SEISMIC STRATIGRAPHY OF THE EASTERN CONTINENTAL SHELF OF THE WEDDELL SEA, ANTARCTICA

by

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A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE MASTER OF ARTS

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ABSTRACT

Interpretation of a 1000 miles of seismic data, covering the eastern part of the Weddell Sea continental shelf, in Antarctica, reveals a passive margin environment adjacent to the East Antarctic craton.

Two major seismic stratigraphic sequences bounded by regional unconformities onlap a highly eroded and possibly faulted acoustic basement. The basement terranes which exhibit high magnetic susceptibilities were tentatively correlated with Mesozoic igneous rocks identified in the Transantarctic Mountains. The older sequence is formed of nearshore deposits which do not contradict the presence of ice on the mainland. The younger sequence, however, exhibits the seismic response expected from glacially derived sediments. The angular unconformity separating both units was probably related to a major (late Oligocene?) drop of sea level, while the erosional event which followed the deposition of the glacial series corresponds to a peak in glacial activity (Pliocene?).

Structural framework of the area can only be approached through broad scale studies involving the complex relationships between the East and West Antarctic subcontinents.
ACKNOWLEDGEMENTS

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Finally a mention goes to the American Women's Group in Paris, whose financial support gave me the opportunity to study in the U.S.A.
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INTRODUCTION

During the austral summers of 1977 and 1979, the Norwegian Polar Institute led two scientific cruises to the Antarctic. The area explored lies offshore of the eastern coast of the Weddell Sea, between longitudes 16°W and 42°W (fig. 1). The geophysical data gathered, representing 1,000 miles of medium penetration seismic tracks and 100 miles of marine magnetic profiles, are the basis of this study. These data are quite unique, since limited exploration of the continental shelf of the Weddell Sea has occurred. An attempt was made in this study to integrate seismic, magnetic, and gravity data to reconstruct the geologic history of the continental shelf of the eastern Weddell Sea.

Both the Weddell Sea and Ross Sea (fig. 1) are part of a series of depressions and basins which naturally divide the Antarctic continent into two subcontinents: East and West Antarctica. These low terranes lie mostly below sea level and are thought to be tectonically related by a large rift zone separating the subcontinents. The two subcontinents exhibit distinct differences. East Antarctica is a continuous cratonic high underlain by some 45 km of continental crust (Demenitskaya, 1960) while West Antarctica is characterized by a series of isolated topographic highs several thousand meters in elevation which are separated by troughs that extend well below sea level (fig. 1, Avzyuk et al, 1964). In addition, West Antarctica is younger, has a total crustal thickness of 30 km (Demenitskaya, 1960) and represents a complex, ancient active margin environment (Dalziel and Elliott, 1973,
Figure 1: Subglacial map of Antarctica
(from Avzyuk et al., 1964)

1. Graham Land
2. Palmer Land
3. Ellsworth Land
4. Marie Byrd Land
5. Byrd Subglacial Basin
6. Crary Trough
7. Queen Maud Land
More is known of the Ross Sea region both offshore and onshore than of the Weddell Sea. An extensive study of DSDP cores and geological data collected from the Ross Sea has led to a detailed knowledge of Cenozoic glacial sedimentation in the basin (Houtz and Davey, 1973; Balshaw, 1981). Since little is known of marine geology in the Weddell Sea, the Ross Sea was chosen as an analog basin in order to calibrate stratigraphic interpretation of the seismic data. The stratigraphic interpretation was based on the seismic stratigraphic techniques developed by Vail et al., (1977) which involve definition of seismic sequences, study of reflection configurations, and study of signal definition.

Anomalies encountered in the magnetic record are indicative of past igneous activity which may be correlated and interpreted with respect to a large tectonic framework. Scarcity of magnetic data eliminated possible extrapolation between magnetic profiles, and thus, the creation of a two dimensional picture of total field intensity. Reduction of the magnetic data was, thus, made on the individual profiles using a simple rule of thumb, known as Peter's Method. Published gravity data (Behrendt, 1966 and 1974), which cover the internal part of the shelf and its associated terranes, were also used to further augment the tectonic interpretation.

This thesis is organized in the following manner. A brief outline of the geology of the Weddell Sea and glacial history of the Antarctic continent is reviewed as basis for interpretation of the geophysical records. The seismic data are presented and their characteristics discussed in terms of seismic stratigraphic concepts. A comparison is then made between seismic events in the Weddell and Ross Seas. Finally,
magnetic and gravity data in the Weddell Sea are integrated with seismic stratigraphy to further define the tectonic framework.
CHAPTER I

GEOLOGICAL FRAMEWORK
GEOLOGICAL FRAMEWORK

Overview of the Geology

Both age and genesis of the Weddell Basin are controversial, since data are limited. Geology of the basin's margins is derived from a restricted number of outcrops located in mountain ranges high enough to rise above the 3,000 m thick ice sheet (fig. 2).

The basin is situated at the western edge of the Eastern Antarctic Precambrian shield. This cratonic area was consolidated during the early Paleozoic by a succession of orogenic episodes - the Beardmore and Ross orogenies - (fig. 3), resulting in a mountain belt which limits the craton on its western sector (Dalziel and Elliott, 1973). The Transantarctic Mountains (fig. 1) form the eastern coast of the Weddell Sea. Further evolution of the craton involved only epeirogenic movements, the latest uplifts having been dated as Cenozoic.

The southern and western coasts consist of late Paleozoic to early Cenozoic strata which have recorded several deformational events. These roughly parallel orogenic belts move towards the Pacific through time (fig. 3). Subduction along the Antarctic Peninsula (fig. 1) continued as late as Eocene time (Weissel et al, 1977). This overall structural pattern is probably simplified. Complexity of the West Antarctic Archipelago seems to indicate, with some probability, movements between the constituent continental block or islands (Scharnberg and Scharon, 1972; Cox and Gordon, 1978; Watts and Bramall, 1980) and eventually between East and West Antarctica (Kellogg and Reynolds, 1977; Alley and Watts, 1979).

Overimprinted on the roughly north-south compressional grain which
Figure 2: Mountain Ranges of the Circum Weddell Sea Region
(from Dalziel and Elliott, 1973)
Figure 3: Tectonic evolution of the Transantarctic Mountains and Antarctic Peninsula, summarizes a review of the literature which is presented in Appendix A.
prevails along the margin of the craton, are the rifting episodes responsible for the breakup of Gondwanaland (Bergh, 1977; Weissel et al 1977; Norton and Sclater, 1979). This Paleozoic supercontinent consisted of all the southern hemisphere continents plus India. It is believed that opening of the Weddell Sea, dated as Mesozoic (Jurassic), by recent paleomagnetic data (LaBreque and Barker, 1981) occurred during one of these rifting episodes.

Since no orogenic pulses were recorded east of the 45°W parallel (fig. 2) after the Triassic to early Jurassic generalized "Gondwanian" episode (Table I), it is likely that the eastern continental shelf of the Weddell Sea is genetically associated with its immediate borderland rather than with the mostly volcanic terranes of the Antarctic Peninsula.

The Crary Trough, which bounds the study area to the west (fig. 1), is part of the series of depression which link the Ross Sea to the Weddell Sea and, thus, represents a portion of the natural boundary between East and West Antarctica. The broad shelf which lies to the west of the trough is thought to be related to West Antarctica (Anderson, 1980). This differentiation within the study area is made on the basis of broad scale structural grain and applies exclusively to basement terranes, rather than to the actual basin infill. Origin of the Crary Trough is believed to be erosional since similar types of features are known in other glacial environments as well as on other parts of the Antarctic Continental margin (Johnson and Vanney, 1976). The trough is bound to the south by the Pensacola Mountains (fig. 1). The large mafic stock which outcrops in the Forrestal Range (fig. 2), the Dufek intrusion, has been determined to extend beneath the trough (Behrendt, 1974). Age of its emplacement is Jurassic and it is
<table>
<thead>
<tr>
<th>AGE</th>
<th>WESTERN QUEEN MAUD LAND</th>
<th>SHACKLETON MOUNTAINS</th>
<th>PENACOLA MOUNTAINS</th>
<th>ELLSWORTH MOUNTAINS</th>
<th>ANTARCTIC PENINSULA</th>
<th>SOUTH SHEETLAND ISLAND</th>
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<tr>
<td></td>
<td>KIRIAN ESCARPMENT</td>
<td>NW of ESCARPMENT</td>
<td>JPAYA</td>
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<td>JURASSIC</td>
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<td>DEVOMIAN</td>
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<td></td>
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<tr>
<td>CAMBRIAN</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BEARDMORE ORDOYEN</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>PRE-CAMBRIAN</td>
<td></td>
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Table 1: Stratigraphic correlations, Weddell Sea Province, Antarctica.
associated with roughly contemporaneous sills and stocks (i.e. Ferrar dolerites) found throughout the Transantarctic Mountains. These intrusives possibly originated during an episode of block faulting and eventual rifting along the Transantarctic front. Their actual relationship in time with the age of the Weddell Sea is certainly significant, but can only be inferred.

The marine geological record of the Weddell Sea is limited to short piston cores (Anderson, 1972) which have usually only sampled Upper Cenozoic glacial deposits. The only nonglacial sediments encountered (i.e. Core G8 - west of Crary Trough) were reworked eolian sands (fig. 4)(Anderson, 1972). They are possibly of Cenozoic age (fig. 1). These same exposures apparently outcrop along the western flank of the trough. The only other nonglacial sediments of upper Cenozoic age recognized in Antarctica are found in the higher latitudes of Seymour Island at the northern tip of the Antarctic Peninsula (fig. 1). Here, outcrops of Paleocene to Eocene marine strata indicate a high energy deltaic environment (Elliott et al, 1975).

Antarctic Glacial And Glacial Marine Sediments

Sedimentation on the Antarctic continental shelf is governed by ice sheets which cover the continent. East Antarctica is thought to be covered by a stable ice sheet with a basal temperature below freezing (dry-base ice sheet of Carey and Ahmad, 1961). The western archipelago, in contrast, is covered by an unstable ice sheet whose basal temperature is above the freezing point (wet-base ice sheet of Carey and Ahmad, 1961).
Figure 4: Locations of Phleger cores and piston cores taken during the 1968, 1969 and 1970. International Weddell Sea Oceanographic Expeditions (from Anderson, 1972)
These two major glacial regimes prevail, at the present, in Antarctica along with the valley glaciers which drain areas of high relief, such as the Transantarctic Mountains.

Ice sheets transport most of their sediment load in their basal portions, as opposed to valley glaciers which bear large quantities of debris englacially (Boulton, 1975). The transportation mode, which is largely a function of the thermal regime of the ice sheet, control glacial-marine sedimentation (Carey and Ahmad, 1961; Boulton, 1972).

Wet-base conditions are the sole criterion for deposition of basal till beneath an ice sheet. The seaward limit of this depositional mechanism is the grounding line of the ice sheet, which cannot extend beyond the continental shelf break. The remaining sediment load is deposited from floating ice. Ice rafting of sediments by icebergs, calved from valley glaciers and ice shelves, transport sediments far seaward of the continental slope. Drainage from dry-base ice sheets is accomplished through outlet glaciers and ice streams. The debris they carry is either deposited at their terminus by floating ice or is transported farther seaward by icebergs.

A study of piston cores from the Antarctic continental shelf indicates that most of the sediments there, were probably derived from the basal zone of the ice sheets (Anderson et al, 1980). Three principal facies are recognized and their characteristics are outlined on Table II. The most widespread facies is basal till, which reflects deposition beneath a grounded, wet-based ice sheet. The other two facies, residual and compound glacial marine sediments, are believed to be ice-rafted sediments reworked in the marine environment either during or after deposition. They reflect either vigorous (residual) or
Table II

Outline of characteristics of Antarctic glacial and glacial marine sediments (after Anderson et al, 1980).

<table>
<thead>
<tr>
<th></th>
<th>Basal Till</th>
<th>Compound Glacial Marine Sediment</th>
<th>Residual Glacial Marine Sediment</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Stratification</strong></td>
<td>None</td>
<td>Crudely to well stratified</td>
<td>Crude or absent</td>
</tr>
<tr>
<td><strong>Texture</strong></td>
<td>Polymodal size distribution. Matrix sorting values range from 2.3 Φ to 3.3 Φ. Matrix is coarse skewed. Individual units exhibit strong textural homogeneity.</td>
<td>Broadly bimodal size distribution with pronounced silt modes consisting of current derived silts and unsorted ice-rafted mode, the relative concentrations of these two components varies with depth in the section.</td>
<td>Poorly sorted sands and gravels whose fine component has been winnowed resulting in coarse skewed grain size distribution.</td>
</tr>
<tr>
<td><strong>Unit Thickness</strong></td>
<td>Minimum thicknesses of 4.5 to 9.0 meters.</td>
<td>Ranges from a few cm to a few tens of meters.</td>
<td>Insufficient data</td>
</tr>
<tr>
<td><strong>Physical Properties</strong></td>
<td>Overcompacted cohesive strengths exceed 2.5 kg/cm.</td>
<td>Often water saturated with cohesive strength of less than 2.5 kg/cm.</td>
<td>Loosely compacted</td>
</tr>
<tr>
<td><strong>Pebble Fabric (vertical plane)</strong></td>
<td>None</td>
<td>Elongate pebbles aligned parallel to bedding</td>
<td>Elongate pebbles showing crude alignment parallel to bedding.</td>
</tr>
<tr>
<td><strong>Pebble Shape</strong></td>
<td>Medial roundness and sphericity. Average of 80% faceted pebbles.</td>
<td>Medial roundness and sphericity.</td>
<td>Insufficient data</td>
</tr>
<tr>
<td><strong>Pebble Striations</strong></td>
<td>Average of 12% abundantly striated.</td>
<td>Average of 12% abundantly striated.</td>
<td>Insufficient data</td>
</tr>
<tr>
<td><strong>Nature of Bounding Contacts</strong></td>
<td>Sharp</td>
<td>Gradational</td>
<td>No contacts penetrated.</td>
</tr>
<tr>
<td><strong>Marine Fossils</strong></td>
<td>None or reworked</td>
<td>Marine fossils with changing diversities and abundances.</td>
<td>Abundant marine fossils</td>
</tr>
<tr>
<td><strong>Distribution</strong></td>
<td>Restricted to continental shelf</td>
<td>Continental shelf to abyssal floor</td>
<td>Occur on shallow portion of continental shelf (above approximately 250 m) and on the continental shelf break.</td>
</tr>
<tr>
<td><strong>Origin</strong></td>
<td>Deposition by grounded ice; lodgement processes.</td>
<td>Deposition from floating ice in low energy marine environment.</td>
<td>Deposition from floating ice in high energy marine environment.</td>
</tr>
</tbody>
</table>
sluggish (compound) current activity.

These three facies can be deposited contemporaneously from a grounded ice sheet and its associated ice shelves and icebergs.

Glacial Chronology in the Antarctic

The Antarctic glacial record has been most recently developed in the Ross Sea area through analysis of marine DSDP cores (Balshaw, 1981). The total section sampled represents 1350 m of glacial and glacial marine sediments spanning the second half of the Cenozoic. It has been shown that most of these sediments were transported from West Antarctica rather than from the entire continent (Barrett, 1975). Thus, the events recorded and correlated through the Ross Sea continental shelf represent evolution of the West Antarctic ice sheet.

A review of the literature summarized in Figure 5 from Balshaw (1981), seems to indicate a possible development of continental ice on East Antarctica by the late Eocene, but excludes the presence of a well developed ice sheet before the Oligocene. An age of 29 MYBP was hypothesized for the formation of the East Antarctic ice sheet. Its formation would then correlate with a major oceanic and climatic cooling (Shackleton and Kennett, 1975), a significant drop in sea level (Vail and Hardenbol, 1979), and finally the establishment of the Circum-Antarctic current (Baker and Burrell, 1977).

The West Antarctic archipelago was, at least in part, glaciated by Oligocene time (Le Masurier, 1972). However, coalescence of the ice shelves surrounding the islands to form the West Antarctic ice sheet did not occur before the mid-Miocene, corresponding to Vail and
Figure 5: Summary of data relative to the glacial chronology of Antarctica
(from Balshaw, 1981)
Abundant: Abundant
Stunted: Stunted
Died out: Died out

Notthofagus Beech Trees on the West Antarctic

WARM WATER SPECIES: No DATA. COLD WATER-SILICOFLAGELLATES on Campbell Plateau

Warming Water Species: No Data. Cold Water-Silicoflagellates on Campbell Plateau

LeMasurier (1972)
Stump et al. (1980)
Rutford et al. (1975)
Drewry (1975)

Uplift of the Transantarctic Mts.

PACIFIC {} 

OPENING OF DRAKE'S PASSAGE

Kemp (1975)
Kemp (1975), Kelgwin et al. (1979)
Kemp et al. (1975)

World-Wide Erosional Unconformities

Kemp et al. (1975), Rona (1975), Moore & Heath (1977), Moore et al. (1977)
Ciesielski et al. (in press) Erosional Unconformity on the M.E.B.

Hayes & Frokes (1975)
Kemp (1965)
Belgren & Haq (1976)
Mercer (1979)
Anderson (1972)

Sea-Level Drop, Blind River New Zealand
Sea-Level Drop, Spain
Patagonian Glaciation
Dry-Base Glaciation, Weddell Sea

Ross Sea Unconformity

Evidence from Tectonic Events Oceanograph Subglacial Volcanoes
Coastal Chart On-Drop Pollen Microflora

Andrews (1972)
Hardenbol's (1979) eustatic sea level drop of 13 MYBP (Balshaw, 1981). The Ross Ice Shelf developed and expanded through the late Miocene and early Pliocene registering peaks of 10 MYBP and 6.6 MYBP. The last extension of the ice shelf produced an extensive erosional surface and was followed by deposition of basal tills (Anderson et al, 1980). The ice is believed to have advanced, then, as far as the present continental shelf break.
Summary

The geological record present under the Weddell Sea continental shelf is poorly known. Bottom sediments have been sampled acknowledging glacial activity on the continent and on the shelf, but nothing is known of the preglacial record and of the entire glacial section. The set of geophysical data which will be interpreted in this study will yield some of these geological unknowns for the easternmost part of the shelf. Since there is no direct geological data to correlate with the geophysics, calibration will be sought in three different domains. Preglacial terranes are bound to be of continental affinity, and will be interpreted through the onshore geology of the circum-Weddell Sea region. Determination of glacial and glacial marine sediments will be made by comparison between the seismic data and the seismic response expected from Antarctic glacial and glacial marine sediments according to the characteristics outlined for these deposits by Anderson et al (1980) (Table II). Finally the chronology determined for the Ross Sea continental shelf by Balshaw (1981) will be used to set a tentative timing of events in the eastern Weddell Sea continental shelf and to compare both depositional environments.
CHAPTER II

SEISMIC DATA INTERPRETATION
SEISMIC DATA INTERPRETATION

Characteristics of the Seismic Data

Figure 6 shows the areas covered by the seismic tracks, which represent the main data base for this study. These areas have been termed Zone A, Zone B and Zone C. Correlation of data between these different zones can only be regarded as tentative. They are located for the most part to the east of Crary Trough and are bound to the south by the Filchner Ice Shelf. Water depths in the Crary Trough range from 150 to over 1,000 m. The location of the lines and the positioning along the tracks are not very accurate, somewhat restraining their interpretation.

The source used was a 4kJ sparker which has a maximum penetration of two seconds..two-way travel time. Further characteristics of the source are given in Appendix B. The data was recorded in analog form and displayed in a 500 ms window centered approximately on the first reflection (sea bottom). Thus, the actual usable section beneath sea bottom has an average of 200 m (300 ms two-way time).

The only processing applied to the data was an onboard analog filter having a 60-400 HZ bandwidth. Since no deconvolution by the source signal was made, the high resolution is limited by the length of the source pulse. A series of synthetic seismograms was generated in order to determine the resolution limit of the source, i.e. the thinnest layer detectable by the seismic system. A discussion of the parameters used for the seismograms is given in Appendix B.

Two interfaces separated by less than the wavelength of the input signal, are very difficult to differentiate. This critical value of 18.
Figure 6: Bathymetric map of the Weddell Sea indicating areas covered by seismic profiles.
one wavelength would be 9.4 m for the 3kJ sparker source and 11.6 m for the 5kJ sparker source, given a velocity of propagation of 1600 m/s. These figures were tested by generating seismograms based on a simple one layer model (Appendix B). In these seismograms, the thickness \( t \) varied, but the velocity of propagation of seismic waves inside the layer \( V_L \) was kept constant at 1600 m/s; \( t \) was given respectively values of 2.5 m, 5 m, 10 m, 15 m and 20 m (fig. 7 and 8). Due to the length of the pulse, layers of less than 20 m, cannot be differentiated.

However, direct comparison of the seismic data with the synthetic seismograms might allow us a much smaller definition. Sparker sources have a relatively large high frequency content, as necessary for high resolution system. It was tempting to test whether a layer smaller than the definition interval could be detectable through frequency analysis of the reflected signal. Although the results for \( t = 2.5 \) m (fig. 7 b) appeared quite satisfactory, this frequency is very dependent on the recording system and the spectrum of the ambient noises.

A last cause for the change of signal shape was investigated by varying the velocity \( V_L \) within a 10 m layer. Three values were used:

\[
V_L = 1450 \text{ m/s} \quad \quad V_L = 1600 \text{ m/s} \quad \quad V_L = 1800 \text{ m/s}
\]

The three responses are shown in Figure 8. Seismogram C \((V_L = 1800 \text{ m/s})\) is very different from the other two. Several reasons may be attributed to these relationships. One of them being that the critical thickness of one wavelength has now become 14.5 m due to the increase in velocity of propagation. This quantity is much larger than the layer thickness adopted in the model (10 m). Another reason might be the interference between the reflection at the layer/'basement' interface and the surface reflection of the ocean-bottom reflection.
Figure 7: Synthetic Seismogram I

Synthetic seismograms and comparison with seismic data.

\[ V_{\text{Water}} = 1450 \text{ m/s} \]

\[ V_{\text{Layer}} = 1600 \text{ m/s} \]

- a. Barren "basement" (homogeneous infinite layer) 5kJ source.
- b. 2.5 m layer overlying "basement" 5kJ source.
- c. 5 m layer overlying "basement" 5kJ source.
- d. 20 m layer overlying "basement" 5kJ source.
Figure 8: Seismogram (II)

Comparison of the response with varying sea floor velocities- 10 m layer overlying homogeneous "basement"

- $V_{\text{Water}} = 1450$ m/s
- $V_{\text{Layer}} = 1450$ m/s, $\rho_{\text{Water}} = 1.3$ g/cm$^3$
- $V_{\text{Layer}} = 1600$ m/s, $\rho_{\text{Layer}} = 2.7$ g/cm$^3$
- $V_{\text{Layer}} = 1800$ m/s
Analysis of the Data

Seismic stratigraphic sequences and units were determined for Zones A and B. Units are defined on the basis of their seismic character while sequences consist of a series of units bounded by major unconformities. Although both areas are separated by at least 200 km, events and units were tentatively correlated from one to the other since their seismic character and "stratigraphic" positions were similar. However, features were mapped separately. A list of the seismic stratigraphic units defined in this study is given in Figure 9. Location of the seismic profiles used in this study are indicated in Figure 10 (Zone A) and Figure 11 (Lagenbanken Area, Zone B).

**Acoustic Basement:**

Acoustic basement is characterized by a highly irregular top surface and by the lack of any organized reflections beneath this surface. All other units onlap this basement toward the continent. The absence of reflections within the acoustic basement has two possible explanation, either the unit has a homogeneous response to seismic waves or the velocity contrast at its top surface is high enough that very little energy can be transmitted into the unit itself.

Acoustic basement was identified on Profiles 77 (fig. 12); V (fig. 13); 81 (fig. 14) and XII, VII and XI in Zone C (fig. 15) (for location refer to Figure 10). Acoustic basement was not recognized in Zone B. The top surface of the acoustic basement dips systematically offshore and outcrops at sea bottom near the continent (fig. 12, 13, 14 and 15). It is the sole seismic unit recognized on profiles XII and VII. Figure 16 shows the approximate offshore limit of acoustic basement outcrop.
Figure 9: Seismic Stratigraphic units seen in Zone A and B
Figure 10: Location map of seismic profiles shot in Zone A and C.
Figure 11: Location map of profiles shot in the Lagenbanken area (Zone B)
Figure 12: Line 77 (Zone A)

Sketch of seismic profile (top) and total magnetic field measurements (bottom). (for location see Figure 10)
Figure 13: Line V (Zone A)

Sketch of seismic profile (for location see Figure 10)
Figure 14: Line 81 (Zone A)

Sketch of seismic profile (top) and total magnetic field measurements along track (bottom) (for location see Figure 10)
Figure 15: Line XI

Sketch of the only seismic profile of Zone C, which transects Crary Trough from west to east (A-A')
(for location see Figure 10)
The most tempting hypothesis would be to extend this basement onto the continent, relate it to rocks found onshore, and consider it to be the true basement.

Interestingly, the area defined on the map Figure 16 corresponds closely to the eastern flank of Crary Trough, where the bathymetry defines a very steep and irregular sea bottom when compared to the western flank. This asymmetry of the trough could be the result of a difference of response of the seabed material to the erosional event which carved the trough. The acoustic basement being much more resistant since it probably represents old sedimentary deposits or nonsedimentary rocks. Another source of evidence is the difference of seismic response seen on profile XI (fig. 15) between the eastern terranes (acoustic basement) and the layered strata seen to the west. The contact between the two is not clear, but the layered unit, most likely younger, seems to onlap the acoustic basement.

Piston cores collected in the vicinity of profiles VII and XII (G3 and 2-20-1, for locations see fig. 4 and fig. 16) were analyzed by Anderson et al., (1981). Both of the cores penetrated basal tills and indicate the presence of at least 5 m of such deposits on the sea bottom. These basal tills lack stratification, have strong textural homogeneity and have polymodal size distribution (Table II). It is therefore unlikely that they will exhibit organized internal reflections on a seismic section. Minimum thicknesses of basal tills are in the 10 m range. The acoustic basement could possibly represent extremely thick units of basal till (up to 200 m). However, there is no evidence as to the existence of basal till units that thick and a closer examination of the sea bottom reflection where the acoustic basement outcrops,
Figure 16: Offshore limit of acoustic basement outcrop.
shows a varying pulse (number of peaks and troughs). Comparison of data with the synthetic seismograms (fig. 7), indicates that the acoustic basement is covered by a thin veneer of sediments of variable thickness (0 to 10 m).

Acoustic basement as defined in the study area represents the sea floor, where the bathymetry indicates an irregular sea bottom. This basement seems more resistant to erosion than overlying sediments and is not likely to represent glacial deposits. It can be tentatively extended onshore. Reduction of total magnetic field profiles recorded along the seismic tracks yielded further information on the nature of this acoustic basement (see section on Magnetic and Gravity Data). In particular these data could support the idea that the acoustic basement seen on the seismic profiles represents the true basement.

Sequence A:

This older seismic stratigraphic sequence contains at least two units, which have been informally named the Draping Unit and Prograding Unit (fig. 9). A regional angular unconformity (υ) bounds the top of Sequence A. Characteristics of both units will be discussed.

- Draping Unit:

The Draping Unit is the oldest recognized sedimentary unit in Zone A and B (fig. 6). It can only be seen at the bottom of Crary Trough where younger sediments have been removed by glacial erosion. The Draping Unit is best illustrated in seismic section 77 (fig. 12).

Seismic signature of the unit observed in Figure 9, indicates a finely bedded unit which drapes the acoustic basement.
Both of these characteristics suggest a low energy environment with fine calastics deposition. The layered appearance represents either chemical changes in the section (difference in carbonate or siliceous content) or changes in grain size (claystone vs. silts). At any rate, the layering is in general enhanced by the composition of the reflections and the pulse length since no deconvolution was applied to this data. The strata have a general depositional dip of $1^\circ$ or $2^\circ$ towards the shelf break, which gives a minimum cumulative thickness of 460 m for this unit.

- Prograding Unit:

  The Prograding Unit was recognized on profiles in Zone A and B (Line 77, fig. 12; Line V, fig. 13; Line 81, fig. 14; Line 80, fig. 18; Line 83, fig. 19; Line 84/85, fig. 20; Line XVIII, fig. 20; Line XVI, fig. 21; Line XVII, fig. 21). The contact between these deposits and the Draping Unit is poorly defined (Profile 77, fig. 12). The Prograding Unit directly overlies the acoustic basement in the vicinity of Crary Trough (Profile 81, fig. 14; Profile V, fig. 13). Its seismic signature, as seen in Figure 9, is best defined on Profile III (fig. 17) where it outcrops on the sea floor.

  The unit is a wedge of tilted strata with a minimum cumulative thickness of 900 m. The dip angles are variable in the range of $2^\circ$ to $3^\circ$ in a seaward direction. They vary from one reflector to another giving a rough "cut and fill" geometry. They also vary along the same reflector with depth. This
Figure 17: Line III, seismic profile (top)
Line VIII, seismic profile (bottom)
(for location see Figure 10)
Figure 18: Line 80 (Zone A)

Sketch of seismic profile (top) and total magnetic field measurements along track (bottom).
(for location see Figure 10)
Figure 19: Line 83 (Zone A)
Sketch of seismic profile. (for location see Figure 10)
Figure 20: Lagenbanken Area (Zone B)

a) Line XVIII, sketch of seismic profile

b) Line 84/85, sketch of seismic profile

(for location see Figure 11)
Figure 21: Lagenbanken (Zone B)

a) Line XVI, sketch of seismic profile
b) Line XVII, sketch of seismic profile

(for location see Figure 11)
variation with depth may not indicate the true attitude of the dipping bed but instead an artifact inherent to the seismic method. Synthetic Seismogram III (fig. 22) illustrates this situation. The model assumes identically dipping layers whose velocities increase with depth. These layers overlie a homogeneous half-space or "basement". Apparent dips on the seismogram decrease with depth. Generally, propagation velocities of seismic waves are known to increase in a more or less regular fashion with depth. It is, thus, possible to explain the occurrence of variations of apparent dip angles along an interface on seismic data purely by data recording (time vs. depth) distortion. Another simple explanation might be a difference between the direction of true dip and the direction along which the data is gathered. Another point in this regard concerns the vertical exaggeration on the profiles. Vertical exaggeration is very large, on the order of 1:10, thus variation in dip angles are largely enhanced. A small variation in the ship's course from the recorded track could cause the strata to appear as though the dip changed.

Figure 23 shows the apparent dip directions seen along the seismic profiles where the Prograding Unit was recognized. The overall picture is that of a northwest regional accretionary wedge of sediments which dips toward the continental shelf break.

On Profile III (fig. 17), two major subunits (A and B) are recognizable and separated by a slight angular unconformity which dips offshore. Subunit B onlaps subunit A. Subunit A
Dip angle varies going downsection when velocity of seismic waves propagation increases with depth.
Figure 23: Map showing apparent dips of the Prograding Unit along the seismic tracks of Zone A. The dips indicate a common northwest direction towards deeper parts of the Weddell Sea.
is interpreted as the Prograding Unit and the angular unconformity as the γ unconformity which defines the top of Sequence A. Seismic character of the Prograding Unit is quite homogeneous consisting of:

- irregular but continuous reflections,
- reflection-free prisms of sediments, and a
- cut and fill geometry.

Values of the apparent dip angles are such that they suggest a depositional rather than a tectonic tilting. The foresets which prograde offshore are indicative of a marginal marine, high energy environment. Although this excludes a purely glacial origin (basal till), it does not deny the presence of ice on the mainland (i.e. glacial outwash). The reflectors within this sediment wedge imply differences in the nature of the deposits, creating acoustic impedance contrasts. The most common interpretation in such environments is the alternation of shales and coarser grained material. A reasonable depositional model is perhaps that of a glacial delta front sequence similar to that illustrated in Figure 24. This model implies less severe glacial conditions than exist today in Antarctica. A good modern analog may be that of the Gulf of Alaska (Bruce Molnia, personal communication). A similar prograding sediment sequence exists in the Ross Sea. The sequence consists almost entirely of late Cenozoic glacial marine deposits (Hayes et al, 1975, fig. 25). The scale of the seismic data in the Ross Sea is very different from that of the Weddell Sea, so the comparison between these two data sets is very problematic.
Figure 24: Sedimentation in a glacial sea. During the summer, mainly climbing ripple-laminated sand is deposited on a prograding delta, while thin silt layers are deposited by a density underflow on the sea bottom. During the winter, these areas are covered with a clay drape. Icebergs, and shore and river ice may raft coarse sediment into the sea during the spring thaw and summer. (from Reading, 1978)
Figure 25: Seismic profile shot in the Ross Sea (USN S-Eltanin, Cruise 51-55A, 1972), at location of DSDP Leg 28, drilling. Showing sequence of tilted strata truncated near sea floor.
The Ross Sea sequence is much poorer in internal reflectors (fig. 25). This might be due to the lower resolution of the data. Also, these reflectors lack the cut and fill geometry, showing greater lateral extension. Finally, the dips seen within this sequence are not systematically directed offshore, rather the deposits seem to follow the structure of the basement.

In summary, the deposits of the Prograding Unit in the Weddell Sea accumulated under essentially marine conditions and represent a marginal high energy environment. The possibility of glaciation on the mainland is not negated. A major erosional event followed the deposition of the Prograding Unit. Its expression in the geologic record is an unconformity of regional extent (\(\mathcal{V}\)) which can be recognized in both Zones A and B. It is a very significant feature since it separates strata which are thought to be nonglacial (Sequence A) from deposits which will now be interpreted as glacial (Sequence B).

**Sequence B:**

The deposits of Sequence B onlap onto the unconformity \(\mathcal{V}\). They are directly in contact with the acoustic basement in the area of Profile V (fig. 13). Sequence B is divided into two units (I and II, fig. 9) which perhaps represents two intricately related facies.

Unit II is well layered with flat bedding planes. These internal reflections seem fairly continuous, although they change in character rather frequently. Unit II suggests the accumulation of large thin patches of sediments. Locally (0090 on Profile 77, fig. 12), hummocky
masses, which are similar to the younger deposits (moraines?), are present within the bedding unit. The configuration can become more complex with superposition of cuts and fills, toward the lateral terminations of Sequence B (0200 on Profile V, fig. 13). Unit I on the other hand has very few internal reflections. The image of occasional surfaces transecting the unit (0100 on Profile V, fig. 13; 1300 and 1400 on Profile 81, fig. 14) and isolated diffraction hyperbolae are the only features observed. The homogeneous and massive seismic character of Unit I is the response which might be expected from basal till units. The hyperbolae would show the presence of internal boulders. There is no ready explanation for the velocity contrast which generates the other reflections in what seems to be an otherwise homogeneous layer. The best interpretation would be that of an erosional surface made by an ice sheet. The ice would overcompact the underlying deposits and create a density contrast with younger sediments, detected by the seismic system.

Contact between Units I and II vary. Unit I overlies Unit II near the mainland, but further offshore, the massive facies of Unit I passes laterally into Unit II (0800 0500 on Profile 80, fig. 18; or eastward on Profile IV, fig. 26). At the shelf break, the Unit II is present. These features are interpreted as being indications of a facies transition from purely glacial deposits, basal tills, at the grounding line of the ice sheet to glacial marine deposits further offshore (fig. 27). The discontinuous nature of the reflectors within Unit II also support this interpretation. Several maps illustrate the geometry of these two units. In Zone A, isochron (isopach in time) maps were made for each unit, indicating the nature of its lateral terminations (fig. 28, and fig. 29). In Zone B relations between both facies are more complex. The distribution
Figure 26: Line IV
Sketch of seismic profile
(for location see Figure 10)
Figure 27: Hypothetical model of glaciomarine sedimentation adjacent to a wet-based tidewater glacier, based on observations of Pleistocene subaqueous outwash and considerations of current dynamics of modern marine deltas. Most of the fresh glacial meltwater rises to surface of the sea as a low density overflow layer. This layer gradually mixes with sea water while being driven by winds and currents. Silt and flocculated clays gradually fall out of suspension. Rapid mixing of fresh water and sea water adjacent to tunnel mouths may produce a high density underflow capable of transporting sand-grade sediment and occasionally coarser material. (from Reading, 1978)
Figure 28: Areal extent and isochrons of Unit II (50 ms interval)
Figure 29: Areal extent and isochrons of Unit I (50 ms interval)
of sediment packages which incorporate mostly deposits of the Unit II facies was studied (fig. 30, fig. 31 and fig. 32). The depositional regime in Zone B, must have been very different from the one in Zone A.

Comparison of the core data (Anderson, 1972; Anderson et al, 1981, fig. 4) with the seismic data is difficult, since the distances separating the core locations and seismic lines are very large (~10 km) and the cores only sampled the sea bed sediments. Most of the cores taken in this area penetrated basal tills with thicknesses of up to 10 m. Units I and II do not always outcrop at the sea floor. They have been locally truncated by an unconformity (T) and overlain by irregular patches of younger sediments. These could be the formations sampled by the cores.

The younger deposits have a very irregular top surface and no internal reflectors. An isochron map of a body of these sediments can be seen in Figure 33. The fan shape seen on this figure is not indicative of any particular depositional environment. The interpretation of these deposits is problematic. They lie at the head of a large canyon which transects the continental slope slightly to the east of Crary Trough.

**Deposits Forming the Continental Shelf West of Crary Trough:**

Profile XI (fig. 15) was shot for the most part in what is assumed to be the West Antarctic part of the Weddell Sea continental shelf. It has a general east-west orientation, transecting Crary Trough at the very edge of the Filchner Ice Shelf, and covers mostly the West Antarctic part of the continental shelf (fig. 10). No correlation was attempted with units identified farther north in Zones A and B, since
Figure 30: Extent of oldest glacial unit seen in the Lagenbanken area (Zone B).
Figure 31: Extent of the middle glacial unit, Lagenbanken area (Zone B)
Figure 32: Extent of the youngest glacial unit, Lagenbanken area (Zone B).
Figure 33: Isochrons of a body of younger deposits (20 ms interval)
little similarity in sedimentary units can be found. The steep eastern flank of the trough is mostly barren of sediments except for small mounds of hummocky material i.e. lodgement tills or relict sediments (?). On the western flank of the trough, well bedded deposits appear truncated almost at the sea floor. Whether these formations correlate with the Prograding Unit is questionable. Their appearance is somewhat different when compared with the Prograding Unit. They lack the cut and fill geometry and the bedding planes which dip at $1^\circ$ or $2^\circ$ to the northwest seem to have undergone some deformation (fig. 15). Their calculated minimum cumulative thickness is 1200 m.

The units overlying this thick sequence dip gently to the west. At the western end of the profile they are cut by canyons (fig. 15). These canyons create time anomalies in their bedding planes which are most probably flat lying. These artifacts created by the seismic method are illustrated in Seismogram IV (fig. 34). A dipping sea bed (first reflector) distorts underlying flat lying interfaces. The higher the velocity contrast at the sea rock interface the greater the distortion.

The internal geometry of the younger units seen on Profile XI can be well defined. They have hummocky top surfaces and rest on uneven surfaces, defining series of little cushions. These units give a reflection free seismic response except at their base where they are very finely laminated. They build up in a fashion similar to the units, believed to be of glacial origin, defined in Zone B. They are complex accumulations of very irregular sediment patches.

**Continental Slope:**

Some idea of the continental slope facies is given by Profile 80,
Figure 34: Synthetic Seismogram IV: Time Anomalies

Flat lying horizons underlying a steeply dipping sea floor show a small apparent dip on the time section.
fig. 19 and VIII, fig. 14. The transition from shelf to slope is not too clear because of the large vertical exageration. It seems, nevertheless, that the top glacial units prograde seawards and become extremely disorganized at the shelf break. Internal deformation such as small faults, erratic reflections an internal deformation can be observed (Profile VIII, fig. 14).

Summary:

Two seismic stratigraphic sequences (A and B) were defined for the eastern Weddell Sea continental shelf and their different features discussed. They represent two very different sedimentary settings, although they might both be associated with glaciation. They are underlain by an acoustic basement which slopes up toward the East Antarctic continent and eventually outcrops at the sea floor. Although this basement seems to have physical properties very different from those of overlying sediments (i.e. velocity of propagation of seismic waves and resistance to erosion), it remains to be shown that it is the true basement.

Overlying unconformably Sequences B and the outcrops of acoustic basement, on the flank of Crary Trough, are patches of hummocky sediments. According to the core data available (Anderson, 1980) these sediments should mostly be basal tills. Any interpretation remains very speculative since the core data was not often directly correlatable to the seismic profiles.

The most important features defined by the seismic data are the two unconformities which bound respectively, the tops of Sequence A (A) and Sequence B (B). These unconformities should provide elements
of information regarding the chronology of the development of the
Weddell Sea's eastern margin. Establishment of this chronology can be
achieved through comparison with worldwide sea level curves (Vail et al,
1977) and through analogy with the Ross Sea, where there is better control
on the timing of events in the Antarctic Region (Balshaw, 1981).
CHAPTER III

TIMING OF EVENTS ON THE EASTERN MARGIN OF
THE WEDDELL SEA, AS SEEN THROUGH THE SEISMIC DATA
Significance of the Oldest Erosional Event Recognized in the Seismic Data

The angular unconformity (fig. 9) which separates Sequences A and B, is of regional extent over the portion of the continental shelf lying east of Crary Trough. It is recognized as far north as offshore Princess Martha Coast (Zone B) and is well defined all over Zone A. This erosional surface seems mostly planar along the profiles covering Zone A. In Zone B (Lagenbanken), seismic coverage is more complete than in Zone A and tighter control in positioning the profiles can be accomplished. This allows for the construction of an isotime map to the unconformity surface (fig. 35). This two-dimensional picture of the surface reveals more irregularities than can been seen in Zone A. A major low trend, running roughly northwest-southeast, is observed (fig. 35).

Offshore of the eastern coast of the Weddell Sea, several canyons transecting the continental slope are outlined by the bathymetry. These canyons are no longer active for they do not show any topographic expression on the shelf. Their age is unknown, but it is possible that they are associated with the erosional surface (Φ). The low trend recognized in Figure 35 could lead to one of these canyons. More tangible support for this interpretation can be found on Profile III (fig. 17). The profile transects the continental shelf, northeast of Crary Trough, in a south-southwest direction beginning at the head of a major canyon (fig. 10). Interpretation of the data (fig. 17) had
Figure 35: Isotime map of the angular unconformity surface, Lagenbanken area (Zone B).
Isotime curves at 20ms intervals.
defined two depositional units (subunits A and B) separated by a slight angular unconformity, which is assumed to be correlative with \( \varphi \). Sub-unit B onlaps onto the Prograding Unit which is represented by subunit A and the unconformity surface dips offshore. If we compare the depths below sea level of \( \varphi \) on Profile 80 (fig. 19), immediately to the west of the canyon (fig. 10) and inside the canyon Profile III (fig. 17) a difference of 200 m (250 ms two-way time) is registered. This difference implies that the canyon was already outlined immediately after the occurrence of the erosional event associated with \( \varphi \). The presence of a later infill of the canyon, represented by subunit B onlapping onto the Prograding Unit, outlines a depositional pattern typically associated with a major fluctuation of sea level (Brown and Fisher, 1980).

The unconformity \( \varphi \) defines a rather planar surface cut by low trends, possibly associated with the canyons seen to transect the continental slope, along the eastern margin of the Weddell Sea. These characteristics favor a wave cut origin for the erosional surface, linking it to a major drop of eustatic sea level. A tentative age can be therefore set for the formation of these erosional features, by comparison with a standard, or world wide sea level curve, Figure 36 (Vail et al, 1977). The deposits of the Prograding Unit, which are truncated by the unconformity \( \varphi \), have been interpreted as nonglacial sediments. However, they do not preclude the presence of ice on the mainland, which is East Antarctica. In view of this, an age of 29 M.Y. B.P. is favored for the unconformity, which relates it to the last Oligocene major sea level drop (fig. 36). This lowering of sea level was also possibly recorded in the Ross Sea (Balshaw, 1980). The difference in thickness between the glacial and glacial marine sediments
Figure 36: World wide sea level curve, Jurassic to Quaternary (from P. Vail et al, 1977)
(200-300 m) which overlies the unconformity \( \psi \) and the deposits (1200 m) which span the assumed same time interval in the Ross Sea, indicate radically different depositional environments. There are two possible interpretations of this difference. One is that the thick Ross Sea section was sampled (Hayes et al, 1975) in the center of the continental shelf, in an area mostly under the influence of the West Antarctic Ice Sheet. The area of study in turn is marginal to the broad continental shelf of the Weddell Sea's West Antarctic sector, and could be primarily affected by fluctuations of the East Antarctic Ice Sheet. Ice sheets on the eastern and western sectors are of different age and nature (Balshaw, 1980). At present the proportion of debris transported to continental shelf by the eastern ice sheet is minimal (Anderson in press). A second interpretation for the differences in thickness of the late tertiary section of the Ross and of the Eastern Weddell Seas resides in the presence of Crary Trough in the area under study. The trough is more than 1000 m deep and when it was formed, a large part of the sedimentary section covering this part of the continental shelf could have been eroded off, considering erosion actually affected even the basement terranes.

Unconformity \( \psi \), a major glacial feature?

In the Ross sea maximum glaciation occurred in the late Miocene and early Pliocene (Balshaw, 1981). Very severe glacial conditions prevailed. A grounded ice sheet reached the continental shelf break, eroding part of the glacial marine section and depositing basal tills. Such condition, similarly, could have existed in the Weddell Sea region. Crary Trough is a feature typical of glacial environments (Johnson and Vanney, 1974): a deep trough parallel to the coast, transecting the
continental shelf. The presence of basal tills outcropping inside the trough, favor a grounding ice sheet as an erosional agent. The ice sheet would have to be in excess of 1000 m thick, nearly as far as the continental shelf break (fig. 6). Referring to the Ross Sea chronology, a tentative age for the formation of Crary Trough would therefore be Pliocene. It is interesting to note that sediments of the western sector of the Weddell continental shelf, seen on Profile XI (fig. 15), are truncated by the unconformity which defines Crary Trough, as are truncated the thick glacial marine deposits of the Ross Sea by this Pliocene event.

The unconformity ψ which was defined as the top of Sequence B, often defines the present sea floor, when it is not overlain by thin patches of sediments. In the area of Profile III it could represent the event which truncated both subunits A and B, and which imprinted on the present sea floor a dip towards the continent rather than offshore. In the area of Crary Trough, the unconformity ψ is the unconformity defining the trough, which was given an age of Pliocene. This would set a maximum time, for the deposition of Unit I and Unit II. If these deposits have been interpreted as glacial and glacial marine, they are probably in part contemporaneous of the erosional event, and as a whole define fluctuations of both the East and West Antarctic ice sheet.

Study of Canyon East of Crary Trough:

The canyon at the head of which Profile III was shot (fig. 10), was also individualized, by the erosional event associated with ψ, as can be observed from the present bathymetry of the shelf in this area.
Other canyons do not have any topographic expression on the shelf. To this singularity can be added some interesting features associated with this canyon which can be observed on the seismic data.

The canyon was carved in deposits of Sequence A (Prograding Unit, Profile III) but also in what are thought to be the younger deposits of Sequence B. The former form a large mound on the western flank of the canyon identified on Profile 80 (1000, fig. 19). The interpretation is only tentative for two reasons, the first being that due to a change in the course of the ship (fig. 10) the profile runs along strike to the Prograding Unit, and thus the angular unconformity has disappeared. The second reason is that the strata which form the mound at shot point 1000 on Profile 80 have bedding planes, which are parallel to the sea floor, rather than flat lying like typical deposits of Sequence B. At shot point 1000 on Profile 80, the sea bottom is extremely rugged and it has been hypothesized that the attitude of the bedding planes was an artifact created by the seismic method and known as time anomalies (Synthetic Seismogram IV, fig. 34). It is, therefore, possible to consider the deposits which form the mound structure seen on Profile 80 (fig. 19) as flat lying.

The occurrence of bedding planes within the glacial units, in such a nearshore position is particularly significant. Oceanographic measurements show the eastern Weddell Sea to have very strong bottom currents (Anderson, 1972) which flow parallel to the coast line and are derived from the circum-polar current. These strong currents contact the shelf break and encounter masses of Antarctic bottom waters which flow sluggishly down the continental slope just east of Crary Trough, in the general area where the mound of glacial sediments sits.
Thus water movements are fairly complex, establishing conditions unique to this zone which often and abruptly change laterally. It is possible that Units I and II might have been locally reworked during deposition by marine currents, which would give them a different seismic character.

An infilled channel, associated with unconformity $\gamma$, is located to the east of the mound (1300 - 1400, fig. 19). On the canyon flank, the sediments of Sequence B are cut by an erosional surface. The reflections become laterally erratic across this surface and fill a wedge shaped body. This feature could have been eroded and later infilled by a subglacial stream. At any rate the channel truncates deposits of Sequence B, and must be younger than the high glacial activity responsible for their deposition.

Conclusions:

The data set represented by the two seismic surveys contains many significant features. A typical seismic stratigraphic column of the eastern continental shelf of the Weddell Sea, can be constructed (fig. 9). Whether it actually reaches the true basement, will be tested in the next chapter.

Intersedimentary features such as unconformities, whether regional or localized, suggest a tentative timing of events and dating of deposits.

Unfortunately, lack of geologic data considerably decreases the confidence in the interpretation of the seismic data. The purely descriptive stage is objective, but the further stage of direct determination of the lithology and stratigraphy can only be considered as tentative.
CHAPTER IV

MAGNETIC AND GRAVITY DATA
MAGNETIC AND GRAVITY DATA

During both their 1977 and 1979 surveys, the Norwegians gathered magnetic data along the seismic tracks. Only sketched versions of the 1979 profiles were made available to us and are the material used in the following interpretation (fig. 37). The gravity data, however, is part of an earlier study (Behrendt, 1962/1966) of the Filchner Ice Shelf. It does not extend into the marine environment. Both these sets of data ideally complement the shallow seismic survey since they give information on the features beneath the sedimentary cover and in particular will allow the determination of the nature and depth of the basement.

Magnetic Measurements:

The total field data was collected continuously with a marine proton magnetometer which had a 1 gamma resolution and was recorded on a paper chart. The display available for this study was of a very small scale: 2 cm = 5 nautical miles = 1 hour, 2 cm = 1000 γ. Its sketchy nature added to the absolute error in determination of the depth to the magnetic source.

Peter's Method:

Peter's "slope method" is a rule of thumb for determining the depth to the magnetic source. It assumes a vertical field measurement and the body generating the anomaly to be a dyke. The first hypothesis can, in general, be accepted in high latitudes where the total field and vertical field are approximately the same.
Figure 37: Values of the magnetic field in the Weddell Sea region and location of magnetic profiles used in this study.
As for the second assumption, an infinite third dimension in the direction perpendicular to the profile, it is quite logical when only two dimensional measurements are available.

The principle of the method, illustrated in Figure 38, is purely geometrical. The maximum slope of the anomaly curve, at the inflexion point, is determined. Two lines with a slope half of this value are then traced tangentially to the anomaly curve at point A and A'. The main parameter(s) is the distance separating these two points, which in the above case is proportional to the depth of burial (h):

\[ s = \alpha h \]

\( \alpha \) ranges between \( \alpha = 1.2 \) (m/\( h \) = 0) and \( \alpha = 2.0 \) (m/\( h \) = \( \infty \)) where \( m \) is the width of the dyke. For an average anomaly the following expression is considered \( s = 1.6 \) h.

Nevertheless, since the basement in the area of survey is rather shallow, the ratio m/\( h \) was considered to be large enough and \( \alpha \) was given a value of 2.0.

**Error Margins:**

In the best of cases the depths determined by the method are accurate to about \( \pm 10\% \) (Peter, 1949). This figure is certainly a minimum considering the quality of the data.

The profiles were enlarged 100%, and the measurements of \( s \) was done on millimetric paper, allowing for an error of \( \pm 1 \) mm for a high frequency anomaly and of \( \pm 3 \) mm for a low frequency anomaly. It represents errors of \( \pm 141 \) m and \( \pm 423 \) m, respectively, in the computed depth of burial. This is within a 10% to 15% margin, considering that depths determined are in the 1000 m and 3000 m range. However, when
Figure 38: Peter's Method, applied to dyke anomaly. (from Peter, 1949)
the basement outcrops, measurements will be absurd since the error margin becomes too large. It is interesting to note that the above figures are dependent on the velocity of the ship. The parameter is not accurately known and has been averaged to 5 knots, which in most cases is the minimum value.

The regional, which was subtracted from the field, (fig. 12, 14 and 19) was chosen in order to approximate as closely as possible the original shape of the dyke anomaly (fig. 38) i.e. symmetrical about the vertical axis. In most cases, the symmetry is not obtained, and two different depths can be determined. The final value was taken to be their average ($\tilde{h}_5$).

**Interpretation of the Data:**

A summary of the results is given in Table III. The horizontal scale of the profiles, which originally was in hours was transformed into meters assuming a velocity of advancement of 9 km/h (5 knots). The combination with the seismic results was done assuming velocities of propagation of longitudinal waves to be 1450 m/s in this water layer and 1600 m/s in the sediment cover (cf. Appendix B). When the acoustic basement is not within the depth of penetration of the system, it is considered to be in excess of 400 m.

Magnetic profiles along Lines 76 and 83 were almost featureless and, consequently, were only considered those along Lines 77, 80 and 81 (figs. 12, 14 and 19). There is no ready explanation for the appearance of profile 76 since Line 76 intersects both 77 and 80 (fig. 10). Its different orientation could certainly be a reason but is not sufficient.
### Table III: Interpretation of magnetic survey

<table>
<thead>
<tr>
<th>ANOMALY NUMBER</th>
<th>5.05 DEPTH TO MAGNETIC SOURCE</th>
<th>WATER DEPTH (FROM SEISMIC DATA)</th>
<th>SEDIMENT COVER OVERLYING MAGNETIC SOURCE</th>
<th>ONE-WAY TRAVEL TIME TO ACOUSTIC BASEMENT</th>
<th>SEDIMENT COVER OVERLYING ACOUSTIC BASEMENT</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>LINE 77</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>1260m</td>
<td>620m</td>
<td>640m</td>
<td>0m</td>
<td>0m</td>
</tr>
<tr>
<td>7</td>
<td>870m</td>
<td>760m</td>
<td>1130m</td>
<td>71.3m</td>
<td>67.5m</td>
</tr>
<tr>
<td>6</td>
<td>1010m</td>
<td>760m</td>
<td>2530m</td>
<td>0.3m</td>
<td>133m</td>
</tr>
<tr>
<td>5</td>
<td>870m</td>
<td>710m</td>
<td>1630m</td>
<td>1230m</td>
<td>210m</td>
</tr>
<tr>
<td><strong>Low F. anomaly</strong></td>
<td>4725m</td>
<td>700m</td>
<td>4073m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td><strong>5.1</strong></td>
<td>530m</td>
<td>700m</td>
<td>0m</td>
<td>143m</td>
<td>227m</td>
</tr>
<tr>
<td><strong>5.2</strong></td>
<td>960m</td>
<td>730m</td>
<td>1180m</td>
<td>115m</td>
<td>144m</td>
</tr>
<tr>
<td><strong>5.3</strong></td>
<td>915m</td>
<td>740m</td>
<td>1130m</td>
<td>1230m</td>
<td>2000m</td>
</tr>
<tr>
<td><strong>5.4</strong></td>
<td>840m</td>
<td>760m</td>
<td>840m</td>
<td>1240m</td>
<td>1800m</td>
</tr>
<tr>
<td><strong>5.5</strong></td>
<td>1000m</td>
<td>755m</td>
<td>2450m</td>
<td>1770m</td>
<td>280m</td>
</tr>
<tr>
<td>4</td>
<td>3000m</td>
<td>700m</td>
<td>2300m</td>
<td><strong>not seen</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td>3</td>
<td>1510m</td>
<td>710m</td>
<td>1280m</td>
<td><strong>not seen (4050m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td>2</td>
<td>2812m</td>
<td>650m</td>
<td>2180m</td>
<td>6000m</td>
<td>6000</td>
</tr>
<tr>
<td>1</td>
<td>3763m</td>
<td>420m</td>
<td>2600m</td>
<td><strong>not seen (4050m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td><strong>LINE 81</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>6.1</strong></td>
<td>610m</td>
<td>340m</td>
<td>2700m</td>
<td><strong>not seen</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td><strong>6.2</strong></td>
<td>751m</td>
<td>340m</td>
<td>10m</td>
<td>10m</td>
<td>10m</td>
</tr>
<tr>
<td>3</td>
<td>770m</td>
<td>400m</td>
<td>3700m</td>
<td>6000m</td>
<td>6000</td>
</tr>
<tr>
<td><strong>2.1</strong></td>
<td>2000m</td>
<td>400m</td>
<td>9300m</td>
<td><strong>not seen</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td><strong>2.2</strong></td>
<td>745m</td>
<td>360m</td>
<td>3850m</td>
<td><strong>not seen</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td>1</td>
<td>3431m</td>
<td>450m</td>
<td>2873m</td>
<td>1500m</td>
<td>2500m</td>
</tr>
<tr>
<td><strong>LINE 80</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>2643m</td>
<td>500m</td>
<td>2130m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td>2</td>
<td>1226m</td>
<td>450m</td>
<td>8300m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td>3</td>
<td>3600m</td>
<td>435m</td>
<td>3120m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td><strong>6.1</strong></td>
<td>1547m</td>
<td>460m</td>
<td>1143m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td><strong>6.2</strong></td>
<td>2683m</td>
<td>2300m</td>
<td>2150m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td>5</td>
<td>1687m</td>
<td>570m</td>
<td>1167m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td><strong>6.1</strong></td>
<td>1744m</td>
<td>580m</td>
<td>1140m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td><strong>6.2</strong></td>
<td>1800m</td>
<td>520m</td>
<td>1270m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
<tr>
<td><strong>LINE 76</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>3809m</td>
<td>1294m</td>
<td>2512m</td>
<td><strong>not seen (4030m)</strong></td>
<td><strong>not seen</strong></td>
</tr>
</tbody>
</table>
As for Line 83 it is in a completely different area.

Most of the anomalies have amplitudes of 200 \( \gamma \) and more, and widths (frequency) of two different orders of magnitude.

- **The Acoustic Basement:**

  The higher frequencies or narrower anomalies seem to be associated with the acoustic basement. Typically they occur when the sediment cover is less than 400 m thick (within the seismic penetration) and they parallel quite strikingly the topography of the basement (anomaly 5' on Line 77, fig. 12). The fit of depths to the magnetic source \( (h_5) \) and to the acoustic basement \( (h_a) \) are, within the margin of error, quite comparable (Table III).

  Magnetic anomalies are generally associated with the presence of igneous rocks which have high enough susceptibilities to create a deviation in the earth's magnetic field. Thus, the acoustic basement identified in the seismic represents the actual basement and can be correlated with the terranes found on the East Antarctic mainland.

  The origin and the type of rock can only be speculated upon. The latest igneous activity recorded in the eastern Weddell Sea dates back to the Middle Jurassic. It is represented by the Dufek Intrusion (Pensacola Mountains) and dolerite stocks which are distributed along the Transantarctic Mountains. Offshore Dronning Maud Land, the Maud Rise (65 S, 30 E) is thought to have a volcanic origin, but no indication as to its age is available (Hinz, 1978). Cenozoic volcanism, was possibly recognized in the seismic data from the Lagenbanken survey on the eastern end of Profile XVIII (fig. 21b). In this area, a volcanic plug
apparently intrudes into quite recent strata.

The surface of the acoustic basement is extremely rugged, indicating an erosional origin which sometimes is enhanced by possible block faulting (fig. 12, Line 77). Two origins are possible for the emplacement of these igneous rocks. A discontinuous source, in otherwise nonmagnetic terranes i.e. stocks or dykes (model assumed to compute depths) which could be related to the block faulting episode, or else a widespread layer of volcanic rocks or igneous basement. This origin is favored when correlating the area with the West Antarctic sector, but in view of the seismic data, seems improbable.

- **Deeper Anomalies:**

  A deeper layer of magnetic sources can be detected at depths of 3000 to 4000 m. They give rise to wider and larger anomalies. There are records of pre-Jurassic volcanic activity on the continent (Kirwan Volcanics, Kirwan Encarpment) which could be related to these anomalies. The magnetic basement indicates an Eastern Antarctic influence for the continental shelf in this area, therefore, a continental equivalent for these anomalies should also be encountered in East Antarctica's history.

**Gravity Data and Interpretation:**

The surveyed area lies beneath the Filchner Ice Shelf and onshore, in the Pensacola Mountains. It represents the extension of the East/West Antarctic boundary, which was covered by the magnetic and seismic data. A composite Bouguer anomaly map (Behrendt, 1962 and 1966) is shown in Figure 39. Its main features are the large 60 Mgal anomaly, which
Figure 39: Bouguer Anomaly Map, Weddell Sea area.
(from: Behrendt, 1962 and 1966)
follows Crary Trough, and the steep gradient (-60 Mgal to +40 Mgal), which outlines the western flank of the Pensacola Mountains. Computa-
tions indicate that the trough is not completely isostatically compen-
sated and confirms the hypothesis that it is a relatively recent
erosional feature (Behrendt, 1962).

Theoretical crustal models were tested (Behrendt, 1974) to account
for the steep east-west gradient farther south (fig. 39). A step-like
discontinuity in the crust or a sharp transition with thinning of
crustal layer from east to west are both compatible with the measured
Bouguer anomaly profile. Thus, the so called boundary between East and
West Antarctica could be as deeply seated as the crust-mantle interface.
It is likely that the zone might extend farther north and parallel the
occurence of mafic stocks along the Transantarctic Mountains. The Dufek
Intrusion, outcrops in the Forrestal and Dufek Massifs (fig. 2) where it
is approximately 8 km thick (Behrendt, 1974), but also extends farther
north.

The surveys inland of the Filchner and Ross Ice Shelves are sparse
and it is not possible to construct either a gravity or a magnetic map
covering the whole area. Nevertheless, there seem to be some similar-
ities between the Ross and Weddell Seas.

A large positive anomaly extends parallel to the front of the
Transantarctic Mountains in the Ross Sea and Ross Ice Shelf (Benett,
1964; Hayes and Davey, 1975). Crustal modeling across this anomaly was
undertaken where seismic data control was sufficient. Although the
authors favor a shallow source of greater density, crustal thinning
similar to the one found in the Pensacola Mountain area could also fit
the model. Intrabasement magnetic sources are also encountered in this
Figure 40: Plate reconstructions of the Southern Continents.

1, 2, 3, 4, 5, 8 and 9: Pre-Mesozoic reconstructions of Gondwanaland.

6 and 7: Cretaceous to early Cenozoic reconstructions of the Southern Ocean.

10 and 11: Pre-Cenozoic reconstruction of the Scotia Arc Region.
area. However, Cenozoic igneous activity is widespread and it is more difficult to differentiate this activity from older stocks such as the Ferrar Dolerites which are found onshore and are of the same age as the Dufek Intrusion.

There could, thus, be a discontinuity in the crust that would define the boundary between East and West Antarctica and would parallel the Transantarctic front from Victoria Land to Queen Maud Land. Whether this discontinuity defines an ancient rift zone or graben is speculative. It seems that the Ross-Weddell depression could have been distinct in early Cenozoic time or prior to it. The existence of a major waterway linking the Pacific and Atlantic Oceans before the opening of the Drake Passage has been suggested (Webb et al, 1974) although there is little concrete evidence.

The magmatic trend which follows the depression extends into South Africa where it does not seem to be associated to such a feature. As for the high gravity anomaly, characteristic of graben structures, it is only unquestionably present in the Ross Sea area. In the Weddell Sea, the gravity high would interfere with the Crary Trough anomaly.

At any rate, in the area covered by this study, a major discontinuity between the eastern and western margins of the Weddell Sea in the area of the Crary Trough seems strongly suggested. The feature could be a rift zone or even a major transform fault, which was active during the second half of the Mesozoic (La Breque, 1980). The eventual displacement of the Ellsworth Mountains block, the morphology of the narrow continental shelf off Dronning Maud Land, the presence of the Crary Trough itself, which is likely to have followed an earlier structural feature are evidence for a rather complex area.
CONCLUSION:

A PICTURE OF THE EASTERN WEDDELL SEA CONTINENTAL SHELF AS SEEN THROUGH GEOPHYSICAL DATA
CONCLUSION

A PICTURE OF THE EASTERN WEDDELL SEA CONTINENTAL SHELF AS SEEN THROUGH GEOPHYSICAL DATA

Because of the peculiarity of the area studied, it was possible through a shallow (200 - 300 m penetration) survey to identify a fairly thick section of sediments (1200 m of minimum cumulative thickness). The presence of a deep erosional feature, the Crary Trough, allows the outcropping of older deposits and even of the basement near the sea floor. The nature of the seismic units defined seems somewhat different from what is known to exist under the Ross Sea continental shelf. This has been attributed to a difference in sedimentary environments. Nevertheless, it is very possible for the western part of Weddell Sea shelf to be more comparable to the Ross Sea.

The acoustic basement was identified as the magnetic basement of the basin, and further as the actual basement of the basin margin. The basement outcrops at the sea floor in the deepest part of Crary Trough and slopes upwards to the mainland, favoring an East Antarctic origin. The high magnetic susceptibility, of the basement indicates its igneous or volcanic origin. Relating this origin to the onshore geology of the Eastern Weddell Sea sector gives a Mesozoic (Middle Jurassic) age for the basement terranes.

The sedimentary section which overlies the basement can be divided into two major seismic stratigraphic sequences separated by a widespread angular unconformity (fig. 9). This unconformity is significant for an approach to the timing of events in the study area. It has a
general planar appearance, which indicates a wave cut origin associated with a major drop in sea level. After comparison with the curve of worldwide eustatic sea level changes established by Vail et al (1977) and information gathered in the Ross Sea (Balshaw, 1981), a tentative age of late Oligocene (29 M.Y.B.P.) was established for it. The lower Sequence (A) of tilted strata would thus not have a direct glacial origin. However, glaciation on the continent is not excluded. The beds dip systematically offshore and form a 900 m thick prograding wedge. They represent marginal deposits along a passive margin. Balshaw (1981) associates the sea level drop at 29 M.Y.B.P. with formation of the East Antarctic Ice Sheet, but establishes the presence of valley glaciers in the high reliefs of the Transantarctic Mountains prior to this.

Deposits of the upper Sequence (B) are mostly flat lying but variable in appearance. They have been hypothesized to be glacial and glacial marine in origin, and each seismic facies was tentatively associated with glacial and glacial marine facies as established by Anderson et al (1980) for the Antarctic continental shelf. Reflection free deposits were interpreted as basal till. Apparently well bedded deposits are thought to represent mostly compound glacial marine sediments. Both facies are intricately associated. The basal tills generally grade away from the continent into glacial marine strata, defining the possible grounding line of the ice sheet. In the northern part of the study area (Zone B: Lagenbanken) the seismic character of the glacial units is slightly different from the ones defined in Zones A and C. Different types of glacial and marine environments between both zones can be suggested.

The top of the deposits of Sequence B (fig. 9) defines a second
unconformity ($\Psi$) which locally coincides with the sea floor. The thin veneer of deposits overlying this unconformity are definitely of glacial origin, mainly basal tills as established by the piston core data which extensively sampled the sea floor in this area (Anderson et al, 1980). The younger unconformity was tentatively given a Pliocene age, corresponding to a major ice advance recorded in the Ross Sea. Crary Trough, a significant morphological feature of the study area,is thought to have been carved by glacial erosion. The presence of basal tills along its flanks (Anderson et al, 1981) strongly supports this origin. The erosional event which formed the trough, is interpreted to be roughly contemporaneous with the peak in glacial activity (Pliocene). As much as 800 m of sedimentary section which can be seen to outcrop at the sea floor on the western flank of the trough (Zone C, Line XI) could have been eroded away. It is suggested that the latter deposits which have a slight western apparent dip, could either be the equivalent of the section found in the Ross Sea or some much older (?) deposits as suggested by the eolian sands encountered in core G8 (fig. 4)(Anderson, 1972).

Comparison of the Ross Sea shelf, DSDP Leg 28 cores, with the eastern coast of the Weddell Sea suggests extremely different sedimentary regimes. Sediments in the Ross Sea have a very definite West Antarctic origin (Balshaw, 1981). However, it is very likely that the area of study in the Weddell Sea is mostly under the influence of the East Antarctic Ice Sheet, a dry-base ice sheet (Crary and Ahmad, 1961), which has much poorer sedimentary output than the wet-base ice sheet of West Antarctica. This could explain in part, the large difference in the thickness (200 m to 1200 m) of post late Oligocene (29 M.Y.B.P.) sedimentary section between
the area of study and the Ross Sea, and eventually between the area of study and the western sector of the Weddell Sea continental shelf.

The tectonic framework of the study area was difficult to approach through the seismic and magnetic data. An overview of the literature (summarized in Appendix A) suggests an extremely complex area located at the transition between very different terranes of East and West Antarctica. Gravity data over the Pensacola Mountains and Filchner Ice Shelf (fig. 39) show a very deeply seated discontinuity between both sectors. However, it is unlikely that this discontinuity can be verified because the data used in this study reflected only the basin fill.

The deposits of Sequence A suggest a passive margin environment, with its borderland subjected only to epeirogeneic movements. As for the presence of Jurassic (?) igneous rocks within the basement, they could be associated with a major block faulting episode recorded all along the Transantarctic front and related or not to the opening of the Weddell Sea.

This study has, thus, mostly emphasized the seismic stratigraphic cross section of the eastern margin of the Weddell Sea. The formations defined are thought to span a long interval of time with major hiatuses. The sedimentary environment seems to be extremely different from what is known in the Ross Sea, and older pre-glacial deposits could be expected with depth, giving further information on the formation and evolution of the Weddell Basin margin.
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REFERENCES


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APPENDIX A

HISTORICAL GEOLOGY OF THE WEDDELL SEA SECTOR

The Continental Surveys which represent the major part of what is known of the Weddell Sea sector were made in the scattered mountain ranges which outcrop through the locally over 3000 m thick ice cover (Veveers et al, 1971; Elliott, 1972; Williams et al, 1972; Dalziel and Elliott, 1973; Craddock and Webers, 1977; Kellogg and Reynolds, 1977; Quilty, 1977; Elliott et al, 1978; Kennett, 1978; Kellogg, 1979; Rowley, 1979; Grew and Marton, 1980; Ford and Kistler, 1980). Figures I and II give the locations of these outcrops in Antarctica and, more specifically, in the area of interest. They give a general idea of the geologic record of both East and West Antarctica and are the sole geological evidence when attempting a geodynamic reconstruction of the area. Correlations from one site to the other can only be tentative, as there are no "linking" terranes. Although it may seem paradoxical with such a poor record, most of the prior geological research in the area emphasized a global tectonic reconstruction of Gondwanaland. The Scotia Arc, West Antarctica, and Weddell and Ross Seas are critical in such schemes. Few if any attempts have been completely successful in integrating all available data.

The old Precambrian polymetamorphic shield outcrops east of the Kirwan Escarpment in Queen Maud Land (fig. 11) and can be found along the Atlantic margin of the continent (Grikurov et al, 1972; Krylov, 1972). Similar terranes, intruded by early Paleozoic igneous rocks, lie farther south, inland from the Transantarctic Mountains.

To the northwest of the escarpment, separated from the basement
Figure I: Subglacial map of Antarctica
(from Avzyuk et al, 1964)

1. Graham Land
2. Palmer Land
3. Ellsworth Land
4. Marie Byrd Land
5. Byrd Subglacial Basin
6. Crary Trough
7. Queen Maud Land
Figure II: Mountain Ranges of the Circum-Weddell Sea Region
(from Dalziel and Elliott, 1973)
complex by the Penck-Jutul Rift, lies the Precambrian undeformed strata of the Ritscher Supergroup (Neethling, 1972a). They are thought to represent an isolated, block-faulted, remnant of the platform cover. The sediments, derived from a westerly source, are indicative of a near-shore environment in an epicontinental basin. This stable region of Precambrian age may extend between longitudes 1\textdegree W and 34\textdegree W (Littlewood Nunataks) as based on geochronologic data (Aughenbaugh et al, 1965). South and west of the escarpment, around the Filchner and Ronne Ice Shelves, one begins to find lower Paleozoic and younger strata which have recorded one or several of what seems generalized deformational events:

(a) Beardmore orogeny (early Paleozoic)
(b) Ross orogeny (probably Upper Cambrian)
(c) Gondwanian orogeny and its local equivalents: Weddell and Ellsworth orogenies (Triassic-Early Jurassic?)
(d) Andean orogeny (Lower to Middle Cretaceous)

A tentative correlation, made by Dalziel and Elliott (1973) is summarized in Table I with some changes due to later investigations (Ford, 1977; Splettstoesser and Weber, 1980; Grew and Marton, 1980).

Other Precambrian rocks are found in the Pensacola Mountains (fig. II) and can be correlative with the upper group of the Ritscher Supergroup (Schmidt et al, 1965; Neethling P. R., 1972a). The lithology and bedding of the 10,000 m thick Patuxent Formation indicate turbidite deposition associated with both acidic (1210 M.Y.B.P) and basic volcanics. Ford (1977) correlates the Patuxent with the Turnpike Bluff Group which outcrops farther north in the Shackleton Range (fig. 11) and postulates the existence of a major seaway along the margin of the craton in
<table>
<thead>
<tr>
<th>AGE</th>
<th>WESTERN QUEEN MAUD LAND</th>
<th>THORN AND SHACKLETON MOUNTAINS</th>
<th>PENASCOLA MOUNTAINS</th>
<th>ELSWORTH MOUNTAINS</th>
<th>ANTARCTIC PENINSULA</th>
<th>SOUTH SHETLAND ISLAND</th>
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<td></td>
<td>KINNARI ESCARPMENT</td>
<td>WHICHAMER FORMATION</td>
<td>JURASSIC INTRUSIONS</td>
<td>ANDESITIC VOLCANIC SERIES (3000m)</td>
<td>LAFAYETTE FORMATION</td>
<td>WILLIAMS POINT BEDS</td>
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<td>JURASSIC</td>
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<td>Dufek Intrusion Gebrrof body</td>
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<td>Stilstones, andesites, alkalic-alkaline volcanic</td>
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<tr>
<td>OAMONIAN</td>
<td>KINNARI VOLCANICS (Basalt lavas w/ minor sediment intercalations)</td>
<td>THORN FORMATION</td>
<td>SANDSTONES, Siltstones, Shales</td>
<td>POLARSTAR Fm. (1500m interbedded slate, argillite)</td>
<td>TRINITY PENINSULA Fm.</td>
<td>EWERS BLUFF FORMATION</td>
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<tr>
<td>OAMONIAN</td>
<td></td>
<td></td>
<td>Silicic Iabammonic interbeds</td>
<td></td>
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<tr>
<td>PENIAN</td>
<td>AMELING Fm. (1500m, sandstones w/shales and conglomerate interbeds)</td>
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<tr>
<td>DEVORIAN</td>
<td>URHJELL GROUP (?)</td>
<td>RIVER SANDSTONE (carbonaceous interbeds)</td>
<td>NEPTUNE GROUP</td>
<td>CRASKE QUARTZITE (2200m)</td>
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<tr>
<td>ROSS Groups (1850m, quartzite, conglomerates and minor mudstone)</td>
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<td></td>
<td></td>
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<td>CRASKE QUARTZITE (2200m)</td>
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<tr>
<td>CAMBRIAN</td>
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<td>BLALOCK BEDS (?)</td>
<td>WIELS FORMATION (Agglomeratic)</td>
<td>MINARET Fm. (0-900m, Marble, minor conglomerates)</td>
<td>CAMBACAPITA Fm.</td>
<td>BASEMENT COMPLEX (?)</td>
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<td>BEARMORE</td>
<td></td>
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<td>GEMINI GROUP</td>
<td>HERITAGE GROUP (700m, Phyllite, limestone, tuff, flows, gray wackes)</td>
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<td>GROMER</td>
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<td>PRE-CAMBRIAN</td>
<td>METAMORPHIC SUPERGROUP (20000m)</td>
<td>RITZSCHER GROUP</td>
<td>LITTLEWOOD VOLCANICS (Acidic)</td>
<td>PATIENT Fm.</td>
<td>TURNIP CORNER BLUFF Fm. (Quartzite, slates)</td>
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<td>Boulder bed, Qtz. Arroite, Arroite</td>
<td>SHACKLETON METAMORPHICS</td>
<td></td>
<td>1000m interbedded shale, turbidites, acidic and basic volcanics</td>
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<td></td>
<td>METAMORPHIC GROUP</td>
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Precambrian times. The seaway might have been 3000 km long, the entire length of the Transantarctic Mountains. The Turnpike Bluff Group is chiefly quartzitic with locally abundant slate but lacks the volcanics and turbidites seen in the Patuxent strata. If the correlation stands, it might indicate a shoreward transition along the cratonic margin (?), as the chemical composition of the Patuxent sandstones seem to favor an "Atlantic" type of margin. Both formations were strongly folded with trends from north-northeast to northeast locally metamorphosed to the greenschist facies. This uppermost Precambrian episode probably corresponds to the Beardmore orogeny recorded elsewhere in the Transantarctic Mountains.

Grew (1980) reports an age of 500-600 M.Y.B.P. (U-Pb and Rb-Sr) for the high grade metamorphic Shackleton complex (amphibolite facies) at Mt. Provender. There is no known contact between these rocks and the Turnpike Bluff Group; whether or not they relate to the same event is still in question.

Overlying unconformably the Shackleton metamorphics are the Blailock Beds. They have been assigned a tentative age by Clarkson et al (1979) on the occurrence of Middle-Cambrian fauna in erratics that are thought to be derived from them.

In the Pensacola Mountains, the Cambrian is represented by the Nelson Limestone, volcanioclastics of the Gambacorta Formation, and the dominantly argillaceous Wiens Formation. These strata were deformed during an event which correlates with the Ross orogeny, and were later intruded by Ordovician granitic bodies. Whether this orogeny was felt in the Shackleton Range is problematic. Blailock Beds are found in homoclines dipping toward the southwest which can either be interpreted
as fold limbs (Clarkson, 1972), as tilted fault blocks, or as depositional talus (Ford and Kistler, 1980). The absence of granitic plutonism, which seems characteristically associated with the Ross orogen, supports the latter.

The Ross event is not recorded in the 13,000 m of Paleozoic rocks which form the Ellsworth Mountains (fig. II). This series, which includes Cambrian carbonates and clastics (Minaret and Heritage Groups), shallow shelf Devonian quartzites (Crashite quartzite), and Permian glacial and post glacial deposits (Whiteout Conglomerate, Polarstar Formation), was highly deformed and subjected to low grade metamorphism during the Triassic (?) at least Permian, Ellsworth orogeny (Dalziel and Elliott, 1973; Splettstoesser and Weber, 1980). The fold trends turn from northwest at the southern end to north near Ellsworth Land; this trend is anomalous when compared to that of the roughly contemporaneous (?) Weddell orogeny (Pensacola Mountains) and "Gondwanian" orogeny (Scotia Arc Sector).

The Permian beds found in the Pensacola, Ellsworth and farther south in the Thiel Mountains are correlative with the Beacon Group formations found in the Ross Sea Sector of the Transantarctic Mountains and in the Theron Mountains (Collinson et al, 1980, fig. II). They are also correlated with the Amelang and Kirwan formations of Western Queen Maud Land (Aucamp et al, 1972). Most of these formations show evidence for a late Paleozoic glaciation in this part of Gondwanaland. East and south of the Pensacola region the strata of the Beacon Group are essentially flat lying, not affected by the Gondwanian orogenies; they represent terrestrial deposits marginal to a stable craton that has been subjected to epirogenic movements since the early Paleozoic (Dalziel and Elliott,
In the Pensacola region itself the Neptune Group, Dover Sandstone (Devonian), Gale Mudstone and overlying strata were deformed into broad open folds.

Following the lower Mesozoic folding episode, little is recorded in the Transantarctic Mountains. In the early Jurassic mafic stocks were emplaced following the trend of the mountains along the depression separating East and West Antarctica (Ferrar Dolerites, Dufek Intrusion). They are believed to be part of an evolving magma (Kistler and Ford, 1979; Neethling, 1972b), and they could indicate an early rifting (?) episode contemporaneous with the break-up of Gondwanaland.

The Dufek Intrusion, which forms most of the Dufek Massif and Forrestal Range in the Pensacola Mountains, has an age of 172 M.Y.B.P. (Kestler and Ford, 1979) and is comparable in size with South Africa's Bushveld Complex. It extends north beneath the Filchner Ice Shelf into the Weddell Sea continental shelf (Behrendt and Drewry, 1979). It is gently tilted towards the southwest, recording what might be a fourth folding event (post Middle-Jurassic).

The Pensacola Mountains lie on the borderline between East and West Antarctica, and geophysical studies in the area have revealed a major discontinuity in the crust which thins toward the northwest of these mountains (Behrendt, 1974). What is certain is that west of this axis there are few terranes which might be related to those of the "Transantarctic" trend; The exception would be Ellsworth Mountains which are anomalously oriented (Schopf, 1969; DeWitt, 1977; Watts and Bramall, 1980) and southwest Marie Byrd Land, Edsel Ford Range, (Wade, 1978) which is believed to have undergone a large amount of displacement relative to East Antarctica (Scharnberg and Scharon, 1972).
Compared to the mono-block structure of East Antarctica, West Antarctica displays more complex structural features. It is composed of relatively high mountains (locally several thousand meters) cut by very deep troughs, which extend well below sea level (fig. 1). If the ice cover were removed and isostatic rebound occurred, there would be an archipelago with four main islands, the Antarctic Peninsula Ellsworth Land, Eights Coast, Ellsworth Mountains and Marie Byrd Land, all underlain by continental crust (Demehitskaya, 1960; Woollard, 1966). These islands seem to have different affinities and they represent a complete puzzle in plate tectonic reconstructions.

To the west of the Weddell Basin lies the S-shaped Antarctic Peninsula: Graham, Palmer, and Ellsworth Lands. Here the basement complex is composed of pre-Jurassic rocks, which have undergone strong deformation and locally some metamorphism (Nordestjold Coast) during the climate of the "Gondwanian" orogeny. Dating is somewhat confused, but the formations considered, i.e. Trinity Bay Series, are believed to be late Paleozoic in age (Dalziel and Elliott, 1973) and the deformational phase to be Triassic (Adie, 1972). Axes of isoclinal folds are parallel to the trend of the Peninsula. These terranes are believed to be correlative with the Andes Basement Complex in South America (Dalziel et al, 1975) and represent a Paleozoic, Lower Jurassic arc-trench complex along the western coast of ancient "Gondwanaland"...i.e. Scotia Sea Region (Suarez, 1976; Dalziel et al, 1981; Smellie, 1981). Much in the post-Middle Jurassic geological record indicates common histories for both areas.

Late Mesozoic sedimentary and volcanic sequences are indicative of an active margin environment, which might have continued as late as
Eocene in the northern part of the Peninsula (Johnson and Vanney, 1976). In Graham Land, Jurassic (?) clastics are mainly composed of conglomeratic rocks. They are interpreted as alluvial fan to marine deposits (Elliott et al., 1978) and are overlain by a thick (up to 3000 m), mostly andesitic volcanic series which outcrops throughout the Peninsula. Farther south in the Lassiter and Orville Coasts, the Middle to Upper Jurassic Latady Formation is mainly composed of siltstone and mudstone of volcanic origin, interbedded with calc-alkaline to silicic (Williams et al., 1972). The lithologic composition of these rocks has been interpreted as a back-arc assemblage (Suarez, 1976). The basin could extend farther east under the Weddell Sea Shelf in prolongation to the Andes Marginal Basin, which takes up most of Tierra del Fuego (De Witt, 1977).

The climax of deformation due to the Andean orogeny occurred in the mid-Cretaceous in the South American Andes (Dalziel, 1975). It highly deformed what has been interpreted as a marginal back-arc basin and its Lower Cretaceous sediment fill (Dalziel et al., 1974). In Antarctica, it seems deformation began earlier, sometime between the Upper Jurassic and Lower Cretaceous, affecting the Latady Formation and overlying volcanics in eastern Ellsworth Land. Folds are isoclinal and overturned towards the Weddell Sea, indicating compressional forces come from the west coast of the Antarctic Peninsula and was probably associated with an east dipping subduction zone. In the northern and central parts of the Peninsula, only block faulting is conspicuous, rejuvenating the pre-Middle Jurassic basement (Katz, 1972; De Witt, 1977). Only Alexander Island bears some evidence of folding on the western limit of the Peninsula. The sedimentary and volcanic rocks of the Lower Cretaceous Fossil Bluff Formation have been deformed into broad open folds with locally some thrusting
towards the east (Bell, 1974).

A widespread low grade thermal event is suggested by dating of metamorphic rocks in Elephant Island (South Shetland Islands; Grikurov et al., 1970), but otherwise little metamorphism is associated with this igneous activity.

Further Cenozoic uplift, rejuvenating old relief from the Transantarctic Mountains to the western continent, and extensive block faulting parallel to the Peninsula shaped its actual morphology and controlled Cenozoic volcanism.

Cenozoic marine rocks have only been recognized on Seymour Island, at the northern tip of Graham Land. Fine gray sands, interbedded with silty clays and cut by sheet-like beds of pebbles, form the bulk of the lithology. It is characteristic of a high energy deltaic environment and has been dated early Cenozoic (Paleocene and Eocene) (Elliott et al., 1975). Faunal similarities with South America and New Zealand, which had previously been recognized in Mesozoic strata (Quilty, 1977), seem to have continued through this period. The age of onset of glaciation on the Antarctic Continent is somewhat problematic.

Plate Tectonics Reconstructions and the Breakup of Gondwanaland

The age of the lithology and intensity of deformation, through space are roughly indicative of an accretionary origin for West Antarctica. Deformation and sedimentation migrated west through time along the edge of the East Antarctic Craton. This theory was brought up at an early stage in the exploration of the Antarctic Continent by Hamilton (1966) and has some supporters (Dietz et al., 1972; Ford, 1972a). Further
studies, both geological and geophysical, tend to modify this simplified picture.

Reasoning when trying to approach the evolution of the region is often circular, as it relies, when no direct data is available, on reconstruction models. These models are predictive. They assume geological events for the different plates involved which have not always been proven.

The theory of a single Precambrian to Paleozoic continental landmass Gondwanaland formed by all southern continents, i.e. Australia, Antarctica, South America, Africa and India was postulated first by Du Toit (1937) and his reconstruction of this paleocontinent is on the whole, still accepted (fig. III). Paleomagnetic data, for example, yields concurrent paleopoles for all continents through the early Mesozoic when they are placed in their postulated "original" position (Creer, 1970; McElhinny, 1970). Data is not continuous through time and does not exclude rapid and limited (〜100 km) motions between blocks (Creer, 1970). Such movements of limited extent are recorded for the Ellsworth Mountains in regard to East Antarctica during the Paleozoic-Late Cambrian (Watts and Bramall, 1980). Post-Jurassic paleopoles are, in contrast, widely scattered and suggest a Middle Mesozoic time for the breakup of the supercontinent.

It is interesting to note that there is no pre-Jurassic data for West Antarctica and relatively scarce pole determinations for the eastern part of the continent. Yet other evidence, i.e. similarity of terranes and fauna between continental blocks and marine paleomagnetic studies, support the Gondwanian theory.
Figure III: Plate reconstructions of the Southern Continents.

1, 2, 3, 4, 5, 8 and 9: Pre-Mesozoic reconstructions of Gondwanaland.

6 and 7: Cretaceous to early Cenozoic reconstructions of the Southern Ocean.

10 and 11: Pre-Cenozoic reconstruction of the Scotia Arc Region.
In the southwest Indian Ocean, age and orientation of magnetic lineations agree with a paleoposition for Dronning Maud Land (fig. 1) against the Mozambique coast of Africa. The initiation of motion between Africa and Antarctica could be as early as Jurassic (Norton and Sclater, 1979). This motion, which must first have been along long transform faults parallel to the Mozambique coast, later became a northeast-southwest drift (Bergh, 1977).

Recent correlations of marine magnetic data also yield a Mesozoic age (\(\sim 165\) m yrs) for the oldest oceanic basement found in the Weddell Sea (La Brecque, 1981). The magnetic lineations defined trend from east-northeast to east-southeast from north to south, giving a convergence point to the west, near the Antarctic Peninsula. This point would indicate the position of the motion's pole of rotation.

The most widely accepted date for the opening of the South Atlantic is the Valanginian, in the earliest Cretaceous. The various rifting episodes taking place in the region of the Weddell Sea during the second half of the Mesozoic are most probably related. They might have had two consequences...the drifting of the Antarctic continent towards its polar position in which it has remained until present (Mc Elhinny, 1973) and the Mesozoic-Cenozoic compressional tectonics which have prevailed along the boundaries of the Antarctic Plate and the now destroyed Aluk Plate in the South Pacific (Weissel et al, 1977; fig. III). Along the southern Antarctic Peninsula, folding began in the early Cretaceous (Dalziel et al, 1975; Quilty, 1977), but an island arc was already present at the time of deposition of the Latady Formation during the middle to upper Jurassic (Williams et al, 1972). The pattern of divergence between the three continental blocks, i.e. Africa, South America and Antarctica, is now in
part obliterated by the Scotia Arc Complex which has a more recent Cenozoic history. It is hypothesized that mafic stocks of Jurassic age in the Transantarctic Mountains might evidence an early stage of rifting between East and West Antarctica (Craddock, 1964; Schopf, 1970), associated with the formation of the South Atlantic and Indian Basins. There are, in fact, two geochemical trends for these Jurassic intrusions; the Karoo trend of South Africa and western Dronning Maud Land and the Transantarctic trend which extends from the Pensacola Mountains into New Zealand (Neethling, 1972b). The former seems to have a less contaminated mantle composition. The evidence from magnetic data in the Weddell Sea and the thinner crust underlying the Ross-Weddell depression are suggestive for such a theory. Another suggestion for the origin of this rifting is given by Dalziel et al. (1975), Suarez (1976) and De Witt (1977), and calls for a back-arc spreading center similar to that found in Southern Andes which developed during the early Cretaceous.

Between the late Cretaceous and the early Cenozoic, breakup occurred in the eastern longitudes of the remaining Gondwanaland Supercontinent. India and Madagascar (Norton and Sclater, 1979), New Zealand and the Lord Howard Rise (Weissel et al, 1977) and finally Australia separated from Antarctica. Rifting in the Southern Ocean initiated 53 M. Y. B. P. (Weissel et al, 1977) and was the first step to the cooling of the Antarctic Continent and consequent glaciation (Kennett et al, 1974).

The evolution of the Scotia Arc region is a key to the plate reconstruction between the three continents, South America, Africa and Antarctica, and ultimately to the relation between East and West Antarctica. Several theories have been proposed (Schopf, 1969; Dalziel and Elliott, 1973; Suarez, 1976; De Witt, 1977) to account for the geological
record and the actual position of these three land masses, but they are largely speculative. First plate reconstructions (Du Toit, 1937; Smith and Hallam, 1970) show an overlap between the Antarctic Peninsula and the Faulkland Plateau, both of which are of continental affinities. This has been tentatively explained either by displacement of West Antarctica in regard to East Antarctica or by intraplate deformation, following separation.

The former hypothesis has been somewhat constrained by paleomagnetic studies which deny large (~10^3 km) displacement between both sub-continents, at least since the emplacement of the Andean intrusions, i.e. Cretaceous (Kellogg and Reynolds, 1977). On the contrary, data from Marie Byrd Land seems to agree with large displacements between continental blocks (Scharnberg and Scharon, 1972; Cox and Gordon, 1978) and supports the idea of a diverse origin for the West Antarctic Islands (Alley and Watts, 1979). Preliminary results from North Graham Land (Hope Bay) support a pre-Late Jurassic counter clockwise rotation, between East and West Antarctica.

The idea of large intraplate deformation could be documented farther south in the Antarctica Peninsula where paleopole determinations favor oroclinal bending as outlined by the change in the trends of folds in the Orville Coast. This could indicate lag along an early tertiary fault system (Kellogg, 1979).

From a geological standpoint, pre-Cenozoic continental connection between the Antarctic Peninsula and South America is evidenced, and this connection forces plate reconstructions to have a position for the Antarctica Peninsula totally different from the one suggested by Smith and Hallam (1970); Dalziel (1973); De Witt (1977); Quilty (1977)(fig. III).
Supportive evidence for all of these models is, in any case, meager. The overall characteristic environment of the area until mid-Cenozoic is that of an active margin. Well documented in South America (Dalziel et al, 1974), it is widely suggested in the Antarctic Peninsula:

- Occurrence of two parallel volcanic and plutonic provinces along the Peninsula during the Jurassic and Cretaceous: a Gabbroic to Quartz Dioritic belt on the west coast flanked to the east by a belt of more potassic, commonly Granodioritic rocks (Lassiter and Orville Coast among other locations, Williams et al, 1972), is indicative of an east dipping subduction zone.

- Jurassic lithology of the Eastern Coast Provinces are characteristic of a back-arc environment (Williams et al, 1972).

- Frequency of intrusives diminish from north to south. The emplacement of these intrusives ranges from syntectonic to post-tectonic supporting the hypothesis of a subduction zone closing from south to north (Dalziel and Elliott, 1973).

- Marine paleomagnetic data from the South Pacific and the Bellinghausen Sea, acknowledge subduction during Cretaceous time which continued into the mid-Cenozoic (Miocene) along the Peninsula (Johnson and Vanney, 1976; Weissel et al, 1977).

Evolution of the volcanic arc is not too clear. Was there a back-arc basin floored with oceanic crust similar to the one documented in Patagonia? Is the folding of the Latady Formation, which is interpreted as a back-arc series, associated to the closure of this basin? There is no proof as to the presence of a mafic crust below the Weddell Sea shelf. But the overturning of early Cretaceous folds in the Lassiter and Orville Coasts involving the Latady Formation is towards the Weddell Sea,
away from the subduction zone (De Witt, 1977).

Final separation between Antarctica and South America and opening of the Drake Passage occurred between 20 and 30 M.Y.B.P. as suggested by magnetic lineations in the western Scotia Sea (Barker and Burrell, 1977). Formation of the South Sandwich Arc is more recent (8 M.Y.B.P.). It lies on Pliocene Oceanic crust created at north-south trending ridge in the eastern Scotia Sea (Weissel et al, 1977).
APPENDIX B

SYNTHETIC SEISMOGRAMS:
A TOOL FOR THE INTERPRETATION OF SEISMIC DATA
SYNTHETIC SEISMOGRAMS:
A TOOL FOR THE INTERPRETATION OF SEISMIC DATA

A computer program SYNSEIS, using P. C. Wuenschel's (1960) scheme's was used to generate synthetic seismograms. Their aim was to test some of the characteristics of the data: source sensitivity to the lithology and to the acoustical parameters, and artifacts through theoretical computations.

Wuenschel's Scheme

The model used is that of a multiple layered media,(N layers and N+1 interfaces) illustrated in Figure I, bound by two infinite media. Only vertical propagation is taken into account; the corresponding wave equation, to be solved for each layer is thus:

\[
\frac{\partial^2 u}{\partial t^2} = c^2 \frac{\partial^2 u}{\partial x^2}
\]

(1)

\( u = \) Particle displacement
\( t = \) time
\( c = \) velocity of propagation
\( x = \) coordinate along the vertical axis

The solution must satisfy boundary conditions, i.e. continuity of the particle velocity (\( \frac{d\bar{u}}{dt} \)) and stress across each interface, and the initial conditions (signature) at the source. The latter is only taken into account at the end through a simple convolution operation. It is assumed first that the source is an instantaneous spike \( \delta(t) \) and the following computation will lead us to the earth's stickogram, or succession in time.
Figure I: Model assumed for synthetic seismogram.
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<th>LAYER NUMBER</th>
<th>INTERFACE DESCRIPTION</th>
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</tr>
<tr>
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<td>(P,C) 2</td>
<td>1</td>
</tr>
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<td></td>
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<tr>
<td>n-th</td>
<td></td>
<td>x=d_n</td>
</tr>
</tbody>
</table>
of the impedance contrasts.

Equation (1) gives a relation between the Laplace transforms of the particle velocity ($\bar{\alpha}$) and that of the stress ($\bar{\sigma}$) of two adjoining layers:

$$\begin{pmatrix} \bar{\alpha}_k \\ \bar{\sigma}_k \end{pmatrix} = \begin{pmatrix} sA & -sB \\ \rho_k c_k & \rho_k c_k A \end{pmatrix} \begin{pmatrix} \bar{u}_k \\ \bar{\sigma}_k \end{pmatrix} = \alpha_k \begin{pmatrix} \bar{\alpha}_{k-1} \\ \bar{\sigma}_{k-1} \end{pmatrix}$$

$$A = \frac{1}{2} (e^{s c_k d_k} + e^{-s c_k d_k})$$

$$B = \frac{1}{2} (e^{s c_k d_k} - e^{-s c_k d_k})$$

$c_k$ = velocity of propagation in the kth layer

$\rho_k$ = density of the kth layer

d$^k$ = thickness of the kth layer

$s$ = Laplace variable

If we consider the entire layered medium, we have the final equation in Laplace domain:

$$\begin{pmatrix} \bar{\alpha}_N \\ \bar{\sigma}_N \end{pmatrix} = \begin{pmatrix} \alpha_N & \alpha_{N-1} & \cdots & \alpha_1 \end{pmatrix} \begin{pmatrix} \bar{\alpha}_0 \\ \bar{\sigma}_0 \end{pmatrix}$$

Practically when using (3) to compute synthetic seismograms, the three following points will be acknowledged:

Constant travel time layers:

$$\tau = \frac{d_k}{c_k} \quad (k = 1, N)$$

The characteristics of the infinite lower medium yields a relation
between \( \tilde{u}_N \) and \( \tilde{\sigma}_N \): \( \tilde{u}_N = \tilde{u}_{N+1} \)

\[
\tilde{\sigma}_N = \rho_{N+1} c_{N+1} \tilde{u}_{N+1} \quad \Rightarrow \quad \tilde{\sigma}_N = \rho_{N+1} c_{N+1} \tilde{u}_N
\]

The infinite upper medium is stress free

\( \sigma_{o-} = 0 \) (stress above the 1st interface is zero)

There are three cases to be considered and these will develop four final equations: (4) (5) (6) (7)

a. Source and detector are at the surface:

Boundary conditions

\[
\begin{align*}
\dot{u}_{o+} & = \dot{u}_{o-} \\
\sigma_{o+} - \sigma_{o-} & = \delta(t) \\
\bar{\sigma}_{o+} - \bar{\sigma}_{o-} & = -1
\end{align*}
\]

\( \delta(t) \) being the time variation of the source

Using relation (3)

\[
\begin{pmatrix}
\bar{u}_N \\
\rho_{N+1} c_{N+1} \bar{u}_N
\end{pmatrix}
= \begin{pmatrix}
\alpha_N & \ldots & \alpha_1
\end{pmatrix}
\begin{pmatrix}
\bar{u}_{o+} \\
-1
\end{pmatrix}
= \begin{pmatrix}
A_{11} & A_{12} \\
A_{21} & A_{22}
\end{pmatrix}
\begin{pmatrix}
\bar{u}_{o+} \\
-1
\end{pmatrix}
\]

which solved for \( \bar{u}_{o+} \):

\[
(4) \quad \bar{u}_{o+} = \frac{A_{22}/\rho_{N+1} c_{N+1} + A_{12}}{A_{11} + A_{21}/\rho_{N+1} c_{N+1}}
\]

The \( A_{ij} \) are polynomials of \( e^{-2\xi t} \) and the results of (4) is also a polynomial of \( e^{-2\xi t} \)

\[
\bar{u}_o = a_0 + a_1 e^{-2\xi t} + \ldots + a_N e^{-2N\xi t}
\]
Going back to the time domain:

\[ q_o = a_o \delta(t) + a_2 \delta(t-2\tau) + \ldots + a_N \delta(t-2\tau) \]

This represents impulses which are separated by a $2\tau$ time intervals. This latter parameter corresponds to the sampling rate of the signal.

b. Source buried at the $a^{th}$ interface: reasoning as above, but setting the discontinuity due to the source at the $a^{th}$ interface.

\[
\begin{align*}
\sigma_a^+ &= \sigma_a^- \\
\sigma_a^+ &= \sigma_a^- \\
\bar{u}_a^+ + \bar{u}_a^- &= \delta(t) \\
\bar{u}_a^+ - \bar{u}_a^- &= 1
\end{align*}
\]

(5) \[ \bar{\sigma}_o = \frac{A_{11} + A_{12}^{i}}{A_{11} + A_{12}} \rho_N^{N+1} C^{N+1} \]

(6) \[ \bar{\sigma}_b = A_{11}^s \bar{\sigma}_o^+ - A_{12}^s \]

A' = \[ \begin{array}{c|c|c|c}
\alpha_N & \ldots & \alpha_{a+1}
\end{array} \]

(7) \[ \bar{\sigma}_b = A_{11}^s \bar{\sigma}_o^+ + A_{12}^s \]

A'' = \[ \begin{array}{c|c|c|c}
\alpha_b & \ldots & \alpha_{a+1}
\end{array} \]

b. Source at surface

c. Detector buried at the $b^{th}$ interface: The variables of interest are now the earth's response at the $b^{th}$ interface:

\[
\begin{align*}
\bar{u}_b \\
\bar{\sigma}_b
\end{align*}
\]

rather than \[ \begin{align*}
\bar{u}_o \\
\bar{\sigma}_o
\end{align*} \]

the signal recorded at the air interface, whether the source is buried or not.

Source at surface

Source buried at $a^{th}$ interface
Program Synseis

The program has four options:

(1) Choice of the value $\tau = \text{T}_\text{AU}$, or half the sampling interval. It might be given $(.5, 1.0, 2.0 \text{ ms in general})$ or it might be computed. The latter case is interesting for some simple models when no convolution by a continuous signature.

(2) The density distribution which can be taken into account or not.

The actual value of the reflection coefficient is:

$$R_{ij} = \frac{\rho_i c_i - \rho_j c_j}{\rho_i c_i + \rho_j c_j}$$

but it is usual in the shallow zone of seismic penetration (3 to 4km) to consider $\rho$ as roughly constant.

(3) Choice of the buried source computation or not.

(4) Choice of the buried receptor computation or not.

The different steps preformed by the whole algorithm leading to the determination of the earth's stickogram are summarized in Figure II. Finally, comes the convolution between the stickogram and the source signature.

A second program SYNSEIS PLOT is then used to plot the data in a wavelet form on the calcomp plotter, at the desired scale.

Complete listings of both programs are in appendix C.

Parameters

The data considered in this study have all been recorded in a marine environment, in water depths generally exceeding 400 m.
Figure II: Simplified organigram of SYNSEIS and SYNSEIS PLOT programs.
MAIN PROGRAMS

INITIALISATION OF WORKING ARRAYS

COMPUTATION OF TRANSMISSION MATRIX FOR MODEL ASSUMED

POLYNOMIAL DIVISION FOR DETERMINATION OF STICKOGRAM

ADJUSTMENT OF STICKOGRAM TO BURIED RECEIVER (OPTIONAL)

CONVOLUTION OF STICKOGRAM BY GIVEN SOURCE SIGNATURE

OUTPUT

SUGRoutines

SOUReC

Transforms layers into time intervals

FILLIN

Transmission matrix for one layer

DUCON

Multiplication of two transmission matrices

VECTOR

Selection of terms in matrices necessary for determination of stickogram

CONVOL

Convolution of two one-dimension arrays

Stickogram (optional) Convolved trace

PLOT OF FILES
The Source

The source used in both surveys was a 4kJ sparker system with three electrodes giving a medium energy acoustical pulse of duration 10 to 20 ms. The system generates energy in the seismic and sonic range by discharging stored electrical energy between electrodes in salt water. The acoustic source bubbles are composed of steam which generate high pressure pulses (fig. III). Their analysis shows:

- Initiation of the steam bubble due to extremely rapid heating of water creates a steeply rising shock wave.
- The pressure then falls exponentially as the bubble expands.
- The pressure lowering and cooling of the steam produces the bubble to collapse or implode, producing a second intense pressure pulse. This gives typically a single bubble pulse as opposed to other marine sources.
- The implosion pulse is followed by a type of "sizzling" presumably resulting from collapse of numerous small bubbles.
- The bubble "lifetime" or period is function of the source energy. It increases exponentially with the energy (fig. IV).

It is interesting to note that the low frequency content of the pulse is provided by the total transient waveform --shock plus implosion-- and is thus also a function of energy. The low-frequency content of the source determines its depth of penetration, as the earth acts as a high cut filter. The 4kJ sparker has a rather shallow penetration 400 to 500 m, but its power of resolution is quite high at these depths.

Unfortunately there was no ready availability for the 4 J pressure signature, to use in the synthetic seismograms. Both 3kJ and 5kJ pulses were used considering the actual one must be intermediary between both.
Figure III: Pressure signature of underwater sparkers

Figure IV: Rayleigh-Willis diagram for representative energy source systems. Plotted for a source depth of 30'.

(from F. S. Kramer et al, 1968)
PERIOD (MILLISECONDS)

3 KJ AT 20'

5 KJ AT 20'

ENERGY IN FOOT POUNDS AT 30' DEPTH

PERIOD (MILLISECONDS)

EQUIVALENT POUNDS OF 60% DYNAMITE AT 30' DEPTH

EQUIVALENT POUNDS OF 60% DYNAMITE AT SHALLOW DEPTH
Quantities Measured

As seen in previously, Wuenschel's scheme solves the vertical transmission equations for the velocity of the particle (\(\dot{u}\)) at the recording depth. The actual value registered by the hydrophones, on the other hand is the acoustic pressure. It can be shown that both the displacement velocity and the acoustic pressure are in phase (fig. V), and follow the same amplitude pattern.

Velocities and Densities

The velocities of propagation of seismic waves is one of the physical parameter which defines the reflection coefficient at the interfaces between each layer, the other one being the density of the medium. They are thus critical data, when building models for synthetic seismograms.

Velocities are not directly determinable with the seismic reflection method and there is no information for the area of study itself. It is generally accepted that in the marine environment velocities of shallow layers, when they are not outcrops are close enough to that of water; the seismic impedance contrasts generating reflections, is then created by density changes.

The velocity in the water layer was taken to be 1450 m/s, this is supported by the results of refraction studies of Dronning Maud Land, in an area contiguous to the survey (Hinz, 1978). Rather than using a density model which yields small contrasts in acoustic impedance,
Figure V: Phase and amplitude relationships between parameters of an acoustic system.

(from N. A. Anstey, 1977)
Particle displacement

Particle displacement

Particle compression

Particle velocity
velocity model was assumed which in the range of depths that were considered, is an acceptable approximation: there would be a $0.13 \text{ m/s}$ time lag or delay between pulses traveling at $1450 \text{ m/s}$ and $1600 \text{ m/s}$ (fig. VI). The ocean bottom velocity, was the latter value. It is the minimum value of refraction velocities gathered in the Ross Sea in a similar environment (Houtz and Davey, 1973) and offshore Dronning Maud Land. It was most important to have a reasonable model for the study of the definition of the signature as it was to be confronted with the actual data.

Other velocities were picked in the $1450 - 2000 \text{ m/s}$ range well within the velocities determined in other sectors for the interval $0 - 500 \text{ m}$. They corresponded to sediments of glacial origin which were sampled on DSDP-Leg 28 cores.

**Interpretation**

Considering the source signal is not the ideal spike -- instantaneous pulse--but rather a long ($20 \text{ ms}$) continuous multipulse signature the actual recording will be a composition of responses, following different raypaths (fig. VI) but which reach the receiver at the same time. The thickness of the "homogenous" layers assumed by the model as well as the depth of source and receiver will define this superposition of pulses. The fact that two elements of the earth's stickogram are separated by a time interval smaller than the length of the signature ($20 \text{ ms}$)will create an overlap of responses deforming the pulse (fig. VII).

The surface reflection is a particularly important point in marine surveys. Both the source and receiver are in the water layer allowing for the different pulses generated within the earth to follow two or more
Figure VI: Seismograms (I)

Comparison of the response with varying sea floor velocities--10 m layer overlying homogeneous "basement"

\[ V_{\text{water}} = 1450 \text{ m/s} \]

\text{a-} \quad V_{\text{Layer}} = 1450 \text{ m/s} \quad \rho_{\text{Water}} = 1.3 \text{ g/cm}^3

\text{b-} \quad V_{\text{Layer}} = 1600 \text{ m/s} \quad \rho_{\text{Layer}} = 2.7 \text{ g/cm}^3

\text{c-} \quad V_{\text{Layer}} = 1800 \text{ m/s}
Figure VII: Some of the possible ray paths and the resulting stickogram. (two-way travel time)
Figure VIII: Seismogram (0): One layer models with 3kJ source
paths. In the model which was used the two of them lie at the same depth (20' or 6.7 m). When a pulse reaches the air-water interface, it is reflected as a pulse of opposite polarity with no amplitude loss. Supposing you have mostly positive reflected pulses incident from the earth, the negative counterparts will eventually "erode" useful information which is in phase with it. This explains why a 5 m layer in the model analyzed with the synthetic seismogram, is not only undetectable but gives a response shorter than the one for a 2.5 m layer. Theoretically a 7.4 m, with a sub-bottom velocity of 1600 m/s would be invisible.
APPENDIX C

SYNSEIS PROGRAM LISTINGS
APPENDIX C

SYNSEIS PROGRAM LISTINGS

1) Main Program generating synthetic data

2) Subroutines called from main program

3) Plotting program
1.2

52 VECT1(1)=TOT(1,2,1)/(CD(1,14,1)*CD(1,14,1))*TOT(1,2,1)
57 DEGV1=NSUM
17 DO 49 I=1,NU2+2
53 VECT2(1)=TOT(I,1,1)*TOT(2,1,1)/(CD(1,14,1)*CD(1,14,1))
62 DEGV2=NSUM
75 CONTINUE
202 FORMAT('TRANSMISSION COMPLETE')
   C
   C POLYNOMIAL DIVISION
   C
0870 VECT1(N)=DNUM*VECT1(I)
0875 N NUM=NUM+2
0880 IF(NUM.NE.0) GO TO 19
0890 WRITE(311)
311 FORMAT('FIRST TERM OF VECT2=* U REAL\)')
0894 NUM=NUM+2
0940 DNUM=VECT2(NUM)
0970 GO TO 14
0975 CALL T(F(1,14,1)+NUM)+VECT1(I)/NUM
0980 DO 55 I=1,NU2+2
0990 VECT2(I)=TOT(I,1,1)*VECT2(I-1)+VECT1(I)
1000 CONTINUE
1010 FORMAT(\'FIRST NON ZERO TERM\)
1015 DO 54 I=1,NU2+2
1020 CONTINUE
1030 WRITE(310)
1040 FORMAT('POLYNOMIAL DIVISION COMPLETE')
1050 C
   C MOVE DOWN TRACE
1080 DEGTOT=NUM-1-DEGV1-DEGV2
1085 IF(DEG=0) GO TO 20
1090 WRITE(6,205)
205 FORMAT('ERROR IN DEGREE OF FINAL POLYNOM')
2060 STOP
2100 IF(DEG=20) GO TO 21
2150 DNUM=VECT1(NUM-NSUM)
2200 DGEV2=NUM-NSUM
2210 NI=NUM-NSUM+2
2220 DO 43 J=1,NI
2230 K=K+1
2240 K=K+1
2250 CONTINUE
2260 CALL T(T(KK,1,1)+DEGV2,NSUM,NSUM)
2310 CONTINUE
1500 FORMAT('EDGE OF PHYSICAL SETTING')
2320 STOP
2330 L=DEGTOT+1
2400 DO 69 J=1,L+1
2450 IF(J.LE.L) GO TO 69
2500 VECT3(J)=TOT(J,1,1)+LL
2550 CONTINUE
2600 FORMAT('EDGE OF LINE')
2650 STOP
2700 LL=LL+1
2750 VECT3(I)=LL
2800 CONTINUE
2850 FORMAT('EDGE OF COLUMN')
2900 STOP
2950 CONTINUE
3000 FORMAT('EDGE OF ROW')
3050 STOP
3100 CALL CONVNL(VECT1,NGV,V,CT,NGVNV)
3150 WRITE(310)
3200 FORMAT('FIRST NON ZERO TERM\)')
3250 CONTINUE
3300 WRITE(310)
3350 FORMAT('EXECUTION TERMINATED')
3400 STOP
SUBROUTINE SQURECOTHX(NX,INX,NSUNX)
C
IF(LX.EQ.0 OR INX.GT.NBLAY) GO TO 600
NX = NX+1
NX = NX + NX(NX)
IF(NX.EQ.NX) GO TO 603
NX(NX) = NX(NX), NX(INX)
C
END

SUBROUTINE CONVOL(X,Y,I,NY)
INTEGER DEGTOT
COMMON/CONVOL, DEGTOT
DIMENSION X(I),Y(I),Z(I)
C
IF(L.GT.IO00) GO TO 503
IF(X(I).EQ.0) GO TO 502
Z(I) = Z(I) + X(I) * Y(I)
CONTINUE
RETURN
C
FURTHER IN DETERMINATION OF STICKOGRAMS
STOP
END

SUBROUTINE OUCON(A,B,Z48,NSZAB,NZFILL,N)
INTEGER NSZAB
FORMAT(1X, NSTAB, 15, NNPAS, NNPAS, 15)
IF(K. GT. 2001) GO TO 512
A(I,J,K) = B(I,J,K) + A(I,L1,ZABC M1)*B(L1,N2)
CONTINUE
RETURN
C
FURTHER IN DETERMINATION OF STICKOGRAMS
STOP
END
SUBROUTINE FILLIN( LL, L, A, TAU, EARTH )

DIMENSION A(2, 2, 2, 2), TAU(2, 2, 2), EARTH(2, 2, 2)

COMMON / EARTH / TAU, NBL, TAU, EARTH

DIMENSION A(2, 2, 2), TAU(2, 2), EARTH(2, 2, 2)

COMMON / EARTH / TAU, NBL, TAU, EARTH

SUBROUTINE VECTOR( NSUMX, NX )

INTEGER H, OPT3, TAB

DATA WORK, ALPHA / 100000, 0.0, 0.0 /, TAB / 500000, 0.0 /,

NSUM = NSUMX * OPT3

COMMON / TRANS/ TOT, VECT1, VECT2, NSUM, NSUMX

DATA WORK, ALPHA / 100000, 0.0, 0.0 /, TAB / 500000, 0.0 /,

NSUM = NSUMX * OPT3

COMMON / TRANS/ TOT, VECT1, VECT2, NSUM, NSUMX

INTEGER H, OPT3, TAB

DATA WORK, ALPHA / 100000, 0.0, 0.0 /, TAB / 500000, 0.0 /,

NSUM = NSUMX * OPT3

COMMON / TRANS/ TOT, VECT1, VECT2, NSUM, NSUMX

INTEGER H, OPT3, TAB

DATA WORK, ALPHA / 100000, 0.0, 0.0 /, TAB / 500000, 0.0 /,
DEFINE FILE 13C5.D004.L.*,MN)
COMMON LID
DIMENSION TRAC(2001)
READ(5,100) MDTW,VM
100 FORMAT(2F8.2)
WRITE6,200) MDTW,VM,TWTB
200 FORMAT(\"DEPTH IN FT\",F8.2,\" VELOCITY IN M/SEC,F8.2,\" TRAVEL TIME TO BOTTOM (M)\",F8.2)
C ALPHA IS COEFFICIENT TRANSFORMING CM INTO INCHES
ALPHA=3.91582
READ(101) SCALE,TAU
101 FORMAT(2E20.12)
WRITE6,203) SCALE,TAU
203 FORMAT(\"SCALE IN TIME CM/1000,F9.2,\" 1/2 OF SAMPLING INTERVAL ZONE RECORDABLE FREQUENCY,1/TAU,F8.2)
XXNM=TWTB+50
MN=XXNM/TAU
IF (MN.LE.0.0) STOP
AXETPS=ALPHA*SCALE/100.
DELTA=1./AXETPS
CALL PLOT(0.0,0.15)
CALL AXIS(0.0,24.*TWO WAY TRAVEL TIME (MS)=24.20.0.,XXNM,DELTA)
CALL PLOT(0.0,12.0,2)
DO 10 J=1,XXNM,1
10 CALL PLOT(J,1.0,AXETPS,J+1.0,AXETPS,J-1.0,AXETPS)
XXNN=XXNM
IF (XXNN.GT.0.0) WRITE6,204) J,XXNN,LID
300 FORMAT(1X.10(E10.4,2X)>CONTINUE
611 FORMAT(XXNN,XXNN,XXNN,XXNN,XXNN)
DO 411 J=XXNN,1,-1
411 CALL PLOT(LID,J,XXNN,LID,J,-1.0,XXNN)
DO 420 I=1,XXNN,1
420 CALL PLOT(I,XXNN,LID,XXNN,TAU)
DO 430 I=1,XXNN,1
430 CALL PLOT(I,XXNN,LID,XXNN,TAU,XXNN)
305 Format(1X.10(E10.4,2X>)CONTINUE
204 Format(1X.10(E10.4,2X>)CONTINUE
20 Format(1X.10(E10.4,2X>)CONTINUE
711 Format(1X.10(E10.4,2X>)CONTINUE
80 Format(1X.10(E10.4,2X>)CONTINUE