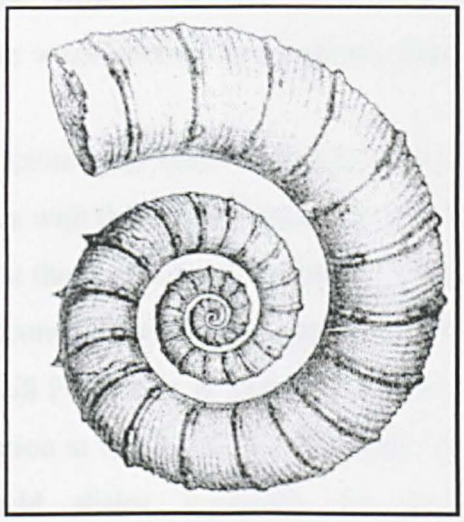


Figure 4.27: Typically loosely coiled example of *Crioceratites*. Ribs can be rounded, straight or slightly curved and carry umbilicle, lateral and ventrolateral spines. (from Arkell *et al.*, 1957)

4.3.1b Crioceratites

Family	Ancyloceratidae (Meek 1896)
Subfamily	Crioceratitinae (Wright 1952)
Genus	*Crioceratites* (Leveillé 1837)

One fragment of an outer whorl approximately 40mm long was collected from Cuesta Los Leones (AMT 19.1). It is evolute in form and likely to be from a very loosely coiled species as no impression of the preceding whorl is obvious on the dorsal surface although this area is slightly damaged. The whorl section is rounded to oval in shape. The fragment displays strong, evenly spaced ribs which are slightly flexuous. The ribs are straight as they cross the venter unlike *Favrella* which has strongly projected ribs and shows a distinctive 'V' shape. There is no groove or smooth area on the venter and the ribs are continuous across this surface. There are three rows of tubercles on the ribs. Strong, spiky ones on every rib at the ventral shoulder, weak, occasional ones in a ventro-lateral position and another set of irregular weak ones at the dorsal rim (Figure 4.23).

The specimen was firstly identified as *Neocosmoceras* (Blanchet, 1922) as it has many affinities with this species including the trituberculate ribs. This would have dated the shales at this location as Berriasian. This age is acceptable as the *Favrella* specimens were found higher in the sequence. *Neocosmoceras* sp. has been found in the Springhill Formation at Lago San Martin, Argentina by Riccardi (1977). The faunal association at this location corresponds quite closely with the fauna found within the Katterfeld shales, especially the bivalves. Olivero (1983) also identified *Neocosmoceras* sp. from Tres Lagunas, Chubut, Argentina, at the same latitude as Coyhaique Alto. However, due to the loosely coiled nature of the form, it is now thought unlikely to be *Neocosmoceras* but almost certainly a heteromorph belonging to the *Crioceras andinum* group (Gerth 1925, Weaver 1931), most probably *Crioceratites* (personal communication Thomson, 1995).

Crioceras sp. was identified by Favre (1908) from Santa Cruz Province, Argentina in association with *Favrella*. *Crioceratites* has been identified by Riccardi (1988) in the

Neuquén (Andean) Basin of West-Central Argentina. It is typically coiled in an open spiral and the whorl section may be oval or subquadrate (Figure 4.27). However the ribs are generally dense, untuberculate and with periodic stronger ribs carrying ventrolateral or umbilical, lateral and ventrolateral spines (Arkell *et al.*, 1957). This would give an upper Hauterivian age.

4.3.1c Summary of ammonite fauna

Ammonites belong to the class Cepholopoda and were entirely marine, their habit ranging from shallow to deep water. They were nektonic carnivores which preyed on small organisms and in turn were preyed on by fish and marine reptiles whose teeth and scales were often found in close association with ammonite remains in the Katterfeld Formation. They relied on camouflage to evade predators and this could be the reason they developed ornament and sculpture in the form of ribs and tubercles, as in *Crioceratites,* to break up the outline of the shell (Black, 1988). Most ammonite species were relatively short lived and geographically widespread due to their nektonic mode of life, making them valuable zone fossils. Unfortunately, due to the uncertainty of the age range of *Favrella,* the only positively identified ammonite in the Katterfeld Formation, and the lack of other datable fossils, it is not possible to constrain the age of the Katterfeld Formation with any certainty. The most likely age for the upper part of the succession where *Favrella* was common is Hauterivian.

4.3.2 Belemnites

In the black shales of the Katterfeld Formation, belemnites were generally found in association with ammonites. Two distinct species of the *Belemnopsis* genera have been identified: *Belemnopsis* aff. *aucklandica* and *Belemnopsis* aff. *extenuatus*. The belemnite most commonly associated with this area, *Belemnopsis patagoniensis,* was first identified by Favre (1908) in association with the ammonite *Favrella,* from Meseta Belgrano, southern Argentina. Literature post-dating Favre's discovery commonly identifies *B. patagoniensis* at most fossiliferous locations in Patagonia. However, the specimens collected in the Aysén Basin, although *Belemnopsis* are not *B. patagoniensis.*

Belemnopsis:	Subclass	Coleoidea (Bather, 1888)
	Order	Belemnitida (Zittel, 1895)
	Suborder	Belemnopseina (Jeletzky, 1965)
	Family	Belemnopseidae (Naef, 1922)
	Genus	Belemnopsis (Bayle, 1878)

The guard of *Belemnopsis* is typically cylindrical, whilst the apical region is generally moderately obtuse (Figure 4.28). The main distinguishing characteristic of *Belemnopsis* is the single, median, ventral groove which normally extends from the alveolus almost to the apex. In early species of *Belemnopsis* the groove is often narrow and moderately shallow but in later *Belemnopsis*, the ventral groove is broad and deep (Stevens, 1965).

The *Belemnopsis* specimens collected in the Katterfeld shales of the Aysén Basin were fragmented and incomplete. The alveolar region was invariably compressed or missing. However, characteristics were defined by amalgamating several examples of the same species.

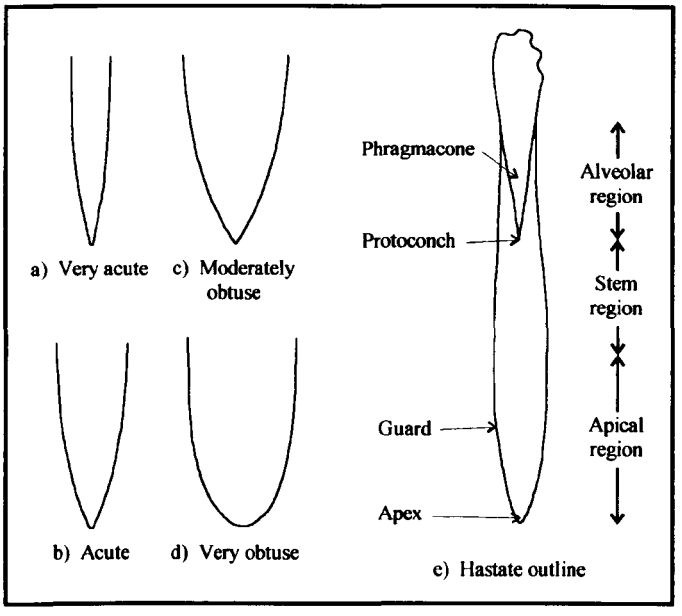


Figure 4.28: a,b,c,d) outlines of the apical regions of several belemnite guards. e) outline of belemnite guard to illustrate terminology, after Stevens (1965).

4.3.2a Belemnopsis aff. *aucklandica* (Hochstetter 1963)

Belemnopsis aucklandica has a slightly hastate (club-shaped) outline with a maximum diameter slightly anterior to the mid-point. It has a prominent, deep, ventral groove which extends from the alveolar region to approximately 20mm from the apex (Howlett, 1989).

The examples collected in the Aysén Basin have a robust, club-shaped guard and are slightly oval in cross section (Figure 4.29). The longest fragment is 52mm long and 14mm at its maximum diameter, however, the apex is missing from this specimen and so it is not possible to determine an accurate length to diameter ratio. The estimated length from apex to protoconch is 75mm. The groove is well-defined, moderately wide (6mm from shoulder to shoulder) but flat. This differs from *B. aucklandica* as it would be expected to be deep. Although the apex is not present the groove is clearly defined, although shallowing, and is likely to extend almost to the tip of the apex. It is likely that this specimen is closely related to, and possibly evolved from, *B. aucklandica* which was common in the Antarctic region during the Kimmeridgian to Tithonian stages.

4.3.2b Belemnopsis aff. *extenuatus* (Yang and Wu 1964)

Belemnopsis extenuatus has a hastate outline with a maximum diameter at the mid-point of the guard. The guard is long and slender. It has a prominent ventral groove which is wide and deep and extends from the alveolar end to within 20mm of the acute apex (Howlett, 1989).

The specimens collected from the study area comprise several fragments and one almost complete specimen (Figure 4.30) although many were observed *in situ.* The guards are conical, thin and elongate in outline and circular in cross-section. The longest specimen collected is 107mm long and 9mm maximum diameter. The apex is acute. The ventral groove is deep but narrow (2mm), rather than wide as would be expected.

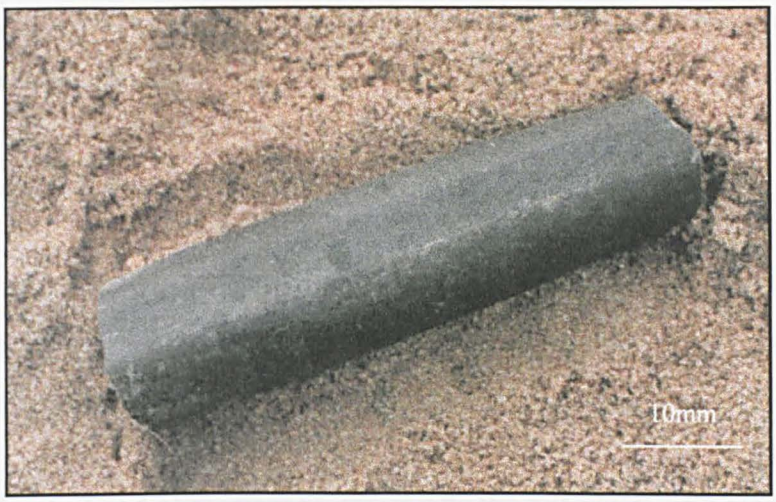


Figure 4.29: Fragment of *Belemnopsis aucklandica* collected from the Katterfeld shales at Baguales Vista (AMT 8). The specimen has a robust, club-shaped guard and a distinctive, wide, flat groove which shallows towards the apex.

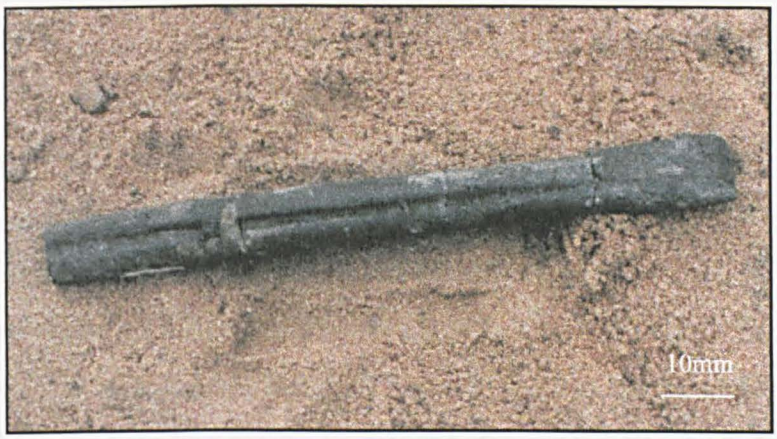


Figure 4.30: An almost complete example of *Belemnopsis extenuatus* collected from Baguales Vista (AMT 8). The elongate, thin guard pinches out near the alveolar region (which is slightly squashed on this example). The groove is deep and narrow and the apex would be acute in form.

The groove is flatter near the apex and pinched out near the alveolar region. This species is also found in Antarctic material (Howlett, 1989) and has an age range of Valanginian to Hauterivian.

4.3.2c Summary of belemnite fauna

Belemnites were carnivores which preyed upon small fish and were in turn preyed upon by predators such as sharks (Stevens, 1965). *Belemnopsis extenuatus* was more commonly found in the Katterfeld Formation than *B. aucklandica.*

The guard (rostra) and phragmacone are usually the only parts of a belemnite that are preserved and often the phragmacone is distorted or missing, due to its fragile nature, as is the case in the specimens from the Aysén Basin. The main features that help identify and distinguish species of belemnites are; outline of the guard, length to diameter ratio, nature of the apex, cross section of the guard at specific points and the surface groove (Stevens, 1965).

The guard of a belemnite is small in relation to the soft tissues of the organism and represents approximately one eighth of the total body length. *Belemnopsis aucklandica* would have attained lengths of above a metre and *Belemnopsis extenuatus* almost two metres (Stevens, 1965). Belemnites probably propelled themselves in the same way as modern squids and the guard was likely to have acted as a counter-weight to achieve horizontal stability for propulsion. Some belemnites are common only in nearshore facies of 10 to 20 metres depth, where it is thought they moved for the breeding season to spawn before dying (Doyle & Bennett, 1995). Dense concentrations of mature forms of belemnite guards are often found and it is thought that this supports the spawning-ground explanation. Río Simpson (AMT 4) had one bedding surface that displayed many more than average belemnites but it is more likely that this was due to some catastrophic event rather than evidence of a spawning-ground, as the guards were of both juvenile and adult forms.

However, belemnites are recorded in deeper water offshore and turbiditic facies in Antarctica (Butterworth *et al.,* 1988) and also in a similar setting in the Magallanes

Basin (Austral Basin) of southern Argentina (Doyle & Bennett, 1995). It is likely that belemnites were close in life habit to neritic/pelagic present day squid (Doyle & Bennett, 1995) and they probably inhabited shelf and deeper water environments. The belemnites found in the Katterfeld Formation shales closely resemble New Zealand forms described by Stevens (1965). However, Challinor (1991) observed that the narrow groove which was found in some of the Aysén Basin belemnites, is consistent with *Belemnopsis* forms of the Tithonian to Berriasian of the Malagasy Republic. Riccardi (1977) recorded *Hibolithes* from southern Patagonia. Urreta and Suárez (1985) identified *B. madagascariensis* from Tierra del Fuego and Ferruglio (1936) noted *Hibolithes* from the Argentine Cordillera. These species of belemnite are all consistent with New Zealand fauna. The evidence of belemnite fauna indicates strong links between southern South America and New Zealand during the late Jurassic but in the Early Cretaceous, genera typical of Tethys (east Africa and Malagasy Republic) appear (Challinor, 1991). Olivero (1987) recorded belemnites from the Lago La Plata Formation near Lago Fontana (Figure 2.11). Although this formation is considered to be made up mainly of pyroclastic rocks, minor fossiliferous, sedimentary intercalations occur throughout the unit. These sedimentary intercalations have previously been considered to be independent stratigraphic units i.e. the Tres Lagunas and Cotidiano Formations. He proposed an age range of Tithonian to Hauterivian for the sediments, based on the fossil assemblage. The examples he collected are very similar to the ones found in the Aysén Basin. The guard is cylindrical, conical and elongate with a narrow, deep ventral groove. He suggests that the specimen is *Belemnopsis* aff. *uhligi* and similar to *B. madagascariensis* as described by Riccardi (1977). It is clear that this is the same species identified as *B.* aff. *extenuatus* in the Katterfeld shales.

4.3.3 Bivalves

The bivalves found in the Katterfeld Formation are all fairly small in size with thin shells, typical of those generally found in black shales. The majority are suspension feeders and mostly shallow burrowers, the exceptions being *Nucula* which is a deposit feeder and *Pleuromya* which burrows to greater depths than the other species found in association (Table 4.1)

Epifaunal	Infaunal (shallow)	Infaunal (deep)	Semi-infaunal
Chlamys	*Astarte*	*Pleuromya*	*Modiolus*
Entolium	*Grammatadon*		
Oxytoma	*Lucina*		
Plicatula	*Protocardia*		
Aetostreon	*Arctica*		
Meleagrinella	*Venus*		
Lima	*Myophorella*		
Pseudolimea	*Nucula*		

Table 4.1: Life modes of the bivalves found in the Katterfeld Formation

4.3.3a Epifaunal bivalves

The major part of the black shale succession of the Katterfeld Formation is sparsely fossiliferous with few bivalves but these generally have an epifaunal mode of life. Many of the epifaunal bivalves are byssate species such as *Oxytoma, Pseudolimea, Chlamys* and *Lima* and would have attached themselves to driftwood, grasses or shell fragments. *Chlamys* and *Lima* preferred secluded locations such as inside root bundles or on the underside of logs, as the organism could not retract fully into its shell and so was easy prey for fishes and other predators (Kauffman, 1969). Some species of *Chlamys* and *Lima* could swim short distances by valve clapping. *Modiolus* is a semi-infaunal, byssate bivalve but it is usually found in slightly agitated waters where it anchors itself by short threads against the currents and close to the sediment surface (Sellwood, 1978). It prefers a hard substrate and was found at only one location of the Katterfeld Formation. An epifaunal, gryph-shaped oyster thought to be *Aetostreon* was found as solitary forms within the shales at a few locations (Figures 4.15 & 4.17). They are thick shelled and inequivalve with a strongly coiled lower valve which provides support in soft sediments. Oysters need a firm surface to anchor to and the soft sediments of the Katterfeld Formation would not have provided many opportunities for colonisation, however, the attachment area of the oyster shells often displayed ridges where they had been anchored to another organism, most likely an ammonite. The most commonly found epifaunal bivalve was *Entolium,* sometimes present in great numbers but it appeared in various concentrations at most locations. *Entolium* is a good swimmer with rounded to ovate, smooth, relatively thin valves and small auricles. It lives independently and its ability for quick movement enables it to

survive a more exposed habitat. *Entolium* can live in a variety of environments from shallow waters subject to strong water movements, to deeper quiet water surroundings (Kauffman, 1969). *Oxytoma, Chlamys, Lima* and *Aetostreon* are all commonly found in shallow shelf environments.

4.3.3b Infaunal bivalves

A few Katterfeld Formation sites have a more varied bivalve fauna. Distinctive, fossiliferous beds at Los Baguales (AMT 3) contain predominantly infaunal, shallow burrowing bivalves such as *Astarte, Arctica, Nucula, Protocardia* and others (Table 4.1). Most infaunal bivalves are smooth or have fine concentric ornamentation which enables rapid burrowing. Coarse, concentric ribs, as are present on *Lucina,* strengthen the valves and stabilises the shell in the substrate, as well as discouraging predation. Shallow burrowers are subject to scavenging and high predation rates especially from gastropods (Kauffman, 1969) which were present in great numbers in the same beds. Shell form depends on the rate of movement through the sediments and the depth of burial. In general, inflated bivalves with rounded to oblate outlines such as *Nucula,* are dominantly shallow, slow burrowers which remain stationary. Deeper burrowing bivalves such as *Pleuromya* are less convex and more elongate in outline. *Myophorella* is a Trigonid which is an active, rapid burrower usually found in a nearshore, high energy environment. Most of the infaunal bivalves found in the black shales are known to occupy relatively soft, unconsolidated substrates in an inner-shelf setting.

4.3.4 Gastropods

Most marine gastropods are benthonic although some can be pelagic. Some are sessile and feed on plankton but others are carnivorous and can bore through shells of other molluscs. Coiled gastropods can show great variation in form and two species of the coiled variety were identified in the Katterfeld Formation, *Natica* and *Bathrotomaria.*

4.3.4a Natica

Natica are thick shelled and globose with a low spire and a large last whorl. They are infaunal and plough through the sediments just below the surface looking for prey. They are carnivorous and are capable of drilling holes in the shells of other molluscs to access the organism (Black, 1988).

4.3.4b Bathrotomaria

The shells of *Bathrotomaria* have a flat base and conical spire. They have existed since the Triassic and originally lived in shallow water but presently inhabit deep water environments where there is less competition from other more advanced organisms (Turek *et al.*, 1988). They are algal grazers.

4.3.5 Trace fossils

The only ichnospecies found in the Katterfeld Formation was *Chondrites*, although the lack of lamination in the sediments would usually be regarded as an indication of high bioturbation and a range of trace fossils would normally be expected.

4.3.5a Chondrites

Chondrites consists of a horizontal to inclined burrowing system exhibiting branching from a main stem. It has been interpreted as the trace of a deposit feeding worm which mined the sediment and represents simple, shallow feeding burrows (Jordan, 1985). The specimens found at Baguales Vista (AMT 8) and Río Simpson (AMT 4) are very small, 5mm - 30mm long and less than 3mm wide with up to three orders of branching. *Chondrites* is regarded as a useful indicator of minimal oxygenation in sediments. It is often the first trace to appear above an apparently anaerobic sequence of muddy sediments, suggesting the animal responsible had a tolerance to low oxygen (Goldring, 1991). *Chondrites* is more common where food was abundant (Goldring, 1995) and the high levels of organic matter found in the Katterfeld shales would suggest that those conditions prevailed at the sites *Chondrites* was found at. Burrows

of other organisms were absent in the black shale deposits and this suggests that the primary food resource at the sea-floor during periods of low oxygen was organic detritus in the sediment and very little was available in the water column (Ekdale, 1985).

4.4 Palaeocurrent data

The only site that provided any palaeocurrent information in the Katterfeld Formation was Río Simpson (AMT 4) where an exposed bedding plane allowed measurements of belemnite guards to be taken. Nearly all of the guards were of a similar diameter but fragmented and so only specimens more than 100mm long were measured, where it was clear in which direction the apex was pointing. The sample size was small but a significant bimodal pattern could be detected (Figure 4.31) which shows the belemnites are orientated in a NW/SE direction.

Elongate organisms such as belemnites usually roll or align themselves parallel to any currents with the apex pointing upstream therefore currents at Río Simpson would have been in a NW/SE direction. Flume studies have shown that the pattern of distribution for cylindrical objects is typical of distribution patterns of belemnite guards seen at Río Simpson at velocities of below 70cm/sec (Dodd & Stanton, 1981), although Goldring (1991) points out that compaction of the surrounding sediments can rotate fossils and create the impression of current formed orientations. Current motion needs to act over a period of time to induce a pattern of least resistance. A firm subsurface is a further requirement because on soft substrate elements tend to become buried (Goldring, 1991). It is also possible that the assemblage at Río Simpson has been winnowed and the smaller guards removed as the remaining guards are of approximately uniform size and mostly fragmentary.

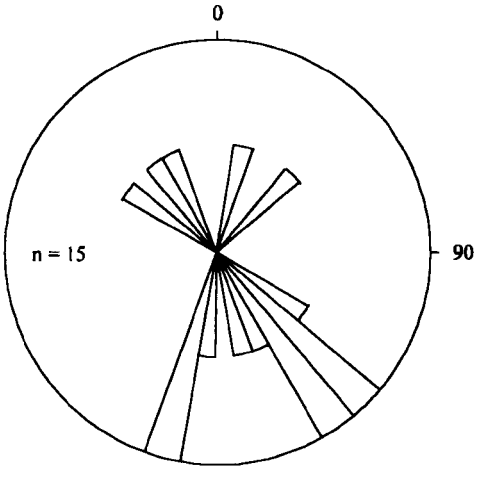


Figure 4.31: Belemnite orientation at Río Simpson (AMT 4). The direction of the apex of the guard was measured in all samples more than 100mm long exposed on a bedding plane of the black shales of the Katterfeld Formation.

4.5 Interpretation of the Katterfeld Formation

The calcareous black shales and mudstones of the Katterfeld Formation are regionally extensive, laterally continuous and usually horizontal and parallel bedded, suggesting prolonged periods of deposition on a relatively flat basin floor. They contain abundant organic material often preserved as large pieces of carbonised wood, indicating the presence of nearby land areas. The black shales are initially homogeneous looking, black to dark brown in colour and generally fissile although at some locations they are more massive and blocky in appearance. The mudstones are usually black but sometimes slightly grey or brown. They occur in thin units interbedded with the shales, sometimes showing laminations but mostly structureless. Some mudstones are fossiliferous but rarely show evidence of bioturbation.

4.5.1 Black shale environments

The term "black shale" has been applied to a range of facies of varied depositional settings and palaeoenvironmental conditions. The usual interpretation of black shales is of marine mudrocks which are rich in organic matter, which is the main cause of the dark colouration. Colouration of shales is influenced by the amount and maturity of organic carbon content which can vary significantly. Sediments containing more than a few percent of immature organic matter are typically more brown than black, however, strata containing less than 2% highly oxidised or thermally mature organic carbon are commonly black (Arthur & Sageman, 1994). Many ancient black shales lose their organic content during thermal maturation or metamorphism and so the definition of black shale by organic content is considered arbitrary and not desirable by a number of authors (Wignall & Hallam 1992, Arthur & Sageman 1994, Wignall 1994).

Black shales have been defined by Tyson (1987) as, "dark coloured, fine-grained mudrocks having the sedimentological, palaeoecological and geochemical characteristics associated with deposition under oxygen-deficient or oxygen-free bottom waters." Oxygen is an essential requirement for metazoan life (Rhoads & Morse, 1971). A reduction in the quantity of dissolved oxygen can lead to a

reduction in species diversity and ultimately to mass mortality. Petroleum geologists noted the relationship between carbon-rich sediments and facies deposited under low oxygen conditions and this led to the assumption that low oxygen conditions resulted in the preservation of organic carbon and the formation of organic-rich sediments (Allison *et al.*, 1995). A number of authors (Demaison & Moore 1980, Pedersen & Calvert 1990) proposed an alternative model that suggests high primary production controls the accumulation of organic-rich facies, which in turn lowers oxygenation. The only common theme in all black shale models is the presence of very low or truly anoxic bottom waters and this has led to an ongoing debate known as the productivity versus preservation controversy.

4.5.1a Productivity versus preservation and water depth

Wignall (1994) noted overwhelming evidence that most black shales accumulate beneath dysoxic or anoxic water, but questioned whether oxygen deficiency is a cause of organic matter preservation or a result of it. The debate centres on whether the presence or absence of oxygen in bottom waters has any control on accumulation of organic matter. The preservation model has received widespread support (Demaison & Moore 1980, Tyson 1987, Wignall 1994), however, there is considerable evidence to suggest that a high organic carbon input to the sediment caused by high surface water productivity may be the main mechanism for the formation of organic-rich sediments. In the productivity model, bottom water anoxia may rise but is considered to be the result of high oxygen demand unrelated to how much organic carbon is preserved (Pedersen & Calvert, 1990). In the preservation model anoxic bottom waters are typically generated by stratification of the water column and the restriction of vertical mixing. Nutrients cannot be recycled back to the surface due to the stratification, with the result that surface water productivity is typically low. High productivity occurs when there is an ample supply of nutrients either by upwelling of deep waters or high influx from river systems. High productivity is typically associated with more dynamic and vigorous circulation (Wignall, 1994). The debate is not resolved with strong argument supporting both models.

Organic matter descending through the water column is subject to oxidation and so greater transit times will favour greater degradation before its arrival at the sea-floor (Wignall, 1994). Generally the transit time is less in shallower water environments, and shallow-water, black shale models have been proposed for several sites in the United States (Savrda & Bottjer, 1993). However, many workers consider that black shales are a deep water facies (Wignall & Hallam, 1991). Most black shales in the geological record accumulated in the deeper parts of epicontinental seas which had poor connections to open oceans (Wignall, 1994). There are few epeiric seas at present but organic-rich sediments are currently found in several distinct settings such as the Black Sea.

4.5.1b Modern analogues and ancient black shales

A major difficulty in interpreting black shale depositional environments is the lack of recent analogues. Depositional analogues for marine black shales most commonly cite large landlocked or silled basins such as the Black Sea. The Black Sea is the most commonly used example of a large, deep, stratified, silled basin. Arthur & Sageman (1994) found the waters below 150 metres to be completely anoxic and devoid of Metazoan life, yet studies of a wide range of ancient black shales show that repeated benthic colonisation events regularly occurred suggesting that dynamic and dysoxic environments were the norm in the older deposits. Pedersen and Calvert (1990) also questioned depositional conditions of the organic-rich sediments in the Black Sea and suggested that bottom waters were oxygenated at the time they were laid down. The controversy as to the conditions in which the Black Sea shales formed and therefore the validity of the Black Sea analogue is not clear.

Bottjer & Savrda (1993) found that black shales could also accumulate in broad, shallow, relatively flat-bottomed epicontinental seaways characterised by warm bottom waters and a high influx of marine or terrestrial organic matter. In these settings oxygen deficiency was most likely generated at the sea-floor in response to intense bacterial oxygen consumption in the sediment. During times of decreased circulation the dysaerobic/anaerobic boundary would have risen above the sediment - water interface whilst during periods of increased water circulation the boundary

would have become stabilised at the sediment-water interface. It is possible that bottom water oxygen levels were sufficient to periodically support epifauna but conditions in the substrate precluded the establishment of bioturbating infauna.

Many epicontinental black shales have characteristics that are inconsistent with those of most modern low oxygen environments. For example, many ancient black shale units are widely distributed and temporally long ranging within epicontinental basins, many units are laminated rather than bioturbated and the fauna is dominated by low diversity assemblages of epifaunal suspension feeders rather than diverse infaunal deposit feeders (Arthur & Sageman, 1994). Some workers concluded that certain widespread black shales were deposited globally and synchronously perhaps within a relatively narrow time period of less than one million years. During the mid to Late Cretaceous, black shales were deposited worldwide in several discrete events. Deposition took place in many situations including oceans, continental basins and shallow-shelf environments. The episodes of black shale deposition have been correlated with high sea-levels and are considered to be related to a warm climate and the resultant stagnant oceans (Francis & Frakes, 1993).

4.5.2 Sedimentology of the Katterfeld Formation

Most fine-grained terrigenous material is transported to basinal settings as aeolian dust and low density suspensions of sediment which is deposited by hemipelagic settling. Weak currents that are present in all seas and oceans keeps the settling rate very low (Wignall, 1994). Hemipelagic settling in basin deposits form a blanket of strata of uniform thickness but the influence of bottom currents usually prevent them from becoming widespread (Wignall, 1994). The black shales of the Katterfeld Formation are extensive and show thicknesses in excess of 200 metres in the study area and 350 metres in Argentina. The formation is probably much thicker as no sampling location showed both top and bottom of the exposure and equivalent formations in the Magallanes Basin have thicknesses of more than 1000 metres. The vast thicknesses suggest that deposition was on a broad, relatively shallow, epicontinental sea bed with a high influx of terrestrial material, but where deposition kept pace with slow subsidence of the basin floor. The basin was probably restricted

with poor circulation. The lack of volcanic horizons in the black shales shows that the sporadic volcanic activity that was evident at the time of deposition of the Toqui Formation had ceased. Deposition of the Katterfeld shales took place in a volcanically quiescent period.

Initially black shales form in the deepest area of basins and gradually expand in extent as oxygen deficient waters rise up through the water column. The most organic-rich sediments preferentially accumulate in topographic lows due to the weak currents but in the Katterfeld shales the organic material showed no such accumulations. Organic material including palynomorphs and woody tissue, were evenly distributed wherever the black shales occurred. Conditions were generally stagnant throughout much of the Katterfeld shale deposition. The general lack of current induced structures suggest a setting below wave base but probably less than 200 metres. At modern sites of anoxic sedimentation, the depth to the top of the oxygen minimum zone is around 150 metres and up to 250 metres in the Black Sea (Whitham, 1993). The upper part of the succession at Cuesta Los Leones did show some thin distal turbidites and shelly storm bed accumulations. This location was very near the suggested eastern border to the Aysén Basin and the entrance to the San Jorge Basin, which was probably the source of the turbidites. A further indication of the proximity of a terrestrial source are the iron-rich concretions that typify depositional environments close to the source area (Sellwood, 1972). They were found in abundance at Baguales Vista although smaller nodules were seen in the shales at Cuesta Los Leones. In contrast, the concretions at Río Simpson were calcareous and probably represent reduced terrigenous influence perhaps due to a deepening phase of the basin (Sellwood, 1972).

The nature of the substrate is an important factor in palaeoenvironmental interpretation because of its influence on the fauna that can colonise the setting. Modern anaerobic sediments are very watery and soft and a precise water/sediment interface is impossible to define (Wignall, 1993). Descending pelagic organisms, such as ammonites and belemnites, would have sunk into the sediments and become instantly buried. Exceptionally soft sediments can result from high organic content and produce impoverished benthic assemblages (Rhoads & Morse, 1971). A decrease in the level of bioturbation caused by a decrease in oxygen levels and subsequent

decrease in benthic fauna, will cause the muddy sediments to become firmer and this is likely to have been the setting for the bedding plane at Río Simpson where belemnites showed indications of alignment. The sediment must have been at least slightly compacted to allow the organisms to become orientated in the direction of current flow.

Lamination is a distinctive characteristic of black shales and results from regular alternation of fabric, grain size or colour. Many black shales are finely laminated but not all are (Arthur & Sageman 1994, Wignall 1994). Lamination can be destroyed by bioturbation or physical processes such as winnowing and redeposition. Wignall & Hallam (1991), found that all black shales show a compacted fabric but not all display laminations. Homogeneous black shales were found at the majority of the Katterfeld Formation sampling sites and imply a more continuous accumulation than the episodic deposition of laminated varieties (Wignall, 1994). The variation in laminated beds was noticeable in the sequences at Los Baguales and Cuesta Los Leones and probably reflects a difference in the rate of deposition of the affected units.

Many black shales are characterised by their fissility. Wignall & Hallam (1991) found that fissility was unrelated to lamination in the black shales of the Lower Oxford Clay but appeared to correlate with the organic richness of the sediments. More recent work (Wignall, 1994) has shown that fissility is likely to reflect environmental conditions prevalent at the time of deposition when large amounts of organic matter and an initially high water content were present beneath anoxic bottom waters.

4.5.3 Palaeoecology of the Katterfeld Formation

Many of the factors which control the distribution of faunal assemblages are depth related, for example, nature of the substrate, nutrient supply and biological competition. Distribution of suspension feeding animals are coincident with this and dominate where large quantities of suspended material and low sedimentation rates occur (Sellwood, 1972). Almost all the bivalves found in the Katterfeld deposits were suspension feeders but the species diversity varied widely between locations within the basin and particularly within horizons at the same locality (Table 4.2).

		Los Baguales (AMT 3)	Baguales Vista (AMT 8)	Río Simpson (AMT 4)	Cuesta Los Leones (AMT 19)	Springhill Formation Riccardi (1977)
Bivalves						
	Nucula	X	X		X	
	Grammatadon	X				
	Modiolus	X				
	Maccoyella			X		
	Oxytoma		X	X		X
	Entolium	X	X	X	X	X
	Chlamys			X		X
	Plicatula	X				
	Myophorella	X				
	Meleagrinella		X			
	Lima				X	
	Pseudolimea			X		
	Corbicellopsis	X				
	Lucina			X	X	X
	Venus				X	
	Protocardia	X				X
	Tancredia	X				
	Isocyprina	X				
	Astarte	X			X	X
	Arctica	X				
	Pleuromya	X			X	X
	Aetostreon			X	X	X
Ammonites						
	Favrella		X	X	X	
	Crioceratites				X	
Belemnites						
	Belemnopsis		X	X	X	X
Gastropods						
	Natica	X			X	
	Bathrotomaria	X				
Brachiopods						
	Lingula			X		
	Discinisca			X		
Vertebrate teeth				X	X	
Trace fossils						
	Chondrites		X	X		

Table 4.2: Comparison of similarity of faunal assemblages between sites of the Katterfeld Formation and the Springhill Formation of the Magallanes Basin.

At the base of the succession, which was seen at Los Baguales, the Katterfeld Formation outcrops as unfossiliferous mudstones overlying a volcanic tuff of the Toqui Formation. The mudstone gradually becomes more fissile upsequence and gastropods and a few thin-shelled bivalves occur within the calcareous mud rocks. Above the mudstones are two distinctive, highly fossiliferous units. The fossiliferous beds are both more than one metre thick. The lower unit contains fairly large shells of shallow burrowing bivalves in a calcareous muddy matrix. The bivalves are articulated and in, or near, their life positions within the sediment. The overlying bed contains a more restricted assemblage of shallow, infaunal bivalves but of a much smaller size and displaying a high dominance and high density of a few species. The boundary between the beds is sharp and the upper contact of the topmost fossiliferous

unit with the overlying black shales is also well defined. The suspension feeding fauna of the fossiliferous beds are typical of an inner shelf setting which require oxygenated and fairly shallow water conditions with a plentiful supply of nutrients (Sellwood, 1972). Large infaunal bivalves, especially the deeper burrowers such as *Pleuromya*, are poorly adapted to low oxygen environments, whereas small or juvenile organisms are relatively more tolerant because of their lower respiratory demands and their more favourable surface-area:volume ratios (Tyson & Pearson 1991, Wignall & Hallam 1991). The abrupt change between the fossiliferous beds is probably a reflection of a decrease in oxygen levels as the basin began to deepen and the larger organisms could not survive in the oxygen depleted waters. The opportunistic, smaller bivalves colonised the sediments and survived for a time until the basin deepened further and became more dysoxic. The absence of the larger infaunal suspension feeders probably also indicates a low environmental energy (Sellwood, 1972). This, together with the lack of any current structures and the life position of the organisms in the muddy matrix further reinforces the argument for a gradual deepening of the basin after a prolonged period of shallower conditions.

Oxygenating events are well known within black shale sequences (Wignall & Hallam 1991, Whitham 1993) but are usually short-lived and occur when storm events temporarily agitate the water column and wash in a benthos which is able to survive for short time periods until the anoxic environment re-establishes itself. This setting most likely accounts for some of the shelly horizons at Cuesta Los Leones, as they were not so well established, but at Los Baguales the fossiliferous beds represent a more substantial colonisation event. Wignall (1994) has found that modern benthic communities require between five and eight years to become established and so the beds at Los Baguales must represent a shallowing event which lasted for a considerable time period. The lower units of the Katterfeld Formation lack pelagic organisms such as ammonites and belemnites (Table 4.2). None were collected in the 70 metre exposure at Los Baguales. It is probable that the waters were not deep enough or oxygen conditions not stable enough for them to live in. It is also possible that as carnivores there was not enough available food to support them at first.

The majority of black shale in the Katterfeld Formation accumulated in a rapidly subsiding basin. Truly anaerobic facies, with no benthic taxa, are very rare in most black shale sequences (Wignall & Hallam, 1991). Under declining bottom-water oxygen levels there is a progressive loss of benthic species and infaunal organisms are driven out onto the surface (Tyson & Pearson 1991, Wignall 1994). At Cuesta Los Leones there was evidence of this on a bedding plane where numerous *Astarte* appeared with their valves open in a "butterfly effect" as they struggled for oxygen on the sediment surface.

The upper beds of the Katterfeld Formation reflect a more typical black shale fauna of low diversity, epibyssate suspension feeders, ammonites and near the top of the succession, belemnites (Table 4.2). At very low bottom-water oxygen levels epifaunal forms and byssally attached taxa are common (Allison *et al.*, 1995). They are characterised by thin shells, flat shapes and reclining habits and are highly adapted to life at the boundary of anoxia. The forms include thin shelled pectins such as *Chlamys* and *Entolium* and the epibyssate bivalves *Oxytoma, Pseudolimea* and *Lima.* With an increase in oxygen levels small bivalves and carnivores appear (Arthur & Sageman, 1994). Epibyssate bivalves were probably pseudoplanktonic and were originally attached to ammonites or the abundant driftwood. This is well known from other black shale sequences and well documented in the Nordenskjöld Formation of Antarctica (Doyle & Whitham, 1991). The thick-shelled, reclining oysters found at Río Simpson and Cuesta Los Leones are indicators of a soft substrate and a more oxygenated environment. However, their overturned position in the sediment and the evidence of attachment to ammonites or logs, suggests they were transported from elsewhere. The brachiopod, *Lingula,* found at Río Simpson, also suggests that the environment may have been turbid at times (Sellwood, 1972).

Fish-teeth and scales and ammonites, which became more frequent upsequence, were found in the upper shales at Río Simpson and Cuesta Los Leones and show that nektonic species were present higher in the water column. Belemnites were not found until near the top of the sequence and this was particularly noticeable at Cuesta Los Leones. Wignall & Hallam (1991), suggest that belemnites were controlled more than ammonites by bottom water conditions and were only present in the more oxygenated

environments. The only trace fossil found in the deeper black shales is *Chondrites* which is accepted as tolerant of the lowest levels of oxygen at the sediment/water interface (Doyle & Whitham, 1991). The fauna is representative of a middle to outer-shelf mud environment (Sellwood, 1972).

Palaeoecology shows that decreasing levels of oxygen in the bottom waters can be recognised from the sediments of the Katterfeld Formation. It reflects the initially dysaerobic conditions which allowed opportunistic colonisation of the sediments by infaunal suspension feeders before becoming predominantly anoxic with episodical dysoxic intervals and colonised by organisms higher in the water column. The fauna of the Katterfeld shales corresponds closely to the fauna found in the more sandy muds of the Springhill Formation in the Magallanes Basin (Table 4.2) and so it would seem that there was some connection between the basins at this time.

4.5.4 Summary of the Katterfeld Formation

The black shales of the Katterfeld Formation form a thick, regionally extensive succession. The parallel bedding and isolated instances of fine-grained turbidites indicate a lack of steep gradients when the muddy deposits were laid down on the bed of an epicontinental sea. Deposition of fine-grained sediment was from suspension onto a flat, sea-bed with a high input of organic debris in the form of palynomorphs and woody material from nearby terrestrial sources, possibly from the San Jorge Basin. Conditions at the sediment/water interface were generally oxygen deficient and unable to support a benthic fauna. The absence of trace fossils except for *Chondrites* in much of the black shale succession shows that oxygenation was minimal. The bulk of the shales show a restricted fauna of mainly epifaunal bivalves and pelagic organisms which accumulated in the soft sediments as they sank after death. An increase in oxygen levels probably due to a shallowing event allowed a more diverse infauna of shallow-burrowing, suspension feeders to colonise the muddy deposits at times.

The upper part of the shales are Hauterivian, dated by belemnites and the ammonite *Favrella*. The base of the succession is possibly Berriasian although the ammonite fauna is inconclusive from the lower parts of the shales. Similar faunal communities collected from the Springhill Formation in the Magallanes Basin are interpreted as Berriasian.

The extensive area of deposition and thick accumulation of shales indicates deposition on a relatively flat, epicontinental sea-bed when volcanic activity had virtually ceased. The poor circulation suggests the basin was restricted but with an opening towards the south and the Magallanes Basin. The faunal similarity between the basins indicates that there was certainly a connection during this time period. The Aysén Basin area was relatively shallow, probably less than 200 metres deep, with weak bottom currents and slow rates of subsidence which kept pace with sedimentation. It was probably wedge-shaped, with the deepest part to the south, in the Magallanes Basin but also stretching east into Argentina.

CHAPTER 5

THE APELEG FORMATION

The Apeleg Formation is the youngest of the three formations of the Coyhaique Group (Table 2.2). It is a clastic marine sequence of sandstones and mudstones which outcrop extensively throughout the study area of the Aysén Basin from Cerro Bayo in the NE (Figure 5.1, AMT 23), to Río Simpson in the SW (Figure 5.1, AMT 4). It extends into the Magallanes Basin of southern Chile where it is equivalent to the Kachaike Formation (Figure 2.6) and into the Río Mayo Embayment of Argentina to the NE of the study area (Figure 1.2). The Apeleg Formation was first defined by Ploszkiewicz and Ramos (1977) from the Apeleg Valley in Argentina (Figure 2.11). The Apeleg Formation corresponds to part of the Apeleg Series described by Russo & Flores (1953) who included it as part of the Tres Lagunas Formation (section 2.2.5). The Apeleg Formation was later recognised by Suárez and De La Cruz (1992) to extend into Chile and to correspond to part of the sediments known then as the Coyhaique Formation. The Apeleg Formation is now recognised as the upper unit of the Coyhaique Group within the Aysén Basin of Chile (see Chapter 2).

5.1 General characteristics of the Apeleg Formation

The Apeleg Formation outcrops extensively throughout the study area (Figure 5.1) usually as small cliffs in river valleys or road cuttings. It has a recorded thickness of up to 1,200 metres in Argentina (Ramos, 1981) but the maximum measured thicknesses within the Aysén Basin study area are, approximately 180 metres at AMT 15 and 160 metres at AMT 23, both near the Argentine border (Figure 5.1).

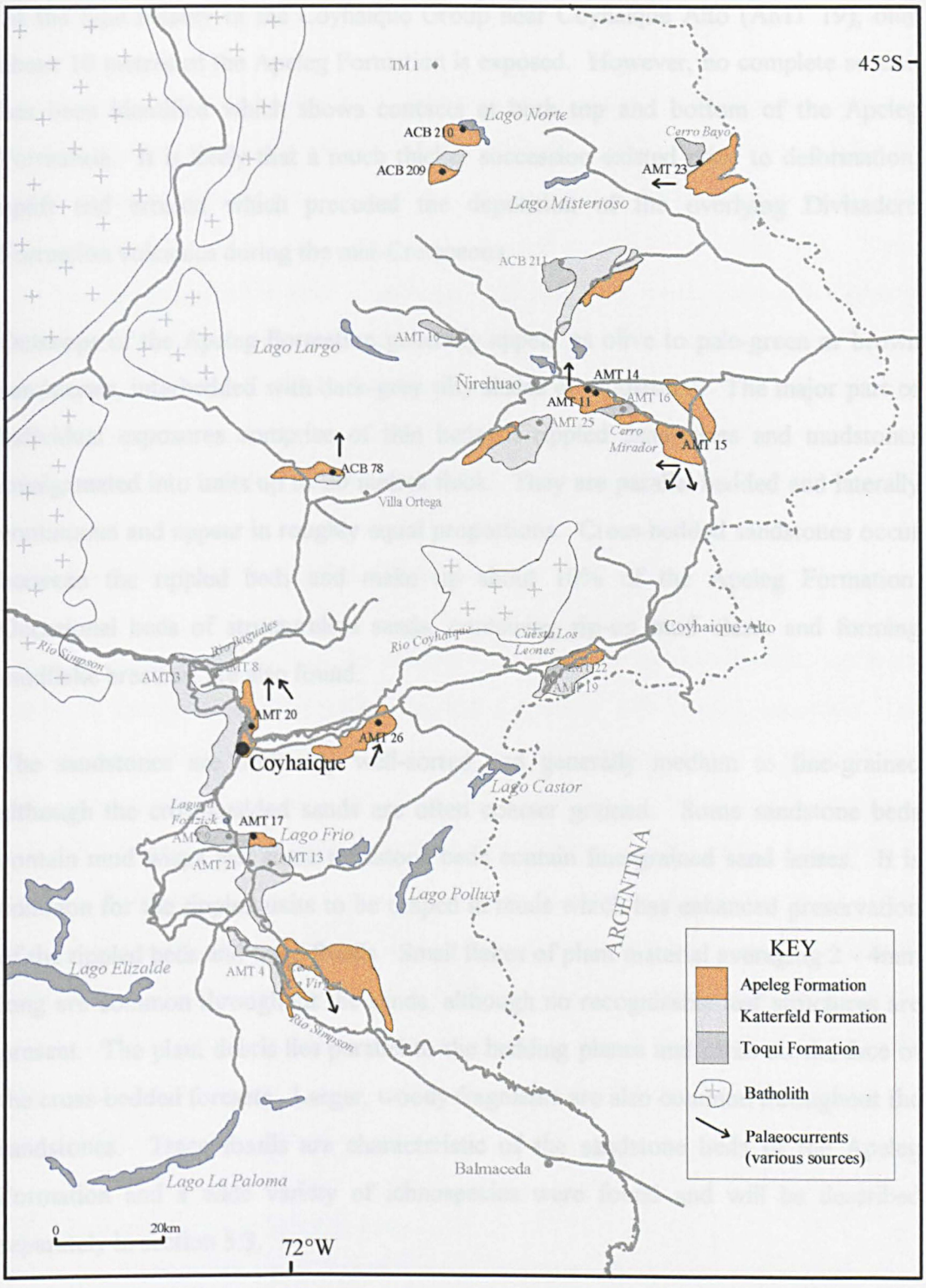


Figure 5.1: The study area highlighting major outcrops and sampling sites of the Apeleg Formation.

Basin Margins: During the Hauterivian to Albian, when the Apeleg Formation was deposited, the Aysén Basin was likely to have been more extensive than that represented by present day exposures. The northern boundary was the Somuncurá Massif (Figure 2.2) whilst a volcanic chain, represented by the present-day Patagonian Batholith, formed the western boundary. The basin extended into Argentina in the east and is recorded near Lago Fontana (Ramos & Palma, 1983). The eastern boundary has not been defined but is likely to interfinger with terrestrial deposits of the San Jorge Basin. To the south the Apeleg Formation extended into the Magallanes Basin where it is equivalent to the Kaichaike Formation.

At the type locality of the Coyhaique Group near Coyhaique Alto (AMT 19), only about 10 metres of the Apeleg Formation is exposed. However, no complete section has been identified which shows contacts at both top and bottom of the Apeleg Formation. It is likely that a much thicker succession existed prior to deformation, uplift and erosion which preceded the deposition of the overlying Divisadero Formation volcanics during the mid-Cretaceous.

Outcrops of the Apeleg Formation generally appear as olive to pale-green or brown sandstones, interbedded with dark-grey silty shales or mudstones. The major part of individual exposures comprise of thin beds of rippled sandstones and mudstones amalgamated into units up to 20 metres thick. They are parallel bedded and laterally continuous and appear in roughly equal proportions. Cross-bedded sandstones occur between the rippled beds and make up about 10% of the Apeleg Formation. Occasional beds of structureless sands, containing rip-up mud clasts and forming mudflake breccias, are also found.

The sandstones are uniformly well-sorted and generally medium to fine-grained although the cross-bedded sands are often coarser grained. Some sandstone beds contain mud wisps and some mudstone beds contain fine-grained sand lenses. It is common for the rippled units to be draped in muds which has enhanced preservation of the rippled beds and trace fossils. Small flakes of plant material averaging 2 - 4mm long are common throughout the sands, although no recognisable leaf structures are present. The plant debris lies parallel to the bedding planes and often on the face of the cross-bedded foresets. Larger, woody fragments are also common throughout the sandstones. Trace fossils are characteristic of the sandstone beds of the Apeleg Formation and a wide variety of ichnospecies were found and will be described separately in section 5.3.

The mudstones are generally grey/brown in colour and frequently occur as brittle shales, especially up-sequence. Few structures are evident apart from occasional mudcracks and sand lenses. Body fossils are rare in the mudstones, as are trace fossils.

5.2 Sampling locations of the Apeleg Formation

Good exposures of the Apeleg Formation occur throughout the study area, usually as small outcrops along river valleys and the sides of road-cuttings (Figure 5.1). A major sequence of the Apeleg Formation was recorded at Cerro Mirador (AMT 15) and a smaller, but important sequence was logged at El Rincon (AMT 20). These two sites are described in detail in sections 5.2.3 and 5.2.4 respectively and a general description of other outcrops is given in section 5.2.5. Contacts of the Apeleg Formation with the Katterfeld Formation at the base (section 5.2.1) and the Divisadero Formation at the top (section 5.2.2), were rarely observed and so location of a particular sequence within the succession is difficult to determine.

5.2.1 Base contact of the Apeleg Formation

The sandstones and shales of the Apeleg Formation conformably overlie the black shales of the Katterfeld Formation displaying a sharp and distinct transition. The contact between the Katterfeld Formation and the overlying Apeleg Formation was only seen in a few places, Río Simpson (AMT 4, Figure 5.1 & Figure 5.2a & b) and near Villa Ortega. The change from the black shales of the Katterfeld Formation is abrupt. It is marked by the appearance of small ripples of very-fine sand which increase in abundance and thickness over a few metres and form lenses of sand in beds up to 20mm thick within the black shales. These lenticular mudstones grade up firstly into flaser beds, and then the rippled sandstones and brown shales of the base of the Apeleg Formation. There is no evidence of interfingering of sediments and the transition is abrupt wherever it is observed.

5.2.2 Top contact of the Apeleg Formation

The Apeleg Formation is unconformably overlain by volcanic, volcaniclastic and sedimentary rocks of the Divisadero Formation. To the east of Villa Ortega the unconformity which marks the contact of the Apeleg and Divisadero Formations is well exposed.

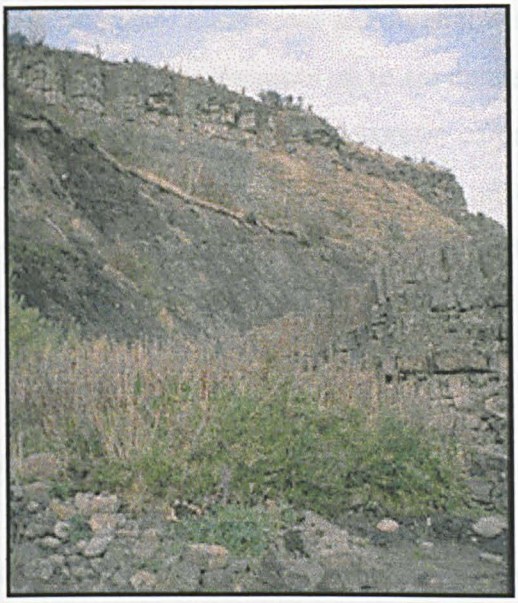


Figure 5.2a: The top of the sequence (approximately 150 metres thick) at Río Simpson (AMT 4) exposes the contact of the Katterfeld shales and the overlying Apeleg Formation sandstones. The abrupt transition occurs over a few metres as the sand content increases within the shales until it changes to the parallel bedded sandstones seen throughout the basin.

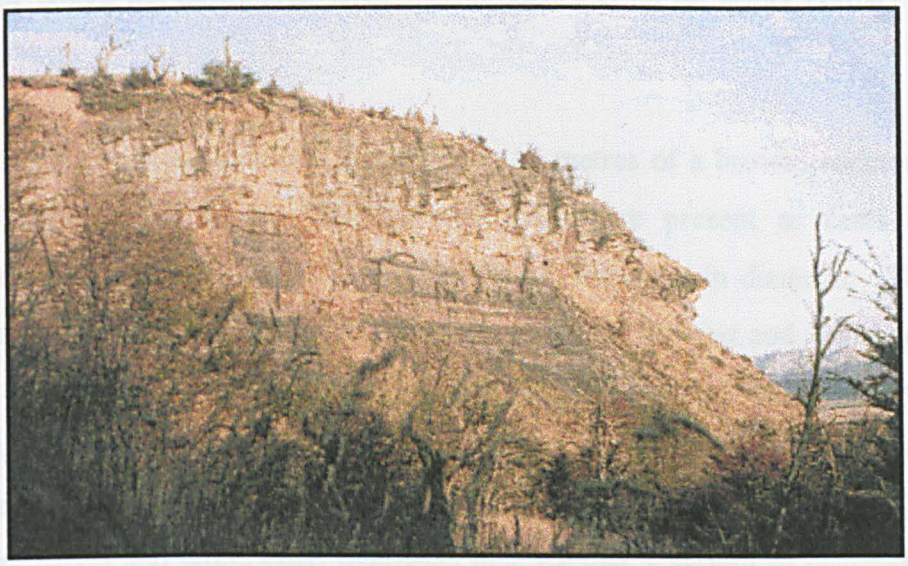


Figure 5.2b: The transitional contact of the Katterfeld shales and the Apeleg sandstones at Río Simpson (AMT 4). Photo shows the top of the exposure seen in Figure 5.2a. The abrupt change over a few metres is typical of many locations where the contact was observed.

The eroded surface of the Apeleg Formation sandstones and mudstones is overlain by a poorly sorted, matrix-supported, pebble-conglomerate up to six metres thick (Figure 5.3). The conglomerate clasts in the overlying sediment consist of angular to well-rounded fragments of sandstone and mudstone derived from the underlying Apeleg Formation, as well as angular rhyolitic clasts. The basal unit of the Divisadero Formation is likely to be a mass-flow deposit. Overlying the conglomeratic diamictite are tuffaceous sediments and a thick succession of ignimbrites.

5.2.3 Cerro Mirador (AMT 15)

The most complete section of the Apeleg Formation was documented at Cerro Mirador (AMT 15). The exposure is the most easterly of a series of gullies which cut through the Apeleg sediments towards the Río Ñirehuao, 10km east of Ñirehuao (Figures 5.4 & 5.5). Almost 200 metres of the Apeleg Formation is exposed in a steep sided gully running southwest from the valley floor to the summit of Cerro Mirador. Contact with the Katterfeld Formation at the base is not exposed but must be quite near the surface as the valley sediments are the black shales typically associated with the Katterfeld Formation. The top of the sequence is the summit of Cerro Mirador where the upper beds have been eroded and so no upper contact is present.

At the base of the measured section (Figure 5.6) is 12 metres of a brown mudstone containing a varied fossil fauna. Most of the fossils are present as casts or impressions and are found in calcareous nodules up to one metre in diameter. The nodules contain many bivalve moulds including *Trigonia, Pleuromya* and *Entolium.* Two species of gastropod are also present, a globose variety likely to be *Natica* and, because of its large size, possibly of the species *ampullispira,* and a small, slender, turreted variety. No further body fossils were found above this level although the abundant trace fossils and bioturbated sediments indicate that a variety of organisms must have colonised the seas and sea-floor. Many of the sandstone beds contain vertebrate teeth and fish-scales as well as a shark fin-spine. The vertebrate remains are associated with the mud-clast units.

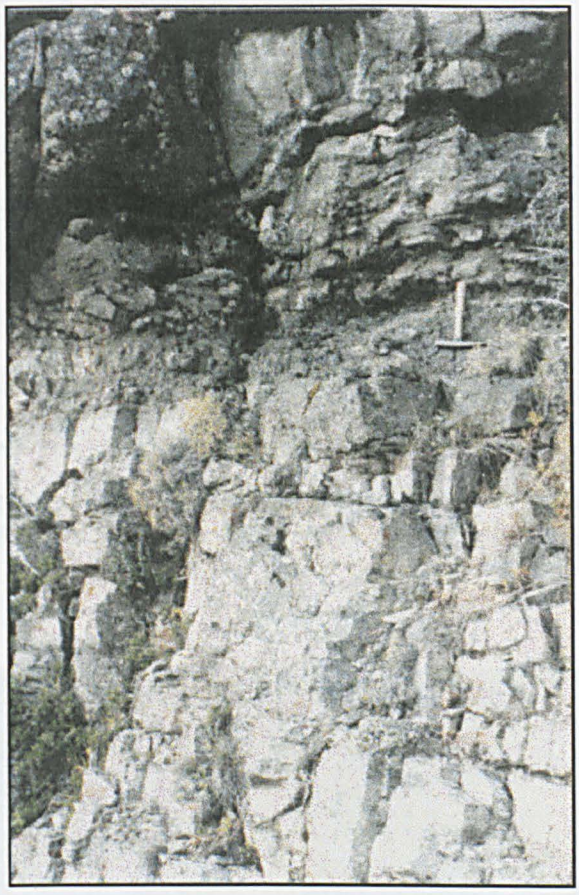


Figure 5.3: The unconformable contact of the Apeleg Formation sandstones and the overlying volcaniclastic succession of the Divisadero Formation east of Villa Ortega (Figure 5.1). The eroded surface of the Apeleg sandstones is overlain by a thick pebble-conglomerate containing well-rounded clasts of the underlying sandstones as well as angular, rhyolitic clasts. The conglomerate is likely to be the product of a mass-flow event. It is overlain by tuffaceous sediments and a thick succession of ignimbrites.

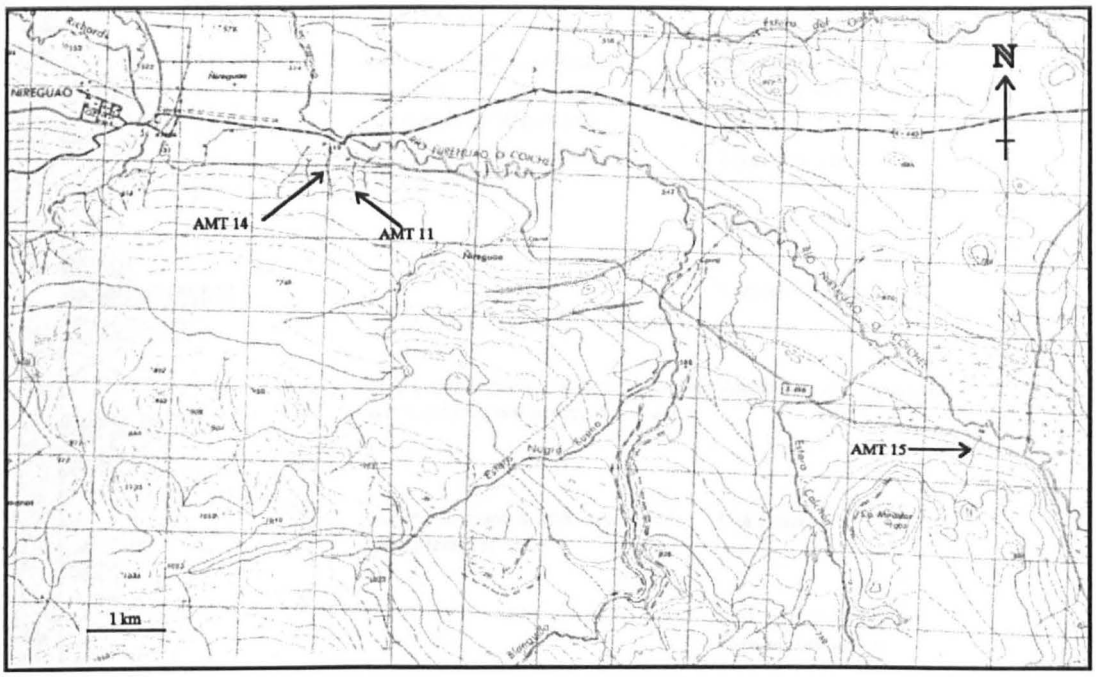


Figure 5.4: Map showing the series of gullies to the east of Ñirehuao where the Apeleg Formation sandstones are well exposed.

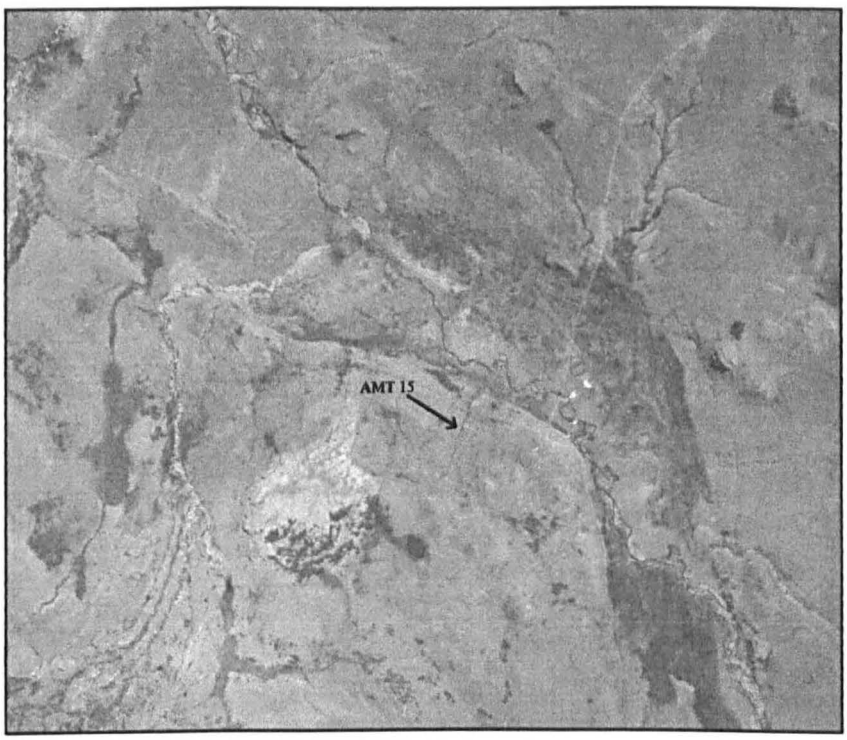


Figure 5.5: Aerial photograph of location AMT 15 where a section of the Apeleg Formation almost 200 metres long is exposed on the flanks of Cerro Mirador.

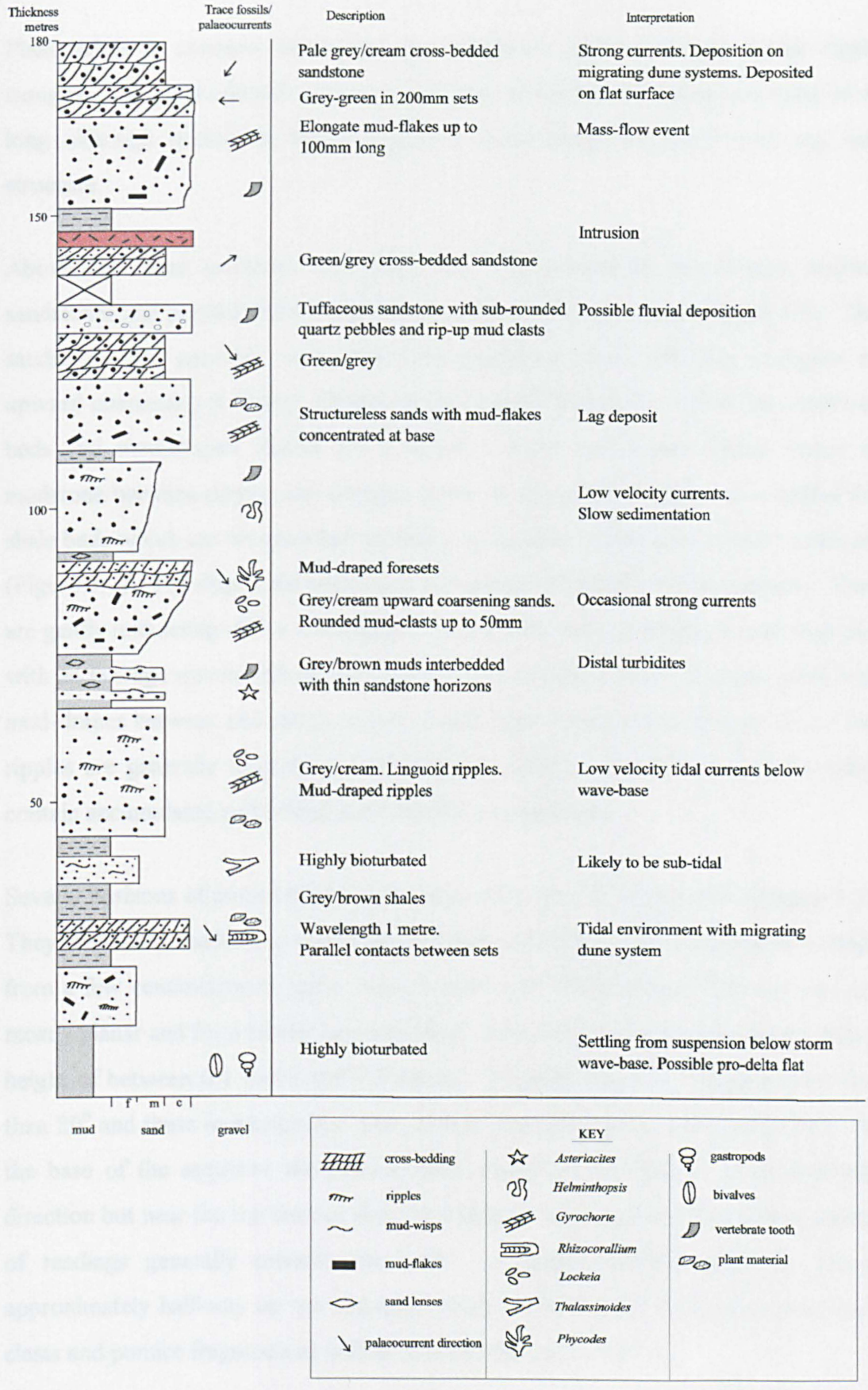


Figure 5.6: Measured section of the Apeleg Formation at Cerro Mirador (AMT 15).

Plant debris is common throughout the succession and particularly in the ripple troughs. The plant material usually occurs as small carbonized flakes averaging 2mm long although occasional longer fragments were found but none with any leaf structure.

Above the basal mudstone are more than 100 metres of interbedded, rippled sandstones and parallel-bedded shales in roughly equal proportions (Figure 5.6). The sandstones are generally well-sorted and medium-grained with few examples of upward coarsening or fining. Mudstone clasts occur frequently within the sandstone beds and mud-draped ripples are common. Some sandstones display wisps of mudstone between ripples and elongate lenses of fine-grained sands occur within the shale beds which can be described as lenticular bedding as the sand content increases (Figure 5.7). The ripples are uniform in appearance throughout the succession. They are gently undulating with a low height to wavelength ratio, averaging 50mm high and with an average wavelength of half a metre. Sets of ripples nestle together often with mud-drapes between and can form sets of more than a metre thick (Figure 5.8). The ripples are generally linguoid and symmetrical. The troughs between ripples often contain accumulated carbonized plant material or mudflakes.

Several horizons of cross-bedded sands occur between the rippled beds (Figure 5.6). They are mostly medium to coarse-grained and well-sorted. The cross-bed sets range from a few centimetres to half a metre in thickness. The contacts between sets are mostly planar and form tabular cross-bedding. Wavelength averages one metre with a height of between 0.1 metre and 0.5 metres. Foresets dip at low angles, often less than 20^0 and these measurements were used to provide palaeocurrent directions. At the base of the sequence the palaeocurrent measurements indicate a southeasterly direction but near the top current directions become less consistent and give a variety of readings generally towards the west. A coarse grained sandstone occurs approximately half-way up the sequence which contains large (>75mm) rip-up mud clasts and pumice fragments as well as sub-rounded quartz pebbles.

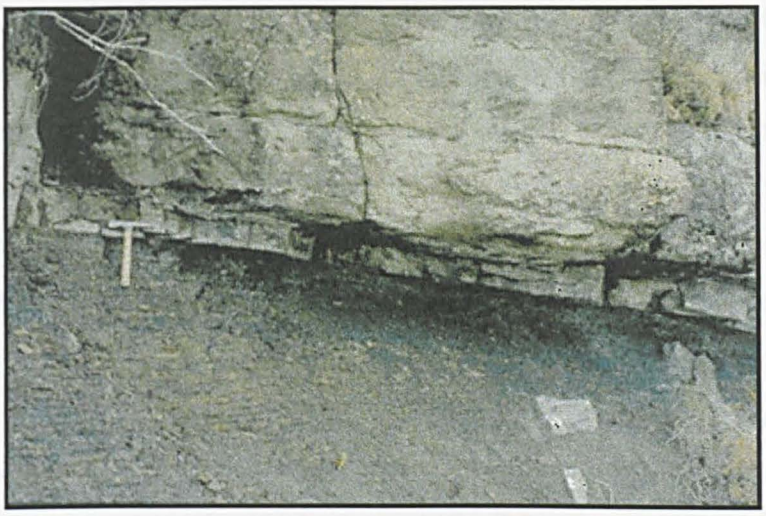


Figure 5.7: Lenticular bedding at Cerro Mirador (AMT 15). The percentage of sand gradually increases up sequence and isolated sand flasers within the shales become more frequent until continuous sand ripples form, which in turn become massive overlying sandstones, as environmental energy increases.

Figure 5.8: Mud-draped, rippled sandstones in the Apeleg Formation.

A wide variety of trace fossils are present throughout the sequence. The alternations of sand and mud beds have resulted in excellent preservation of the ichnofossils. The shale units generally do not show any preserved traces or evidence of bioturbation however, many of the sandstone beds are intensely bioturbated. In the bioturbated sandstones the disruption is very uniform with no indication that the top or the bottom of the unit is more bioturbated than the rest of the bed. *Gyrochorte* appear throughout the sequence associated with the rippled sandstones, their trails frequently following the line of the ripple troughs (Figure 5.9). *Lockeia,* the dwelling burrows of bivalves, are common and frequently found on the same surfaces as *Gyrochorte.* The lower part of the sequence contains the horizontal, lined burrows of *Rhizocorallium* and *Thalassinoides* which are absent from the upper units. Starfish traces (*Asteriacites lumbricalis*) occur in convex hyporelief on the base of a thin, rippled sandstone bed at 75 metres (Figure 5.10) and isolated examples of *Phycodes* and *Helminthopsis* are found in the sandstones. Trace fossils of the Apeleg Formation are discussed in detail in section 5.3.

5.2.4 El Rincon (AMT 20)

At El Rincon a shorter sequence of the Apeleg Formation displays the same consistent lithology seen across the study area (Figures 5.11 & 5.12). The measured sequence is just under 70 metres thick (Figure 5.13). At the base of the sequence a grey, fissile shale with no evidence of fossils grades up into a browny, softer shale containing sand lenses. The sand lenses increase in frequency and the bedding becomes lenticular eventually grading into a green, medium-grained, highly bioturbated sandstone. The rippled sandstone beds throughout the sequence are very uniform in appearance, slightly asymmetrical and with rounded crests. The average height of the ripples is 50mm and the average wavelength is 80mm. Often, ripples 'nestle' into each other with narrow horizons of mud between producing flaser bedding.

At the top of the sequence, at least 10 metres of cross-bedded sands are exposed. The contacts between beds are occasionally curved but mostly planar producing both trough and tabular cross-bedding. The palaeocurrent measurements on the topmost beds indicate a current direction towards the north and northwest (Figure 5.13).

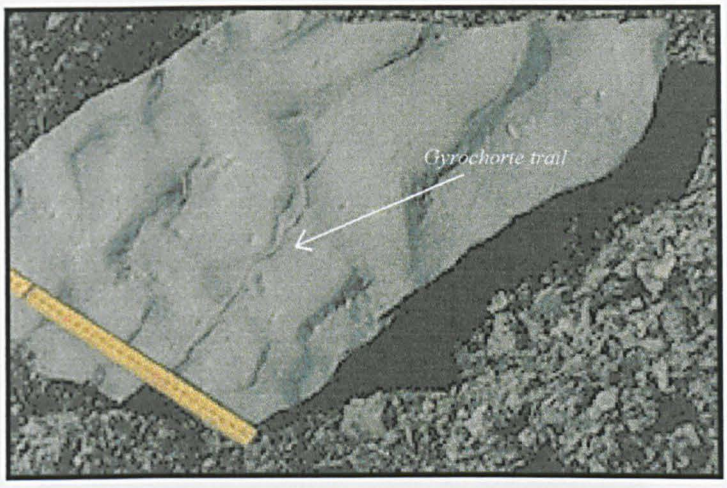


Figure 5.9: *Gyrochorte* trails following ripple troughs were commonly observed suggesting the organism mined the sediments for nutrients at the site of maximum accumulation.

Figure 5.10: *Asteriacites lumbricalis* traces on the underside of a thin, rippled sandstone bed at AMT 15.

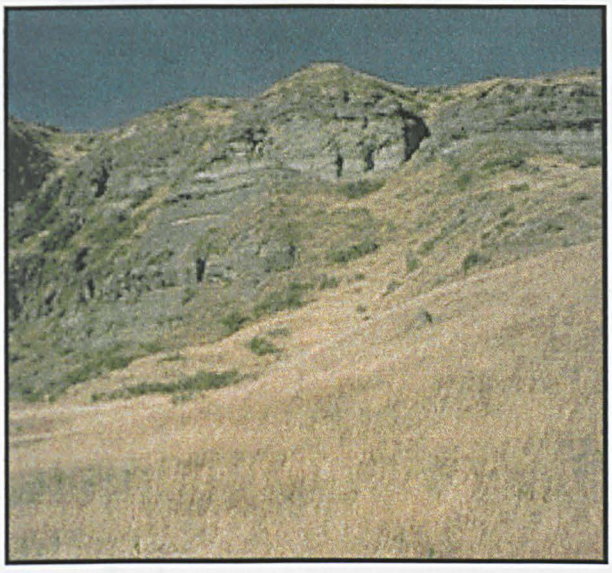


Figure 5.11: The outcrop of the Apeleg Formation at El Rincon (AMT20). The black shales of the underlying Katterfeld Formation are covered by vegetation and the contact is not exposed. The cliff is approximately 100 metres high.

Figure 5.12: The top of the Apeleg sequence at El Rincon (AMT 20). The sandstones and shales have the same uniform appearance seen throughout the study area. They are greenish/grey in colour, parallel bedded, and well-sorted. The top units are cross-bedded but the units near the base of the picture shows disrupted bedding and some evidence of fluid escape structures. Height of outcrop approximately 20 metres.

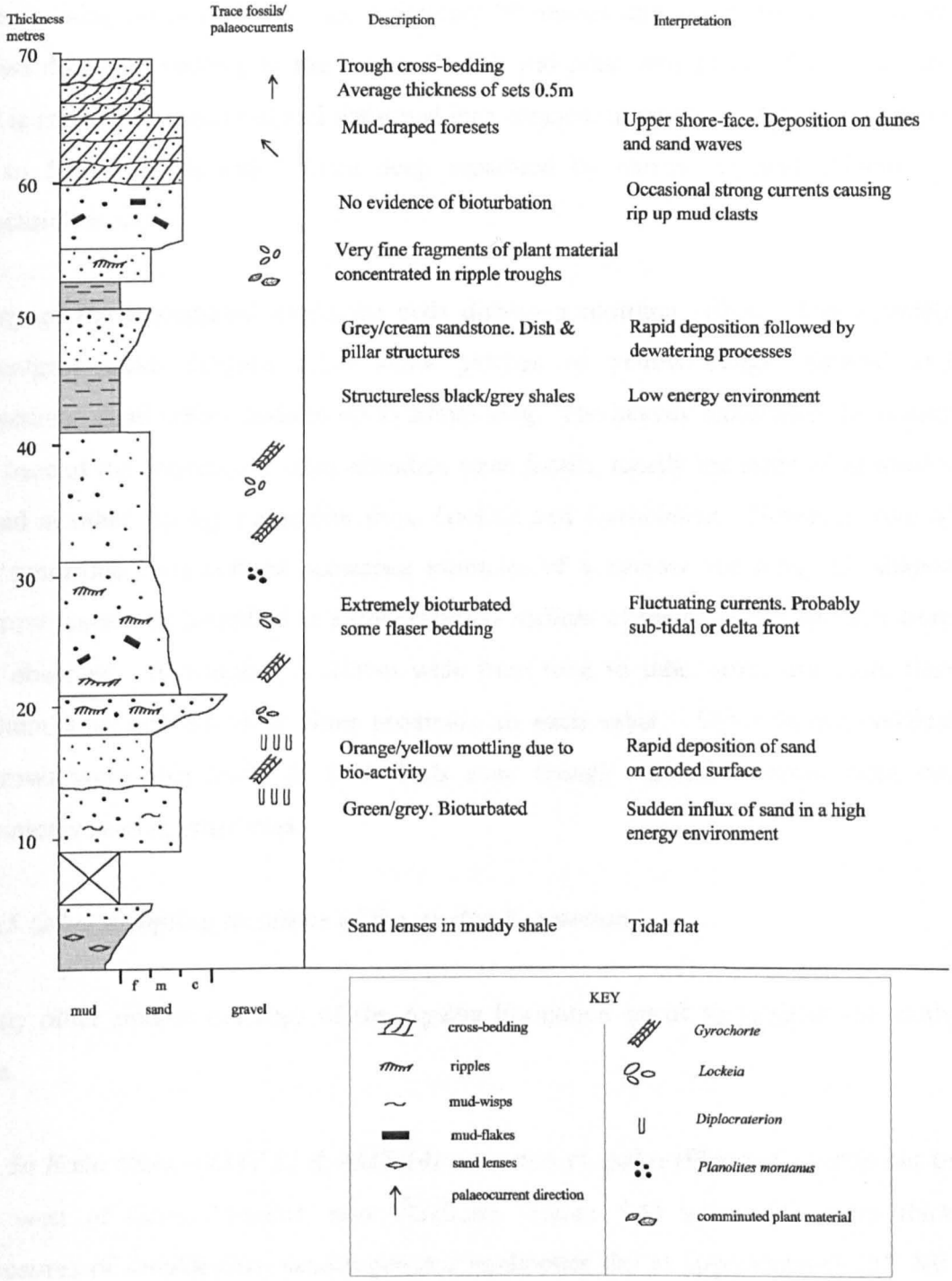


Figure 5.13: Measured section of the Apeleg Formation at El Rincon (AMT 20).

A coarsening upwards bed at approximately 50 metres into the measured sequence shows disrupted bedding at the base with dish and pillar structures. The sandstone bed is irregularly contorted and deformed into concave upwards sandstone structures up to 500mm wide and 250mm deep separated by narrow vertical columns of structureless sands.

Many of the bioturbated sandstone beds display a mottling effect. The normally green/grey sands (Figure 5.12) show patches of yellow/orange material and sometimes small yellow nodules up to 20mm long. The heavily bioturbated beds near the base of the sequence contain abundant trace fossils, mostly the same ichnospecies found at other Apeleg Formation sites, *Lockeia* and *Gyrochorte*. However, two of the sandstone beds contain numerous examples of a narrow but long 'U' shaped burrow tentatively identified as *Diplocraterion habichi* as the expected spreiten were not observed. Although only 10mm wide from tube to tube, some are more than 110mm long and found in close proximity to each other. Other larger, vertical burrows were also found in these beds even though vertical burrows were not commonly seen at other sites.

5.2.5 Other sampling locations of the Apeleg Formation

Many other smaller outcrops of the Apeleg Formation occur throughout the study area.

5.2.5a Baño Nuevo (AMT 11 & AMT 14): A series of gullys (Figure 5.1) crop out to the west of Cerro Mirador, near Ñirehuao (Figure 5.4) where 20 metre thick exposures of fossiliferous, sandstones and mudstones dip at approximately 20° NE. At the base of this succession the sandstones are thinly bedded, fine-grained with small irregular ripples and with ripple cross-bedding. The intercalated mudstones are brown initially, becoming dark-grey to black higher up the sequence. Both the lithologies at these sites contain fossils. Body fossils are concentrated in centimetre thick beds of sandstones between the rippled sandstones. The sandstones are bioturbated and contain comminuted plant material in the ripple troughs. Calcareous nodules occur within the mudstones and these contain a varied assemblage of fossils

including bivalves, gastropods and ammonites. The bivalves found are, *Chlamys octoplicatus, Entolium argentinium, Panopea* sp., *Pinna patagoniensis* (Figure 5.14a), *Pleuromya* sp., *?Protocardia* sp., *Pterotrigonia* sp. and *Steinmanella* sp. The most abundant bivalve is *Entolium argentinium.* The bivalves are generally found unbroken and articulated although many appear as casts only, indicating diagenetic dissolution of the shell material. Two types of gastropod are found within the nodules, *Turitella* sp., a high spired, spiral shaped specimen with up to nine whorls, found as a mould only, and *Natica* sp., a globose specimen with a small spire (Figure 5.14b). The ammonites found are all examples of *Favrella americana* (Figure 5.14c). Only one specimen of juvenile *F. americana* has been identified. It appears as a negative mould approximately 18mm in diameter. It is involute with straight ribs and two rows of tubercles on the shoulders.

5.2.5b Cerro Bayo (AMT 23): At Cerro Bayo (AMT 23, Figure 5.1) approximately 160 metres of the Apeleg sandstones and shales are exposed on a SW facing hillside (Figure 5.15a). The lower 45 metres comprise of interbedded sandstones and shales which are highly bioturbated. The sandstones are generally cream/grey in colour, fine to medium-grained and contain occasional calcareous concretions. The sandstones contain abundant amounts of carbonaceous plant material generally less than 2mm long and many examples of fossilised wood more than one metre long, some with *Teredo* borings. The trace fossils *Gyrochorte* and *Lockeia* are common throughout the complete sequence but at the lower level, vertical, lined burrows of *Ophiomorpha* are present. The mudstone content increases and lenticular bedding and flaser bedding are common for more than 40 metres. Rare, irregular and discontinuous mudcracks are present in the thin mudstone layers. Approximately 100 metres up sequence some laterally discontinuous shell beds up to 0.4 metres thick contain fragments of the distinctive trigonid bivalve *Steinmanella* and complete but poorly preserved specimens of the ammonite *Favrella.* Other small, articulated bivalves such as *Tancredia* and *Astarte ?erephyla* and a high-spired, turreted gastropod are also found in these discontinuous beds. The sandstones at the top of the section are medium to coarse-grained and light grey in colour.

Figure 5.14a: The bivalve *Pinna patagoniensis* found embedded in an ammonite at AMT 11. *Pinna* usually bury the lower part of their shell in the sediments and attach byssal threads to sediment particles to stablise themselves against a shifting substrate or currents. This specimen preferred to embed itself in the aperture of a recently deceased ammonite.

Figure 5.14b: The gastropod *Natica*. A globose species with a small spire which scavenged on the sediments and whose present day ancestors are carnivores feeding on other molluscs.

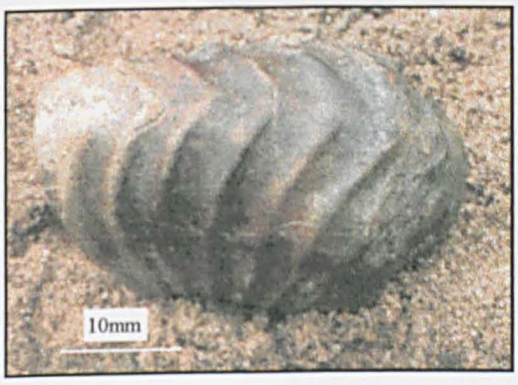


Figure 5.14c: The distinctive chevron pattern of the ammonite *Favrella americana*. Many fragments of this species were found in calcareous nodules in the lower beds of the Apeleg Formation. It has been dated as lower Hauterivian (Riccardi, 1988).

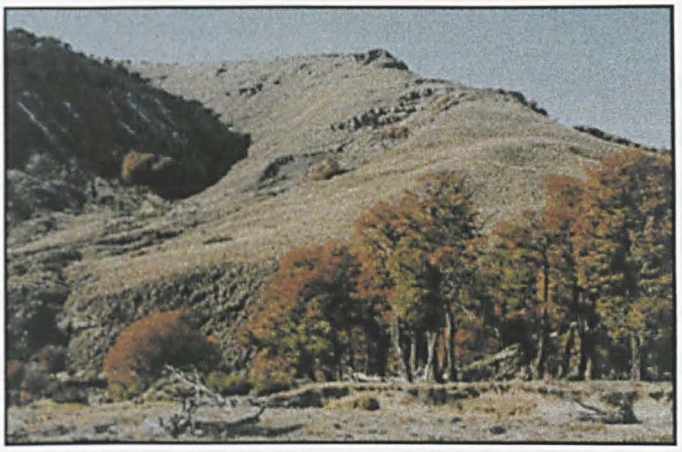


Figure 5.15a: The Cerro Bayo locality (AMT 23). Outcrop almost 200 metres high.

Figure 5.15b: Minor ripples occur at right angles to the more major dune ripples.

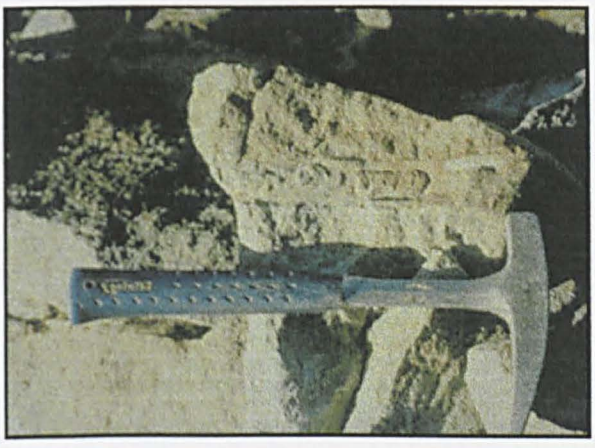


Figure 5.15c: The horizontal trace fossil *Rhizocorallium* was commonly found parallel to bedding planes. Spreiten structure infilling the area between the burrow walls can be seen.

The beds are frequently cross-stratified with palaeocurrent direction variable but generally towards the west and southwest. Many of the dune-ripple beds display minor ripples at right angles on the front face of the dune (Figure 5.15b). At the top of the sequence the horizontal trace fossil *Rhizocorallium* (Figure 5.15c) and *Gyrochorte* and *Lockeia* are very common. At this level the traces are generally horizontal and parallel to bedding planes with no vertical burrows such as *Ophiomorpha.*

5.2.5c Río Simpson (AMT 4): The outcrop of the Apeleg Formation at (AMT 4) is exposed above the black shales of the Katterfeld Formation in the deep gorge of the Río Simpson (Figure 5.2a). The contact between the two parallel-bedded formations is clearly seen but inaccessible. However, a faulted area west of Cerro La Virgen has brought down to a lower level an exposure of the Apeleg sandstones. The sandstones are identical to outcrops of the Apeleg Formation seen at locations across the Aysén Basin. They are green/grey in colour, generally medium to coarse grained and well sorted. Some beds are rippled and some contain rip-up mud-clasts. All of the sandstones have a high organic content, usually small flakes of carbonaceous plant material and frequently large pieces of fossilised woody material. Many of the beds are highly bioturbated and trace fossils are abundant. The most frequently observed trace is *Gyrochorte* often found in association with *Lockeia* (Figure 5.16a). The horizontal trace *Thalassinoides* and the sometimes vertical burrow *Ophiomorpha* (Figures 5.16b & 5.16c) are also common, although not in the same beds. The short, horizontal burrow *Planolites beverleyensis* was found in association with horizontal *Ophiomorpha irregulaire* burrows in some of the sandstones.

Figure 5.16a: Sinuous trails of *Gyrochorte comosa (Gy)* left by the organism as it exploited the nutrient rich layers in the ripple troughs. The almond shaped outlines of *Lockeia (L)* are also visible and near the bottom centre of the photograph, a broken *Lockeia* pod is sitting in an indentation made in the sediments.

Figure 5.16b: *Thalassinoides,* the branching, feeding and dwelling burrows of crustaceans. It is a characteristic trace of coherent substrates where the burrow walls do not need supporting.

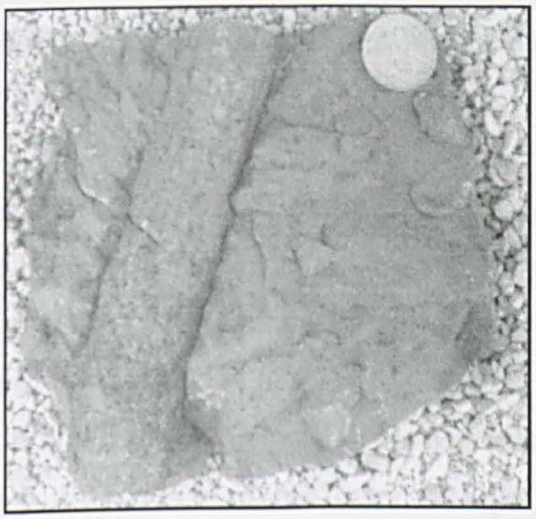


Figure 5.16c: An *Ophiomorpha irregulaire* horizontal burrow, similar to *Thalassinoides* but the top of the burrow is lined with faecal pellets to provide support in unstable sediments.

5.3 Trace Fossils in the Apeleg Formation

Trace fossils are biogenic structures that originated through the locomotion, feeding, burrowing, or resting activities of organisms. Trace fossils can be classified in several ways which are discussed further in section 5.5 however, a behavioural scheme is used in the systematic descriptions of this section which uses the following commonly recognised categories;

Cubichnia - resting traces, Repichnia - crawling traces, Pascichnia - grazing traces, Fodinichnia - feeding structures and Domichnia - dwelling structures.

SYSTEMATIC ICHNOLOGY

5.3.1 Ichnogenus *Asteriacites* (von Schlotheim, 1820)

Asteriacites lumbricalis (Seilacher, 1953)

Asteriacites lumbricalis were found in large numbers but in only one location, at Cerro Mirador (Figure 5.1, Location AMT 15). They occur in convex hyporelief, on the base of a thin (less than 50mm thick), fine-grained, rippled, sandstone bed, which overlies a bioturbated mudstone. The resting traces are well preserved, but the only other traces that occur with them are ophiuroid marks and some vague flute casts. The diameter of the specimens of *Asteriacites lumbricalis,* range from 1mm to 20mm and some of the impressions are 2 - 3mm deep. Nearly all of the casts show pentameral symmetry with little or no distortion of the position of the arms (see Figure 5.10).

Discussion: *Asteriacites lumbricalis* are regarded as resting traces of Asterozoa, produced by asteroids or ophiuroids (Häntzschel, 1975). The excellent state of preservation of the impressions of *A. lumbricalis,* is due to the sudden burial of the cubichnia by an influx of fine-grained sand. No evidence of body fossils were seen, but it is likely that the animals were smothered by the sudden, rapid sedimentation event. Modern asteroidea cannot escape from sediment more than 50mm thick (Clarkson, 1979). The range of sizes of the impressions and the undisturbed nature of the position of the arms, indicate that this is an autochthonous assemblage, most likely

of asteroids. Only as a result of catastrophic smothering events are starfish preserved whole, as their remains disarticulate easily and rapidly after death. Asterozoa often occur in vast numbers on shallow marine shelves in present day settings (Goldring & Stephenson, 1972) but also live at bathyal and abyssal depths (Clarkson, 1979). The Cerro Mirador environment was likely to have been similar to present day shelf settings, with slow sedimentation but also experiencing periodic influxes of coarser materials. Asteroidea are entirely marine organisms. They are very sensitive to influxes of fresh water and so cannot colonise estuaries or tolerate the close proximity of fresh water rivers (Clark, 1968). Most modern starfish are voracious predators and feed on mollusca, especially bivalves although others are deposit feeders and ophiuroids are suspension feeders. The sediments at Cerro Mirador reflect abundant food supplies of organic matter and there is ample evidence of the presence of bivalves to support the asteroidea.

5.3.2 Ichnogenus *Diplocraterion* (Torell, 1870)

Diplocraterion are abundant at El Rincon (Figure 5.1, Location AMT 20). They appear as U-shaped burrows with parallel limbs. The vertical burrows are perpendicular to the bedding planes and although often only one of the tubes is exposed it is possible to see concave upwards spreiten in some of the examples that occur on the face of outcrops (Figure 5.17). The average length of each burrow is 75mm and the average width of each tube is 15mm. The distance between parallel tubes averages 20mm. On many beds the top of the burrow has been eroded off prior to deposition of successive beds but on one surface it is possible to see a funnel shaped opening at the bedding surface.

5.3.2a Diplocraterion habichi (Lisson, 1904)

A finer grained bed 0.5 metres thick, at the same location, contains good examples of the ichnospecies *Diplocraterion habichi.* The burrows occur as thin, long, vertical U-shaped tubes. The longest measures 108mm with each tube having a diameter of 4mm. The parallel tubes abut each other and so there is no spreiten visible between the tubes.

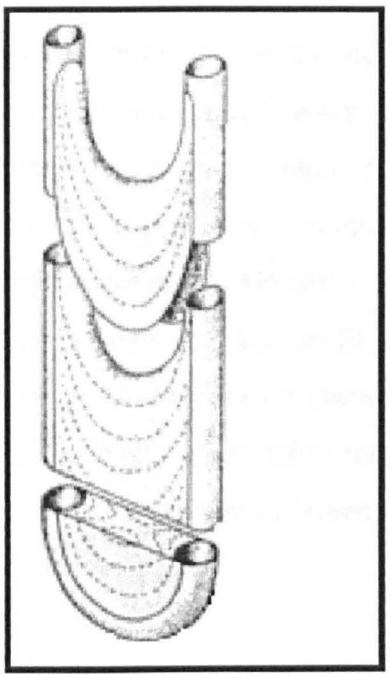


Figure 5.17: *Diplocraterion* showing spreite infill between tube walls. The example shown is protrusive which indicates an erosional environment causing the organism to burrow downwards to maintain a constant depth below the sediment surface. *Diplocraterion habichi* occur at El Rincon (AMT 20). The burrows are much longer and narrower than the usual *Diplocraterion* burrows and they do not have the distinctive spreiten.

The sandstone face exhibits fairly dense occurrences of *Diplocraterion habichi* with average distances between burrows of 50mm. El Rincon is the only location that *D. habichi* was found at.

Discussion: *Diplocraterion* is recognised to be the domichnion of a suspension feeding animal, probably living in an area of high wave energy (Häntzschel, 1975). *Diplocraterion habichi* indicates a relatively high energy environment, with insufficient sedimentation of fine-grained particles to support deposit feeders resulting in suspension feeders becoming predominant (Heinberg & Birkelund, 1984). The erosive horizons on which the *D. habichi* were seen at El Rincon, are typically found in association with occurrences of *Diplocraterion* (Häntzschel 1975, Heinberg & Birkelund 1984, Dam 1990a). The length of the tubes reflects their interpretation as protective shelters against a high energy environment with unstable sea-floor conditions and low rates of sedimentation.

5.3.3 Ichnogenus *Gyrochorte* (Heer, 1865)

Gyrochorte comosa (Heer, 1865)

Gyrochorte traces are found throughout most of the measured sandstone successions of the Apeleg Formation within the Aysén Basin, but are particularly prevalent at Cerro Mirador (Figure 5.1, Location AMT 15). They are invariably associated with the rippled sandstones. *Gyrochorte* appears as straight to winding burrows up to one metre in length. The burrows are found in epirelief as plaited ridges with biserially arranged braids of sediment and a median furrow, on the upper surfaces of bedding planes. Some specimens show a corresponding counter-expression on the underside of beds where the burrow appears as a groove with a median ridge. This feature was only definitely observed in the thinner beds (less than 30mm thick). On several of the thicker beds the suggestion of a corresponding groove could be seen, sometimes offset from the ridge on the upper surface. The specimens collected show a consistent width of burrow of between 5 and 8mm with only one particularly thin example of 3mm width. Burrow height varies between 1 and 3mm above the bedding planes. The burrows meander across the bedding planes, sometimes following the troughs of

ripples (see Figure 5.9) and at other times crossing the crests. A *Gyrochorte* burrow may intersect another without displacing either burrow and details of the underlying burrow structure are still discernible (Figure 5.18). At other times the traces are seen to run parallel for several centimetres. Some examples show a sharp, angular change of direction rather than the usual sinuous trail (see Figure 5.16a). Often a burrow will merge out from, or in to, the surrounding sediments rather than appear or disappear abruptly, as though the organism dipped under the surface of the bedding plane.

Associated structures: Ovate pods, similar to the *Lockeia* ichnogenus are intimately associated with many examples of *Gyrochorte* (Figure 5.19). The pods are up to 20mm long and 10mm high, with pointed ends giving an "almond" shaped outline. They appear in convex epirelief and do not show any surface detail. The structures occasionally disrupt the burrow rather than appear at the side or end and also commonly occur on the underside of bedding where *Gyrochorte* appears on the upper surface.

Several examples of *Gyrochorte* show a series of parallel striations or ridges crossing the burrow and extending for up to 40mm either side (Figure 5.20). This is possibly an example of the ichnogenus *Syringomorpha* (Nathorst, 1886). The ridges are thin (1 - 2 mm wide) and are in contact with each other. The ridge like striations do not always make contact with the *Gyrochorte* burrow but often appear in the near vicinity. On some examples the ridges run parallel to each other but on others they fan out from a single depression in the centre.

Discussion: *Gyrochorte* is well documented from Dorset, England (Hallam 1970, Fürsich 1975), from Yorkshire, England (Powell, 1992) and from East Greenland (Heinberg 1973, Heinberg and Birkelund 1984 and Dam 1990a, 1990b). *Gyrochorte* trails are thought to be produced by tunnelling through sediments by an organism such as a worm or gastropod. For many years it was thought that the grooves appeared on the upper surface and were the trails of grazing organisms. It was not until much later that it was proposed that the burrow was actually a tunnel structure with an epichnial expression. Hallam (1970), concluded that the burrow was made by subsurface tunnelling just below the sediment-water interface. This theory was discounted by

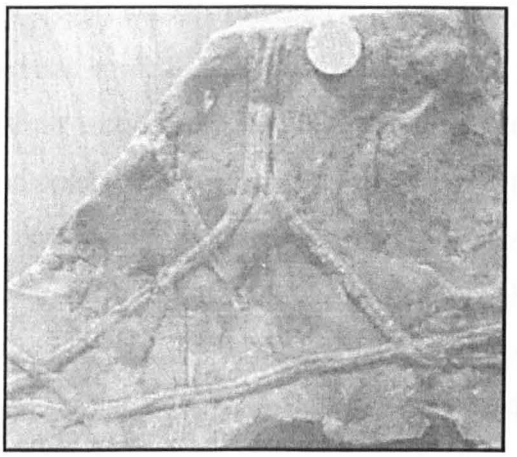


Figure 5.18: *Gyrochorte comosa* tracks running parallel to each other but sometimes crossing. The organism burrowed through the sediments and lifted previous trails rather than ploughing through and disrupting them. The lower trail is therefore the younger.

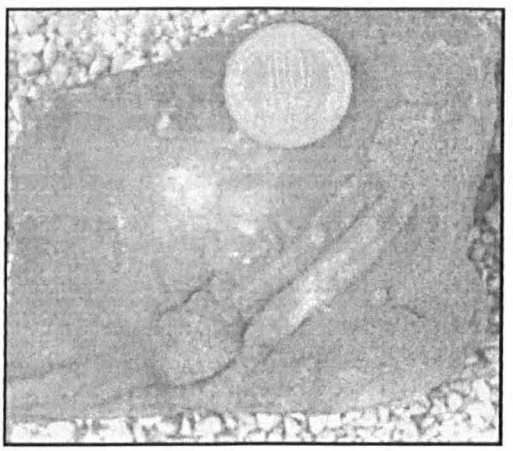


Figure 5.19: Twin tracks of *Gyrochorte* with *Lockeia* sitting centrally. The association of *Gyrochorte* and *Lockeia* at many locations was notable and leads to the suggestion that both tracks were made by bivalves.

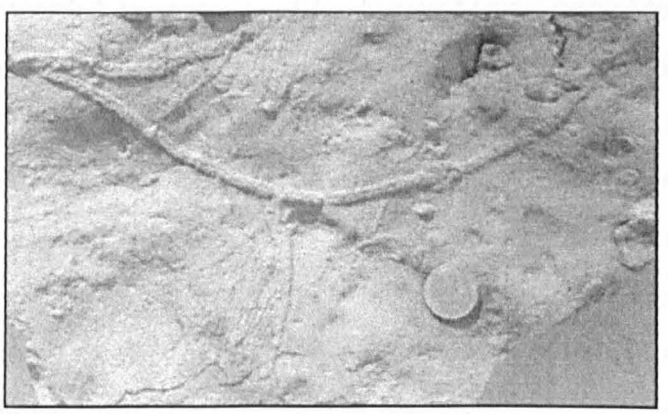


Figure 5.20: Striated traces were occasionally found in conjunction with *Gyrochorte* trails, usually where the track changed direction. It is possible they could be classed as *Syringamorpha* but there is a suggestion they were made by some surface activity of the *Gyrochorte* organism.

Heinberg (1973) who had studied examples of *Gyrochorte* from East Greenland. He noted that the vertical depth of the burrow can extend for up to 160mm and at such thicknesses it would be unlikely to produce a surface ridge. The sediments from East Greenland are extremely mica-rich, fine sandstones in which the orientated mica flakes highlight the internal construction of the burrows clearly. The examples from the Aysén Basin do not show this depth of vertical burrowing although there is a clear link on the thinner beds, between the *Gyrochorte* ridge on the upper surface of the bedding plane, and the groove on the lower surface. Heinberg (1973) concluded that *Gyrochorte* was produced by an elongate organism moving obliquely through the sediments probably at an angle of up to 40° to the horizontal. This would correspond to the offset examples of ridges and grooves found on some of the examples in the present study. The non-disruption of intersecting ridges that were clearly seen on many specimens collected in the Aysén Basin, were also observed by Hallam (1970). He proposed that the lower track is the younger one and that the first track made, is uplifted rather than ploughed through by succeeding *Gyrochorte*, so preserving the original structure of the first trace.

The association of *Gyrochorte* with the pod shaped structures similar to *Lockeia* isp. is notable. Some examples show the pod "sitting" on top of the *Gyrochorte* trail, but more commonly at the side or end. Other authors such as Hallam (1970) and Hakes (1977) recorded occurrences of *Lockeia* and *Gyrochorte* (described as a ladder trail in the Hakes, 1977 paper) from the same beds but did not comment on their association. Powell (1992), however, did note the intimate association of the two ichnospecies in the topmost beds of the Scarborough Formation of Yorkshire. He deduced that the pod represented a resting trace and was therefore likely to have been made by a mollusc rather than a worm. The striated traces also seen near *Gyrochorte* trails could belong to the *Syringomorpha* ichnogenus. Powell (1992) also describes fan shaped ridges near the bend of the *Gyrochorte* trace, suggesting sand was fanned away from the burrow by scilia or appendages as the animal changed direction. Pollard (personal communication) has noted this association from other collections of *Gyrochorte* burrows and has inferred that it is caused by some surface activity of the organism which constructed the *Gyrochorte* burrows. The striking association of the

pods and the trails suggest that the *Gyrochorte* burrows are indeed made by a mollusc rather than a worm type organism actively digging its way through the sediments. The merging of the trails in and out of the surface layer, further suggest that the organism exploited organic rich layers at differing levels as it travelled along. *Gyrochorte* is considered to be in the pascichnia category as it combines the activities of deposit feeding and locomotion. The organism was likely to be an opportunistic animal which was able to colonise mobile substrates and exploit nutrient levels below the sediment-water interface.

Gyrochorte occurs throughout the Aysén Basin wherever the Apeleg sandstones are found. The repeated occurrence of the trace at differing levels show that the organism was able to rapidly colonise mobile substrates even when a sudden influx of volcaniclastic sands occurred. *Gyrochorte* has been shown to be present in a number of different environmental settings but in the Aysén Basin it appears on ripple bedded sandstones with a range of other trace fossils which indicate shallow, tidally influenced, marine conditions with abundant food supplies.

5.3.4 Ichnogenus *Helminthopsis* (Heer, 1877)

Helminthopsis isp. was observed at Cerro Mirador and Galpos (Figure 5.1, Locations AMT 15 and AMT 26 respectively). It occurs on the surface of fine-grained sandstones. *Helminthopsis* isp. appear as irregular, smooth, meandering trails, 1 to 2mm wide, which never cross or branch. The trails are of a consistent width with raised marginal ridges (Figure 5.21).

Discussion: *Helminthopsis* are believed to be the traces of deposit feeders such as polychaete annelids. The feeding traces are probably combined with locomotory traces and can therefore be considered grazing trails (pascichnia). Pascichnia commonly follow surfaces parallel with the sea-floor if not on the sediment surface and *Helminthopsis* is produced by a well-organised annelid systematically mining the sediments at the water-sediment interface (Bromley, 1996). Most grazing traces occur in relatively quiet depositional environments below storm-wave-base (Tucker, 1982).

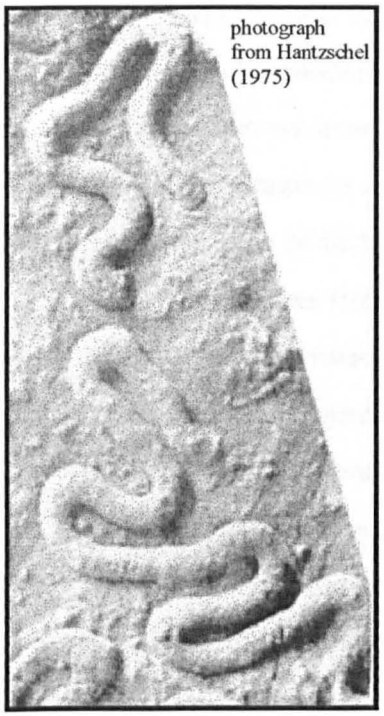


Figure 5.21: Meandering trails of *Helminthopsis* produced by an annelid type organism mining sediments at the water/sediment interface.

5.3.5 Ichnogenus *Lockeia* (James, 1879)

(=*Pelecypodichnus* Seilacher, 1953)

Abundant examples of the almond-shaped, pod-like trace, *Lockeia* were found at Río Simpson (AMT 4), Cerro Mirador (AMT 15) and El Rincon (AMT 20, Figure 5.1) as well as most other locations that the Apeleg sandstones were found. They are usual in fine to medium-grained sandstones which are interbedded with mudstones. *Lockeia* commonly occurs on the soles of bedding planes in convex hyporelief preserved in the mudstone layers, but sometimes on the tops of bedding surfaces as elongate pods in convex epirelief. There is a wide range of sizes from 2 to 40mm in length and up to 20mm in width, with a height of up to 20mm, however it is usual for the pods to all be of a similar size in any one specific location (Figure 5.22). The surfaces are smooth and evidence of ridges or ornamentation is absent despite having a wide range of samples from different locations. The pods are pointed at both ends and pinch in around the perimeter of the trace where it meets the bedding surface (Figure 5.23 & 5.24). This gives the appearance of sitting in a depression. Often the pods break off leaving a small connecting channel to the sandstone body and a sunken outline where the pod once sat. The density of the occurrence of *Lockeia* traces varies considerably. Some surfaces, usually the top surfaces of beds, only show occasional random examples while the base of the beds often show numerous examples (Figure 5.22). In general no preferred orientation of the cubichnia could be detected although occasionally individual traces are aligned behind each other suggesting repositioning of an individual organism along a line of specific direction (Figure 5.24).

Discussion: There has been much discussion as to the nomenclature of this trace. Some authors (Hakes 1977, Wright & Benton 1987) consider *Lockeia* to be a *nomen oblitum* (unused name) and prefer to use *Pelecypodichnus*, as the name *Lockeia* was out of use until the 1970s. However, Häntzschel (1975) clearly states that *Lockeia* is the form that should be used and is the name used in the present study.

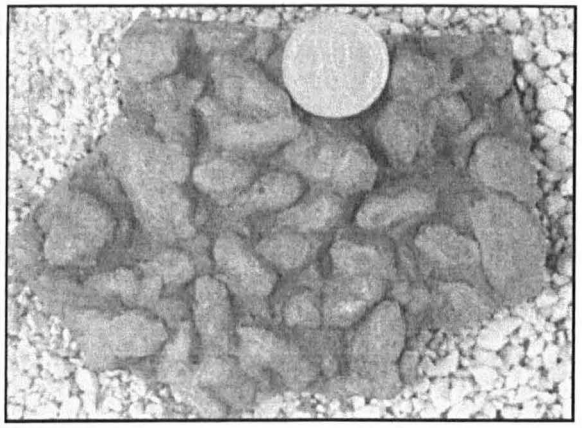


Figure 5.22: Similar sized examples of *Lockeia* on the base of a sandstone bed.

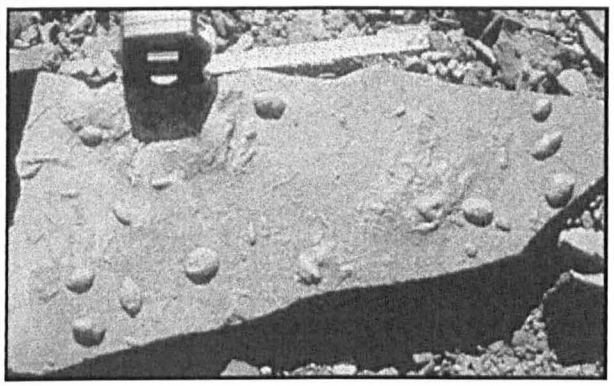


Figure 5.23: Surface expression of *Lockeia* clearly showing the almond shaped pods. The surfaces are smooth with no evidence of ridges.

Figure 5.24: Regular impressions made by the *Lockeia* organism as it followed a specific pathway across the sediments. The characteristic pods, pointed at both ends, are often found nestling into the surounding sediments, sometimes breaking off and leaving a depression.

Lockeia is considered by most authors to be cubichnia of bivalves. No body fossils were apparent in any of the traces but as Hallam (1970) noted, this is usual due to dissolution of shelly material. Wright and Benton (1987) describe sandstone facies in Staffordshire where they noted an association of bivalve moulds on the upper surface of sandstone slabs and *Lockeia* on the lower surface. They deduce that the *Lockeia* traces were the product of the bivalve *Eotrapezium* which burrowed through the sands to the underlying clay horizon. In the Aysén Basin, *Lockeia* type traces are intimately associated with *Gyrochorte* trails and possibly the result of a mollusc resting trace. *Lockeia* occurs with a wide range of trace fossils and so is a poor indicator of environmental conditions. Dam (1990b) allocates *Lockeia* to the *Cochlichnus* ichnocoenosis which reflects a well aerated, medium to low energy, shallow water environment with a good supply of food.

5.3.6 Ichnogenus *Ophiomorpha* (Lundgren, 1891)

Ophiomorpha irregulaire (Frey, Howard and Pryor, 1978)

At Río Simpson (Figure 5.1, Location AMT 4), several examples of *Ophiomorpha irregulaire* were seen. They typically appear as oval, sinuous, horizontal burrows in convex epirelief. The average diameter of the cross-section of the burrow is 25mm and the average height is 10mm. The base of the burrow is slightly flattened and the top is lined with muddy pellets containing black flecks of comminuted vegetation. The burrow walls are irregularly pelleted which gives it a distinctive "knobbly" appearance. The pellets have a conical or pointed shape, are 1 to 2mm long, and spread out into the surrounding sandstone producing "flame-like" structures. The burrow fill appears to be of the same composition as the host sandstone (see Figure 5.16c).

Discussion: *Ophiomorpha irregulaire* is regarded as the dwelling structure of a suspension feeding organism such as a thalassinidian shrimp (Frey & Howard, 1970). Comparisons of recent and ancient structures show that the trace has considerable value as a facies indicator. They are widely interpreted as indicators of marine littoral and sub-littoral environments (Kennedy & Sellwood 1970, Häntzschel 1975). However, more recent publications (Frey *et al.* 1978, Pollard *et al.* 1993) point out

that *Ophiomorpha* type burrows can be made by a variety of organisms under a varying range of environmental conditions from nearshore lagoons and estuaries, to offshore deeper-water sediments. *Ophiomorpha* can also be found in non-marine facies (Pollard *et al.*, 1993). The horizontal burrows of *Ophiomorpha irregulaire* commonly occur at lithological boundaries. The smooth floor and the pelleted roof lining reflects the need for stronger support from above than from below. The pelleted wall lining is likely to be related to substrate coherence, with the more unstable environments requiring more support from the mammalated lining. Burrows constructed in soft, watery sediments may be generally similar to those constructed in loose shifting sands. Cohesiveness of the substrate being more important than the texture of the sediments (Frey *et al.*, 1978). The drawn-out "flame-structures" of the burrow walls, clearly seen at the Río Simpson location, are thought to reflect soft sediment deformation of the muddy wall lining. The depositional environment of the sandstones containing *Ophiomorpha irregulaire* was likely to have been a normal marine, inner-shelf with shifting sands and unconsolidated muds.

5.3.7 Ichnogenus *Phycodes* (Richter, 1850)

Phycodes palmatus (Hall, 1852)

One example of *Phycodes palmatus* was collected from Cerro Mirador (Figure 5.1, Location AMT 15). It is preserved in convex hyporelief as a horizontal, semi-circular structure of bundled burrows emanating from one point and spreading out into a gentle curvilinear, or fan shaped feature. The individual tubes are 2 to 4 mm wide and 15 to 30mm long. They thicken towards the exterior part of the burrow and become up to 8mm wide. The tunnels tend to merge into each other and so some of the burrow walls are not distinct. The burrow fill is a fine to medium-grained sandstone, the same as the surrounding host rock.

Discussion: *Phycodes* is produced by a deposit-feeding worm-like animal (Häntzschel, 1975). It systematically mined forwards and then retraced its steps and made a new tunnel but Dam (1990a) suggests the burrow could have been made by any one of a number of organisms. *P. palmatus* is a much finer species than many of the other *Phycodes* species and it is likely to have been produced by some form of

annelid. It has been reported from the high Andes of Argentina (Mangano & Buatois, 1994) where it was found in conjunction with seven other species of *Phycodes.* The environment at Cerro Mirador was likely to have been well aerated with an abundant food supply to support deposit feeders. The rippled sandstone bed *Phycodes* was found in suggests a medium energy shelf environment with some tidal influence.

5.3.8 Ichnogenus *Planolites* (Nicholson, 1873)

5.3.8a Planolites beverleyensis (Billings, 1862)

Planolites beverleyensis was found in the medium to coarse-grained, grey/green sandstone of the Apeleg Formation at Río Simpson (Figure 5.1, Location AMT 4). The burrow consists of short cylindrical or sub-cylindrical segments up to 20mm long and 4 to 10mm in diameter (Figure 5.25). It occurs in convex hyporelief on the base of bedding planes and often fades into the sandstone at the end of the burrow. The burrow runs parallel to the bedding plane, is straight to gently curved in form and does not branch. The sediment fill of *P. beverleyensis* is structureless and tends to merge with the surrounding sediments.

5.3.8b Planolites montanus (Richter, 1937)

Planolites montanus was found at El Rincon (Figure 5.1, Location AMT 20). It is smaller in size and generally has a more curved burrow than *Planolites beverleyensis.* It averages 1 to 2mm in diameter and sometimes has short branches. The burrow fill is a more muddy sand than the surrounding sediment but is structureless. The surrounding sediments are heavily bioturbated.

Discussion: *Planolites* probably represents active backfilling of sediment in a burrow created by a worm like deposit feeder (Dam, 1990a). Howard and Frey (1984) and Pollard *et al.* (1993), noted the appearance of *Planolites* around *Ophiomorpha* burrows and this concurs with a similar association at Río Simpson (AMT 4). The *Planolites* association with *Ophiomorpha* suggests a low energy environment with relatively slow sedimentation. Dam (1990b) indicates that this could be a bioturbated shelf or distal part of a subaqueous fan delta.

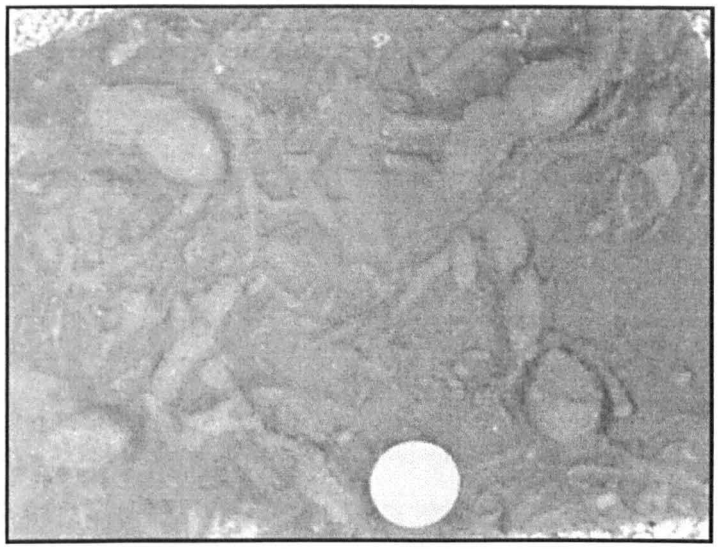


Figure 5.25: Short cylindrical burrows of *Planolites beverleyensis* on the base of a bedding plane. It is seen in association with *Lockeia* The *Lockeia* organism followed a preferential route through the sediments and created a trail of resting pods. The environment was likely to have been of low energy with relatively slow rates of sedimentation.

5.3.9 Ichnogenus *Rhizocorallium* (Zenker, 1836)

Rhizocorallium was found at a number of locations but most prominently seen at Cerro Bayo and Cerro Mirador (Figure 5.1, Location AMT 23 & AMT 15). It is found in a horizontal position along bedding planes. It is a simple U shaped burrow with protrusive spreiten. The arms of the burrow run parallel and are about 20mm apart and 5 to 10mm thick (see Figure 5.15c). The burrows average 250mm in length but were probably longer than this as very few were complete. The sandstone beds that *Rhizocorallium* were found on were fine to medium-grained and contained much comminuted plant material.

Discussion: *Rhizocorallium* are interpreted as fodinichnia, the burrows of deposit-feeding animals, by Dam (1990a). The organisms often incorporate faecal pellets into the walls of the burrow. The most likely trace makers are crustaceans or annelids. Environmental indications suggest a quiet, calm, inner-shelf environment when organic rich material was deposited but this was regularly interrupted by storm events which buried the organic material.

5.3.10 Ichnogenus *Scolicia* (de Quatrefages, 1849)

Scolicia appears as a sinuous burrow up to 120mm long and 30mm wide with a depth of 20mm. It is convex with a flattened top surface. The burrow fill is of a coarse to medium grained sand but shows no structure. The burrow occurs on the base of a bedding plane.

Discussion: Only one example of *Scolicia* was present at La Fraille (Figure 5.1, Location AMT 13). It was found in association with *Gyrochorte comosa* and *Lockeia* although these appeared at a lower level in the sequence. *Scolicia* is regarded as the feeding trail of burrowing gastropods by Häntzschel (1975), although Dam (1990a) regards large polychaete or a holothurian to be the most likely trace makers. *Scolicia* appears in the Rhizocorallium ichnocoenosis of Dam (1990b) and suggests a well aerated environment in an inner-shelf setting characterised by short periods of low energy.

5.3.11 Ichnogenus *Thalassinoides* (Ehrenberg, 1944)

Thalassinoides occurs frequently in the finer sandstones and interbedded mudstones at Cerro Mirador (Figure 5.1, Location AMT 15). It is found as horizontally and vertically branching cylindrical burrows. It forms large branching systems with smooth walls, often in Y or T shaped formations. At the point of branching a node or enlargement is often discernible (see Figure 5.16b). The diameter of the burrows range from 15 to 40mm but are usually of a consistent thickness within the burrow system. The walls of the burrows are thinly lined in patches of darker material but the burrows themselves are of the same material as the surrounding sandstones.

Discussion: Häntzschel (1975), asserts that *Thalassinoides* traces are undoubtedly the feeding and dwelling burrows of crustaceans. The thinly lined walls of the burrows are characteristic of coherent substrates where wall reinforcement is unnecessary (Howard and Frey, 1984) and this corresponds to the interbedded muds and sands at Cerro Mirador. In certain instances *Thalassinoides* has a very similar appearance to *Ophiomorpha* and a gradational connection can be made (Myrow, 1995). *Thalassinoides* is nearly always found in fine-grained substrates where no linings are needed, whereas *Ophiomorpha* is more frequently found in unconsolidated substrates. *Thalassinoides* presence usually indicates a hardground surface. It has been found in a varied range of environments from tidal flats to shore-line settings and at depths of up to 300 metres (Myrow, 1995). Fürsich & Heinberg (1983) found that *Thalassinoides* often occur as an isolated ichnospecies whereas Myrow (1995) found it in association with *Chondrites*. The palaeoenvironment where *Thalassinoides* was most abundant at Cerro Mirador was likely to have been of low energy and probably intertidal or subtidal. Thalassinoides is found in the *Curvolithos* ichnocoenosis of Dam (1990b) which occurs in the distal portion of subaqueous fan delta environments.

5.4 Lithology of the Apeleg Formation

In hand specimen the Apeleg sandstones appear pale-green to brown in colour, well-sorted, occasionally coarse-grained but usually medium to fine-grained and generally well cemented. Their appearance is remarkably consistent wherever they are found throughout the basin.

In thin section the sandstones are again consistently uniform in their attributes. Fifteen samples were studied in thin section. The examples were selected to give the widest available range of stratigraphical and geographical distribution. Grain size in the study samples vary from coarse sand to very-fine sand size (1mm - 0.05mm) with occasional granule size grains. The sandstones are generally well sorted and within each specimen the size range is very consistent. The overall shape of the grains are angular to sub-angular with occasional sub-rounded grains and these are usually the volcanic clasts. Quartz overgrowth is common and forms a euhedral outline to many of the grains. There is very little matrix present in any of the samples studied.

The composition of the Apeleg sandstones is uniform. The most abundant component of the samples are lithic clasts. These average approximately 50% of each sample. Mostly they are very-fine grained volcanic clasts probably andesitic in composition. They are generally structureless and many show alteration to chlorite. A smaller percentage of the lithic clasts are metasediments, which include quartzites and muscovite schists, and a few mudstone intraclasts. Feldspars comprise approximately 25% of the sample. The majority of the grains are plagioclase although there is a very small amount of microcline in some specimens. The feldspars are extensively altered to clays and the mica to sericite. Quartz is the next most common constituent, comprising approximately 10% of each specimen although in some samples it forms up to 25%. The quartz is of two types, monocrystalline and polycrystalline which appear in approximately equal proportions. The monocrystalline grains are unstrained, some contain vacuoles and many are embayed. Some of the polycrystalline quartz grains have sutured boundaries and are aligned in a preferred direction which is characteristic of a metamorphic source, possibly a mica schist. The polycrystalline quartz is generally more rounded than the other grains.

Some specimens showed some inter-penetration of the quartz grains although no physical squashing or fracturing. Glauconite is found in many of the samples, though not all and comprises approximately 10% of the rock. The grains occur as sub-rounded individual pellets with a distinctive pale-green colouration in plane-polarised light and a deeper, grainy, olive-green in crossed-polarised light. Some of the grains show a brown limonite rim, resulting from oxidation of the ferrous iron. The sandstone specimens also contain minor quantities of muscovite and opaque minerals and some contain distinctive cellular tissue of plant matter.

5.5 Interpretation of the Apeleg Formation

The sediments of the Apeleg Formation were deposited under marine conditions to the east of a volcanic chain. The sandstones and shales are regionally extensive. They occur in all parts of the study area of the Aysén Basin and extend eastwards into Argentina as well as southwards where they are equivalent to the Kachaike Formation in the Magallanes Basin. The Apeleg Formation is characterised by its uniform appearance and narrow range of lithologies and sedimentary structures. The considerable thickness of the sediments, especially in Argentina and near the Argentine border, indicates a long term depositional environment with constant energy levels and little local variation in depositional processes.

The Apeleg Formation conformably overlies the black shales of the Katterfeld Formation with an abrupt vertical transition over a few metres. The two formations do not interfinger at any of the sites visited in the present study although Ramos (1981) and Ramos & Palma (1983), report that in the Lago Fontana area of Argentina (Figure 2.11), there are no exposures of the Apeleg Formation near the border possibly due to non-deposition and, to the east, the Katterfeld and Apeleg Formation sediments interfinger laterally and overlie Upper Jurassic volcanic rocks. The discrepancy between the two areas suggests that although receiving the same sediments, environmental conditions differed. In Argentina the interfingering of the different lithologies have been interpreted as characteristic of a transgressive deltaic system (Ramos & Palma, 1983). However, at all the contact sites visited in the present study the transition is abrupt and conformable indicating that the change from the low energy environment of the black shales of the Katterfeld Formation to the higher energy environment represented by the Apeleg Formation, was regionally immediate and synchronous and not due to any gradual, transgressive event. A similar sharp transition between the palaeogeographically equivalent Río Mayer and Kachaike Formations at 49^0S (Riccardi, 1988) suggests a regionally significant event over a large area of the supercontinent.

After a long period of relatively stable environmental conditions during the Hauterivian (dated by the ammonite *Favrella*) and possibly Barremian and Aptian (no

dateable fossil evidence), in which the Apeleg sediments were laid down, volcanic activity re-commenced. The Divisadero Formation comprises subaerial volcanic and sedimentary rocks unconformably overlying the Apeleg Formation in the north of the region but apparently having a conformable contact with the Apeleg sediments in the south, near Lago La Paloma (Figure 1.4) (Bell & Suárez, 1997). The base of the Divisadero Formation is probably diachronous with recorded ages of between Hauterivian and Albian (Suárez & De La Cruz, 1994a).

5.5.1 Interpretation of the depositional environments of the Apeleg Formation

The sediments of the Apeleg Formation can be separated into three main lithological categories which represent the environmental conditions prevalent at the time of deposition; rippled sandstones with interbedded mudstones and shales, cross-bedded sandstones and massive sandstones containing rip-up mudclasts.

5.5.1a Rippled sandstones with interbedded mudstones and shales: The majority of the Apeleg sediments comprise of interbedded rippled sandstones and mudstones varying in thickness from centimetres to several metres thick. The mudstones occur in thin beds between the rippled sandstones, sometimes as mud-drapes on the rippled surface or in the ripple troughs forming flaser and lenticular bedding. The ripples are uniform in scale and appearance and often amalgamated into sets several metres thick.

Mudrocks are common in the Apeleg Formation representing up to 40% of the lithology present. Because the mudrocks are easily weathered they are poorly exposed at most sites. The majority of the mudstones are interbedded with the rippled sandstones. Shales are more common higher in the sequence. The mudstones are generally structureless and massive with no visible laminations. The massive nature may be due to depositional mechanisms where the sediment was viscous, containing little water at the time of deposition, as in debris or mudflows (Tucker, 1981) or, it may be due to bioturbation effects where the original laminations have been destroyed by burrowing activities of organisms as they mined the sediments for nutrients. Except in the basal beds at Cerro Mirador (AMT 14) there are no body fossils in the mudstones or shales and, although trace fossils are abundant in the

sandstones, none are found within the mudrock units. Never-the-less bioturbation is the most likely reason for the lack of lamination as there is ample evidence of biogenic activity in the other beds. The horizontal, parallel nature of the bedding and absence of any evidence of slumping features or palaeoslopes where mass-flows could have originated, indicates that deposition was likely to have been on the relatively flat sea floor and therefore unlikely to have been the result of debris flows.

The linguoid shape of the sandstone ripples and the uniform, well-sorted nature of the sands, result from transportation by moderate tidal currents over a period of time (Boggs, 1987). Mud drape deposition is favoured by time-velocity asymmetry, such as ebb and neap tides, and low strength tidal currents (Johnson & Baldwin, 1986). Where mud deposition is intermittent with ripple migration, mud can be deposited in ripple troughs and streaks of mud will occur between ripple sets, forming flaser bedding (Tucker, 1981). Flaser bedding suggests deposition under fluctuating hydraulic conditions. Periods of current activity, when transportation and deposition of rippled sands takes place, alternate with periods of quiescence when mud is deposited (Boggs 1987, Ricci Lucchi 1995). Lenticular bedding forms when mud dominates and isolated sand lenses occur. Both types of bedding are common in tidal and sub-tidal environments (Reineck & Singh, 1980) and also in marine delta-front environments where fluctuations in sediment supply and current velocity are widespread (Boggs, 1987).

Mudcracks are rare but found in the mud-drapes between ripples at Cerro Bayo (AMT 23). The irregular and discontinuous nature of the cracks suggest that they are syneresis cracks which formed due to shrinkage in subaqueous muds. They commonly occur in thin mudstones interbedded with sandstones (Boggs, 1987) as found in the Apeleg sediments. The cracks are unlikely to be due to subaerial exposure, as no evidence of features associated with emergence such as rainspots, are present and cracks due to desiccation by intermittent exposure are polygonal and more sharply defined (Ricci Lucchi, 1995) than those found at Cerro Bayo. Nodules are common in the lower part of the Apeleg sequence at Baño Nuevo (AMT 11 & 14). The nodules are ellipsoidal and contain a range of fossils. Some nodules contain only one specimen, usually an ammonite, whilst others contain an amalgamation of

many small bivalves. The fossils are uncrushed and this would suggest that the sediment was soft and uncompacted at the time of formation, indicating early diagenesis (Tucker, 1981).

The lower beds of the Apeleg Formation at Baño Nuevo (AMT 11 & 14) are fossiliferous but it is a distinctive normal to restricted marine fauna. The abundance of the fossils indicate that there was ample oxygen and nutrients to support a mixed fauna which allowed predators such as the ammonites to be able to survive. The articulated and unbroken nature of the shells show that energy levels were fairly low. It is probable that sediment deposition resulted from settling out of mud from suspension and transportation of sand by weak currents, below stormwave base, in an offshore shelf environment. Although many examples of *Favrella americana* were found it was very rare to find the juvenile form. Only one other example of the juvenile form of *F. americana* was found and that was in the Katterfeld shales at Coyhaique Alto.

Although body fossils were rare in the majority of the sandstone units, vertebrate teeth were present throughout the Apeleg sandstones, indicating that there was ample fauna to support predators. The teeth have been identified as from laminid sharks but a fin spine from another species, not a laminid, was also collected. Laminid sharks range from Tithonian to Recent. Modern laminids are mainly pelagic but can frequent shallow waters (Sanson, 1997, personal comm.).

5.5.1b Cross-bedded sandstones: The cross-bedded sandstones are more common higher in the Apeleg sequence. The tabular and trough cross-bedding seen at most sites are the product of the migration of large-scale ripples or small dunes. Ripples on the stoss side of the dunes indicate that the currents were tidal (Blatt, 1982), although palaeocurrent directions show a predominantly unimodal flow direction with no evidence of herringbone cross-stratification. This would suggest that either the flood, or the ebb, current was dominant. The sinuous bedforms, mud drapes and the smoothing of the dune crests by flow in the opposite direction to dune migration, points to a tidal environment (Ricci Lucchi, 1995). The absence of hummocky cross-bedding or any other evidence of storm activity indicates that deposition was below

wave base (Reineck & Singh 1980, Ricci Lucchi 1995) probably between 30 and 100 metres.

From the palaeocurrent measurements that were taken from the foresets of the cross-bedding, the overall impression is that there is no preferred current direction but many variations both locally and regionally. González-Bonorino and Suárez (1995) recorded more than 70 measurements from the Cerro Bayo locality (AMT 23), in the northeast of the Aysén Basin, and they were able to deduce that there was a strong unimodal current direction to the west and south-west which later changed to a northerly direction. Most commonly documented palaeocurrents are unimodal due to a number of factors which can modify data, but enhancement of tidal currents by storms is the most common process to affect current direction (Johnson & Baldwin, 1986). The present study did not provide enough data to allow a meaningful interpretation of palaeocurrent direction, although the general picture is of a random pattern of distribution, whilst at Cerro Mirador (AMT 15), south of Cerro Bayo, current directions indicate sediment transportation generally towards the southeast. Tidal current variation can be characteristic of a narrow gulf setting where an ebb dominated current is enhanced by additional unidirectional currents of fluvial, oceanic or meteorological origin (Bridges, 1982). However, the most common cause of a polymodal distribution of palaeocurrents is oscillatory flow of currents typically found in a shallow shelf environment (Tucker, 1981).

5.5.1c Massive sandstones with rip-up mudclasts: Homogeneous sandstone beds containing rip-up mudclasts make up a small percentage of the Apeleg Formation. They are generally structureless except for a few fluid escape features. De-watering structures are not common but were recorded at El Rincon (AMT 20) and La Fraille (AMT 13). At both sites the fluid escape features occur as dish and pillar structures with the dish structure more pronounced. Dish and pillar structures are indicative of rapid deposition and form by escape of water during consolidation of the sediments. They are common in deltaic and shallow marine deposits (Boggs, 1987).

The intraclasts found in the massive sandstones have the same lithology as the interbedded mudrocks and were caused by occasional strong currents ripping up

flakes of the mudstones and incorporating them into the mass-flow deposit. Mudflake clasts are also found on the foresets of the cross-bedded sand units and infilling shallow channels and troughs between dunes.

5.5.2 Interpretation of the lithology of the Apeleg sandstones

The Apeleg sandstones are classed as physically and compositionally immature lithic arenites. They contain less than 15% fine-grained matrix and with a rock fragment content in excess of the feldspars. The identifiable plagioclase occurs as partially altered phenocrysts as well as individual crystals. The immaturity of the sandstones indicate that they are very close to their source rocks. They are well-cemented and show a moderate degree of compaction which results in a lack of matrix and low porosity. The inter-penetration of quartz crystals indicates that the grains underwent a degree of compaction and solution of grains at points of contact. This is the first stage of pressure solution and suggests that the loading stresses were moderate as, with more intense pressures, the grains become sutured and grain boundaries cannot be defined (Blatt, 1982). Quartz overgrowth is common and the source of the silica for the overgrowths is likely to be from the dissolution of grains due to the compaction. Chlorite occurs as a replacement mineral with many grains showing chlorite rims. The absence of chlorite rims at grain to grain contacts show that the replacement post dates the compaction event. The chlorite presence is likely to be responsible for the distinctive green appearance of the Apeleg sandstones although glauconite could also account for some of the colouration. Occasionally some sandstone samples contain a calcitic cement which infills the pore spaces.

Glauconite is an indicator of exclusively marine conditions and usually forms in fairly shallow waters and in areas with low sedimentation rates (Tucker 1981, Adams *et al.* 1984). Glauconite is considered to have an organic origin, usually as faecal pellets and sometimes as aggregates of crystals. Glauconite minerals grow initially in pores and fissures of foram tests and faecal pellets at the sediment-water interface, in slightly alkaline marine waters and where the oxic-anoxic boundary is located close to the surface (Winn, 1994). At present glauconite is being formed on many continental shelves at depths of a few tens to hundreds of metres (Tucker, 1981). Recent work

has shown that although glauconite is a reliable indicator of marine settings it can also occur as a product of inorganic processes resulting from the replacement of feldspars (Chaudhuri *et al.* 1994, Amorosi 1995). It can also be present in re-worked sediments as an allochthonous deposit. The pelleted nature of the glauconite and the absence of any evidence of stages of glauconitization of feldspar grains in the Apeleg sandstones indicates the likelihood of an organic origin for the grains, but they are probably allochthonous within the detrital sandstones.

5.5.3 Interpretation of the trace fossil assemblage

Trace fossils reflect animal behaviour rather than actual body remains and can therefore be used to compliment environmental interpretations gained from palaeontological and sedimentological methods. Ichnology reveals important information that cannot be gained from usual palaeontological studies. It is rare for trace fossils to be subject to transportation or reworking. Furthermore, their occurrence in unfossiliferous rocks can show that organisms were present when their absence would possibly indicate an unfavourable environment for fauna. Trace fossils can also represent soft bodied fauna that would not otherwise have been represented in the fossil record because they lack hard parts. This could be a large part of the biocoenosis. Dodd and Stanton (1981) estimate that up to 70% of the species in modern marine benthic communities are soft bodied and have little potential for preservation.

Several types of classification schemes for ichnofossils have been in vogue at one time or another. Trace fossils have been categorised by; i) morphological - biological parameters (Seilacher, 1953), ii) position within the sediment (Seilacher 1964, Martinsson 1970) and iii) behavioural patterns (Ekdale *et al.*, 1984).

A classification relating a trace fossil to the organism that made it (scheme i), is difficult to quantify as there can be many variations. The same trace may be made by several organisms or, the same organism can make a different trace under differing environmental conditions. A refinement to scheme ii), was introduced by Martinsson (1970), categorising fossils on their position relative to the original depositional

surface. It uses purely descriptive terminology with no interpretation of the trace or trace producer. It is only the position of the trace fossil in the sediment that is considered and it has to be supplemented by further analysis and interpretation of trace fossils in general (Häntzschel, 1975).

The most commonly used scheme is behavioural classification (Scheme iii) and is used in the present work. This scheme allows ichnocoenoses to be related to water depth and so provide an ichnofacies interpretation (Table 5.1). This is a useful quality as water depth reflects marine palaeoenvironments such as water energy, food supply and sedimentological characteristics.

Ichnospecies	Classification	Originator	Ichnofacies	Trophic Class
Asteriacites lumbricalis	Cubichnia	Asteroid	Cruziana	Deposit feeder
Diplocraterion habichi	Domichnia	Annelid	Skolithos	Suspension
Gyrochorte comosa	Pascichnia	Bivalve?	Cruziana	Deposit feeder
Helminthopsis isp.	Pascichnia	Annelid	Cruziana	Deposit feeder
Lockeia isp.	Cubichnia	Bivalve	Cruziana	Suspension
Ophiomorpha irregulaire	Domichnia	Crustacean	Skolithos	Deposit feeder
Phycodes palmatus	Fodinichnia	Crustacean	Cruziana	Deposit feeder
Planolites beverleyensis	Pascichnia	Annelid	Zoophycos	Deposit feeder
Planolites montanus	Pascichnia	Annelid	Zoophycos	Deposit feeder
Rhizocorallium isp.	Fodinichnia	Crustacean	Cruziana	Deposit feeder
Scolicia isp.	Pascichnia	Polychaete	Skolithos	Deposit feeder
Thalassinoides isp.	Fodinichnia	Crustacean	Cruziana	Deposit feeder

Table 5.1: Trace fossils found in the Apeleg Formation showing; behavioural classification, likely originator of the trace, depth related ichnofacies and trophic class.
(Adapted from Dam 1990a & 1990b)

Behavioural patterns of benthic organisms are controlled by environmental factors and although individual trace makers have evolved through time, basic behaviour has not. Similar behaviour patterns can be made by a variety of different organisms and it is possible to compare categories of behaviour. Since the 1980s classification schemes using ichnofabric analysis have become popular and have replaced ichnofacies analysis in facies interpretations to some extent. Ichnofabric is sedimentary rock fabric

resulting from biogenic reworking (Ekdale *et al.*, 1984). It therefore includes all bioturbation, including trace fossils recorded as "mottling", not just identifiable ichnotaxa. This categorisation is useful when analysing sediments in vertical section or core. Recent papers (Droser & Bottjer 1989, Taylor & Goldring 1993) use quantitative methods to provide a bioturbation index which allows comparative studies to evaluate the extent of bioturbation and assign a descriptive grade to the ichnofabric. This classification frees ichnotaxa from their confined associations with ichnofacies as some ichnogenera such as *Planolites* or *Chondrites*, cross ichnofacies and appear in more than one facies association (Goldring, 1995). The present study uses the more traditional ichnofacies interpretation as not enough data is available to make useful comparisons using the ichnofabric classification scheme.

Trace fossils are best preserved in alternating layers of sands and shales or mudstones. These are the lithologies that are found in the Apeleg Formation throughout the Aysén Basin. A distinctive feature of the formation is the almost complete absence of body fossils in the sandstones associated with trace fossils. If it were not for the ichnofossil evidence these environments would have been interpreted as unfavourable for living organisms. It is not unusual for interbedded shales and sandstones to exhibit a lack of body fossils as diagenetic processes tend to produce acidic porewaters under these circumstances which dissolve out body fossils. The diagenetic process often actively enhances preservation of trace fossils (Dodd & Stanton 1981, Pemberton *et al.* 1990). However, the Apeleg Formation is nearly always bioturbated with abundant examples of trace fossils wherever the formation outcrops, indicating that the basin was colonised by a variety of animals. By studying the trace fossil sequence throughout the basin it is possible to detect subtle changes in the Aysén Basin environment at the time of deposition of the Apeleg Formation sands and muds during the Hauterivian, dated by the *Favrella*, and the Lower Albian, when the Apeleg Formation was overlain by the Divisadero Formation ignimbrites (Table 2.2).

The ichnofossils which indicate the deepest level found in the Aysén Basin are *Planolites* isp. and *Helminthopsis* isp. *Planolites* is found in two forms and in most abundance at two sites (Table 5.2).

	western sites		eastern sites	
Ichnospecies	**Río Simpson**	**El Rincon**	**Cerro Mirador**	**Cerro Bayo**
Asteriacites lumbricalis			X	
Diplocraterion habichi		X		
Gyrochorte comosa	X	X	X	X
Helminthopsis isp.			X	
Lockeia isp.	X	X	X	X
Ophiomorpha irregulaire	X			
Phycodes palmatus			X	
Planolites beverleyensis	X			
Planolites montanus		X		
Rhizocorallium isp.			X	X
Thalassinoides isp.	X		X	

Table 5.2: Ichnocoenoses at the main sampling sites.

At Río Simpson the smaller *Planolites montanus* is the common form whilst at El Rincon *Planolites beverleyensis* is more dominant. *Planolites* pascichnia burrows are easily overlooked and are often more abundant than would at first appear (Howard & Frey, 1984). They are probably responsible for a considerable amount of the bioturbation that is evident at Río Simpson and El Rincon. Under the ichnofabric classification *Planolites* is considered a facies-crossing ichnotaxa by Goldring (1995) as the burrow is usually common where more food is abundant. *Planolites* is found in association with *Ophiomorpha* isp., *Gyrochorte comosa* and *Lockeia* isp. This ichnocoenosis is typical of the *Skolithos* ichnofacies of Pemberton *et al.* (1990) although no *Skolithos* are present. This probably reflects the absence of high energy events as *Skolithos* commonly occur in storm beds. The ichnofauna that are present indicate a lower shoreface facies. The low diversity is common in this facies and suggests a limitation of oxygen within the sediments. The sediments consist of fine to medium grained sands interbedded with shales. There is a high content of organic matter and carbonised plant detritus which, although providing ample food for the ichnofauna probably also caused a lack of oxygen and so limited the diversity of organisms able to colonise the sediments. The *Ophiomorpha irregulaire* indicates

that the sediments were probably unconsolidated and unstable therefore the burrows required extra support by lining with faecal pellets.

At El Rincon *Planolites montanus* is found with *Lockeia* isp. and *Gyrochorte comosa* which exhibit the typical, striated, rib-like pattern possibly caused by some surface activity of the organism. The thinly bedded shales and sandstones are highly bioturbated many beds showing yellow mottling. A thin bed (0.5 metres thick), near the top of the sequence, contains many narrow, U-shaped dwelling burrows (domichnia) of *Diplocraterion habichi* where the tubes provided shelter in unstable sediments. This is the only ichnospecies found in this bed. Goldring (1995), noted that *D. habichi* is associated with colonisation of the top of storm event units. *D. habichi* is well known from the Jurassic and Lower Cretaceous sediments of East Greenland and is indicative of a high energy environment with insufficient sedimentation of fine-grained particles to support deposit feeders (Heinberg & Birkelund, 1984). In Greenland it is associated with transgressive surfaces and omission suites (Dam, 1990b). The bed containing *D. habichi* at El Rincon is a fine to medium-grained, pale grey-green sandstone which has a sharp contact with the overlying coarse-grained, rippled sandstone. It is likely to represent a storm event within the sandstone/shale sequence when a vast amount of loose, fine-grained sands were washed in and colonised by the *Diplocraterion habichi* organism before being covered by coarser sediments. The environment returned to calmer, deeper, waters and *Lockeia* isp. and *Gyrochorte comosa* organisms re-established themselves.

The most complete vertical section was recorded at Cerro Mirador. More than 120 metres of interbedded sands and shales of the Apeleg Formation were examined for the trace fossil assemblages. By looking at a complete vertical section it is possible to see discrete changes in the environment over time. At the base of the section there are about 12 metres of bioturbated mudstone before approximately 10 metres of medium to coarse sandstone with several thin horizons (averaging 70mm thick) of a mudflake breccia. Above the mudflake horizons the sandstones become finer grained and show frequent mud wisps. The rippled surface sandstones contain a high percentage of organic material and show some evidence of cross-bedding. Along these bedding surfaces the horizontal trace fossil *Rhizocorallium* isp. is found.

Rhizocorallium burrows belong to deposit feeding animals such as crustaceans who exploited the organic detritus which accumulated under quiet conditions. The horizons of mudflake breccia and rippled sandstones indicate that these conditions did not always prevail but regular storm events occurred which buried the organic material and ripped up clasts of the mudstone layers which were obviously consolidated at that time. *Rhizocorallium* is also recorded at Cerro Bayo (AMT 23) (Table 5.2). Interbedded shales and cross-bedded, rippled sandstones with numerous mudflake beds, are exposed for more than 30 metres above the *Rhizocorallium* horizons. The dominant ichnofossils in these sandstone beds are *Gyrochorte comosa* and *Lockeia* isp. *Rhizocorallium* does not appear in these beds and it is possible that a change to a higher energy environment restricted the colonisation of the sediments, which are still rich in organic matter, to the highly adaptable *Gyrochorte* organism which was able to take advantage of the highly mobile, sandy substrates deposited under a tidal regime. A further change occurs as the sandstones gradually give way to thicker beds of shales and mudstones reflecting quieter conditions. The horizontal and branching burrows of the trace fossil *Thalassinoides* isp., become common. This trace indicates that the sediments were fairly consolidated as it does not line its burrow walls, a trait that is characteristic of coherent substrates. The environment was likely to have been sub-tidal at this time. The quiet conditions continued and *Asteriacites lumbricalis* are found on the sole of a thin, fine-grained sandstone bed in convex hyporelief. They are preserved as negative impressions in the mudstone layer they had been resting in, the association with *Thalassinoides* providing evidence that the sediments were consolidated but soft enough to leave traces in. The fine-grained sandstone that covers the Asterozoa is probably the result of a sudden influx of material, perhaps from a distal turbidite, which buried them instantly. An increase in energy level of the palaeoenvironment is indicated by the predominance of rippled surfaces in the heterolithic sands, with *Gyrochorte* and *Lockeia* again becoming the dominant ichnofauna but also appearances of *Phycodes palmatus*. This ichnocoenosis suggests a well-aerated, medium-energy, shelf environment, subject to tidal influence.

5.5.4 Summary of the Apeleg Formation

The abrupt transition from the black shales of the Katterfeld Formation to the rippled sandstones and silty-shales of the Apeleg Formation indicate a sudden influx of coarser material and a change to higher energy conditions. The layered nature of the contact with no evidence of interfingering suggests a widespread regional event as environmental conditions changed simultaneously throughout the basinal area studied. Ramos (1981) studied the Apeleg Formation in the Lago Fontana area of Argentina which is a little north of the present study area, at 44° LS. He observed that there are no exposures of the Apeleg Formation to the west (the area of the Aysén Basin) possibly due to non-deposition, however to the east the Katterfeld shales and Apeleg Formation sediments interfingered laterally and lie conformably on Upper Jurassic volcanic rocks. Furthermore he regards the lower member of the Apeleg sediments to be a continental facies containing fresh-water fauna.

This work has found no evidence of fresh-water fauna in the Apeleg Formation at any of the sites visited. González-Bonorino and Suárez (1995) also suggest a possible fluvial origin for the upper units at Cerro Bayo (AMT 23). The suggestion appears to be based on the appearance of beds of coarse-sand containing a low percentage of mud but with carbonaceous partings, which they interpret as a braided, fluvial system. The mudflake horizons are interpreted as forming in a tidal-plain environment where inter-channel muds dried out and mudflakes were then deposited in the bottom of channels (González-Bonorino & Suárez, 1995).

From the present study it is apparent that the Apeleg Formation sandstones across the study area generally have a high mud content and a high proportion of plant detritus. The mudflake breccias are also common to most outcrops of the Apeleg Formation, not only at Cerro Bayo. It is much more likely that they formed in a subaqueous environment where strong currents ripped up the consolidated muds and incorporated them into the foresets of cross-bedded sand units as well as the channel areas. A further indication that the area is not terrestrial is the appearance of a normal marine fauna of bivalves and ammonites which were collected less than fifteen metres from the top of the sequence. The bivalves *Tancredia* and *Astarte* are typically found as

shallow burrowers in shallow-water, near-shelf environments (Clarkson, 1979). It could be argued that the fluvial setting occurred above this level but the ichnofaunal evidence is consistent with a normal marine, shallow-shelf environment (Table 5.1 & Table 5.2) from the base of the section, continuously up to the top. Hallam (1975), found *Gyrochorte, Thalassinoides* and *Rhizocorallium* characteristic of a shallow sub-tidal regime with no evidence of tidal flat emergence, in sandstones in Germany. He found them comparable with traces in modern sediments of up to 40 metres depth.

Organic detritus including large pieces of woody material confirms that the area was near a terrestrial source, but no firm evidence of braided river systems were seen. At Cerro Mirador (AMT 15), a little further south than Cerro Bayo, the presence of starfish traces shows that for at least some of the time, strictly marine conditions prevailed as asteroidea cannot tolerate even brackish water. The glauconite content of the sandstones is a further indicator of marine conditions and this was found in most of the thin sections studied. It also contributes to the distinctive green colouration of the Apeleg Formation sandstones which is a characteristic feature throughout the basin. It is therefore most unlikely that the Cerro Bayo locality was fluvial.

It is possible that the Aysén Basin was slightly deeper and of a lower energy environment in the southwest, at Río Simpson (AMT 4), where the trace fossils *Ophiomorpha* and *Planolites* are found (Table 5.2). At El Rincon (AMT 20), to the north of Coyhaique, energy levels were much higher, reflected by the presence of *Diplocraterion*, and the evidence of strong tidal currents and frequent influxes of sediments. The sediments at El Rincon were heavily bioturbated in places, especially the beds that *Diplocraterion habichi* were found in. Dam (1990b) found the long tubes that the organism produced were protection against a high energy environment and unstable conditions, with low sedimentation rates and periods of non-deposition, possibly in a lower offshore setting. The high density of bioturbation also indicates that the rate of biogenic re-working exceeded the rate of sedimentation (Bromley, 1996).

The locations nearer to the Argentine border are much thicker than those to the southwest of the basin but the lithologies remain constantly similar. The main body of the Apeleg Formation comprises a heterolithic facies of rippled sandstones and mudstones. These sediments suggest alternating tidal currents which deposited the uniform, fine to medium-grained sands, and low energy periods when muds were allowed to settle out from suspension. The thick, uniform successions of sandstones and mudstones suggest that energy levels were constant over fairly long periods of time. At times the currents were sufficiently strong enough to rip up the consolidated mudstones which created the mudflake breccias, and form the cross-bedded sand waves or dunes. The sandstones were deposited as migrating sand ridges on flat surfaces with flat interbar areas between. Deposition in the interbar areas was slow and this allowed organisms to colonise the areas resulting in the significant bioturbated areas that were found. There was very little evidence of slumping. The absence of other storm features suggests that deposition was below wave-base probably at depths of between 30 and 100 metres. The suggestion is reinforced by the ichnofacies associations which would favour a shallower depth of up to 50 metres.

CHAPTER 6

MODELS OF BASIN EVOLUTION

Although the Aysén Basin has been described as a backarc basin in the literature (Skarmeta 1976, Baker *et al.* 1981, Haller & Lapido 1982, Niemeyer *et al.* 1984) the characteristics of the basin of the study area do not seem to fit the established models of basin evolution associated with a subducting oceanic plate below continental lithosphere. This chapter will set out the characteristics of the Aysén Basin before summarising the various established models for basin evolution in similar plate tectonic settings. These include; intra-arc, retroarc-foreland and backarc basins (Figure 6.1). The final section will discuss the Aysén Basin in the context of the established models.

6.1 Introduction

Volcanic arcs are generally arcuate or linear belts typically exceeding 1,000 km in length and ranging from 50 to 250 km in width, which parallel subduction zone trenches. The trenchward side of the arc is characterised by a sharply defined volcanic front of active volcanoes. A diffuse zone of volcanism may extend more than 200 km behind this alignment with most volcanism concentrated within 100 km of the volcanic front, so the boundary from intra-arc to backarc setting is indistinct (Smith & Landis, 1995). Forearc and backarc basins are located adjacent to the arc platform, intra-arc basins are located on the arc platform, either immediately in front of, or behind the active volcanic front. Many backarc basins originate with rifting within the arc platform and are therefore intra-arc basins during the early stages of development (Smith & Landis, 1995).

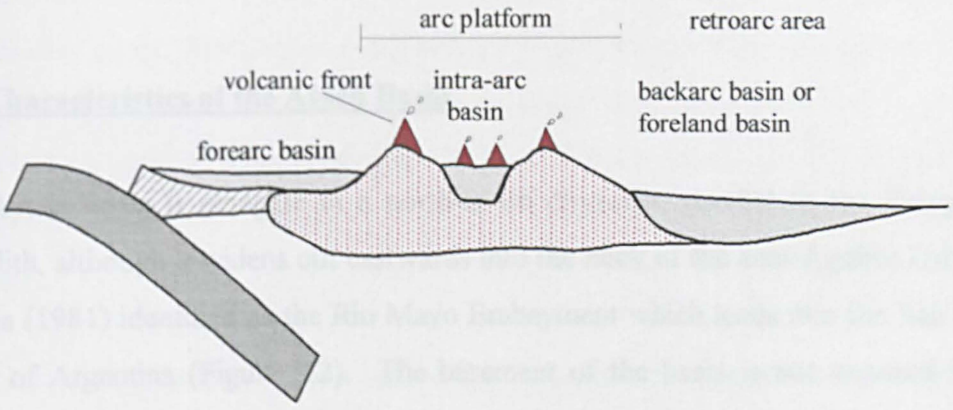


Figure 6.1: Diagramatic cross-section through a convergent plate margin showing general terminology. Intra-arc basins form within active and typically high standing arc platforms. Arc volcanoes are typically dispersed over a wide zone, but most active volcanoes are aligned along a distinct volcanic front on the trenchward side of the arc. In an extensional setting the retroarc area can develop into a backarc basin as oceanic crust is generated at the basin floor. In a compressional setting, or when structural thickening of the upper crust occurs, a wedge shaped foreland basin will form in the same geographical location (adapted from Smith & Landis 1995).

Difficulties in classifying convergent margin basins are well documented (Flint *et al.* 1993, Smith & Landis 1995) and highlight the ambiguities that can arise when defining forearc, intra-arc and backarc basins and their fill. Basins containing volcanogenic material may, more generally, be referred to as arc related basins, especially where spatial relationship to the arc platform is not clear (Smith & Landis, 1995).

6.2 Characteristics of the Aysén Basin

The Aysén Basin is elongate in a north-south direction, parallel to the Patagonian batholith, although it widens out eastwards into the neck of the area Aguirre Urreta & Ramos (1981) identified as the Rio Mayo Embayment which leads into the San Jorge Basin of Argentina (Figure 2.2). The basement of the basin is not exposed in the study area but outcrops near Lago General Carrera (Figure 1.3) and at about 50°S, show it to be metamorphosed Palaeozoic sediments (Niemeyer *et al.* 1984, Ramos 1989). The basin fill begins with the late Jurassic volcanics of the Ibañez Formation which are conformably overlain by marine sediments of the Coyhaique Group. The basin was infilled by the mid-Cretaceous and the sediments are capped by sub-aerial volcanics of the Divisadero Formation (Table 2.2). The sedimentary sequence of the Coyhaique Group shows little evidence of tectonic activity. The regionally extensive shales and sandstones are virtually horizontal and parallel bedded. There is no evidence of slumping or major faulting at any of the sites investigated in this study, neither is there any recognised interfingering of sedimentary units. The three formations of the Coyhaique Group in the study area are all marine. There is no evidence of dominantly terrestrial deposition in any of the sequences although abundant woody organic material was present throughout the Katterfeld Formation shales.

At the base of the Coyhaique Group, the Toqui Formation shows calcareous and sandstone beds intercalated with volcaniclastic material. The Toqui Formation has been dated as lower Berriasian in age by fossil evidence from the calcareous units (Covacevich, De La Cruz & Suárez, 1994). However, in the thick succession of black shales of the Katterfeld Formation, there is only very rare evidence of volcanic

activity. The only example found was at Río Simpson (Figure 1.4, AMT 4) where a thin, discontinuous horizon (less than 0.5 m thick) within the shale sequence, has been identified from thin section as an airfall volcanic tuff. A vertical swarm of basaltic andesite dykes which are intruded into the Katterfeld shales at Baguales Vista (Figure 4.9) have recently been dated at 61Ma (±0.4Ma) (Petford & Turner, 1996) and are therefore associated with the magmatic event at the end of the Cretaceous period. The contact of the Coyhaique Group sediments with the Patagonian batholith to the west is sharp and intrusive. There is no evidence of mylonitisation or shear zones which would indicate strike-slip removal of the sediments and no sediments of the Coyhaique Group have been reported either to the west of, or within, the Patagonian batholith.

The thickness of the sedimentary units is difficult to estimate as nowhere in the study area was there a complete measurable section of the Coyhaique Group from base to top. Either the base units were not exposed or the top unit had been eroded off. However, there was no obvious differences in formation thicknesses across the basin and the impression is that the sediments have a uniform thickness from west to east.

6.3 Intra-arc basins

Intra-arc basins are most common where a broad forearc region has developed over continental crust as in the Andes but can be found on other crustal types. Intra-arc basins are located within or including the arc platform and are associated with both intra-oceanic and continental-margin arcs. These elongate basins typically follow volcanic lines and are filled with mainly non-marine fluviatile and lacustrine sediments dominated by volcanic constituents (Allen & Allen, 1990). Intra-arc basin is therefore a descriptive term carrying no genetic connotation other than association with a volcanic arc and convergent plate margin (Smith & Landis, 1995).

Intra-arc basins can be volcano bounded or fault bounded or a combination of both. Some volcano bounded basins include coalesced aprons of volcanic and sedimentary debris that accumulate in the low areas among volcanoes on the arc platform. Significant thicknesses of basin fill seldom form on relatively high standing continental

margin arc platforms. Other volcano bounded basins are linear arc parallel troughs between volcanic ridges on the arc platform. Because these basins are between volcanic ridges it might be argued that they be termed forearc or backarc basins. These troughs are on the arc platform however, and are spatially separate from forearc and backarc basins. They are also not associated with rifting and formation of new ocean crust as would be the case in backarc basins (Dickinson 1974, Smith & Landis 1995).

The relationship between intra-arc extension and backarc basin development is not fully understood (Smith & Landis, 1995). Many backarc basins are initiated by extension within the arc and many intra-arc basins may have evolved into backarc basins if extensional stresses had been larger or deformation had continued longer. Extensional intra-arc basins may therefore represent incipient or arrested backarc basins (Smith & Landis, 1995).

6.4 Retroarc foreland basins

Foreland basins are defined as the sedimentary basins lying between the front of a mountain chain and the adjacent craton. Dickinson (1974) formally introduced the term and proposed two classes of foreland basin; i) peripheral foreland basins, which develop against the outer arc of the orogen and are associated with continent - continent collision, and ii) retroarc foreland basins which are situated behind the magmatic arc and are associated with subduction of oceanic lithosphere.

The majority of foreland basin development is related to thin-skinned thrust belts where sedimentary cover rocks are folded and thrust above undeformed basement rocks. Less commonly, upper crustal shortening and structural thickening triggers subsidence. Foreland basins are wedge shaped in cross section with the thickest margin proximal to the thrust belt (Jordan, 1995).

Retroarc foreland basins have a dominant formative mechanism of flexure and develop on continental lithosphere associated with subduction of oceanic lithosphere (Dickinson 1974, Bally & Snelson 1980, Allen & Allen 1990). They are large-scale,

long lived features which can be 100s of km wide and 1,000s km long and can contain kilometres of sediments (Jordan, 1995). The Andes of South America contains some of the best known examples of retroarc foreland basins.

Foreland basins typically contain strata deposited in non-marine to shallow-water depositional environments. Strata accumulate at rates ranging from 0.1 to 1m per 1,000 years (Jordan, 1995). Long term accumulation of sediments in a foreland setting requires tectonic subsidence. The horizontal shortening produces crustal thickening which triggers a series of responses. Isostatic adjustment to the transfer of mass produces a high topography within a thickened region and a down-flexed moat adjacent to this region (Allen *et al.* 1986, Jordan 1995). The mass is transferred by simultaneous erosion of the topographic highs and sedimentary infilling of the depression. Isostatic compensation is strictly true only for weak lithosphere. Typical continental lithosphere has lateral strength and isostatic disequilibrium is compensated for over a broad region rather than just below the thrust belt. Models of flexure basins are unrealistic because they are designed for homogeneous materials, whereas lithosphere and crust have variable compositions and inherited zones of weakness (Jordan, 1995). Therefore, true subsidence and uplift patterns in and adjacent to basins, differ from the models.

Submarine and shallow marine siliciclastic strata dominate retroarc foreland basins (Allen & Allen, 1990). Long term progradation is due to progressive migration of the fold and thrust belt. Progradation places shoreline deposits over marine strata, fluvial deposits over shoreline facies and coarse-grained over fine-grained fluvial facies. Eustatic sea level is the prime control on whether a retroarc basin is marine or non-marine (Jordan, 1995).

Sediment supply to a retroarc foreland basin is strongly asymmetric. Much more sediment is supplied by the thrust belt than by the low relief distal margin. Sediment may be supplied by a few point sources or the supply may approximate a line source. At a point source a sedimentary lobe with facies corresponding to a delta or a large alluvial fan accumulates. Between point sources linear coastlines may form (Jordan, 1995). The nature of the drainage system is reflected in the composition of the

sandstones in the foreland basin. The average foreland basin sandstone is rich in quartz, has very little feldspar and contains abundant lithic fragments, reflecting provenance in the thrust belt (Schwab, 1986).

Basement uplifts and thin-skinned thrust belts commonly form contemporaneously. In retroarc forelands this style of deformation is much less common than thin skinned thrusting. The typical result is a set of elongate and variably orientated, non-marine basins. Because volume of sediment source from the uplifted basement blocks is small, such basins are commonly underfilled (Jordan, 1995).

6.5 Backarc basins

Backarc basin supersedes the term inter-arc basin or marginal basin previously used by Karig (1971).

Backarc or marginal basins are semi-isolated basins or series of basins lying behind the volcanic chains of island-arc systems (Karig & Moore, 1975). Karig (1971) proposed that backarc basins initiate by crustal extension, producing first rifts and then new ocean crust by seafloor spreading. Backarc basin depositional environments on oceanic crust are deep marine except along their margins where fluviatile, coastal and shallow marine depositional environments may exist. Initially volcaniclastic wedges produced from the arc, interfinger with a background of pelagic clays. Backarc extension is less common on continental crust because the arc commonly becomes compressional (Allen & Allen, 1990).

Over 75% of the world's backarc basins are found in the circum-Pacific region, mostly on the western side of the Pacific Ocean (Marsaglia, 1995). Backarc basins are amongst the least understood of all the basins associated with subduction processes because modern counterparts are entirely submarine. It is generally accepted that backarc basins are the result of extensional and spreading processes associated with subduction of oceanic lithosphere at convergent margins. However, no one theory adequately explains the formation of all backarc basins. The

extensional origin of the basins is indicated by the presence of normal faults and the high heat flow characteristics which are commonly found (Marsaglia, 1995).

The first models proposed by Karig (1971) were divided into three categories; i) active ii) inactive with high heat flow iii) inactive with normal heat flow. Toksoz and Bird (1977), added a further category of, iv) undeveloped. This was originally included in Karig's "inactive with normal heat flow" category. Uyeda and Kanamori (1979) incorporated these previous models into a classification system for modern systems at subduction boundaries.

i) Active basins display characteristics of spreading centres with a topographic high parallel to the trend of the trench and island-arc and have a high heat flow. Shallow earthquakes are common (e.g. Mariana Basin).

ii) Mature basins are considered to be the same as Karig's "inactive with high heat flow" category. These basins show evidence of past spreading and have broad heat flow anomalies with elevated temperatures. Active basins may develop into mature basins as continued spreading creates more sea-floor and convection heats a broader region. After spreading ceases, thermal effects persist and give a broader heat anomaly (e.g. Sea of Japan and North Fiji Basin).

iii) The most common inactive basins have normal heat flow and they are generally old, representing the last stages of basin evolution. After convection and spreading stop, a mature basin loses its thermal energy with time and heat flow returns to normal (e.g. South Fiji Basin and West Philippine Basin).

iv) The final category of undeveloped basins was included in category iii by Karig (1971). They have typical basin morphologies but without the high heat flow or other evidence of spreading. Generally they occur behind young island-arcs where subduction has not progressed long enough (e.g. Aleutian and Caribbean subduction zones).

Uyeda and Kanamori (1979) found that subduction alone was not sufficient for back-arc opening although it was found to be necessary. Uyeda and Kanamori (1979) also discovered that changes in the downgoing slab due to age differences could cause spreading when older, denser crust was subducted, or closure of the basin when younger crust was subducted below an established basin (Dalziel, 1981). Earthquake studies also showed a significant difference in the mode of plate motions at the plate boundaries. Uyeda and Kanamori (1979) compared the Marianas arc and Chilean arc and found that the stress state in the backarc area differed, with compression occurring in the Chilean type and tension in the Mariana type. Further, they concluded that continental arcs by definition have no backarc basins and Uyeda and Kanamori included the Peruvian and Chilean arcs within this group, although some backarc areas such as in the Sumatra and Java arc have shallow water (<100m) cover over continental crust.

6.6 Discussion

Aguirre Urreta & Ramos (1981) describe the area around Lagos Fontana and La Plata and southwards to Coyhaique (i.e. Rio Mayo Embayment), as an intra-arc basin with a western volcanic arc and an eastern outer volcanic arc. The intra-arc basin was filled with turbidites, fossiliferous shales and deltaic deposits. Ramos & Palma (1983) recognised two stages of basin fill. In the east they found extensive fluvial systems and terrigenous deposits originating from the craton, whilst to the west they record intermittent calk-alkaline volcaniclastic sediments. To the south the volcanism disappeared while to the north the volcanics formed a unified arc. The basin fill that Aguirre Urreta & Ramos report in the Rio Mayo Embayment was not evident in the sediments of the Aysén Basin area of this study. There was no evidence of turbidites except at Cuesta Los Leones (Figure 1.4, AMT 19) where some very weak distal turbidites were recorded and no indication of the deltaic fluvial systems that are so prominent in the east of their study area. The usual basin fill of an intra-arc area is non-marine but all the formations in the Coyhaique Group are unambiguously marine. Furthermore, it would be expected that an intra-arc basin would display frequent horizons of volcanic deposits within the sediments but such layers do not appear within the shales of the Katterfeld Formation. It appears that subaerial volcanism

ceased at the end of the Berriasian when the pyroclastic units of the Toqui Formation were deposited and the basin underwent a period of volcanic quiescence until the mid-Cretaceous.

It is possible that the Aysén Basin could be considered an intra-arc basin that failed to develop into a backarc basin. Many, though not all, backarc basins originate by rifting of the arc platform followed by the generation of new oceanic crust (Karig, 1971). By Uyeda & Kanamori's classification scheme the Aysén Basin would appear in the fourth category of an undeveloped backarc basin, as extension ceased before oceanic crust formed. Fossil evidence indicates that the Aysén Basin was consistently shallow marine and never developed into a deep marine environment. No evidence of high influxes of volcaniclastic sediments from the volcanic arc, as would be expected in a backarc setting, are found in the Coyhaique Group. The lack of evidence of folding and thrusting in the Aysén Basin study area is also inconsistent with a backarc origin to the Coyhaique Group sediments.

The Rocas Verdes/Magallanes Basin of southern South America (Figure 2.2) however, can be considered to be a true backarc basin. Dalziel (1974) has found it to be analogous to the present Japan Sea area which is classified as category two, a mature basin. In the Late Jurassic to Early Cretaceous, basaltic magmas were intruded into the continental crust behind a volcanic complex which consequently moved westward, creating a backarc basin between a volcanic island arc and the South American continent (Dalziel 1974, Biddle *et al* 1986, Wilson 1991). The backarc basin formed during a single event of rapid extension. The floor of the basin consists of large lenses of mafic rocks, representing the upper part of an ophiolite assemblage, and an intrusive gabbro unit which is cut by sheeted dykes. This sequence is overlain by a thick unit of pillow lavas. These assemblages clearly indicate an ocean floor origin. The strips of ocean floor are separated by elongate blocks of older continental crust and therefore the basin must have formed by extension and intrusion.

In the Central Andes Basins, which includes the Neuquén Basin, subduction occurred but areas of ophiolitic rocks are unknown. The basins contain flysch-type sediments

with andesitic detritus and mafic extrusives which indicates a more limited extensional setting than in the Magallanes Basin. The most noticeable difference however is that the deformation in these basins was much less intense, consisting of block faulting and gentle warping.

In Uyeda and Kanamori's (1979) classification of present day subduction boundaries, continental arcs, by definition have no backarc basins. Suárez (1984) disagrees and suggests that this might be correct for modern arc systems but ancient systems such as in the Andes did produce backarc systems as shown in the Magallanes Basin, where oceanic rocks are evident. Suárez (1984) proposed that this model should be classified as the Patagonian type. Suárez also interprets the basins in the coastal cordillera of Peru as an elongate backarc basin, where the continental crust thinned but did not rupture and high density material can be found at high levels within the continental crust immediately beneath the basin. He termed this type of basin Early Andean. The Aysén Basin would be included in Suárez's Early Andean model as it formed on thinned continental crust which did not rupture to allow an oceanic basin to form.

A third possibility of origin for the Aysén Basin is as a retroarc foreland basin. Retroarc basins form as a result of flexure and therefore the sediments do not experience the high deformation that are present in backarc basins and this would fit in with what is seen in the Aysén Basin sediments. Foreland basin deposits are wedge shaped in cross-section with the thickest margin nearest the thickened crust. The sediments in the Aysén Basin do not show any distinct regional variations in thickness from west to east across the basin. However, further east in Argentina, Ramos (1976) has shown that the sediments are thicker than any that are exposed in Chile, which is contrary to what would be expected in a foreland basin setting. This could be explained by the argument that continental lithosphere has a lateral strength and the loading has been spread over a broad region. It is another possibility that the intrusion of the Patagonian batholith has disrupted the sediments near the proximal edge of the foreland basin although there is no evidence that these have been removed by strike/slip movements.

6.7 Summary

The Aysén Basin sediments were apparently laid down on a virtually flat surface which subsided gently and where basin margins never became steep enough to promote slumping or mass flow deposits. The study area sediments are entirely shallow marine and there is no sign of any tectonic events or deepening of the basin. The few intruded dykes have been shown to have occurred after closure of the basin and so there is no evidence for the widespread extension that was present in the basins further south. The absence of volcanic horizons also suggests that volcanism had ceased at the time the Coyhaique Group sediments were being laid down in the Aysén Basin.

The Aysén Basin does not fit neatly into any recognised categories of basin formation. The evidence points to a gently subsiding basin which formed due to limited extension where the continental crust thinned but did not rupture. The Aysén Basin therefore appears to fall between an undeveloped backarc basin and a retroarc foreland basin, having some of the attributes of both.

CHAPTER 7

LOWER CRETACEOUS PALAEOGEOGRAPHY OF SOUTHERN SOUTH AMERICA

7.1 Tectonics and structural development of the study basins

Evolution of the basins along the western coast of southern South America was influenced by events that occurred while the area was part of Gondwanaland. Subduction along the edge of the supercontinent built up a broad, accretionary margin system that formed the basement of Patagonia. Deformation associated with the subduction (Table 2.1) imposed a structural grain on the basement rocks and regional extension in the Late Triassic led to the formation of a number of NNW trending rift basins which followed the basement grain. During the early stages of extension the fault-bounded basins were small and isolated but during the Late Jurassic the areas affected by normal faulting enlarged and became interconnected, creating a composite backarc system along the Pacific margin. After the initial shared history, development of the basins varied in the amount of extension, subsidence and consequent basin fill that each received. Continued extension during the Early Cretaceous ultimately led to the opening of the South Atlantic Ocean when faulting and volcanism declined in Patagonia (Uliana & Biddle, 1987). A regional compressional event in the mid-Cretaceous resulted in simultaneous elimination of the basins along the length of the South American continental margin which Dalziel (1986) suggests was caused by a global increase in sea-floor spreading rates which significantly increased the convergence rate along the Pacific margin of South America. However, the intensity of deformation in the backarc areas varied significantly along the length of the Cordillera.

In the south, the site of the Rocas Verdes/Magallanes Basins underwent considerable extension which developed into a marginal basin with an oceanic floor. During the Early Cretaceous it received volcaniclastic turbidites in the far south and a fluvio-deltaic and marine sandstone (Springhill Formation) in the Magallanes Basin. The sandstones were overlain by a thick succession of black shales of the Río Mayer Formation, before mid-Cretaceous compression caused the closure of the basin and created the Magallanes fold and thrust belt. Deformation was extremely intense in places and the sedimentary infill was tightly folded and thrust continentward.

In the Neuquén Basin, extension resulted in a series of normal faulted, half-grabens but never led to the development of oceanic crust. Subsidence was related to a thermal sag and a marine embayment formed. Deposition in the backarc area was essentially devoid of volcanics and produced a thick sequence of mainly marine clastic sedimentary rocks alternating with carbonate and evaporite units. The structural effects of the mid-Cretaceous compressional event range from uplift with no horizontal shortening to gentle or moderate folding, but are much less severe than in the far south of South America.

The San Jorge Basin, to the east of the Aysén Basin also formed due to extension during the Late Triassic. Northwest to southeast trending grabens formed due to normal faulting but became infilled with dominantly continental sediments. As normal faulting ceased in the Early Cretaceous, the rift system evolved into a broad interior sag related to thermal subsidence. The basin became infilled during the mid to Upper Cretaceous probably due to uplift of the source area.

To the east of the Aysén Basin, in Argentina, Ramos & Aguirre-Urreta (1994) describe the Rio Mayo Embayment as an intra-arc basin and the basin fill, the Tres Lagunas Formation, as a turbiditic series containing andesitic volcanics and breccias which was succeeded by a deltaic system that pro-graded the basin from the east. They proposed that in the area of Lagos Fontana and La Plata the intra-arc basin occurred between an external volcanic arc and a volcanic interior arc. Further north the intra-arc basin disappeared and volcanism expanded to create a unified arc, whilst to the south the external volcanism disappeared completely.

In the Aysén Basin of Chile there is no evidence of faulting or graben formation. Extension was limited to thinning of the crust and, like the Neuquén Basin, no rupture of the ocean floor to produce mafic rocks. The sediments all reflect a gently subsiding basin with very rare indications of distal turbidites or evidence of steep margins. The likely tectonic setting was of a volcanic arc in the west, with the Patagonian batholith forming in response to easterly subduction along the western margin of Gondwanaland. To the east of the volcanic arc was a shallow, elongate depression caused by limited extensional tectonics which created a marine basin. The Aysén Basin was eliminated in mid-Cretaceous times by vertical uplift rather than horizontal closure as in the Magallanes Basin. The elimination of the basin involved no development of penetrative fabrics and in places the overlying volcanics of the Divisadero Formation are conformable.

7.2 Palaeoecology of the Aysén Basin

Palaeoecology examines the relationship between fossils and the associated sediments and the relationship between the various taxa of a fossil community. An integrated approach using palaeontology and sedimentology can recognise community stasis and changes in the palaeoenvironmental habitat. Most conventional palaeoecological interpretations come from data on body fossils and sedimentary analysis, but ichnology can provide an important understanding of biological processes and their relationship with the sediments they are found in. Ichnofossils are unique in that they not only represent the morphology and life habit of the trace making organism but also the physical characteristics of the substrate. They are closely linked to the environmental conditions prevailing at the time of construction and are therefore a useful tool in palaeoecological analysis.

7.2.1 Palaeoecology of the Toqui Formation

The Toqui Formation occurs at the base of the Coyhaique Group. It is a laterally and vertically variable succession which can be subdivided into calcareous, arenaceous and pyroclastic lithologies. The strata represent a transgressive succession which in places has been found to be intercalated with volcanic rocks of the Ibañez Formation. The Toqui Formation shows a distinct variation in its lithological expression from NE to SW across the study area. The extensive oyster banks are only found in the north and NE of the region whilst along the eastern margins of the study area the Toqui Formation is represented by fossiliferous, calcareous sandstones. South of Coyhaique the Toqui Formation appears as a bedded limestone, interbedded with sandstones and volcaniclastic strata.

7.2.1a Palaeoecology of the oyster banks

The overall picture of the bioherm environment is one of persistent moderate to high energy, in warm seas that could support a suspension feeding community. Influxes of fresh water from nearby land areas brought abundant nutrients in the form of organic

material but also caused fluctuations in salinity levels which restricted the faunal community that was able to colonise the bioherm.

At Estero Pedregoso a volcanic conglomerate provides the stable substrate that all flat lying oysters require as an anchor point (Figure 7.1). A mudstone channel deposit which cuts through the conglomerate contains articulated *Nucula* sp. bivalves which were buried rapidly. The few *Pinna* sp. that were found confirm that the sediment had some sand content to support their semi-infaunal mode of life. Like *Nucula*, they prefer a sheltered, stable environment. The sandy mudstone also provided some species of ostracods which confirmed a shallow environment. The undulating nature of the deposit, abundant organic material, and unidentifiable vertebrate teeth, indicate that this was a channel which cut through the coarser conglomeratic bed below (Figure 7.1).

All oysters explicitly like warm, shallow water, no deeper than 80 metres. Flat lying oysters such as *Exogyra* require a firm and stable substratum to anchor onto. If the floor is too soft or muddy, oyster larvae cannot find suitable attachment areas. Oysters are therefore confined to a specific environment where waves and currents are too weak to shift sand and too strong to allow mud to settle (Stenzel, 1971). The thick-shelled epifaunal bivalves have an extremely high tolerance to salinity fluctuations which may have occurred near the edge of the basin as fresh water rivers diluted the normal marine conditions. Once the oyster community had colonised the sediments, succeeding generations used their predecessors as anchor points enabling thick sequences of bioherms to be built up. The oyster shells were often encrusted by serpulids and algae which helped to bind the oyster mounds together. The crevices between shells were filled with the debris from the oysters and predators feeding on the shells and provided a new habitat for boring organisms such as *Lithophaga* bivalves (Figure 7.1).

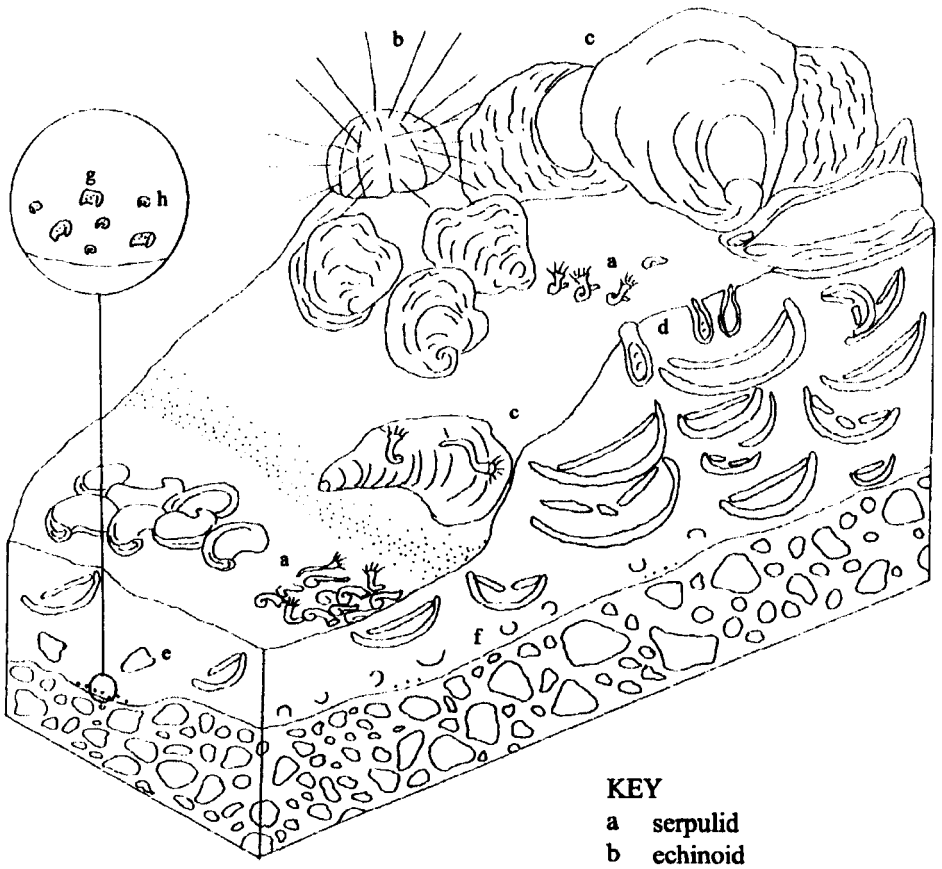


Figure 7.1: Diagrammatic representation of the faunal community of the oyster bioherm at Estero Pedregoso during the Toqui Formation period.

The oyster bioherms found in the north of the basin area marked the margin of a shallow basin which ran NW/SE. Probable salinity fluctuations prevented other organisms colonising the build-up and resulted in low species diversity and large individuals as the oysters exploited their niche in the environment. Each generation of oysters used the previous one as a fixation point to establish themselves and eventually the mature oysters could not obtain enough nutrients and became smothered. Regular cycles occurred which stopped development but as conditions improved the oyster community again became established. The bioherm eventually died probably due to inundation by sediments from a terrestrial source as currents waned.

7.2.1b Palaeoenvironment in the southwest of the basin

At Foitzick (AMT 9), the base of the Toqui Formation overlies a volcanic breccia of the Ibañez Formation. The lower bed is a poorly sorted conglomerate which contains large, angular, rhyolitic boulders which have not been transported far from their source. They were deposited on a sub-aqueous volcanic apron which received debris from the steep flanks of an intermittently active volcano (Figure 7.2). The lack of any finer-grained matrix suggests fairly strong currents which would have removed the smaller particles. Tectonic activity and deposition of the angular boulders ceased although strong currents continued to transport and deposit material along the volcanic shoreline. The boulder bed acted as a base for a reefal system and organisms such as corals, coralline-algae, echinoids and bivalves including oysters colonised the topography (Figure 7.2). The fauna present indicates a shallow sea of normal salinity and a high energy environment. Regular influxes of volcanic sands transported by the coastal currents, smothered the organisms and prevented the reef from growing, but as conditions stabilised the volcanic debris once again became colonised. Volcanic activity became less regular and enabled carbonate sediments to accumulate in an open platform environment when the bedded limestones were laid down. Periodic storms washed in terrigenous material and sporadic volcanic events which marked the end of the previously widespread igneous activity, produced air-fall tuffs. The fining-up sequences of well-sorted shell beds at the top of the section reflect a high energy environment with strong E/W currents which winnowed the shell fragments.

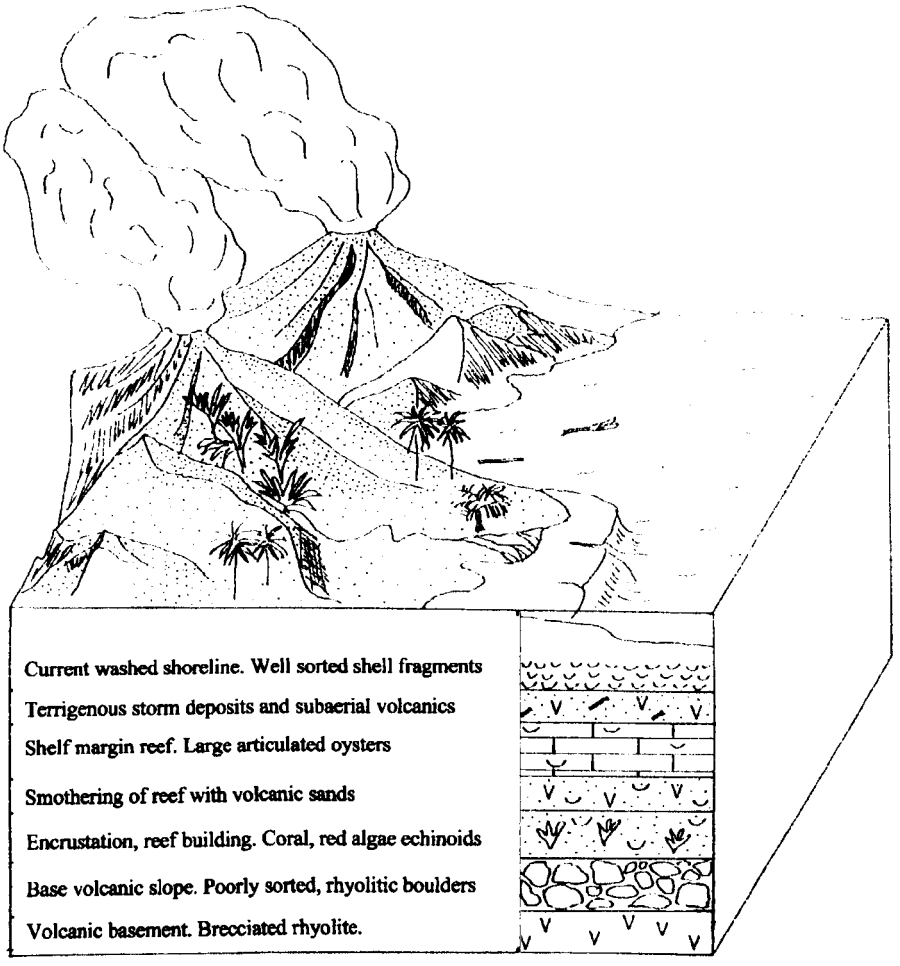


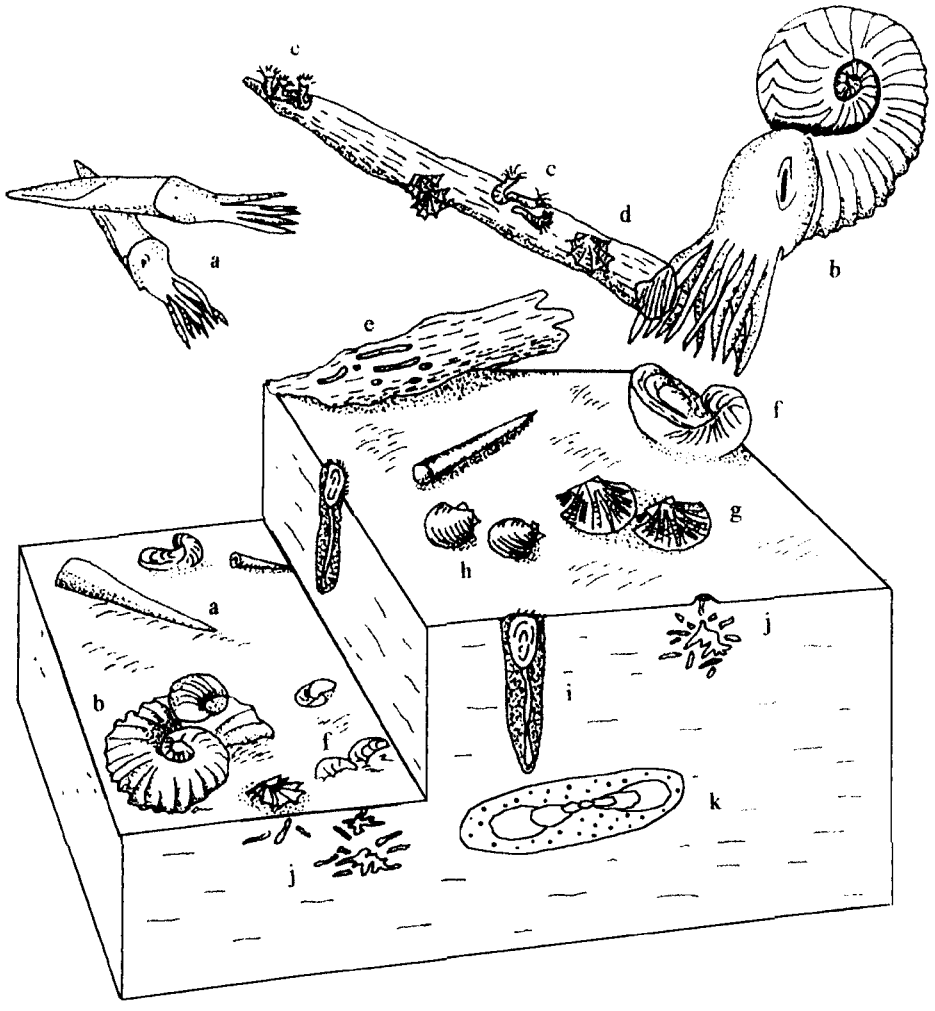
Figure 7.2: Representation of the depositional environment in the south-west of the basin when the Toqui Formation bedded limestones, sandstones and volcanics were being deposited.

Palaeoecological investigations show that a marine transgression covered the Ibañez Formation volcanics during the Berriasian (dated by ammonites). A shallow, warm sea which experienced strong currents and a high energy environment covered most of the study area. At the margins of the basin, salinity fluctuations encouraged the growth of large oyster bioherms, calcareous sands were washed in by distal turbidites and occasional storm events ripped up the nearby oyster banks and redeposited them as storm beds. Deeper in the basin reefs formed along the shorelines of active volcanoes which intermittently produced sub-aerial deposits.

7.2.2 Palaeoecology of the Katterfeld Formation

The calcareous black shales and mudstones of the Katterfeld Formation are regionally extensive, laterally continuous and usually horizontal and parallel bedded, suggesting prolonged periods of deposition on a relatively flat basin floor. Isolated instances of fine-grained turbidites indicate a lack of steep gradients. The shales indicate high organic productivity in the surface waters and also contain large pieces of carbonised wood, indicating the presence of nearby land areas. The black shales are initially homogeneous looking, black to dark brown in colour and generally fissile although at some locations they are more massive and blocky in appearance. The mudstones are usually black but sometimes slightly grey or brown. They occur in thin units interbedded with the shales, sometimes showing laminations but mostly structureless.

Conditions at the sediment/water interface were generally oxygen deficient and unable to support a benthic fauna. The general assemblage is dominated by nekton and plankton but also includes epifaunal bivalves such as *Oxytoma, Chlamys* and *Entolium* that were attached to floating vegetation and wood that floated above the anaerobic sea-floor (Figure 7.3). The presence of gryph-type oysters and the brachiopod *Lingula* indicate that the substrate was likely to be soft although not always completely anaerobic. The absence of trace fossils except for *Chondrites* in much of the black shale succession shows that oxygenation was however, usually very low (Figure 7.3). The bulk of the shales show a restricted fauna of mainly epifaunal bivalves and pelagic organisms which accumulated in the soft sediments as they sank after death.



KEY

a *Belemnopsis extenuatus*
b *Favrella*
c *Annelida*
d *Oxytoma*
e *Teredo*
f *Aetostreon?*
g *Chlamys*
h *Entolium*
i *Lingula*
j *Chondrites*
k Calcareous concretion

Figure 7.3: Diagramatic representation of the Katterfeld Formation palaeocommunity during deposition of the organic-rich mud rocks

Fossiliferous horizons did occur within the black shales, particularly near the base of the sequence. The fauna of these beds were mostly infaunal suspension feeders. They show that the basin had become shallower with a consequent increase in oxygen levels and the introduction of weak currents which replaced the stagnant conditions of the oxygen deficient environment.

The generally anoxic conditions of the basin suggests deposition in a restricted basin which kept pace with slow thermal subsidence. The shallowing events near the base of the succession indicate that the subsidence was more erratic at first but became stabilised at a steady rate for the major part of deposition of the sequence. Volcanic activity had almost ceased, and apart from a few thin, tuffaceous intercalations there is no evidence for the volcanic chain recorded by Ramos & Palma (1983). The extensive area of deposition and the thick accumulation of shales indicates the Katterfeld Formation was laid down on a relatively flat, epicontinental sea-floor with a restricted opening to the sea, probably towards the south.

7.2.3 Palaeoecology of the Apeleg Formation

The Apeleg Formation conformably overlies the black shales of the Katterfeld Formation with an abrupt vertical transition over a few metres. The two formations do not interfinger at any of the sites visited in the present study. The transition is abrupt and conformable indicating that the change from the low energy environment of the black shales of the Katterfeld Formation to the higher energy environment of the Apeleg Formation, was regionally synchronous and not due to any gradual, transgressive event. The well-sorted, rippled volcaniclastic sandstones and shales were deposited as offshore tidal sandbars and ridges in a marine shelf environment. These sediments suggest alternating tidal currents which deposited the uniform, fine to medium-grained sands, and low energy periods when muds were allowed to settle out from suspension. The thick, uniform successions of sandstones and mudstones suggest that energy levels were constant over fairly long periods of time. There was very little evidence of slumping and most deposition was below wave base. The ammonite *Favrella wilckensi* was found in abundance at the base of the succession and has been used to date the lower beds as Upper Hauterivian.

The diverse trace fossil assemblage found throughout the Apeleg sandstones reflect a wide variety of behavioural categories including dwelling, resting, feeding and locomotion and were produced by a variety of organisms which left no body fossil evidence in the sediments. Traces such as *Asteriacites, Gyrochorte, Thalassinoides, Planolites and Rhizocorallium* (Figure 7.4) are typically found in a low-energy, off-shore environment above storm wave base. The assemblage is characterised by a mixed association of vertical, inclined and horizontal feeding and locomotory structures mostly constructed by mobile deposit feeders. The *Diplocraterion* and *Ophiomorpha* ichnofossils (Figure 7.4) are more typical of a shallow, higher energy facies.

Palaeoecological investigations show that the Apeleg Formation sediments were transported by tidal currents and deposited below wave base on an extensive and broad epicontinental shelf. The ichnocoenoses indicate that the fauna was controlled by water depth and water oxygenation. The rippled sandstones of the inner shelf deposits reflect a high diversity and a wide variety of feeding and behavioural categories and indicates a well-aerated, overall low-energy environment with high concentrations of nutrients. The shoreface and sandwaves on the inner shelf experienced a well-aerated, high-energy and shallow environment.

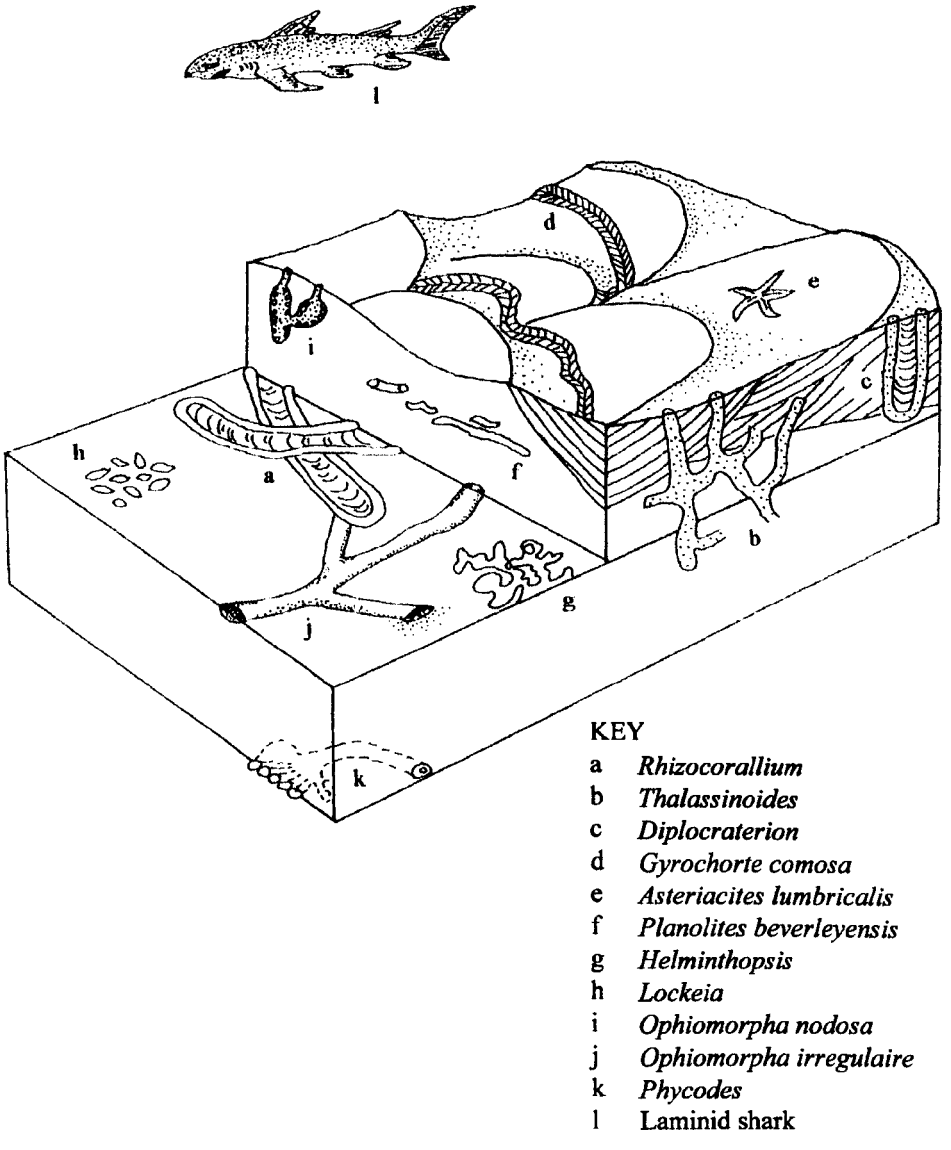


Figure 7.4: Schematic trace fossil assemblage found during Apeleg Formation period of deposition

7.3 Palaeogeography of the Aysén Basin

7.3.1 Links to other areas

Previous studies have suggested that the Aysén Basin was connected to other basins. Riccardi (1988) considered the Aysén region to be a northerly extension to the Magallanes Basin whilst Ramos (1989) identified an intra-arc basin at the same latitude, which he named the Rio Mayo Embayment. The palaeoecological study of this work has shown that whilst there are undisputed links to the adjacent basins, the Aysén Basin was a separate basin and probably only connected into other basins for short periods at times of sea-level highstands. The general transgressive pattern of the southern basins is in agreement with a global rise in sea-level during the Tithonian and which reached a peak in the Hauterivian (Riccardi, 1988).

7.3.1a Magallanes Basin

The most convincing evidence of connection to the southern basins are the similarity of the faunal assemblages. The Springhill Formation of the Magallanes Basin contains the same bivalves, ammonites and belemnites found in the Katterfeld shales of the Aysén Basin. The Springhill Formation is a silty, marine sandstone overlain by the Rio Mayer Formation which is a thick sequence of black shales. The Rio Mayer Formation contains the ammonite *Favrella* which was common in the Katterfeld Formation of the Aysén Basin. The Springhill Formation fossils have been dated as Berriasian from ammonites. It is most likely that the Aysén Basin was too shallow in Berriasian times to be connected to the Magallanes Basin but as gradual subsidence occurred, a link developed to the south and before Hauterivian times the fauna of the southern basins was able to colonise the deeper water environment of the Aysén Basin.

7.3.1b San Jorge Basin

The San Jorge Basin to the east of the Aysén Basin contains continental deposits. Inconclusive palaeocurrent data suggests it may be the source of most of the sedimentation of the study area. Fitzgerald *et al.* (1990) described several horizons of marine biota within the lacustrine shales at the western margin of the San Jorge Basin and interprets this as inundation of coastal lakes from a connection to the west during the Hauterivian, a time of sea-level highstand.

7.3.1c Neuquén Basin

There is no obvious link to the Neuquén Basin to the north of the Aysén Basin. The ammonites described by Leanza & Leanza (1973) from the Neuquén area are not the same as those found in the southern basins. Aguirre-Urreta (1993) found that during the Berriasian the Neuquén ammonites shared several genera with those from the Magallanes area but during the Valanginian to Hauterivian there was a sharp change and the southern genera like *Favrella,* become strongly endemic and disconnected with those to the north. *Favrella* has been recorded from as far south as the Antarctic Peninsula (Thomson & Farquharson, 1984). The Neuquén ammonites become more related to genera from further north and represented by faunas with Mediterranean affinities. Aguirre-Urreta (1993) found that by the Barremian, the Magallanes Basin ammonites show strong similarities with coeval faunas of South Africa. Although the Andean faunas of southern South America are quite well known, the centres of origin and migration patterns are still unclear Aguirre-Urreta (1993).

7.3.1d South Africa and Antarctica

Studies in the Aysén Basin have not discovered any ammonite fauna from Barremian times but there are other indications of a possible link with South African fossil faunas. Belemnites and bivalves which include trigonias and oysters have been shown to have a close connection with fauna from South Africa and Madagascar. Species of trigonia found in the present study in the Aysén Basin include *Megatrigonia rogersi* and *Steinmanella herzogi.* Reyes (1970) found these and other species of trigonia at

the Cuesta Los Leones location and he was able to show that they were identical to South African species. He found that *Megatrigonia* was abundant in the mid-Cretaceous sediments of the east and south of South Africa although it was more commonly found in Tithonian deposits in India. *Steinmanella* was also well represented in the marine strata of South Africa and Reyes (1970) assigned an age range of late-Valanginian to early-Hauterivian to the Chilean trigonia assemblage. He also pointed out that trigonias in northern Chile and Peru have a different form to those from southern Chile and were not associated with South African species. Antarctic trigonids as well as other fauna found in Antarctica are very different to those found in southern South America even though both areas experienced similar arc-related processes and a closer faunal similarity would be expected (Crame, 1984).

The link between southern South American and South African faunas is further enhanced by the common occurrence of gryphaeid oysters which are not present in Antarctic fauna. The faunal difference between the areas is thought to be a response to some form of climatic gradient (Crame, 1984). Riccardi (1977) also commented on the close similarity of the gryphaeid oyster *Aetostreon,* to Lower Cretaceous forms from South Africa and to *Gryphaea balli* which was recorded previously from the Coyhaique area, and proved to be connected to South Africa and Madagascar. *Aetostreon* was found in the Springhill Formation of Argentina and is common in the Katterfeld Formation of the Aysén Basin.

The belemnites found in the Katterfeld Formation shales closely resemble New Zealand forms described by Stevens (1965). However, Challinor (1991) observed that the narrow groove which was found in some of the Aysén Basin belemnites, is consistent with *Belemnopsis* forms of the Tithonian to Berriasian of the Malagasy Republic and eastern South Africa. Riccardi (1977) recorded *Hibolithes* from southern Patagonia. Urreta and Suárez (1985) identified *B. madagascariensis* from Tierra del Fuego and Ferruglio (1936) noted *Hibolithes* from the Argentine Cordillera. These species of belemnite are all consistent with New Zealand fauna. The evidence of belemnite fauna indicates strong links between southern South America and New Zealand during the late Jurassic but in the Early Cretaceous, genera typical of Tethys (east Africa and Malagasy Republic) appear (Challinor, 1991).

Belemnopsis is unknown from most of the Indo-Pacific region in the Early Cretaceous but some species derived from *B. aucklandica* are found in the Antarctic Peninsula from Valanginian to Hauterivian (Doyle & Howlett, 1989).

It is clear from the faunal evidence that an Austral realm developed in the Late Jurassic and migration of marine organisms continued into the Cretaceous. By the mid-Cretaceous the marine connections were enhanced by rifting and production of oceanic crust but Antarctica and Australasia moved southwards and into a cooler climate zone (Doyle & Howlett, 1989). Shelf seas separated Antarctica from the main oceanic areas of the east Pacific border and by the late Valanginian to early Hauterivian, endemic species groups developed in southern South America and Antarctica and Australasia.

7.3.2 Palaeogeographic development of the Aysén region

Easterly subduction in the Aysén region during the Middle to Upper Jurassic resulted in the Ibañez Formation calc-alkaline volcanics and plutonics. Voluminous rhyolitic, subaerial volcanism occurred in the backarc area and formed the basement to the Aysén Basin. Towards the end of the Jurassic the western arc separated from the South American continent, the backarc area became marine and the Coyhaique Group sediments were laid down as the basin subsided.

Figure 7.5 shows the evolution of the Aysén Basin of the study area from the Late Jurassic until the end of Coyhaique Group sedimentation in the Hauterivian to Barremian. The Toqui Formation was variable and discontinuous across the basin from SW to NE. Intermittent volcanic activity produced andesitic tuffs near the margins of the basin. During the Berriasian the calcareous beds of the Toqui Formation covered most of the study area and the volcanic tuffs indicate active volcanoes in the west and northeast of the basin (Figure 7.6). The northeast margin of the basin is defined by the thick oyster bioherms found near Ñirehuao (Figure 7.5). To the northeast in Argentina, the Tres Lagunas Formation sands and sporadic limestones were being laid down in a more volcanically active setting (Figure 7.6).

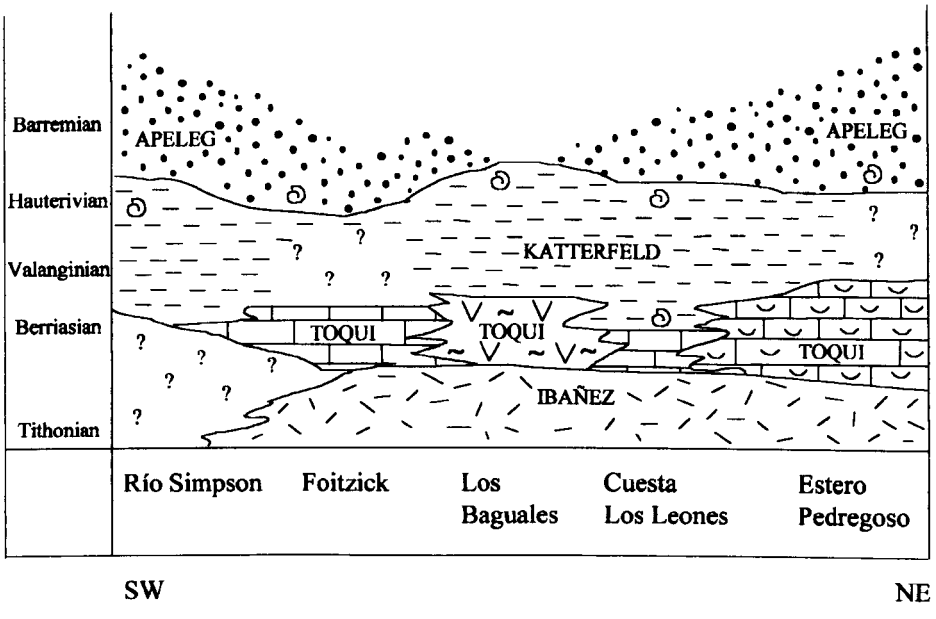


Figure 7.5: Evolution diagram across the study area of the Aysén Basin from SW to NE. Stratigraphy constrained by datable ammonite fauna.

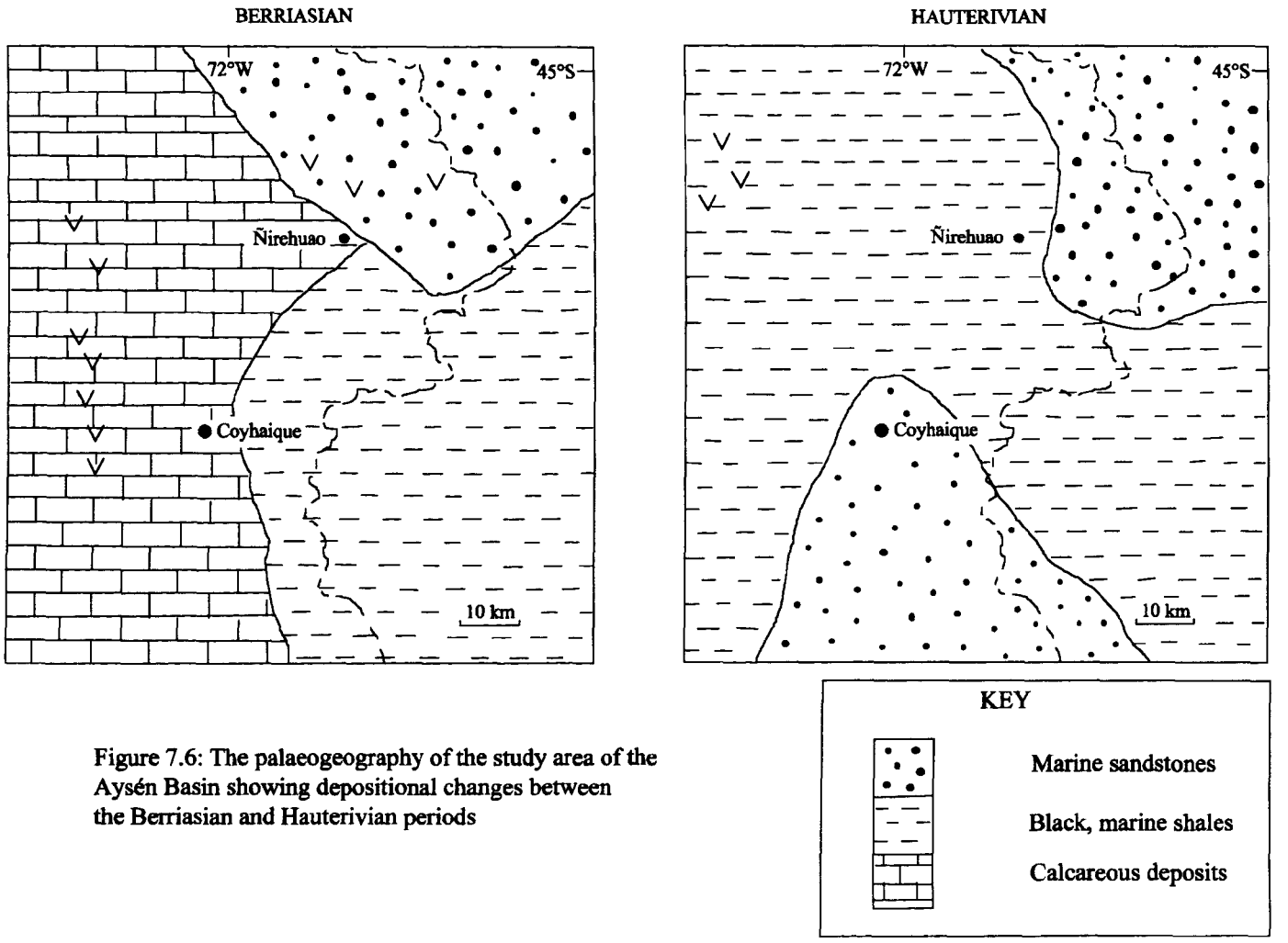


Figure 7.6: The palaeogeography of the study area of the Aysén Basin showing depositional changes between the Berriasian and Hauterivian periods

As the basin subsided in the Berriasian the irregular topography of the Aysén Basin allowed black shales to form in the deeper restricted areas, dated by Berriasian fossils (Figure 7.5). Gradually the entire basin area became covered by a shallow stagnant sea and the uniformly thick succession of black shales were deposited. By the onset of the Hauterivian the black shales of the restricted basin had become regionally extensive. No interfingering of sediments are found (Figure 7.5) and this reflects the gradual subsidence of the basin.

The sharp transition from restricted black shales to high energy, rippled sandstones during the late Hauterivian (Figure 7.5) marked the opening of the basin from the south and east, to normal marine conditions. The change was synchronous and regionally extensive across the whole of the basin area (Figure 7.6).

7.4 Palaeogeography of southern South America

The initial history of all the southern basins of South America is related to the breakup of Gondwanaland. Triassic to Late Jurassic intracontinental rifting, spreading and subsidence (Table 2.1) culminated in the opening of the South Atlantic in the mid-Cretaceous. The crustal extension resulted in a series of northwest to southeast trending, fault-bounded grabens and troughs. Along the Pacific margin of South America, subduction produced a magmatic arc along the continental margin and Late Jurassic to Early Cretaceous marine basins which received sediments from the volcanic arc.

The Aysén Basin was segregated from the Neuquén Basin to the north by the Somuncurá Massif (Figure 7.7). It most probably received sediments from rivers flowing west from the San Jorge Basin. During the Late Berriasian a southerly connection to the Magallanes Basin formed but this was restricted and as slow subduction continued in the Aysén Basin, a vast, shallow, mainly dysoxic sea covered the area. In the Hauterivian sea-levels rose to their highest peak and a short-lived connection formed between the Aysén Basin and the San Jorge Basin.

During the Late Jurassic, the site of the Rocas Verdes/Magallanes Basins underwent considerable extension which developed into a marginal basin with an oceanic floor (Figure 7.7). Late Jurassic subsidence was followed by a marine transgression which covered the southwest margins of the Deseado Massif. During the Berriasian the marginal basin received volcaniclastic turbidites and a fluvio-deltaic and marine sandstone (Springhill Formation) in the Magallanes Basin (Figure 7.7). Volcanic activity developed in the southern volcanic-arc and rhyolitic flows, and pyroclastic rocks accumulated along the rear flank of the arc (Riccardi, 1988) whilst in the Aysén Basin there was a period of volcanic quiescence from the end of the Berriasian into the Barremian and Aptian.

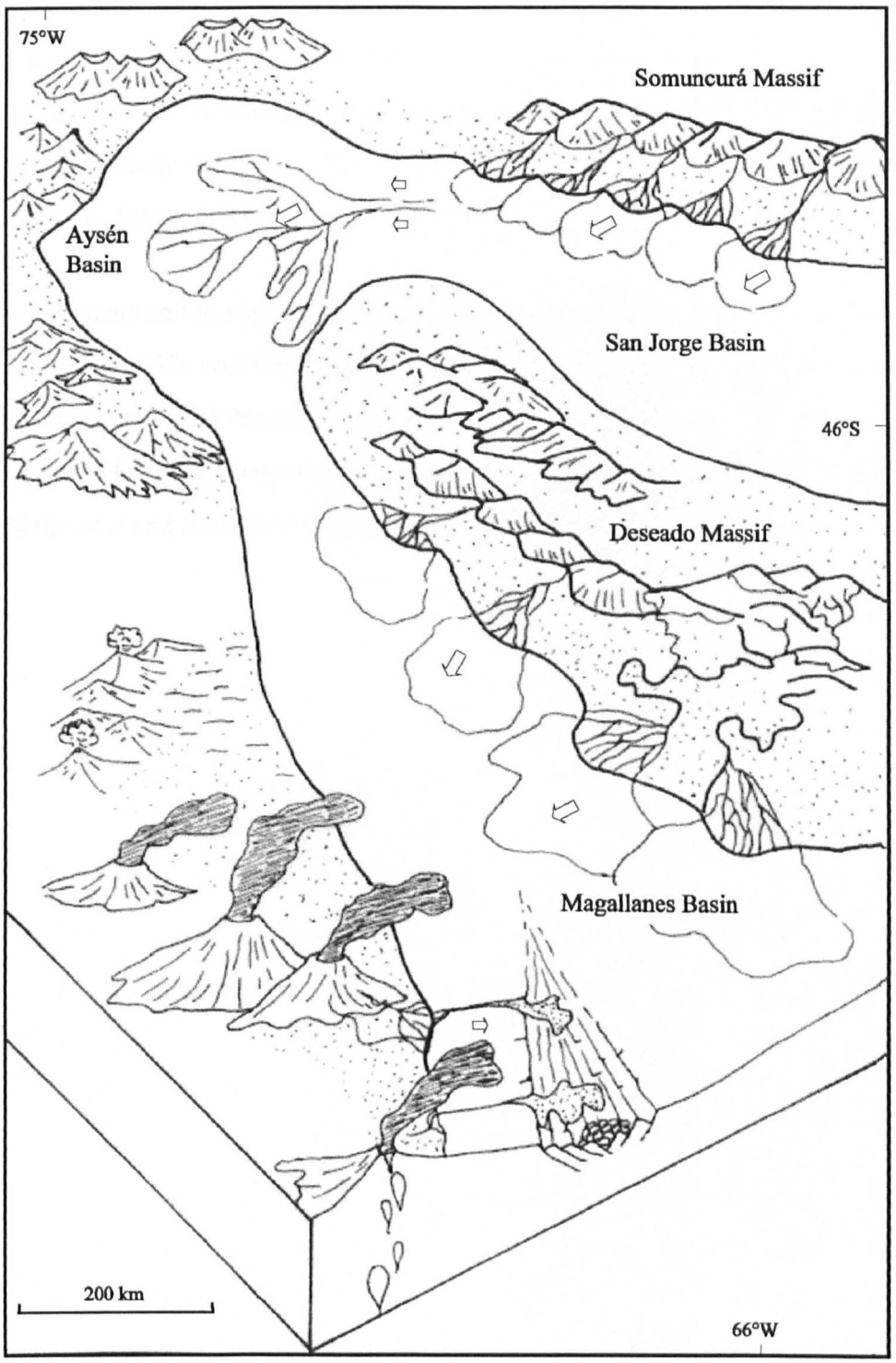


Figure 7.7: A general interpretation of the palaeogeography of southern South America during the Early Cretaceous

During the Hauterivian to Barremian the restricted area of the Aysén Basin abruptly opened up to normal oceanic tides and currents with deposition in a shallow, near-shore environment. A similarly sharp transition between the stratigraphically and palaeogeographically equivalent Rio Mayer and Kachaike Formations (Table 2.2) of the Magallanes Basin suggest this was a regionally significant event.

Marine sedimentation in the Aysén Basin ended with uplift and erosion in Barremian to Aptian times. This was followed by extensive Aptian-Albian calc-alkaline volcanic activity which produced the unconformably overlying Divisadero Formation. In the south, mid-Cretaceous compression thrust the sediments of the Magallanes Basin towards the east and resulted in the closure of the basin.

CHAPTER 8

CONCLUSION

This dissertation has used palaeoecological and palaeoenvironmental information to produce an interpretation of the Aysén Basin during the Lower Cretaceous when the sediments of the Coyhaique Group were deposited. The systematic sampling and logging of the fieldwork has provided the foundation for the palaeogeographic interpretation of the Aysén Basin area and a more general representation of southern South America.

It is well established that the basins along the west coast of South America resulted from subduction of an oceanic plate below the continental lithosphere of the South American plate during the Triassic to Jurassic. Extensional rifting produced a series of linear basins separated by pre-Jurassic basement ridges trending NW-SE. The Aysén Basin formed at this time and received marine sediments and volcaniclastic material from the rising volcanic arc. Previous reconstructions of the Aysén region have used extrapolations from the surrounding basins to deduce palaeoenvironmental conditions during the Lower Cretaceous, when the sediments were deposited. The Rio Mayo Embayment was presented as a distinct basin isolated from both the adjacent San Jorge and Magallanes Basins whilst other work showed the Aysén Basin as a link between the Magallanes and San Jorge Basins.

Although other workers have looked at the volcanic sequences in the Aysén Basin this study is the first to investigate the palaeontological and sedimentological characteristics of the Coyhaique Group sediments and present an integrated analysis of the findings as a palaeogeographical interpretation.

8.1 Summary of the new research findings

At the base of the Coyhaique Group, the Toqui Formation comprises a variable succession of calcareous, arenaceous and pyroclastic lithologies. The oyster banks found in the north of the study area indicate the northeast margins of the Aysén Basin during the Berriasian. Further along the coastline volcanic sands were washed in by turbidity currents and pyroclastic material comprising tuffs and volcanic breccias were interbedded within the marine deposits. In the southwest of the study area the Toqui Formation comprises discontinuous occurrences of bedded limestones, sandstones and pyroclastic rocks. The reef fauna and shelly deposits indicate a warm, shallow sea less than 100 metres deep, and a high energy shoreline at the edge of an intermittently active volcano. Fossil evidence points to an Upper Berriasian age for the Toqui Formation deposits.

The black shales and mudstones of the Katterfeld Formation form a thick, regionally extensive succession. The parallel bedding and isolated instances of fine-grained turbidites indicate a lack of steep gradients when the muddy deposits were laid down on the bed of a relatively shallow, epicontinental sea. There was a high input of organic material from nearby terrestrial sources likely to have been the San Jorge Basin. The low diversity of fauna shows that water circulation was generally restricted but occasional oxygenating events occurred which allowed a more diverse infauna to colonise the soft sediments. The fossils of the Katterfeld shales of the Aysén Basin correspond closely to the more sandy facies of the Springhill Formation of the Magallanes Basin, which has been dated as Berriasian. However, the ammonite *Favrella* was found in the upper part of the sequence and is considered to be Lower Hauterivian. The faunal similarity between the basins indicates that there was certainly a connection during this time period. Similarity of fossil assemblages of the southern basins to South African fauna, show that by the late Valanginian to early Hauterivian, endemic species-groups with a South African affinity had developed. The Aysén Basin area was relatively shallow, probably less than 200 metres deep, with weak bottom currents and slow rates of subsidence which kept pace with sedimentation. It was probably wedge-shaped, with the deepest part to the south, in the Magallanes Basin but also stretching east into Argentina.

The Apeleg Formation shows an abrupt transition from the black shales of the Katterfeld Formation to rippled sandstones and silty-shales which indicate a sudden influx of coarser material and a change to higher energy conditions. The layered nature of the contact with no evidence of interfingering suggests a widespread regional event as environmental conditions changed simultaneously throughout the basinal area studied. The suggestion that the Apeleg sediments are continental is not borne out in the Aysén Basin of the study area. Unambiguously marine fossils and trace fossils characteristic of a shallow-shelf environment are found throughout the sequence. The prograding delta setting described in the Rio Mayo Embayment does not extend into the area of the present study although the source of the sediments is probably rivers flowing from the San Jorge Basin in the east. These sediments suggest alternating tidal currents which deposited the uniform, fine to medium-grained sands, and low energy periods when muds were allowed to settle out from suspension. The thick, uniform successions of sandstones and mudstones suggest that energy levels were constant over fairly long periods of time. There was very little evidence of slumping and most deposition was below wave base, possibly at less than 50 metres depth. The ammonite *Favrella wilckensi* was found in abundance at the base of the succession and has been used to date the lower beds as Upper Hauterivian. No stratigraphically useful fossils were collected from higher in the succession and so the precise upper age of the formation cannot be determined although the overlying Divisadero Formation is thought to be Aptian. The change from the quiet, restricted environment of the Katterfeld shales to the higher energy environment of the Apeleg sandstones, which represent offshore tidal sandbars, is the result of a regionally extensive and synchronous environmental change related to the opening up of the restricted basin to normal ocean tides and currents.

Analysis of models of basin formation have shown that the Aysén Basin does not fit neatly into accepted models but falls between an undeveloped backarc basin and a retroarc foreland basin, having some of the attributes of both. Unlike the Rocas Verdes Basin in the south, extension in the Aysén Basin ceased before the seafloor ruptured and oceanic crust formed. In the Neuquén Basin in the north, ophiolitic rocks are also unknown but block faulting and folding is present as well as flysch

deposits which are not found in the Aysén Basin. The Aysén Basin sediments were apparently laid down on a virtually flat surface which subsided gently and where basin margins never became steep enough for slumping or mass flow deposits. The study area sediments are entirely shallow marine and there is no sign of any tectonic events or deepening of the basin. The evidence points to a gently subsiding basin which formed due to limited extension where the continental crust thinned but did not rupture.

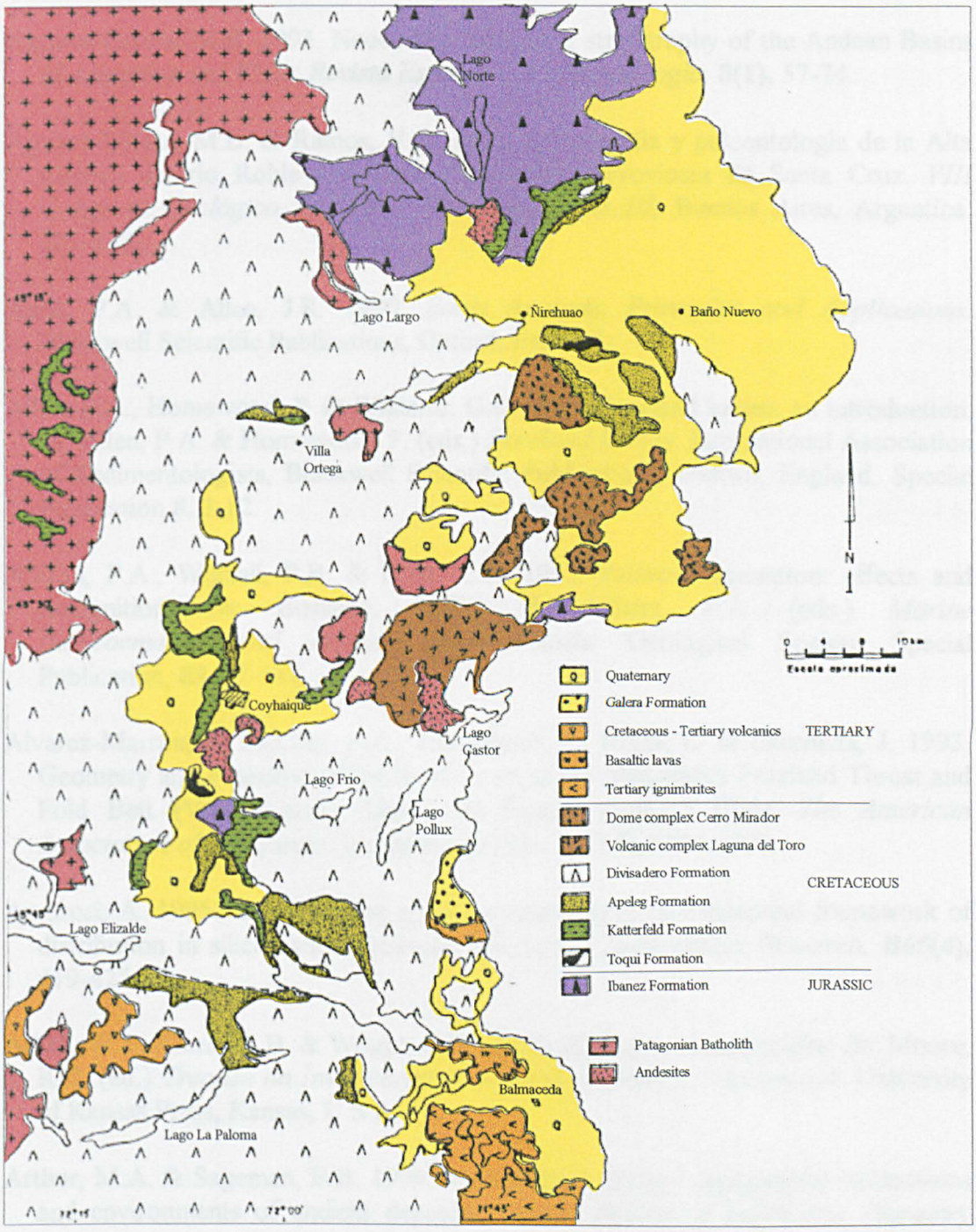
The palaeogeography of the Aysén Basin can be interpreted from the information gathered from studies of the Coyhaique Group sediments and structural and tectonic style of the basin. The Toqui Formation was laid down in a marine transgression which covered the volcanic basement rocks of the Ibañez Formation during the Tithonian to Berriasian. In the southwest of the study area reefs formed in a warm, shallow sea along the shorelines of active volcanoes. The northeasterly margin of the basin occurred near the Argentine border where oyster banks formed. Further east in Argentina, equivalent formations are more pelitic in nature and some small, restricted, deeper basins encroached into the study area near Cuesta Los Leones. To the northeast of Ñirehuao, near Lago La Plata in Argentina, equivalent successions are composed of calcareous sands and volcanic tuffs. By the beginning of the Hauterivian a restricted basin had formed over much of the study area and the black shales of the Katterfeld Formation were laid down over a vast area. Fossil assemblages indicate a marine connection to the deeper Magallanes Basin to the south. During the Upper Hauterivian the basin opened up to normal marine conditions and probably received sediments from rivers flowing from the San Jorge Basin in the east. Marine sedimentation ended in Barremian to Aptian times when calc-alkaline activity associated with a major phase of magmatism produced the overlying sedimentary and silicic volcanic rocks of the Divisadero Formation.

8.2 Significance of the new research findings

This work has shown that the Aysén Basin did not share a common geological history with adjoining basins as previously surmised but was a distinctive basin. It was almost certainly segregated from the Neuquén Basin to the north which although not developing into a true backarc basin with oceanic crust emplacement, underwent much greater tectonic and structural deformation than the Aysén Basin. Faunal comparisons show that the two areas were not connected. The Rocas Verdes and Magallanes Basin in the south experienced rapid extension and ophiolite emplacement which was not found in the Aysén region. The Magallanes Basin and Aysén Basin were connected by a restricted opening during the Late Berriasian or Early Hauterivian and shared a similar faunal community related to South African genera. The Aysén Basin developed as a result of gentle subsidence, probably due to thermal relaxation related to a period of volcanic quiescence, from the Berriasian until at least the Late Hauterivian, possibly Early Barremian. The tectonic setting, palaeoecology and palaeogeography of the basin suggest a shallow epicontinental environment.

This research has provided important new evidence that shows that the Aysén Basin is a distinctive unit with geographically restricted and well-defined characteristics. It has provided information on energy levels, nutrient supply and oxygen availability which reflects environmental conditions throughout the basin. This work fills in an important missing part of the palaeogeography of southern South America during the Lower Cretaceous period.

Appendix A



Based on a map by De La Cruz *et al.* (1994)

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