THE SEDIMENTOLOGY OF HOLOCENE
PRYDZ BAY: SEDIMENTARY PATTERNS
AND PROCESSES

by

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November, 1997
STATEMENT

All analyses, arguments, and conclusions presented in this work are original, except where acknowledged in the customary manner.

Dennis Franklin
Institute of Antarctic and Southern Ocean Studies
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December 1996
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Dennis Franklin
"On the whole, I'd rather be in Philadelphia"

W. C. Fields, last words.

"No problem is too big to run away from"

Snoopy the Dog
# Table of Contents

Statement ...................................................................................................................... ii
Authority of Access ...................................................................................................... iii
Quotes ............................................................................................................................. iv

TABLE OF CONTENTS ................................................................................................... V
LIST OF FIGURES .......................................................................................................... VI
LIST OF TABLES ............................................................................................................. VII
LIST OF PLATES ............................................................................................................. VIII
ACKNOWLEDGMENTS .................................................................................................... IX

ABSTRACT .................................................................................................................... X

CHAPTER 1 - INTRODUCTION ....................................................................................... 1
Geology ........................................................................................................................... 8
Previous work .................................................................................................................. 11
Rationale .......................................................................................................................... 11
Content of Appendices .................................................................................................... 13

CHAPTER 2 - LITERATURE REVIEW .......................................................................... 15
Facies ............................................................................................................................... 15
Processes of the Glaciomarine Environment .................................................................. 17
Facies of the Glaciomarine environment ....................................................................... 45
Results of Previous Studies .......................................................................................... 49
Pleistocene glacial history of the Vestfold Hills ............................................................. 59

CHAPTER 3 - MATERIALS AND METHODS ............................................................... 61
Sample Collection ......................................................................................................... 61
Analyses .......................................................................................................................... 69

CHAPTER 4 - CROSS COASTAL PROCESSES .......................................................... 75
Glacial ............................................................................................................................... 75
Aeolian ............................................................................................................................. 77
Fluvial ............................................................................................................................... 95
Wave mobilisation ........................................................................................................ 100
The evolution of sand characteristics with transport ...................................................... 101
Chapter Summary ......................................................................................................... 101

CHAPTER 5 - INSHORE PROCESSES ....................................................................... 104
Recent Glacial History .................................................................................................. 108
Sub-surface Geology - Coastal Broad Peninsula ........................................................... 110
Sedimentological Processes ......................................................................................... 113
Offshore sediments ....................................................................................................... 125
Chapter Summary ......................................................................................................... 125

CHAPTER 6 - OFFSHORE PROCESSES .................................................................... 129
Description and Distribution of Sediments in Prydz Bay .............................................. 130
Processes ....................................................................................................................... 146
Facies of Prydz Bay ....................................................................................................... 169
Prydz Bay Sediment Budget ......................................................................................... 174
Chapter Summary ......................................................................................................... 181

CHAPTER 7 - CONCLUSIONS .................................................................................... 183
The sedimentary processes in Prydz Bay ...................................................................... 183
Facies Associations ...................................................................................................... 186
Palaeoclimate Implications................................................................. 186
Conclusions ..................................................................................... 187

BIBLIOGRAPHY ............................................................................ 189

APPENDIX A - LOCATIONAL DATA .................................................. 205
APPENDIX B - GRAIN SIZE DATA .................................................... 218
APPENDIX C - GEOCHEMICAL DATA ............................................. 230
APPENDIX D - HEIDEMANN VALLEY CORE LOGS ....................... 234
APPENDIX E - PRYDZ BAY BOTTOM PHOTOGRAPHS .................. 245
APPENDIX F - KROCK CORE LOGS ............................................... 258
APPENDIX G - OTHER PAPERS ARISING ....................................... 259

LIST OF FIGURES

FIGURE 1.1 - PRYDZ BAY LOCATION AND BATHYMETRY .................. 2
FIGURE 1.2 - GEOMORPHOLOGY OF PRYDZ BAY .......................... 3
FIGURE 1.3 - MAJOR FEATURES OF THE VESTFOLD HILLS ............. 5
FIGURE 1.4 - THE LAMBERT GLACIER ........................................ 7
FIGURE 2.1 - SEDIMENT CLASSIFICATION ..................................... 47
FIGURE 2.2 - DISTRIBUTION OF ACOUSTIC SEA FLOOR PROVINCES .52
FIGURE 3.1 - OFFSHORE GRAVITY CORER ....................................... 62
FIGURE 3.2 - VAN VEEEN GRAB SAMPLER .................................. 63
FIGURE 3.3 - ECKMANN GRAB SAMPLER ..................................... 63
FIGURE 3.4 - AEOLIAN SAND TRAP DESIGN AND TOWER CONFIGURATION .68
FIGURE 4.1 - GROUNDED ICEBERG SHOWING DEBRIS RICH LAYER .78
FIGURE 4.2 - GSM PROPORTIONS AND GRAIN SIZE DISTRIBUTIONS FOR 93017 & 94004 .79
FIGURE 4.3 - SNOW AND DEBRIS ACCUMULATING IN SNOWDRIFT ON ANCHORAGE ISLAND .81
FIGURE 4.4 - DEBRIS RICH SNOWDRIFT IN THE VESTFOLD HILLS .81
FIGURE 4.5 - SMALL DEPOSIT OF AEOLIAN SAND IN LEE OF BOULDER .82
FIGURE 4.6 - DEBRIS LADEN SEA-ICE ........................................... 82
FIGURE 4.7 - AEOLIAN SAND TRAP LOCATIONS .............................. 83
FIGURE 4.8 - MONTHLY AEOLIAN DEBRIS TRANSPORT .................. 84
FIGURE 4.9 - MONTHLY AEOLIAN DEBRIS YIELDS FOR SITE 4 .......... 85
FIGURE 4.10 - THREE HOURLY WIND VELOCITY ........................... 87
FIGURE 4.11 - ANNUAL AEOLIAN SEDIMENT TRANSPORT AGAINST HEIGHT .90
FIGURE 4.12 - SITE 4 GRAIN SIZE DISTRIBUTIONS BY MONTH .......... 91
FIGURE 4.13 - DEBRIS TRANSPORT V AVERAGE WIND SPEED AND MAXIMUM WIND GUST .94
FIGURE 4.14 - SEDIMENT FAN BELOW SNOW DRIFT ....................... 96
FIGURE 4.15 - CROSS SECTION AND GRAIN SIZE DISTRIBUTIONS FOR SNOWMELT FAN .98
FIGURE 4.16 - MEAN GRAIN SIZE V SORTING FOR AEOLIAN SANDS .99
FIGURE 4.17 - SORTING V SKEWNESS FOR VESTFOLD HILLS SANDS .99
FIGURE 4.18 - MEAN GRAIN SIZE V SKEWNESS FOR VESTFOLD HILLS SANDS .99
FIGURE 4.19 - CROSS COASTAL SEDIMENT PATHWAYS ................. 102
FIGURE 5.1 - COASTAL BROAD PENINSULA MAPPING AREA AND OFFSHORE SAMPLE LOCATION .105
FIGURE 5.2 - ONSHORE SAMPLE LOCATIONS - COASTAL BROAD PENINSULA .107
FIGURE 5.3 - SURFICIAL SEDIMENT DISTRIBUTION - COASTAL BROAD PENINSULA .109
FIGURE 5.4 - COASTAL BROAD PENINSULA DRILL HOLES AND SUBSURFACE GEOLOGY 112
FIGURE 5.5 - HYPOTHETICAL ICE DYNAMICS IN HEIDEMANN VALLEY ... 116
FIGURE 5.6 - FORMATION OF BEACH RIDGES AND BENCHES .......... 119
FIGURE 5.7 - HEIDEMANN VALLEY PROFILE .................................. 119
LIST OF TABLES

<table>
<thead>
<tr>
<th>Table</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Table 1.1</td>
<td>GEOLOGICAL HISTORY OF THE VESTFOLD HILLS</td>
<td>9</td>
</tr>
<tr>
<td>Table 2.1</td>
<td>SUMMARY OF ECHOSOUNDER PROVINCES</td>
<td>53</td>
</tr>
<tr>
<td>Table 2.2</td>
<td>PRYDZ BAY ACOUSTIC UNITS</td>
<td>56</td>
</tr>
<tr>
<td>Table 3.1</td>
<td>WORKING SOLUTIONS FOR MOLYBDATE BLUE SPECTROPHOTOMETRY</td>
<td>71</td>
</tr>
<tr>
<td>Table 3.2</td>
<td>RESULTS OF RADIOMETRIC CARBON DATING</td>
<td>74</td>
</tr>
<tr>
<td>Table 4.1</td>
<td>DAILY AEOLOGICAL DEBRIS TRANSPORT RATES</td>
<td>89</td>
</tr>
</tbody>
</table>
TABLE 6.1 – DATA USED FOR THE STUDY .................................................................................. 130
TABLE 6.2 – RANGES OF GRAIN SIZE PARAMETERS FOR AEOLIAN DEBRIS ........................................ 164
TABLE 6.3 – MASS OF SEDIMENTS ACCUMULATING IN PRYDZ BAY ............................................. 176
TABLE 6.4 – SEDIMENT TRANSPORT VALUES IN THE LAMBERT GLACIER ............................... 179
TABLE 7.1 – FEATURES OF THE MAIN FACIES OF PRYDZ BAY .................................................. 187A

LIST OF PLATES

PLATE E1 – BOTTOM PHOTOGRAPHS – SITE 1 .......................................................................... 246
PLATE E2 – BOTTOM PHOTOGRAPHS – SITE 2 .......................................................................... 247
PLATE E3 – BOTTOM PHOTOGRAPHS – SITE 2 .......................................................................... 248
PLATE E4 – BOTTOM PHOTOGRAPHS – SITE 3 .......................................................................... 249
PLATE E5 – BOTTOM PHOTOGRAPHS – SITE 3 .......................................................................... 250
PLATE E6 – BOTTOM PHOTOGRAPHS – SITE 4 .......................................................................... 251
PLATE E7 – BOTTOM PHOTOGRAPHS – SITE 4 .......................................................................... 252
PLATE E8 – BOTTOM PHOTOGRAPHS – SITE S5 & 6 ................................................................. 253
PLATE E9 – BOTTOM PHOTOGRAPHS – SITE S7 & 8 ................................................................. 254
PLATE E10 – BOTTOM PHOTOGRAPHS – SITES 9 & 10 ............................................................ 255
PLATE E11 – BOTTOM PHOTOGRAPHS – SITES 11 & 12 .......................................................... 256
PLATE E12 – BOTTOM PHOTOGRAPHS – SITES 13 & 14 .......................................................... 257
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Material being deposited in Prydz Bay is glaciogenic, aeolian, and biogenic. Terrigenous material is entrained by basal glacial erosion and introduced to the sea via iceberg rafting, meltout, and rollover in the periphery of Prydz Bay where iceberg drift tracks are determined by ocean currents. It accounts for the deposition of about 8 Mt a\textsuperscript{-1}.

Aeolian processes transport debris from coastal oases into the marine environment in the summer or onto fast sea-ice during the winter. Sea-ice breaks out and releases the debris into the water column as it melts within a few hundred kilometres of the source, accounting for approximately 1 Mt a\textsuperscript{-1} of sediment.

The skeletons of phytoplankton settle to the sea floor directly or via the faecal pellets of predacious zooplankton. Although production is geographically uniform, the deposition of the resultant ooze depends on bottom currents concentrating the fine material in deep water basins (eg. Amery Depression). Approximately 0.75 Mt of such material is deposited annually.

The deposition of the remains of calcareous organisms is important in water depths above 400 m, mainly on the shelf break, due to a shallow Calcite Compensation Depth and Lysocline. Approximately 0.5 Mt of calcite is deposited annually.

Sediments are reworked by iceberg turbation between 200 and 720 m water depth producing iceberg turbate, an admixture of the current sediment supply and underlying relict tills. Elsewhere, the sediment is a Siliceous Mud and Ooze (SMO) which broadly represents the contemporary sediment supply.

Sedimentation rates are at the lower end of the range of Antarctic rates: between 0.08 mm a\textsuperscript{-1} and 0.16 mm a\textsuperscript{-1}. Were the sediments sourced from the Lambert Glacier, sedimentation rates would be an order of magnitude higher,
indicating that the source of the glacial constituent is smaller, probably the glaciers along the coast. As such, sedimentary changes in Prydz Bay during the Holocene are indicative of small scale climate change rather than any large scale change required to affect the Lambert Glacier catchment.

The sub-surface sedimentology offshore from the Sørsdal Glacier indicates that a significant marine influence existed in the area during the last glacial maximum, and the Lambert Glacier was not grounded in the Svenner Channel during this period. The sub-surface geology of Heidemann Valley in the Vestfold Hills also suggests that ice extent during the last glacial maximum was not as extensive as previously thought, with glacial ice not extending as far as the coast.
Prydz Bay

The continental margin of Antarctica has three large embayments: the Ross Sea, the Weddell Sea, and Prydz Bay, each associated with significant ice shelves. Prydz Bay (Figure 1.1) is a triangular embayment in East Antarctica (Hambrey et al. 1991) situated on the coast between latitudes 68°E and 80°E. It is bordered by Mac. Robertson Land and the Amery Ice Shelf to the southwest and by Princess Elizabeth Land to the south east. Its maximum dimensions are approximately 400 km east-west and 300 km north-south.

The continental shelf of Antarctica exhibits typical high latitude morphology being deeply incised by glacial activity, producing coast parallel and coast normal troughs (Vanney and Johnson 1985) with intervening banks and depressions often surrounded by submarine moraines (Domack et al. 1989). Prydz Bay is deeper (more than 500 m) (Dunbar et al. 1985) than shelf areas around other continents due to factors such as isostatic adjustment to continental ice loading, sediment starvation, and periodic erosion by advancing ice sheets (Vanney and Johnson 1985, Quilty 1985, Cooper et al. 1991). This pattern can be explained by inner shelf erosion and outer shelf deposition with lithospheric flexure due to ice loading being small in comparison (Ten Brink and Cooper 1992).

Within Prydz Bay (Figure 1.2) one large basin, the Amery Depression, reaches depths of greater than 750 m and has a north trending extension, the Prydz Channel, with depths in excess of 500 m. The Four Ladies Bank in the north east and Fram Bank in the north west of the bay are shallow features of around 200 m depth. The Amery Ice Shelf in the south west is a large floating ice shelf fed by the Lambert Glacier. Two very deep (>1000 m) features (Lambert and Nanok Deeps) are located in front of the Amery Ice Shelf. A crescentic plateau, the Nella Rim, surrounds the Nanok Deep (Quilty 1985).
Figure 1.1 - Antarctica and Prydz Bay location diagram and Prydz Bay bathymetry based on AGSO data set.
Figure 1.2 - Geomorphology of Prydz Bay - showing major features. [after Quilty 1985]
These elongate features probably extend from well beneath the ice shelf in a north-easterly direction extending some 100 km north of the current ice shelf edge. A large, fault controlled channel, the Svenner Channel, trends to the north-west of the ice shelf, parallel to the coastline at a distance of about 40 km. Stagg (1985) described the sea bed as generally smooth throughout the bay, except in the southwest where it has been scoured by the Amery Ice Shelf. Closer investigations reported herein reveal a more complex situation. Prydz Bay lies at the oceanward end of a major graben, the Lambert Graben, which is now occupied by the Lambert Glacier and Amery Ice Shelf. The Lambert Graben extends some 700 km towards the south. It has bedrock depressions up to 5 km deep onshore and possibly 10-12 km deep offshore in Prydz Bay (Cooper et al. 1991, Stagg 1985). Basement is composed of Precambrian igneous and metamorphic rocks (Tingey 1982, 1991, Cooper et al. 1991), overlain by Lower Cretaceous and older continental rift strata, including a series of [probably] glacially dominated sediments in what are described as six acoustic units by Cooper et al. (1991).

The Vestfold Hills

The Vestfold Hills (Figure 1.3) occupy an ice free area of land of ~ 400 km² on the eastern side of Prydz Bay on the Ingrid Christensen Coastline between latitudes 68°22'S and 68°40'S and longitudes 77°49'E and 78°33'E. Most of the sea floor within 5 km of the coast is less than 25 m deep (Tucker and Burton 1987).

The Vestfold Hills are the most easterly and largest of a series of rocky coastal outcrops that extend north-east of the Amery Ice Shelf (Adamson and Pickard 1986b), bounded to the north west by Prydz Bay, to the east by the continental ice sheet, and to the south by the Sørsdal Glacier. The lines of hills consist of three peninsulas, separated by fjords and valleys trending westwards from the edge of the continental ice sheet, sub-parallel to the Sørsdal Glacier. Mule Peninsula to the south is bounded to the south by Crooked Fjord, the Sørsdal Glacier, and Chelnok Lake.
Figure 1.3 - Major features of the Vestfold Hills.
The structural depression which forms Crooked Fjord is partially filled by Crooked Lake. Marine Plain, a significant site of Pliocene marine sedimentation, is on the northern shore of Crooked Fjord in the vicinity of Burton Lake. Mule Peninsula is separated from Broad Peninsula by Ellis Fjord which trends in an east-north-easterly direction from the sea to ice sheet. The structural depression continues towards the ice sheet but swings around to trend north-east/south-west. Here, Druzhby Lake and Zvesda Lake fill the lower points of the valley. Approximately central in Broad Peninsula and parallel to the Ellis Fjord / Druzhby system is a relic fjord called Death Valley. Death Valley contains a series of hypersaline lakes, including Deep Lake and Lake Stinear. Broad Peninsula is bounded to the north by Long Fjord which extends east/west from the sea to the ice sheet. The northern peninsula, Long Peninsula is bounded to the north by the sea and is bisected by the north/south trending Tryne Fjord. To the north of the Vestfold Hills are two rocky outcrops, the Wyatt Earp Islands (68°21.5'S, 78°32'E) and Walkabout Rocks (68°22S,78°33'E) (not shown on Figure 1.2).

Lambert Glacier / Amery Ice Shelf (LGAIS)

The LGAIS occupies a major graben structure (Wellman and Tingey 1976, Kurinin and Grikurov 1982, Federov et al. 1982 ) which extends into Prydz Bay. This system (Figure 1.4) is the largest glacial system in the world and has been estimated to drain up to 20% of the East Antarctic ice sheet. While the figures vary as to the amount of ice that passes through the system, amounts in the order of 30 Gt a\(^{-1}\) (Allison 1979) have been mooted. According to oxygen isotope studies on an ice core from the Amery Ice Shelf, up to 40% loss of ice mass is by basal melting close to the grounding zone (Morgan 1972). One third of this mass is replaced by the basal freeze accretion of saline ice seawards of the grounding zone where the ice decouples from the bed and begins to float (Robin 1983). Most of the basal debris is believed to be lost very close to the grounding zone while what is left is frozen in by bottom accumulation of saline ice and therefore is probably transported to the edge of the ice shelf and beyond (Drewry and Cooper 1981).
Figure 1.4 - The Lambert Glacier and Amery Ice Shelf - Structural glaciological map based on Landsat imagery. [after Hambrey 1991]
While the LGAIS is the major glacial influence in Prydz Bay, there are a series of smaller outlet glaciers along the Ingrid Christensen Coast, draining a much smaller area, probably only the coastal fringe. These include the Sørsdal Glacier, the Ranvik Glacier, the Polar Record Glacier and the Publications Ice Shelf. Relatively little is known about these glaciers and their importance to glaciogenic sedimentary processes in the study area.

GEOLOGY

The geology of the Prydz Bay area is characterised by large areas of complexly deformed, high grade, polymetamorphic rocks (Collerson and Sheraton 1986, Tingey 1991). Specifically, Archaean granitic basement of massive to poorly foliated gneisses is exposed in the southern Prince Charles Mountains and the Vestfold Hills (Oliver et al. 1982). Archaean metasediments and Proterozoic pegmatites, granites, migmatite, amphibolite, slate, shale, quartzite, banded iron formations (Stagg 1985, Hambrey 1991) are all intruded by a series of Archaean to Proterozoic tholeiitic dykes and unmetamorphosed granites and pegmatites of probable Cambrian age.

Younger sediments near Beaver Lake (Tingey 1982, Ravich & Federov 1982, Stagg 1985, Tingey 1991) in the Prince Charles Mountains are part of the Permian and Triassic Amery Group. These sediments consist of 250 m of conglomerate, gritstone, siltstone and shale; 900 m of rhythmically bedded sandstone, siltstone, claystone, and coal; and 200 m of arkose, coarse to fine sandstone, and claystone. The youngest sediments occur in Prydz Bay where Cainozoic glacial sediments overlie Cretaceous redbed sandstones (Hambrey et al. 1991). Quaternary sediments form a flat lying sequence ranging in thickness from a few metres in inner and western Prydz Bay to nearly 250 m in the outer or eastern parts of the bay. Of these, the uppermost few metres consist of “Holocene diatom ooze and diatomaceous mud with a minor ice rafted component” overlying Miocene to Quaternary massive to stratified diamicrite (Hambrey et al. 1991)
The geological and crustal evolution of the Vestfold Hills (Table 1.1) has been a complex history of deformation of very old sediments and the intrusion of various igneous rocks. It is characterised by an assemblage of gneissic lithologies from both igneous and sedimentary protoliths (Collerson and Sheraton 1986, figure 2.2). These gneisses have granulite facies mineral assemblages.

<table>
<thead>
<tr>
<th>Event</th>
<th>~ Age (Ma)</th>
</tr>
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<tbody>
<tr>
<td>Extrusion of protoliths of the Tryne metavolcanics; deposition of the Chelnok supracrystals, intrusion of mafic igneous units.</td>
<td>&gt;3000</td>
</tr>
<tr>
<td>High-grade metamorphism and deformation ($D_1$), partial melting of Tryne metavolcanics and emplacement of protoliths of the Mossel Gneiss</td>
<td>c. 3000</td>
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<tr>
<td>Intrusion of mafic dykes.</td>
<td></td>
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<tr>
<td>Emplacement of dioritic to granitic intrusives (Crooked Lake Gneiss and localised partial melting of older units, deformation ($D_2$))</td>
<td>2400-2500</td>
</tr>
<tr>
<td>Formation of steeply dipping mylonitic shear zones ($D_3$)</td>
<td></td>
</tr>
<tr>
<td>Intrusion of high Mg tholeiitic dykes</td>
<td>c. 2400</td>
</tr>
<tr>
<td>Intrusion of dolerite dykes (Group I)</td>
<td>c. 1800</td>
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<tr>
<td>Deposition of sediments, emplacement of granitic intrusives</td>
<td></td>
</tr>
<tr>
<td>Intrusion of abundant dolerite dykes (Groups II and III)</td>
<td>c. 1375</td>
</tr>
<tr>
<td>Deposition of sediments including pelites and psammites.</td>
<td></td>
</tr>
<tr>
<td>Emplacement of tonalitic to granitic intrusives, deformation of granulite facies metamorphism of rocks in the Rauer Islands and south along the coast of Prydz Bay, high-pressure metamorphism in the south-western part of the Vestfold Hills</td>
<td>c. 1100</td>
</tr>
<tr>
<td>Intrusion of anorogenic granites and pegmatites.</td>
<td>c. 650-400</td>
</tr>
<tr>
<td>Intrusion of alkaline dykes</td>
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<tr>
<td>Rifting and separation of Antarctica from India, Madagascar, and Africa</td>
<td>c. 115-80</td>
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Table 1.1 - Geological History of the Vestfold Hills - Rauer Islands [after Collerson and Sheraton 1986, Table 2.1]
There has been a complex polyphase deformational history including the intrusion of an intensely concentrated mafic dyke swarm (Black et al. 1991, Lanyon et al. 1993). Three distinct east-west trending gneissic units have been mapped (Crohn 1959, Ravich 1960), although subsequent geological mapping continues to highlight the complexity of the geology and the difficulty in its interpretation.

The more recent geology of the Vestfold Hills is typical of many ice free coastal areas where the older basement rocks are covered with a veneer of Cainozoic glacial and marine sediments. The oldest sediments in the Vestfold Hills are the marine deposits of Pliocene age in the Marine Plain area (Adamson and Pickard 1986a, Quilty 1992). These sediments contain diatomaceous sands, a thin glacial unit and opal cemented units. There is a diverse invertebrate fauna, and cetacean fossils are also found in these sediments.

Sediments deposited by a series of Quaternary glaciations are common throughout the hills. The Vestfold Glaciation (c. 20–14 ka) resulted in glacially polished, striated and fractured rocks, roches moutonnée and ice pluck features which occur across the hills at all elevations (Adamson and Pickard 1986a). During this glaciation, the ice sheet reached at least as far as the offshore islands which have morainal ridges. The predominantly north-east trending ice movement stripped and plucked the Archaean basement. Other minor valley glaciation occurred in the north east of the hills and in the vicinity of Druzhby Lake (Adamson and Pickard 1986a).

The Vestfold Hills may have been covered by up to 1000 m of ice during the Vestfold Glaciation in the late Pleistocene. During the Holocene the ice sheet melted back allowing the land to rebound isostatically, now at a rate of 1 to 1.5 mm a\(^{-1}\). During this overall retreat there were at least two minor readvances, one building moraine ridges on Broad Peninsula, and the northwards surge of the Sørsdal Glacier in the so called "Chelnok Glaciation" (Adamson and
Introducion

Pickard 1984). Sedimentary deposits in Heidemann Valley have been linked to these two advances by Hirvas et al. (1993).

PREVIOUS WORK

The study of the sedimentology of Prydz Bay and the surrounding areas is in its infancy. There have been relatively few studies conducted in the area, and until the mid 1980's, none of these had been particularly comprehensive. In 1873, the *HMS Challenger* Expedition collected a single sample from the north east of Prydz Bay. In 1982/83, the Australian National Antarctic Research Expeditions (ANARE) conducted a marine geoscience cruise (Quilty 1985, Stagg 1985). The Ocean Drilling Program (ODP) drilled five holes in a section extending to the north west of Davis Station in the 1987/88 (Barron et al. 1991), the very important results of which are reviewed in Chapter 2. In 1990/91, grab and core samples were collected during the Australian Antarctic Marine Biological Ecosystem Research (AAMBER) Expedition undertaken by ANARE in the *RSV Aurora Australis* (Franklin 1991,1993). In 1992/93 the Krill and Rock (KROCK) voyage of the *Aurora Australis* collected grab, core and dredge samples (O'Brien et al. 1993). This thesis and associated publications report some of the results from the KROCK voyage and the results of a sampling program conducted over the summer season of 1993/94 wherein approximately 250 samples were collected in the Vestfold Hills and surrounding coastal areas. Subsequent programs have been conducted but are not reported in this thesis. These include a marine geoscience cruise of the *RSV Aurora Australis* and a number of studies undertaken by Russian scientists, the data from which are not readily available, due to disputes over ownership (Pers comm Dr. Phil O'Brien).

RATIONALE

While the study of polar sedimentology is well established, there has been little attempt to apply methods and techniques from other areas to the modern sediments of Prydz Bay. This has been, in the main, because of the paucity of
sampling programs undertaken there. With the advent of the Antarctic Co-operative Research Centre in Hobart in 1992, a focussed effort to study the sedimentary record in Prydz Bay commenced, with a view to investigate climate change over the last five million years. Because the globe is currently in an interglacial period, it is very difficult to determine what sedimentary processes occur during periods of glacial maximum. In addition, the action of advancing glaciers and ice sheets tends to remove the traces of the previous interglacial sedimentary patterns. On a more local scale, the peripheral advance and recession of the Antarctic ice sheet during the Late Quaternary is poorly understood (Domack et al. 1989).

The problem facing workers in Prydz Bay is to develop an understanding of sedimentary processes occurring during interglacials as well as to infer, from secondary evidence, those processes which occur during glacial maxima. Of most use is the location of areas where sediments from previous interglacials are preserved. The study of such locations would allow workers to trace the changes which have occurred in the sediments due to the changing climate conditions.

While the previous programs have provided in excess of 200 sampling sites, they have mainly been reported in isolation. This thesis will reinterpret the data from these studies in light of the material collected by the author over the last few years and will:

a. Identify the processes eroding, transporting, reworking and depositing debris in the marine environment in Prydz Bay.

b. Describe the facies resulting from the interaction of these processes.

c. Identify the structural and textural indicators of those processes in the resultant sediments and identify the key indicators that can be used to identify temporal and spatial changes in sediment source location.
d. Develop a preliminary sediment budget for Prydz Bay and the Ingrid Christensen Coastline, against which data gathered in Prydz Bay in the future can be compared.

CONTENT OF APPENDICES

Appendix A – Locational Data
Appendix A provides locational data for all samples collected for this study, as well as material collected by others and used for this study.

Appendix B – Grain Size Data
Appendix B contains tables with the gravel/sand/mud proportions and sand grain size distributional data. Not all samples were analysed by all methods. For instance, sediments mostly consisting of fines were not analysed for sand grain size statistics.

Appendix C – Heidemann Valley Drill Hole Core Logs
Appendix C contains core logs for the eight drill holes taken in Heidemann Valley.

Appendix C – Prydz Bay Bottom Photographs
Appendix E contains bottom photographs collected by Prof Pat Quilty.

Appendix E – Trace Element Geochemical results
Appendix E contains the results of trace element geochemistry.

Appendix F – KROCK Core Logs
Core logs for KROCK GC 29 and GC 30
Appendix G – Papers Arising

Appendix H contains other papers which have arisen from this study and have already been published or are in the process of being published. These include the description, for the first time, of preserved parts of the algal Order Parmales in the sediments of Prydz Bay (Franklin and Marchant 1995, Appendix G) and secondly, a study of the $^{14}$C reservoir effect in Prydz Bay (Franklin et al. In prep).
CHAPTER 2 - LITERATURE REVIEW

Introduction

Research on the sedimentology of Prydz Bay is in its infancy. Even though HMS Challenger collected a sample to the north of Prydz Bay in 1873, very little attention was paid to Prydz Bay until the 1980’s when systematic seismic and sedimentological sampling commenced. Nonetheless, there is a wealth of data on the sedimentology of the polar regions, from both the north and south, and this section will review the literature on glacial and glaciomarine sedimentology. It concentrates particularly on Antarctic sedimentology and Prydz Bay.

FACIES

Nicholas Steno introduced the term “facies” to the geological literature in 1669, but it was first used in the modern sense by Amanz Gressly in 1838 (Boggs 1987). Since this time, the term has been applied in many different ways and, consequently, the usage has become confused, with extended meanings being applied even to non-stratigraphic usage such as metamorphic facies.

Definition.

Hambrey (1994) defines a sedimentary or lithofacies as a body of sediment or rock with specified characteristics. One useful definition of sedimentary facies is found in Whitten and Brooks (1972), who describe facies as

“...The sum total of features such as sedimentary rock type, mineral content, sedimentary structures, bedding characteristics, fossil content, etc which characterise a sediment as having been deposited in a given environment. Facies which are particularly characterised
by rock type are referred to as lithofacies, whereas those especially characterised by their fauna are called biofacies."

Sedimentary environments and sedimentary facies are different with environments being determined by physical characteristics such as physical, chemical and biological parameters. These parameters result in the deposition of a body of sediment having particular structural and compositional properties known as facies. This study will use the term “Facies” to mean -

rocks/sediments deposited by the action of a discrete set of processes resulting in an identifiable sediment distinguishable from other sediments.

The Glacial and Glaciomarine Environment

The importance of the glacial environment can be emphasised by its extent. Glaciers currently cover 10% of the earth’s land surface, mainly as large ice sheets in Greenland and Antarctica. This figure may have reached as much as 30% during the Quaternary glaciation and as a result glacial deposits are widespread (Reading 1978).

The glacial environment is, sensu lato, a composite environment including fluvial, aeolian, and lacustrine environments, and may include parts of the shallow marine environment. Glacial deposits make up a small part of the stratigraphic record as a whole. Precambrian to Late Cainozoic glacial deposits are reviewed in Anderson and Molnia (1989) and include deposits of probable glacial origin in the Huronian in North America and the Griqualand West and Transvaal Basins of South Africa (Visser 1971); in the Hamersley Basin of Western Australia (Trendell 1981); Late Precambrian deposits from all continents except Antarctica (Hambrey and Harland 1981); and a large amount of work on Palaeozoic to Late Cainozoic deposits from across the globe.
The glacial environment is confined to areas with more or less permanent accumulations of snow and ice. Glaciers are formed by accumulations of snow above the snow line where snow does not entirely melt away in the summer.

Definitions of the glaciomarine environment all centre on the interaction of glacial ice or sea ice and the sea. It includes all those areas where sediment is deposited in the sea after release from glacier ice or sea ice (Dowdeswell and Scourse 1990), or where glaciers reach the sea and influence sedimentation (Hambrey 1994). The processes involved in glaciomarine sedimentation are complex with a variety of sediment sources being affected by glaciological, oceanographic, and biological processes.

In Prydz Bay, any discussion of sedimentation or sedimentary processes and the resultant facies must be framed around the premise that these two environments are contiguous. To date, different workers have used the terms “glacimarine”, “glacial-marine”, “glacio-marine”, and “glaciomarine”. The term “glaciomarine” will be used herein unless quoting other usages. The definition of the term “glaciomarine environment”, for the sake of this study, is:

The glaciomarine environment incorporates all areas, both terrestrial and marine, where ice or cryogenic processes are involved in the erosion, transportation, deposition, and reworking of sediments.

In Antarctica, the glaciomarine environment extends from the continental ice sheet where cryogenic processes erode and entrain sediments, to the north of the polar front as far as the influence of floating icebergs which raft sediment continues.

**PROCESSES OF THE GLACIOMARINE ENVIRONMENT**

**Introduction**

The glacial environment is divided into a number of constituent and distinguishable zones, each with particular characteristic sedimentary facies.
Discussions of glaciomarine processes and sediments in the literature are usually based on events in these zones. Edwards (1978) divided the glacial environment into the “subglacial” zone which is influenced by contact with the bottom of the glacier, the “supraglacial” zone which is the upper surface of the glacier, the “ice-contact” zone around the margin of the glacier, and the “englacial” zone within the glacier itself. Environments around the margins of the glacier influenced by melting ice but not in direct contact with the ice are the “proglacial” environment which includes the “glaciofluvial”, “glaciolacustrine”, and “glaciomarine” environments. The area extending beyond and overlapping the proglacial environment is the “periglacial” environment (Boggs 1987).

The glaciomarine environment has been variously described. Reading (1978) described it in terms of the sediment deposited. It has also be described by proximity to the glacial sedimentary source (Molnia 1983, Eyles and Miall 1984), by glacial type (Anderson et al. 1983), and by major setting (Hambrey 1994) and by the sediments deposited (Powell 1980).

Reading (1978) identifies sediment types based on grain size parameters and sedimentary structure. Eyles and Miall (1984) discuss the glacial-marine environment in terms of the facies deposited, which they divide into proximal, medial and distal facies.

Hambrey (1994) discusses the glaciomarine environment by dealing separately with fjord environments and then follows on with a discussion centred around differences in sedimentary processes. All of these approaches are valid but the close relationship of environment, process, and sediment requires an integrated approach. Any environment can be considered as the sum of the processes, both static and dynamic, which constitute that environment. It is the response to the processes that is recorded in the stratigraphic record and it is these that need to be codified.

Entrainment processes cause sediment to be supplied to the medium which will transport it to the general area of deposition. They include, but are not limited
to, subglacial, aeolian, and fluvial erosion and transport, englacial transport, and sea-ice freeze in.

Other important processes identified are slumping, debris flow, and turbidity currents (Anderson et al. 1980b, Pudsey et al. 1988). Transport and depositional processes are those which transport and release entrained sediment to the marine environment. They include sea-ice and iceberg rafting and subsequent meltout of sediment, subglacial fluvial processes, vertical and autochthonous accumulation of biological remains, aeolian transport, and sediment gravity flows.

Reworking processes affect the nature of the sediments after they are deposited on the sea-floor. They include the mixing of sediments by iceberg and biological activity and the winnowing of fines by bottom currents.

A number of other factors affect the distribution and deposition of material in the glaciomarine environment. These include, but are not limited to, continental shelf physiography, biogenic productivity, water mass circulation and sea ice fluctuations (Domack et al. 1991a).

For this section, the processes will be segregated into two main groupings, the biological and physical processes. The biological processes are those associated with the action of living organisms; productivity, in situ accumulation of bioclastic material, and bioturbation. The physical processes are all the rest of the processes, those associated with the erosion, transport, and subsequent reworking of clastic material by moving fluids (ice, water, and air).

**Biological processes**

**Introduction**

Deposition of the remains of once living organisms occurs throughout the marine part of the glaciomarine environment. Calcareous and siliceous organisms which lived in the water column of lakes and the ocean settle
directly to the bottom as part of the marine snow, or are predated and settle in faecal pellets.

The \textit{in situ} deposition of biogenic material occurs in the sub-marine zone on the sea floor. Calcareous and siliceous benthos die and their remains accumulate in the sediment. While avoiding UV light in the top 10 m, within the photic zone, most organisms are represented. Below the photic zone, plants are absent. The accumulation of calcite is dependent, also, on the water chemistry. If the water is undersaturated with respect to CaCO$_3$ then calcite which is deposited will be dissolved. Dissolution of calcite is discussed in detail later. Carbonate plants and siliceous phytoplankton are restricted to the photic zone.

\textbf{Biogenic silica}

Diatoms bloom in the surface waters of polar seas in the summer after cryophilic diatoms are released from melting sea ice and act as an inoculum (Krebs 1983), and later settle through the water column. Nutrient availability affects faunal and floral growth so facies may vary with nutrient levels (Heinrich \textit{et al.} 1992). Sea ice plays a major role in the control of surface productivity by providing stable microenvironments and allowing phytoplankton blooms during the summer sea ice breakup. This process will be discussed in detail later.

Sediment trap experiments in the central and western Ross Sea suggest that nearly all biogenic opal delivered to the mid and deep water column consisted of unbroken diatom tests of the genus \textit{Nitzschia}, many of which settled within faecal pellets (Dunbar \textit{et al.} 1991). Gersonde and Wefer (1987) estimate that up to 50\% of biogenic silica particles are incorporated in the faecal pellets of \textit{Euphausia superba} (Antarctic krill), and other zooplankton such as \textit{E. crystallorophias} (Crystal Krill). Dunbar \textit{et al.} (1991) reported large spatial variation in the magnitude and style of vertical biogenic sediment flux. This variation may be due to many things such as hydrography (Scharek \textit{et al.} 1994).
and the limited availability of silicon (Nelson and Tréguer 1992) and iron (Brandini 1993).

In Prydz Bay this style of accumulation of biosiliceous material results in the deposition of Siliceous Mud and Ooze (SMO) which is one of the most widespread modern surficial sediment types, and is the most abundant surficial sediment (Anderson et al. 1984). Many of the deep glacial troughs of the Antarctic continental shelf are currently being filled with SMO. Dunbar et al. (1989) conclude that biogenic deposition dominates sedimentation in McMurdo Sound. In Bransfield Strait, the total flux of debris reaching sediment traps deployed in the study of Gersonde and Wefer (1987) was in the order of 2.9 to 4.3 g m$^{-2}$ day$^{-1}$, of which up to 60% was biogenic silica.

**Biogenic carbonate**

Extensive carbonates are now forming at high latitudes in the seas off Europe, Canada, and Antarctica (Rao 1996). Polar carbonates in the past formed contemporaneously with glaciation during the last glacial maximum (Taviani et al. 1993) and continued later during the rise of sea level (Wilson 1988) and during the subsequent retreat of glaciers to their present position (Domack 1988, Rao et al. 1995, 1996).


Among the many controls on bio-carbonate accumulation rates is high turbidity as it is an environmental stress for CaCO$_3$ organisms. As a result, areas with high carbonate accumulation rates are therefore probably not affected by significant turbidity (Domack 1988). Other major controls on the genesis of
bio-carbonates include saturation levels of CaCO₃, and the rate of carbonate formation (Rao 1996).

Saturation levels of CaCO₃ in the water, when low as they are in Antarctica, restrict the growth of carbonate organisms or dissolve their skeletons after death. Finally, the rate of accumulation of carbonate (relative to terrigenous accumulation) is a factor in the formation of carbonate sediments.

**Biogenic rafting**

Many species of benthic macroalgae use rocks on the seafloor as anchors and use buoyant gas filled vesicles to maintain an upright position. Gilbert (1984) showed that the alga *Fucus vesiculosus* (L.) was able to transport stones to which it was attached. In addition, algae attached to stones may be frozen into sea-ice. Gilbert (1990) argues, that while it is difficult to quantify the relative importance of this process, it is important to recognise that it occurs because it may be possible for algal rafting of large stones to occur over great distances. If this is the case, and cobbles are then deposited in fine sediments, it may be possible to misidentify such a sediment as having been influenced by ice rafting.

**Bioturbation.**

Living organisms such as molluscs, crustaceans, worms (Hambrey 1994) and foraminifera (Franklin 1991, 1993) which are sparsely disseminated through the sediment play an important role in the mixing of that sediment (Jones and Jago 1993). Such mixing or bioturbation is common in glaciomarine sediments and can be recognised by mottling in sediment cores and by the presence of infaunal organisms. Much of the structureless sandy mudstones and diamictites in cores from McMurdo Sound have been interpreted as once stratified sediment that has been bioturbated (Hambrey 1994).

Nearly all of the Late Cretaceous to Quaternary strata from ODP Leg 119 cores were completely bioturbated with ichnofabric (sedimentary rock fabric resulting from bioturbation (Ekdale 1984)) indexes commonly around 6.
Bioturbation occurs at varying depths in the sediment with Broecker and Peng (1982) demonstrating bioturbative activity to a depth of 7 cm in deep sea cores. Droser and Bottjer (1991) identify a zone of high bioturbation down to ~ 8 cm (below sea floor) and a zone of reduced biological activity down to ~ 20-35 cm in Leg 119 cores.

**Summary**

Biogenic processes on the Antarctic continental shelf are currently resulting in the deposition of biogenic siliceous muds, not only as SMO in deep and/or protected basins, but as a component of all glaciomarine sediments. These biosiliceous materials are sourced mainly from phytoplanktonic productivity in the surface waters in the summer, constrained, to a major degree by sea ice breakup in the early summer. Siliceous skeletons settle directly, or via zooplanktonic faecal pellets, to the sea floor. Calcareous sediments are accumulating in shallow areas near the coast and around the continental shelf break where nutrient levels are high and post-mortem dissolution of CaCO₃ is minimal. Bioturbation is a common process which can homogenise surface sediments to a depth of 8 cm and may act down to depths of 35 cm. Minor rafting of rock clasts by buoyant algae has been observed but is unlikely to play an important role other than locally.

**Physical processes**

**Aeolian processes**

Aeolian processes are those associated with the erosion, transport, deposition and reworking of debris/sediment by the wind. In some areas aeolian deposition is significant (Dunbar et al. 1985). For instance, aeolian dust is a major sediment source in the deep ocean basins with up to 80% of accumulated material in some areas (Pye 1994, Nickling 1994).

Terrigenous material is entrained by the wind via both the erosion of rocks by wind blown sand and dust (loess) and also by the entrainment of previously
deposited sediments by saltation. As a result, these processes are only important in areas of exposed basement or sedimentary material.

Bagnold (1941) identified three distinct modes of aeolian transport: suspension transport which moves small particles of 50 \( \mu \text{m} \) (Gilbert 1990) or between 60-70 \( \mu \text{m} \) (Bagnold 1941); saltation of large particles (60-1000 \( \mu \text{m} \)) and; surface creep of large particles (>500 \( \mu \text{m} \)) (Gilbert 1990).

Silt, sand, and gravel (cobbles up to 128 mm diameter) have been located up to 5 km from their likely source (Gilbert 1990) where they were trapped in sastrugi, implying that transport could have continued to greater distances. Under ideal conditions of unlimited sediment supply, aeolian transport rates may reach 1 Mg h\(^{-1}\) (mega grams per hour) per metre width at wind speeds of 16 m s\(^{-1}\) (Bagnold 1941).

The movement of sediment by the wind can be described with the same mathematical expressions used to describe water transport of sediments. Air works as a fluid and the amount of sediment and maximum grain size moved is dependent on the fluid (wind) strength and density (discussed later).

In Antarctica, aeolian processes are not generally important but may be locally important. The addition of wind-blown material to the top of glaciers is dependent on the availability of ice-free areas of rock and or sediment. Sediments can be introduced to the supra-glacial zone directly by the wind, or via the meltout from snow-drifts. Significant aeolian deposition to the top of sea ice (Clark and Hanson 1983) occurs during the winter months when sea ice covers the sea adjacent to areas where rocks and pre-existing sedimentary deposits are located.

Barrett et al. (1983) concluded that aeolian sand initially deposited onto sea-ice is the major source of sediment on the floor of western McMurdo Sound. When sea ice is not present during the summer, aeolian sediments are delivered directly to the inshore marine waters.
The major sources of debris are ice-contact sediments exposed by isostatic rebound (1-1.5 mm a\(^{-1}\) in the Vestfold Hills, Adamson and Pickard 1986a), glacial retreat, and fluvial and lacustrine sediments exposed during periods of low water (Molnia 1983).

During winter, aeolian transport is arguably more effective because winds are stronger. The temperature is colder, resulting in denser air, although this effect is probably negligible. In addition, during the winter, much of the land surface which is not covered with permanent ice is covered with snow drifts, reducing the area of rock and soil exposed to the wind. As a result, this process is only important locally, and at certain times of the year, after snow fields have melted and before the sea-ice breaks out, and then only in the vicinity of ice-free areas.

Sediments entrained by wind are typically very well sorted, fine grained, with individual grains displaying surface textural characteristics such as surface smoothing and rounding (Nickling 1994), although the final grain size distribution of aeolian sediments is probably strongly dependent on both the source material and the climatic regime. The action of wind erosion is also evident in the formation of facets on exposed rocks (ventifacts).

To summarise, aeolian erosion, transport, deposition, and reworking of debris/sediment is a locally important process which moves material up to coarse sand size (or larger in some instances) from areas of exposed rock and sediments. This material is deposited directly into the marine environment in summer, onto sea ice in the winter, and onto glaciers throughout the year.

Mathematical models.

Dingler et al. (1992) reviewed the literature to determine the most appropriate mathematical models for the prediction of aeolian sediment transport. The major shortcoming with the theoretical work conducted to date is that it has been developed for dry, reasonably fine sediments, such as those on beaches. Also the presence of salt in the sands may significantly affect the applicability
of the theoretical mathematical models of wind driven sediment transport (Dingler et al. 1992).

Dingler et al. (1992) contend that the entrainment of sand is a function of the velocity of the wind, such that

$$q = CU_*^3$$

where \(q\) is the transport rate (g cm\(^{-2}\) s\(^{-1}\)), \(U_*\) is the shear velocity, and \(C\) is a grain size dependent constant. Most workers agree that transport rate is related to the cube of the shear velocity.

Because the theoretical models so far developed are of use only in environments of fine grained, dry, salt free sediments, this study is not able to make use of them. The Vestfold Hills, and in particular, Heidemann Valley, contain sediments of greatly diverse grain size, from boulders to very fine diatomaceous material which have been deposited in a coastal marine environment, and glacial sediments that have been submerged in marine waters. As such, there are significant amounts of marine salts, and certainly water, incorporated in these sediments. For these reasons, the application of the theoretical models of aeolian wind transport without qualification is not practical.

**Sea ice processes**

In the Arctic, the littoral (coastal/shallow coastal) environment is the only one in which the incorporation of sediments into sea ice is important (Gilbert 1990), and this is probably also the case in Antarctica. Sediment is incorporated into sea-ice when aeolian and fluvially transported sediment is deposited on top of it, when the sea-ice is in contact with the bottom sediments and sediment is frozen in, when fine grained suspended sediments are frozen into anchor ice during its formation, and when anchor ice rising through the water column scavenges material in suspension.
Freeze-in occurs in areas where coastal waters shoal to less than the maximum seasonal sea ice thickness and is best developed where there are broad intertidal flats. Sediment loads can reach as much as 670 g l\(^{-1}\) in the Arctic (Gilbert 1990). Debris rich bands which form at low tide are interfingered with clear bands forming at high tide and there is no sorting of sediments in the process with grain size distribution of sea ice debris closely reflecting that in the underlying sediment (Gilbert 1990).

Fine grained suspended debris can be incorporated into growing sea ice, when frazil and anchor ice formed in the initial stages of freeze up, scavenge suspended material from the water column or sea floor. This results in a uniformly darkened layer (Gilbert 1990). Sediment loads have been measured in the 10's of grams l\(^{-1}\) (Gilbert 1990) in the Arctic, and up to 1.6 g l\(^{-1}\) (Barnes et al. 1982).

Anchor ice formation is best developed where temperature conditions allow ice to form and accumulate on the seafloor. For instance, areas which are exposed at low tide may become sufficiently cooled to allow the ice to form on the substrate when the tide rises. When the buoyancy of growing patches of this ice is sufficient, it will lift from the surface and carry with it sediments from the sea floor and these can eventually be incorporated into sea ice (Gilbert 1990, Reimnitz et al. 1987).

Sea ice transport of terrigenous material in Antarctica is poorly described and is probably dependent on the presence of a clastic shoreline to act as a debris source. Nonetheless, sea ice transports and deposits sediments in a similar manner to icebergs, but on a smaller scale. Observations of sediment concentrations in sea ice in Antarctica are few but average particle concentrations of sediment in sea ice in the Beaufort Gyre, Arctic ocean, is much higher (40 mg l\(^{-1}\)) than in sea water (0.8 mg l\(^{-1}\)) (Reimnitz et al. 1993).

Sea ice, once decoupled from the shore, is driven by wind more than ocean currents, as sea ice is only, at most, a few metres thick compared to icebergs with hundreds of metres draft.
Melt rate for sea ice is also different as sea ice is saline and therefore melts at a lower temperature than does glacial (fresh) ice. Sea ice meltout involves the establishment of pools on the top of the sea ice which subsequently melt through the ice and drain the surrounding area (Gilbert 1990, and references cited) as well as the melting of sea ice from the bottom up.

Sediments arising from sea ice rafting form drops, dumps, and aggregates. Grain size ranges from clays to boulders (most commonly at the finer end of that range) decreasing in size distally from the source. Sorting is poor to good and clasts may be angular to rounded, and possibly striated. Fabric is poorly developed and may contain transported faunal and floral material. The transport distance of these sediments from the source is less than for iceberg rafted sediments and deposits can occur in any depth of water.

Sea ice not only transports sediment, it also plays a major role in controlling surface productivity. Sea ice meltwater creates an environmentally stable zone which is quickly exploited by phytoplanktonic blooms (Smith and Nelson 1986a&b). Areas which undergo seasonal fluctuations in sea ice cover therefore contribute significant quantities of biogenic detritus. This process may be important in controlling the siliceous facies distribution as well as influencing the carbonate facies by the advection of planktonic productivity to the benthos (Domack 1988).

Sea ice scouring is similar to iceberg scouring but operates at much shallower depths and is confined to shallow water (Drewry 1986), although below pressure ridges it may occur to depths of 30 m (Hambrey 1994).

In summary, sea ice is an effective debris transport mechanism. Debris deposited on, or incorporated in, sea ice is moved by the prevailing wind until the sea ice melts and releases the debris load. Sea ice melts at a lower temperature than glacial ice and most of the entrained debris is probably released relatively close to the source. Sea ice is also an important agent in the phytoplanktonic cycle because, as it melts, it provides a stable
microenvironment which is exploited by blooming phytoplankton in the early summer.

**Glacial processes**

Glacial advance and retreat is affected by global and local climatic variations. The retreat of northern and southern hemisphere ice sheets during the Late Pleistocene and early Holocene is probably due to quite different causes. In the northern hemisphere the most recent retreat was probably caused by Milankovich summer insolation while the Antarctic ice sheet retreat was probably a response to the resultant sea-level rise and global warming (Harris *et al.* 1996, and references cited).

Antarctic ice sheet response to global climate fluctuations may not occur in an obvious fashion. A 5°C increase in global temperature may result in short term mass accumulation rather than ablation due to the increased precipitation resulting from temperature increase (Domack *et al.* 1991b). Domack *et al.* (1991c) discusses the possibility of the advance of some East Antarctic outlet glaciers during the Hypsithermal, a period of warm conditions identified in palaeoclimate records between 7 and 4 ka (Pielou 1991).

Glacial processes include erosion beneath the ice sheet and outlet glaciers, deformation of previously deposited sediments beneath moving ice, incorporation of debris in glacial ice and the subsequent transport of that debris to the coast, and finally the deposition of sediments beneath, and adjacent to moving glacial ice. Transport and deposition subsequent to iceberg calving is discussed later.

In Antarctica about 99% of the continent is ice covered and subglacial erosion is the primary erosive process (Anderson *et al.* 1983). Glacial erosion includes several discrete processes including abrasion and fracture of structurally homogenous rock, and fracture of jointed rock, followed by the entrainment of resultant debris into the basal zone of the glacier by pressure-melting regelation and direct freeze on (Anderson *et al.* 1983, Hambrey 1994). The erosive
effectiveness of a glacier is controlled by the thermal regime, with wet-based
glaciers being very effective and cold dry glacial regimes being incapable of
significant erosion, transport, or deposition (Anderson et al. 1983 and
references cited).

Abrasion is the production of fine grained material by the scoring of bedrock
by debris laden ice (Hambrey 1994). Factors affecting the degree of glacial
abrasion are the presence of basal debris and the velocity of the glacier. Less
important is ice-thickness (affecting downwards pressure), presence of water
(water reduces the downward pressure), the relative hardness of basal debris
and bedrock, and the size and shape of basal debris particles. The relative
importance of these factors varies according to the thermal regime of the
glacier, with cold glaciers tending to be less effective agents of erosion
(Hambrey 1994). In Antarctica, erosion at the base of moving glacier ice
provides most of the debris that is transported from the continent (Hambrey et

Glacial erosion can be identified after the event by major landforms such as
glacial troughs, and cirques, medium scale landforms such as roches
moutounnée, and small scale features such as polished and striated pavements
(Hambrey 1994). Another indicator is erosion surfaces with reverse
(shoreward) gradients (Anderson et al. 1994). At the grain scale, evidence of
glacial erosion can be seen in striations on large clasts.

The glacial processes to be discussed here are those which are directly related
to the action of glacial ice.

As outlet glaciers increase in size, and the grounding lines move to seaward,
previously deposited sediments are "bulldozed" by the grounded ice. This
produces moraines and laterally transposes the results of previous cycles of
sedimentary deposition.

When a glacier moves over an unconsolidated sediment, there is a coupling
between the glacier and the underlying strata (Boulton and Jones 1979,
Boulton 1979, Alley et al. 1986) which leads to deformation within the sub-glacial sediments (Hart and Roberts 1994). This deformation is enhanced in wet based glaciers by high pore water pressure. Such deformation has been described as subglacial glacio-tectonic deformation (Hart and Boulton 1991) and it occurs in a layer below the ice sheet known as the deforming layer. Strain in this layer varies longitudinally with extensional strains up-glacier and compressive strains at the margin.

Subglacial deformation occurs in three ways (Hart and Boulton 1991); meltout at the ice-sediment interface, deforming sub-glacial sediment moving along with the ice, and changes in the thickness of the deformable layer. These processes produce a sediment termed a soft bed till. Features which result from sub-glacial deformation include small scale folds and overturned features for low shear regimes to cataclastic shear zones with attenuated folds, augen, and boudinage in high shear regimes (Hart and Roberts 1994).

Glacial deformation of sediments also occurs remote from the glacier, particularly in the proglacial environment where large scale compressional folds and listric thrusts are produced (Hart and Boulton 1991, and references cited) and form push moraines. Other ways to form moraines have been described by Fitzsimmons and Colhoun (1995) who identify “ice-cored” moraines formed by the upwarping of debris rich bands in glacial ice.

The zone of erosion beneath a glacier is followed downstream by a transition into a depositional regime beneath actively moving ice (Hambrey 1994). Pressure melting allows material to be released from the debris rich basal zone in areas where the pressure from the ice overburden is lessened, as in areas immediately upstream from basement irregularities, or in depressions (Hambrey 1994). In the process known as lodgement, melted out debris is plastered to the glacial bed and progressively built up (Hambrey 1994) on both basement rock and older sediments. Sediments so deposited are known as lodgement tills if the ice is actively flowing and subglacial meltout till if stationary (Anderson
et al. 1983) and are characterised by internal shear structures and imbricated clasts.

During sublimation, ice is vaporised directly without passing through a liquid phase (Hambrey 1994). The process is of little importance in temperate areas but is a common process in Antarctica (Shaw 1977) and results in the accumulation of any sediment that was entrained in the ice.

Antarctic glaciers are depauperate in debris load in comparison to Arctic glaciers (Anderson et al. 1980). Anderson et al. (1980) surveyed a total of 370 glaciers and identified 4 which carried visible debris and a further 3 sightings of debris laden ice were also made. Two of the first four samples mentioned possessed thick (>15m) debris rich zones with distinct sediment bands interlayered with clean blue ice. This suggested pressure melting regelation processes of debris incorporation. The debris in one of these bands ranged from 4.7% to 7.8% and the sediment displayed a bimodal size distribution. This distribution was quite different to common tills which are usually unsorted; however, the bimodality approximated that produced by mechanical abrasion experiments in the laboratory conducted by Rogers et al. (1963). Of all of the observations, debris rich layers varied in thickness from 3-15 m with one sample comprising 5-15 cm thick layers in shear zones. The average total sediment content was 6.4%. Debris in Antarctic glacial ice in the Vestfold Hills measured ~ 6.4% in the basal zone of the ice sheet (Fitzsimmons and Colhoun 1995).

Most sediment from warm based glaciers and ice shelves is lost by meltout before icebergs calve (Elverhøi and Roaldset 1983, Dunbar et al. 1985). Drewry and Cooper (1981) showed that bergs derived from large ice shelves contain little basal debris-rich ice.

Antarctic glaciers supply sediment directly to the sea by subglacial sedimentation or by ice rafting of material entrained in icebergs or floating ice shelves and ice tongues (Dunbar et al. 1985). Debris can melt out of glacial ice either subglacially or supraglacially. This process is most active at the snouts of
receding or stagnating glaciers or ice sheets (Hambrey 1994). Supraglacial debris melts out as a result of melting of surface ice. Debris accumulates and becomes unstable, whereupon it is liable to flowage (Hambrey 1994), particularly in glaciers with a high bed gradient and with abundant meltwater (Boulton 1968, 1970). Sediment may also be deposited at the base of ice cliffs which terminate at the shoreline. Although ice cliffs are uncommon in the Vestfold Hills, they are maintained by thermal erosion or wave action (Chinn 1990, Fitzsimmons and Colhoun 1995). Cliffs form when basal ice velocities are significantly lower than surface velocities and where ice is removed by processes that generate or maintain a cliff form, including erosion of the ice face by wave action (Fitzsimmons and Colhoun 1995). Chinn (1986, 1991) argued that when the semi-rigid zone of a glacier (a brittle layer ~ 20 m in thickness) is grounded, it obstructs ice flow and determines the formation and position of a cliff, convex, or ramped terminus.

To summarise, debris is produced by moving glacial ice as it abrades the rock surface. This debris, and previously deposited sediment is incorporated into basal glacial ice via freeze-in, regelation, and the incorporation of deformed subglacial sediments. Other debris is incorporated into glacial ice by aeolian and colluvial, and fluvial (infrequent in Antarctica) processes adding debris to the top and sides of glaciers. This material is transported with the moving ice as debris rich basal ice and is deposited in a variety of ways downstream. Depositional processes include meltout beneath the glacier or beneath floating glacier tongues or ice shelves forming lodgement tills in the first instance and meltout and waterlain tills in the second instance.

**Glaciofluvial and fluvial processes**

This section deals with all processes in which water is the agent of erosion, transport and deposition. These processes are either glaciofluvial, when the water is operating in close association with glacial ice, or fluvial when rivers and streams are involved.
Glaciofluvial processes have been comprehensively studied by Powell and colleagues over the last 20 years, with studies of tidewater glaciers and glacial/glaciomarine environments in the temperate, wet based glaciers of Alaska (Powell 1981, 1983a&b, 1990, Powell and Molnia 1989, Cowan and Powell 1988, 1990, and Simenstad and Powell 1988).

Glaciofluvial processes are important in areas where glaciers tend to be wet-based. Glaciofluvial deposition occurs supraglacially (accumulation of fines in crevasses, pools and streams on the ice surface), englacially, subglacially (in tunnels etc.), marginally (around glacier margins), and proglacially (downstream of the glacier) (Hambrey 1994). A diverse collection of landforms results from glaciofluvial deposition and these are documented in Hambrey (1994, Chapter 5).

Glaciofluvial erosion removes material from below glaciers. Tunnel flow involves the flow of water and sediment/water admixtures through submarine tunnels in glaciers and or sub-glacial streams, which, because they vent sub-aquatically are always full of water. Hydrodynamically they can be modelled using full pipe flow conditions (Powell 1990). Conditions of fluid flow in tunnels result in efficient movement of sediments within the tunnel, and this is particularly so when grain sizes are not uniform.

When these sediment rich flows of glacial meltwater are discharged from the end of the glacier from subaqueous tunnel vents into the marine or lacustrine environment (Powell 1990) they are best modelled as jets or plumes. The subsequent behaviour of these jets and plumes depends on the hydrological conditions existing in the water body receiving the discharge. Differences in the density and temperature of the two water masses may result in the discharge being less dense and overriding the standing water mass, or more dense resulting in the discharge flowing along the bottom. Both of these circumstances cause different patterns of sedimentation which are described in Powell (1990).
In many instances more sediment is removed from a glacier by meltwater than by ice (Cowan and Powell 1988, Powell 1983a&b, Powell and Molnia 1989). In glaciers where meltwater is significant, water is derived from supraglacial melt which reaches the glacier bed via crevasses etc, although some comes from basal melting and valley side runoff (Hambrey 1994). The water then flows either in ice tunnels, sheet flow at the ice/bedrock interface, through the underlying sediment, or in channels in the glacier bed. Sediment is carried by meltwater as either suspended load or bedload and the suspended sediment load is related to discharge rate. Bedload is the material that is saltated or rolled along the bed (Hambrey 1994, Powell and Molnia 1989).

Modern sediment input by meltwater into the marine environment is considered to be minimal in the Antarctic (Dunbar et al. 1985). The limited dilution of the biogenic component of sediments by terrigenous mud emphasises the absence of meltwater derived sedimentation (Domack 1988, Rao 1996).

Because temperatures in the study area are above 0°C for only a few weeks every year and the glacial system is dry-based rather than wet based, fluvial processes are not very important. Nonetheless water does flow in the coastal areas of Princess Elizabeth Land and does move sediment with it. Van Autenboer (1962, 1964) stated that despite severe climate and continuous sub-zero temperatures ice and snow can still melt near contacts with rocks heated by high absorption of solar radiation. Throughout the Vestfold Hills, both snow and windborne dust accumulate together in the lee of hills during the winter. When the summer nival melt commences in about December, the dust in the snow is carried downslope into lakes or fjords and sometimes directly to the sea. Where dust is moved to the sea or fjord shore before sea ice breaks out, those sediments may be deposited to the top of the sea ice, but is most likely to be moved down the ubiquitous tidecracks to beneath the sea ice and will then be available to freeze in to the next year's sea ice.

On occasion, spectacular fluvial events have occurred. The collapse of an ice dam in the Vestfold Hills in 1987 released ~ 1.1 x 10^6 m³ of water (Gore 1992).
This failure, and a more recent one in 1990 seems not to have caused any appreciable effect on the geomorphology of the downstream path. Gore (1992) suspects this is because the dam failure may have released the water slowly. Regardless of this, Gore (1992) describes a significant alluvial channel contained within a deep valley downstream of the position of the ice dam and concludes that, judging by the volume of diamicton removed from the channel, fluvial erosion below failed ice dams may be an important mechanism of sediment redistribution.

In summary, glaciofluvial processes are common in temperate, wet based glaciers which do not occur in the study area and will therefore not be further considered. Large scale fluvial action is also not common, but individual catastrophic fluvial sediment transport associated with the failure of ice dams may be important agents in the movement of sediment. Water flows associated with the summer nival melt also move small amounts of sediments.

**Iceberg processes**

Icebergs are probably the most obvious agents of sediment transport of coarser sediment to more than 1000 km offshore in the northern hemisphere (Clark and Hanson 1983). Iceberg processes are those in which floating icebergs are the main agent of erosion, deposition and reworking. These processes include iceberg calving from glaciers and ice shelves, drift, meltout of debris load, and iceberg turbation of sediments.

Determining the relative importance of iceberg related sedimentation is hampered by the lack of reliable observations of the number of icebergs which carry visible sediment (Gilbert 1990). The best observations suggest a few percent of bergs carry sediment (Anderson et al. 1980) but such approximations are of limited utility as the majority of the bulk of icebergs is invisible underwater.

Iceberg calving in the temperate glaciers of Alaska occurs by three mechanisms - fracturing of glacial ice above tide level with the berg falling into the sea,
fracturing below water level with the berg rising to sea level, and large sheets shearing off the glacier front and toppling forward into the sea (Powell 1983b). In Antarctica, these processes probably also occur but the majority of iceberg calving occurs at floating glacier or iceshelf fronts and rarely involves more than the relatively passive break off of large tabular bergs. The rate at which this occurs is determined by the glacier velocity and determines the rate at which sediment laden ice is supplied to the marine environment (Dowdeswell and Murray 1990).

The major factors controlling sedimentation from icebergs include the concentration and distribution of debris in the ice, the iceberg melt rate, the ambient water temperature and the iceberg velocity, stability, and drift track (Dowdeswell and Murray 1990). In the Weddell Sea, mean water temperatures vary between -0.5°C in the summer to -1.88°C in the winter, but never exceed 0°C (Foldvik et al. 1985a).

Dispersed sediment in an iceberg can be concentrated on the ice surface due to melting from the top. Indeed, the sediment load in or on an iceberg may accelerate the melting process due to differences in albedo. Even a light coating of dust can greatly reduce the albedo of floating ice (Gilbert 1990), and references cited. Surface melting of icebergs can occur concurrently with basal freeze on, and as a result sediment will be concentrated at the top of the berg.

Vorren et al. (1983) suggest that such concentrations of relatively dense sediment on the top of an iceberg may make it unstable and contribute to its overturning. Overturning can, therefore, release large amounts of sediment in one event. This sediment spreads into a cloud (Gilbert 1990, fig 13) and is separated by grain size as it falls through the water column. Upon reaching the bottom, it deposits as a graded bed (Gilbert 1990) or, if a significant current is acting, as a laterally graded bed that may resemble turbidity current deposits or deposits from a turbid plume of subglacial meltwater.

The grain size of iceberg rafted sediments can range from clay (<2 μm) to boulders (>256 mm) with the average grain size decreasing distally from the
source (Clark and Hanson 1983). Sediments are generally poorly to very poorly sorted with angular to rounded grains possibly displaying facets and striated surfaces. Fabric in such sediments is poorly developed. These sediments can be found in all water depths and laterally to the edge of the area of iceberg influence, usually the Polar Front. (Gilbert 1990)

The distribution of ice rafted material seems to be related to water mass distribution and water temperatures. Surface and sub-surface temperatures on the shelf are almost always below 0°C and this severely limits the melt rate of icebergs and the rate of debris release (Domack 1988).

The keels of drifting icebergs which touch, penetrate, and plough through sea floor sediments generate characteristic curvilinear furrows called iceberg scours (Woodworth-Lynas and Guigné 1990). Iceberg scours have been reported from a number of higher latitude continental shelf areas in the northern hemisphere and Antarctica (Dowdeswell et al. 1993, and references cited) and have been reported as overwriting the ice sheet record in areas between 500 and 350 m water depth near Anvers Island (64.5°S 63.5°W) (Pudsey et al. 1994). In Antarctica, this process may occur to depths of 500 m (Hambrey 1994, Domack et al. 1989), however Dunbar et al. (1985) state that reworking of shelf sediments by glacial ice is of minor importance in the Antarctic because relatively little of the shelf is sufficiently shallow to be scoured by sea ice and icebergs. Conversely, Woodworth-Lynas and Guigné (1990) consider this is a major glaciomarine process. Severe reworking may result in the generation of iceberg turbates (Vorren et al. 1983) or palimpsest deposits (Barnes 1987), which can be identified by the resultant formation of low angle faults extending up to 6 m below the scour incision surface, large displacements, sub-horizontal thrust faults, high angle faults and large berms parallel to the scour. In addition, disharmonic, disarticulated folds with a well developed fracture cleavage, slickensided surfaces may be in evidence (Woodworth-Lynas and Guigné 1990).
To summarise, icebergs release debris into the marine environment by direct meltout, or by the concentration of dispersed sediment on the surface due to insolation and the subsequent overturning of the iceberg and dumping of the debris load. Iceberg keels cause scours where their keels contact the sea floor and result in the turbation of sediments between about 200 and 500 m water depth.

**Oceanographic processes**

The physical processes that control the structure and chemistry of the continental shelf water column will influence the nature of sedimentation because some of the sedimentary processes involve vertical or horizontal transport of sediment through the water column.

Ocean currents, which are of prime importance to sedimentation on the Antarctic continental shelf (Jacobs 1989), have the effect of resuspending and remobilising sediment, and redistributing finer fractions (Hambrey 1994). They redistribute sea ice to create polynyas which give rise to variability in biological productivity. Biological debris resulting from that productivity is transported to the sea floor. Lateral movement of water masses through which debris falls results in the decoupling of areas of high productivity with underlying sea floor (Jacobs 1989).

Bottom currents are produced by various means. Differences in density, salinity, and water temperature produce geostrophic currents. Tidal variation and wind action at the surface also produce bottom currents.

Marine current activity along the continental shelf edge and upper continental slope in the form of geostrophic currents produce mean bottom current speeds near the continental shelf break in the order of 6-7 cm s\(^{-1}\) in the Weddell Sea (Foldvik *et al.* 1985b), and 5-9 cm s\(^{-1}\) near the Ross Ice Shelf (Pillsbury and Jacobs 1985). Maximum current speeds near the Ross Ice Shelf exceeded 40 cm s\(^{-1}\) (Pillsbury and Jacobs 1985). These currents scour sediments leaving residual glacial marine sediments (Anderson *et al.* 1983). Resuspension of fine
grained sediments by bottom currents is pervasive and results in the movement of biogenic and fine grained terrigenous material from shallow exposed areas to shelf basins (Dunbar et al. 1989). Bottom currents in Prydz Bay are typically around 10 cm s\(^{-1}\) and never exceed 25 cm s\(^{-1}\) (Hodgkinson et al. 1988, 1991a, 1991b).

The circulation of the deep Weddell Sea is dominated by two cyclonic gyres, the Weddell Gyre and another gyre off the Ronne Ice Shelf (Foldvik et al. 1985a). The Weddell Sea gyre circulation is believed to extend to the sea floor but mean speed in the abyss is weak at ~1 cm s\(^{-1}\) below 400m (Foster and Middleton 1979). A similar cyclonic gyre is acting in Prydz Bay (Smith et al. 1984) however it is yet to be quantified in similar detail.

Fluctuations in current velocities are correlated with changes in grain size distribution, with regions of high activity being reflected in the winnowing of fine grained, lighter fractions (Allison and Ledbetter 1982) and downcore grain size variation has been used to infer changes in bottom currents through time (Pudsey 1992). The intensity of bottom currents can be inferred from textural and compositional analysis of the surface sediments. Grain size distribution can provide important information about suspension of sediments, bottom transport mechanisms, and maximum velocities of bottom currents (Dunbar et al. 1985). From particle transport experiments, Southard et al. (1971) and Singer and Anderson (1984) inferred relative bottom current strength from grain size distributions. These experiments demonstrate that grain size distributions can be used to identify winnowed sediments (lag deposits), sediments subject to bedload transport conditions, and basinal deposition of fines out of suspension. Further predicts current strengths based on the flume experiments of Southard et al. (1971). A more extensive explanation is available in Dunbar et al. (1985).

On the outer portion of the continental shelf near Adélie Land, impinging Circum-polar Deep water (CDW) has effectively sorted relict diamict into coarse to medium sands and gravels and has redistributed the biogenic detritus.
(Domack 1988). This process is aided by bioturbation (Singer and Anderson 1984) which resupplies the sediment surface with fines from deeper in the sediment. Fine grained material moved in this manner may end up within the basinal Siliceous mud and ooze deposit as it is transported shelfward. Sediment budget studies have determined that the terrigenous content of siliceous basin sediments could have been derived entirely from this source (Dunbar et al. 1985)

**Turbidity current processes**

Sedimentary deposits from gravity flows are common on the continental shelf of Antarctica (Wright et al. 1983). The relief on the Antarctic continental shelf is sufficient to support common mass movements of sediments through slumping, debris flows and turbidites (Anderson et al. 1984). Accordingly, mass flow processes are likely to be key sedimentary agents on the continental shelf due to the great depth, absence of meltwater runoff and lack of a wave dominated coastal zone (Dunbar et al. 1985). Reworking by subaquatic mass movement includes a continuous spectrum of processes (Hambrey 1994) such as sliding (the displacement of sediment mass along a slip plane), slumping (displacement by internal folding), debris "flowage" (movement of sediment as a slurry) and turbidity "flowage" (sediment in suspension). Sediment distribution patterns in the Weddell Sea show a large turbidite fan complex extending across the northeastern portion of the Weddell Sea basin.

Turbitides on the shelf are generated by post glacial readjustment of local relief features on the shelf. Storm related currents may also be responsible for graded sand depositions seen in cores from the continental shelf. Whatever causes them, the currents are sufficient to prevent the upper divisions of the Bouma sequence (B-D) from being deposited since sands have not been seen to grade into silts anywhere on the shelf. Instead, overlying units are siliceous muds which are in hydrodynamic disequilibrium with the underlying graded beds. The finer fractions of the turbidity currents are probably deposited in the deeper quieter parts of the basin and shelf slope (Domack 1988).
Some of the sediment gravity flow deposits are probably related to areas of high biogenic accumulation as evidenced by the presence of biogenic detritus in those parts of the turbidite units adjacent to the shelf banks where such biogenic material is found (Domack 1988). Such turbidity currents may be initiated by the slumping of unstable piles of biogenic debris.

**Tidal processes**

Tidal currents are likely to affect the composition of sediments where the water depth is sufficiently shallow to allow significant tidal streams to develop, such as in areas where the stream is constricted.

Of all coastal areas in the world, Antarctica is the most poorly served with respect to sea level and tidal measurements. As at 1985, records existed for only 3 locations near Prydz Bay. At Mawson Station (63° S 64° E) two, one month long, records were taken in 1956 and at Mirnyy Station (66.5° S 94° E) a record was taken of tidal variations for the period November 1956 to January 1957. There have been no records taken from within Prydz Bay (Lutjeharms and Stavropoulos 1985). Numerical tidal simulation of the Ross Sea, however, shows that periodic tidal currents drive steady barotropic circulations in the order of $10 \text{ cm s}^{-1}$ along the sides of several topographic features formed by the combined relief of the seabed and the ice shelf base (MacAyeal 1985). MacAyeal (1985) suggests that these tidal currents may trigger large scale oceanic convections that control the heat and mass exchange between the ice shelf and the ocean.

Irish and Snodgrass (1972) described four tidal constituents that propagate in a clockwise direction in the Weddell Sea in a mixed diurnal and semi-diurnal regime, and are subject to seasonal change with a winter reduction in amplitude (Foldvik and Kvinge 1974, Foldvik et al. 1985b) of about an order of magnitude (Barber and Crane 1995). The diurnal component is up to three times larger than the semi-diurnal component (Foldvik et al. 1985b). Regardless of the lesser energy levels during the winter, Pillsbury and Jacobs
(1985) reported significantly higher energy levels in the winter period of sea ice formation along the Ross Ice Shelf. The large seasonal variation in the power of the inertial component in the deep ocean is most likely due to differences in the variability and magnitude of Weddell Sea Bottom Water flow.

**Calcite dissolution processes**

The amount of dissolved calcite in the water generally decreases with increasing depth and decreasing temperature (Rao 1996). The lysocline represents the level below which CaCO$_3$ begins to be dissolved by reaction with the seawater but organisms can still precipitate CaCO$_3$ to produce shells, and is the depth below which dissolution effects are first seen (Broecker and Peng 1982). Above the lysocline, the water is saturated with respect to CaCO$_3$ and it is in this zone that calcite organisms are most common because the precipitation of CaCO$_3$ is readily achieved.

At greater depths, the Calcite Compensation Depth (CCD) represents the level below which the water is undersaturated with respect to CaCO$_3$ and calcite which falls below this level is quickly dissolved. Broecker and Peng (1982) identify this depth as the depth below which the CaCO$_3$ content of sediments is $<20\%$. The dissolution process can greatly affect the resulting sediment.

Saturation of CaCO$_3$ is low in high latitude seas but it occurs in sufficient concentrations to form extensive carbonates (Rao 1996) and the calcite compensation depth (CCD) around Antarctica has been documented as being much shallower than in the rest of the world's oceans. It is generally considered to be in the vicinity of 1500 m in depth adjacent to the continent (Anderson 1975, Kennett 1966, Milam and Anderson 1981, Quilty 1985, Franklin 1993). Because Prydz Bay is not as freely connected to the Southern Ocean as the areas where CCD studies have been conducted, there has been interest in the role of the CCD in Prydz Bay (Franklin 1993). Quilty (1985) concluded that the CCD is not a factor in the distribution of foraminifera in areas shallower than 1500 m.
Problems with radiogenic carbon dating

The use of radiocarbon for dating late Quaternary samples has been a well established dating technique for more than 40 years (Libby 1955). It provides a fundamental research tool for establishing the geochronometric framework in the stratigraphic record of the last ca. 40,000 years (Gordon and Harkness 1992). There are, however, specific problems with using this technique in the Antarctic. This study highlights some of these problems, particularly those that relate to dating of marine sediments in Prydz Bay, Antarctica.

Since the existence of $^{14}$C reservoirs was first recognised, corrections have commonly been used to account for them. The oceanic reservoir is better understood than many and results from a lag in the carbon dioxide exchange between the oceans and atmospheric carbon reservoirs (Skirrow 1975, Gordon and Harkness 1992, Broecker and Peng 1982). Oceanic $^{14}$C residence time is up to 95 times atmospheric $^{14}$C residence times (Harkness 1979). This results in the depletion of $^{14}$C in the oceans and marine organisms relative to concurrent terrestrial organisms, making those marine organisms appear older (Gordon and Harkness, 1992). Additional variations are due to the non-uniform mixing in the oceans with differences in $^{14}$C concentrations occurring both laterally and vertically (Gordon and Harkness, 1992). In the Antarctic, the $^{14}$C reservoir effect is particularly strong with apparent ages of living marine organisms varying between 0 (Gordon and Harkness 1992) and 1770 ybp (Stuiver et al. 1981). Surface water $^8$ $^{14}$C values are between -18 and -292 %o (Omoto 1983 - fig 1). The oldest published age for surface water is 2860 years (Omoto 1972, Yamasaki et al. 1977).

Gordon and Harkness (1992) reviewed a diverse range of modern Antarctic $^{14}$C analyses and because of the extensive variation concluded that $^{14}$C dating in the Antarctic is not as useful a tool for dating as in other parts of the world. While the data Gordon and Harkness (1992) presented strongly suggest that an Antarctic-wide correction factor cannot be applied to ancient samples, when these data are considered by single species, or groups of related species, collected in a particular area, in some cases apparent ages are less disparate. In
particular, sessile species show the least disparity. Variations in methodology, collection, preparation, and analysis as well as the stated geographic variation in the reservoir effect itself do however make comparison of analyses from different studies difficult.

Dates of surface material collected from the Mac. Robertson shelf have proved to be anomalously old due to contamination of surface sediments by older material derived from outcropping Mesozoic sediments (Harris et al. 1996). Other dates from the George V Coast (Domack et al. 1991a) also show very old dates for surface material. Domack et al. (1991a) suggest caution in the interpretation of $^{14}$C dates in Antarctica due to both the $^{14}$C reservoir effect and reworking affects.

FACIES OF THE GLACIOMARINE ENVIRONMENT

Non-genetic classification of sediments

The identification of sediments in this work follows the method used by Hambrey (1994) which is a slight modification to that used by Moncrieff (1989). The modification involving the maximum proportion of gravel in diamict reduced from Moncrieff's 80% to 50% was to allow compatibility with the ODP's definition for diamict and conglomerate/breccia (Hambrey 1994). The classification table is reproduced in Figure 2.1 below. The classification is presented here as a ternary plot and will be used later in the thesis. Of note, the vertical axis representing the percentage of gravel is presented as a logarithmic scale to account for the overwhelming dominance of the fine grained fractions.

Genetic Classification of Glaciogenic and Glaciomarine deposits.

The classification of terrestrial glaciogenic sediments and glaciomarine sediments is discussed at length in Hambrey et al. (1991) and is summarised here. That classification is adopted for this study and definitions are provided
below. Those sediment types encountered in this study but not identified by Hambrey et al. (1991) will be defined separately.

These definitions of sediments represent the sediments supplied to an area of deposition but do not include sediments resulting from modification or reworking. Nor does the classification adequately address the importance of biogenic component of sediments. The following classification is proposed for sediments in Prydz Bay and borrows from Hambrey et al. (1991), Dreimanis (1988), Domack (1988), Anderson et al. (1980), and Rao (1996).

**Lodgement Till.**

Dreimanis (1988) defines lodgement till as “deposited by plastering on of glacial debris from the sliding base of a moving glacier by pressure melting and/or other mechanical processes.” Hambrey et al. (1992) interpret massive diamict as lodgement till described as “non-stratified muddy sandstone or sandy mudstone with matrix supported clasts (comprising about 1-20% of rock) with occasional shells and diatoms.”

**Melt-out Till**

Dreimanis (1988) defines melt-out till as being “deposited by slow release of glacial debris from ice that is not sliding or deforming internally.”
Figure 2.1 - Classification based on Hambrey (1994) but presented as a ternary plot. Note that the vertical axis is logarithmic to allow for the fine grained nature of the sediments. C = Conglomerate, D = Diamict, M = Mud, S = Sand, cp = clast poor, cr = clast rich, d = dispersed dropstones, g = gravelly, i = intermediate, m = muddy, s = sandy.


**Waterlain Till**

Waterlain till is deposited by a continuous rain out of basal debris from a floating glacier tongue or ice shelf. This usually occurs close to the grounding line and there is little or no reworking by bottom currents. The marine [biogenic] component is negligible. The weakly stratified diamict of Hambrey et al. (1992) interpreted as waterlain till is described as “[similar to] massive diamictite but with diffuse or whispy stratification[, and] some bioturbation and slumping. [It is] diatomaceous in part, with occasional shells.”

**Proximal glaciomarine sediment**

Proximal glaciomarine sediment is transitional with waterlain till and contains a high proportion of ice rafted material (~ 1% - 5% gravel). The background marine sediment is probably derived from plumes of sediment carried in glacial meltwater (minor in Antarctica) and from biogenic accumulation. Hambrey et al. (1992) interpreted “well-stratified” diamictite as proximal glaciomarine/glaciolacustrine sediment and described it as “As [for] massive diamictite, with prominent but generally discontinuous and often contorted stratification. Clasts dispersed with occasional dropstone structures. Significance [sic] diatom component and common shells.”

**Distal glaciomarine sediment**

This sediment is mainly marine sediments of terrigenous origin brought to the site in suspension and sediments of biogenic origin. The biogenic component, which may be dominant, is principally diatomaceous. The ice rafted component is minor (< 1% gravel).

**Iceberg Turbates**

Iceberg turbates are the result of turbation of sediments by icebergs as discussed previously. As icebergs are capable of causing turbation to great depths in the sediment (10’s of metres) such reworked sediments can comprise
many disparate sediment types, from terrestrial glaciogenic to distal glacimarine to biogenic.

**Facies of the glacio-fluvial environment.**

Powell (1981, 1983a, 1983b) recognised five distinct facies associated with 11 glaciers in the Glacier Bay National Monument, Alaska. These were formulated into a sedimentary facies model based on processes operating in a temperate tidewater glacimarine setting. These five lithofacies comprised a morainal bank, diamicton, iceberg-zone mud, marine outwash mud, and outwash deltaic and braided stream lithofacies. Powell (1981) further identified three facies associations indicative of rapidly retreating ice fronts, slowly retreating ice fronts, slowly retreating or slowly advancing icefronts, turbid outwash fjord, and shallow water environments distal from ice fronts. Using these associations Powell (1981, fig 5) produced a hypothetical glacimarine sediment section.

In Prydz Bay, glaciofluvial processes are not encountered, and nor, therefore, are the facies associated with them so further discussion of them will not be undertaken.

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**RESULTS OF PREVIOUS STUDIES**

**The Challenger Expedition**

In the 1870's *HMS Challenger* undertook a world wide cruise collecting marine biological and geological samples. As part of the Heard Island to Melbourne leg, collected one sample to the north-east of Prydz Bay from a depth of ~3092 m (1675 fathoms). This sample was described as follows;

"From the depth of 1675 fathoms, the dredge bought up many kinds of rocks and pebbles, some of them showing distinct marks of glaciation, and many of them having a coating of peroxide of manganese on that part which had projected above the mud when lying on the bottom. The
rocks belonged to the following lithological types: granites, quartziferous diorites, schistoid diorites, amphibolites, mica-schists, grained quartzites, and partially decomposed earthy shales."

"From the foregoing description it appears that the deposits forming at the most southerly points reached by the Challenger are composed chiefly of continental debris carried into the ocean by the floating ice of these regions, and that this material makes up less and less of the deposit as the distance from the Antarctic continent increases until it almost disappears at about lat. 46° or 47°S. The nature of the rock fragments dredged in these latitudes conclusively proves the existence of continental land certainly of considerable extent within the Antarctic Circle." (Murray and Renard 1891)

The sample contained many rocks and pebbles with attached biota. Quartz grains were sometimes rounded and covered with limonite, and comprised granitic rocks containing orthoclase, plagioclase, quartz, and black mica; amphibolite with large grains of green hornblende and quartz; metamorphic quartzite speckled with black mica; fine grained micaceous sandstone passing to a schist; and red sandstone. The biota included foraminifers (Globigerinidae, Textulariidae, Astrorhizidae, Lituolidae, Lagenidae, and Rotalidae), gastropods, echinoderms, bryozoa, radiolaria, sponge spicules, and diatoms.

**MV Nella Dan**

The two relevant studies conducted on the *MV Nella Dan* are those reported by Quilty (1985) and Stagg (1985). As some of the data from these studies are used in this thesis, the methods used will be described here.

Quilty’s study concentrated on the distribution of foraminifers and sediments. Pipe dredges, gravity cores and grab samples were collected on an opportunity basis. Samples were sieved within hours of collection. Standard sieve fractions examined comprised all of those coarser than 125 μm. In addition, the percentage CaCO₃ by mass was determined as described in Chapter 3.
Stagg’s (1985) study of the structure and origin of Prydz Bay and the Mac. Robertson Shelf was based on ~5000 km of three fold and sixfold digital seismic data and 8000 km of magnetic and bathymetric data. He concluded that the Prydz Bay basin, which occupies an area of at least 40000 km$^2$, contains at least 5 km thickness of sedimentary rocks. The basin forms a failed rift at a triple or four armed junction and the sedimentary pile is little disturbed by folding and faulting.

He identified two groups of seismic sequences, the first consisting of Permian to Late Jurassic or Early Cretaceous strata of continental origin, and the second upper group, of Cretaceous and Cainozoic shallow marine sediments. He identified three major unconformities in the basin, which he interpreted as the Triassic or early Jurassic rift onset, Early Cretaceous margin breakup, and mid Miocene to early Pliocene glacial advance.

**3.5 kHz Echosounding**

3.5 kHz echosounder profiles were obtained during the 1982 voyage of the *MV Nella Dan*. These data were first interpreted by Stagg (1985) and later reinterpreted by O’Brien (1992) immediately prior to the KROCK sampling program. Some of these data were made available for this study. O’Brien (1992) identified 7 sea floor provinces (Figure 2.2, Table 2.1) in Prydz Bay. O’Brien’s (1992) interpretation encompasses the following major points

1. Prydz Channel is typical of troughs that cross high latitude shelves and was probably formed by fast flowing ice streams crossing the shelf and delivering [subglacial] debris to the shelf edge, causing the formation of a large fan.
Figure 2.2 - Distribution of sea floor provinces [after O'Brien 1992]
<table>
<thead>
<tr>
<th>Province</th>
<th>General Features</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Shallow areas Mounds 2-8 m high, 40 m across</td>
<td>Jagged, hashy appearance</td>
</tr>
<tr>
<td>2</td>
<td>Smooth sea floor smooth</td>
<td>Large ridges 50 m high, 10 km across. Inshore ridges are asymmetric with steeper sides offshore.</td>
</tr>
<tr>
<td>3</td>
<td>Smooth sea floor smooth</td>
<td>Irregular ridge and swale morphology 25 m high, 3 km wide. Swales are clearly defined with steep sides.</td>
</tr>
<tr>
<td>4</td>
<td>Smooth sea floor smooth</td>
<td>The Lambert Deep, defined by sharp bends in the channel.</td>
</tr>
<tr>
<td>5</td>
<td>Smooth sea floor smooth</td>
<td>Large ridges and troughs. Valleys contain depressions to 1000 m and deeper. Valleys contain depressions to 1000 m and deeper.</td>
</tr>
<tr>
<td>6</td>
<td>Smooth sea floor smooth</td>
<td>Adjacent to the Lambert Deep.</td>
</tr>
</tbody>
</table>

Table 2.1 - Summary of echosounder provinces from O'Brien (1983)
2. The Svenner Channel does not cross the shelf, indicating that the eastern and western sides of the Lambert Glacier behave differently.

3. The inner shelf deeps and outer banks of Prydz Bay probably form because of isostatic depression of the crust by the ice sheet and by subglacial erosion during repeated glacial advances.

4. The shelf deeps probably mark the position of maximum glacial erosion during stadia, whereas banks were constructed by subglacial deposition.

5. The jagged, small scale features in Provinces 1 and 3 and shallow areas of Provinces 4 and 7 are confined to shallow areas of the bay and are randomly oriented indicating that they are iceberg gouges.

6. Shallow parts of the continental shelf not cut by these features and which are surrounded by “nick point and shoal” morphology mark the shallow limit of ice gouging.

7. The extreme depth of iceberg gouging (720 m) could be due to icebergs increasing their draft as a result of rolling, and by the probable gouging during a sea level lowstand indicating that the Amery Ice shelf retreated from the edge of the continental shelf before or during a sea level rise.

8. The large scale ridges in the Amery Depression are probably large scale flutes or drumlins produced by subglacial moulding of the till at the glacier sole, and their disappearance into smooth sea floor downstream is attributed to the reduction of effective pressure at the glacier bed.

9. The extensive asymmetric ridges crossing Provinces 1 and 2 resemble morainal banks and are probably grounding line moraines.
In the study area adjacent to the Ingrid Christensen Coastline the provinces which were encountered were Provinces 2, 5, and 6. Of particular interest, the sediment lens identified by O'Brien (1992, fig 14) is seen in many adjacent tracks and is interpreted as extending over 200 km parallel to the coast. O'Brien's Province 6 is interpreted as crystalline basement and pre-Holocene subglacial till that has been incised during a series of glacial fluctuations in the coastal outlet glaciers draining the Ingrid Christensen Coast. P6 is interpreted as sub-glacial tills deposited during glacial maximum or maxima previous to the last glacial maxima, and subsequently draped with glaciomarine sediments for at least last 25000 a. With respect to Holocene sediments, P2 is interpreted as an area of present deposition of siliceous muds and oozes which are forming a drape over subglacial tills and landforms emplaced by a more extensive Lambert Glacier grounded to the edge of the ice shelf at some time in the past.

ODP, 1988

During the austral summer of 1987-88 a transect of five shallow holes was drilled across Prydz Bay as a part of the Ocean Drilling Program (ODP) Leg 119 (Cooper et al. 1991) onboard the drillship JOIDES Resolution. The five holes vary from 98 (Hole 743) to 487 m (Hole 739) in penetration (metres below sea floor - mbsf).

Cooper et al. (1991) and Cochrane and Cooper (1991) reported on the reinterpretation of previous seismic studies of Prydz Bay (Stagg 1985). Stagg (1985) defined several acoustic units on the continental shelf (PS.1 to PS.6) and continental slope (PD.1 to PD.5). Cooper et al. (1991) confirmed the general configuration of those acoustic units but reinterpreted the stratigraphy based on the newer higher resolution seismic and drilling data. The interpretations offered by Cooper et al. (1991) are in Table 2.2.

Cooper et al. (1991) have interpreted these units to represent graben formation within the metamorphic cratonic rocks prior to the Permian, followed by a long period of terrestrial basin sedimentation resulting in the deposition of fluvial and deltaic redbed sandstones, siltstones, and eventually coals. An unknown
thickness of this material was probably eroded from the inner shelf by grounded ice sheets and was redeposited, together with younger rocks in the glacigenic sequences that have moved the shelf edge to seaward since the late Eocene.

<table>
<thead>
<tr>
<th>Units</th>
<th>Age</th>
<th>Lithology</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>PS.1</td>
<td>Late Miocene to Holocene</td>
<td>Marine: diamictite and diatom ooze</td>
<td>Glacial tills, topset beds for PS.2A</td>
</tr>
<tr>
<td>PS.2A</td>
<td>Late Eocene – early Oligocene to Holocene</td>
<td>Marine: massive and friable diamictite</td>
<td>Glacial, prograding sequences</td>
</tr>
<tr>
<td>PS.2B</td>
<td>Early Cretaceous and younger (?)</td>
<td>Non-marine: sandstone, siltstone, coal.</td>
<td>Fluvialite and alluvial plain</td>
</tr>
<tr>
<td>PS.3</td>
<td>Early Mesozoic (?)</td>
<td>Non-marine sediments (?)</td>
<td>Alluvial Plain (?)</td>
</tr>
<tr>
<td>PS.4</td>
<td>late Palaeozoic (?) to early Mesozoic (?)</td>
<td>Non-marine: siltstone and redbed sandstone.</td>
<td>Preglacial, fluvial basin</td>
</tr>
<tr>
<td>PS.5</td>
<td>Precambrian (?) [sic]</td>
<td>Metamorphic rocks</td>
<td>Basement</td>
</tr>
<tr>
<td>PS.6</td>
<td>Precambrian (?) [sic] and Mesozoic (?)</td>
<td>Metamorphic and intrusive rocks.</td>
<td>Intruded basement.</td>
</tr>
</tbody>
</table>

Table 2.2 – Prydz Bay acoustic units – interpretations [after Cooper et al. 1991]

Hambrey (1991) discussed the structure and dynamics of the Lambert Glacier and concluded that although some workers suggest that the Antarctic Ice sheet may have been subject to periodic surging (Hollin 1969, Budd and McInnes 1978), there is no structural evidence to support it. This means that if surging does occur it occurs at the same time throughout the whole system or that it occurs at a period greater than the residence time of the ice in the system. Hambrey (1991) also concluded that because distinctive rock types are known in the Lambert Basin, and on the evidence of ice flow lines, not much ice from the western side of the Lambert Glacier reached the Prydz Bay drill sites. Jenkins and Alibert (1991) conversely contend that petrographic studies of rock clasts in the glacial diamictons of Prydz Bay indicate that the predominant ice supply was from the Lambert Glacier with no compelling evidence of clastic supply from the Vestfold Hills and Larsmann Hills.
Solheim et al. (1991a&b) described the geotechnical properties of glacigenic sediments and interpreted the stepwise downcore increase in pre-consolidation stress of the sediments as representing periods of increased action of grounded glaciers covering the entire shelf.

**AAMBER 2, 1990/91**

Thirty three samples were collected during the austral summer in early 1991 as part of the Australian Antarctic Marine Biological Ecosystem Research (AAMBER) II cruise. The samples were collected on an opportunity basis and the results published in Franklin (1991, 1993) These results are re-interpreted within this study.

**Stratigraphy of Prydz Bay**

Hambrey (1991) produced a stratigraphic summary of the 5 ODP Drill holes which is reproduced below for the uppermost units. A generalised stratigraphy, based mainly on the drilling from Site 742, of the glacially associated sediments is as follows.

**Unit I**

Unit I (0 to 5.4 mbsf) comprises unconsolidated diamict overlying firm diamicton and is divided into two sub units. The uppermost sub-unit IA consists of 7 cm of soupy olive diatomaceous sand/silt. Its principal constituents are quartz-feldspar sand (70%), diatoms (20%), and clay (10%) and has a Quaternary age. Sub-unit IB is a soft, grey, massive diamicton, the principal constituents of which (sand, silt, and clay) are present in roughly equal proportions.

**Unit II**

Unit II is a soft, normally consolidated sandy mud with 10-20% diatoms and a few shells. Both units I and II have a Quaternary biostratigraphic age and a Holocene age is considered most likely.
At site 742 Unit II (5.4 to 115.2 mbsf) is a dark grey homogenous, massive diamictite with up to 15% gravel. Poor recovery characterises this interval.

**Unit III**

Unit III was very poorly recovered, but granite and gneissic rock fragments and the drilling behaviour suggest the same lithology as sub-unit 1B at Site 740, namely a probably overconsolidated boulder rich diamicton. Its age is unknown but a Pleistocene age is suggested.

**Unit IV**

Unit IV. Consists of preglacial sand silt and conglomerate of probable Albian age.

**Other areas.**

Although the glacial history of West Antarctica is currently the subject of debate (Webb and Harwood 1987, 1991, Clapperton and Sugden 1990), seismic profiles of the continental shelf of the Antarctic Peninsula reveal an architecture similar to Prydz Bay, with basal sediments in seaward accreting sediment wedges bounded by widespread unconformities. A prolonged period of glacial erosion and deposition has resulted in the present deep and rugged topography on the inner shelf (Anderson et al. 1991). A history based on stratigraphy from the CIROS-1 core from the Ross Sea also broadly confirms the general Antarctic shelf sedimentary models presented by Hambrey (1991) although the authors are cautious about extrapolating the glacial chronology from McMurdo Sound to the rest of the Antarctic Continent (Barrett et al. 1991). Sediments from George V continental shelf (Domack et al. 1991a) also closely resemble Prydz Bay sediments.

**The Vestfold Glaciation**

Glacially polished, striated and fractured rocks, roches moutonées, and ice plucking occur across the hills at all elevations. These features are consistent with ice flow over the rock surfaces but there is no evidence to suggest that they were produced by any glacialiation older than the most recent in the terminal Pleistocene. Adamson and Pickard (1986a) attribute these features to the Vestfold Glaciation which is the local name for this Late Wisconsinan expansion of the Antarctic ice sheet. The ice sheet reached at least to the offshore islands where there are moraine ridges. Indicated ice movement is remarkably uniform and indicates that the ice moved out in a WNW direction.

**Minor Valley Glaciation**

Adamson and Pickard (1986a) note an absence of glacial striae which parallel existing valleys. In the area of relatively high relief in the north east of the hills, a minor valley glacier appears to have flowed down a small cross valley in the late Holocene.

**Holocene marine sedimentation**

The large number of saline lakes in the Vestfold Hills, as well as the abundance of fossil molluscs and other marine biota indicate that marine sedimentation was contemporaneous with and concurrent with Holocene deglaciation of the Vestfold Hills (Adamson and Pickard 1986a).
Holocene deglaciation

During the Holocene, the ice sheet melted back to its present position and exposed the hills, the land emerged from beneath the ice, the sea invaded the low lying areas, and the land rose in isostatic adjustment. There is no clear spatial pattern of the retreat, however, indications suggest that the long periods of negligible retreat punctuated by episodes of very rapid melt (Adamson and Pickard 1986a).

Holocene Glaciation

In the South east corner of the hills there is a northwards directed set of glacial striae superimposed on pre-existing WNW striae. The northwards moving ice was a minor readvance after the retreat of the ice sheet. This minor advance, the Chelnok Glaciation, left few traces although minor drapes of till occur near its northern limit and a set of linear moraines occur, all within 1 km of the Sørsdal Glacier which may represent lateral moraines associated with the advance between 3000 and 1500 a.
CHAPTER 3 - MATERIALS AND METHODS

This study considers data collected from more than 500 sites in Prydz Bay, including sediment samples from grabs and cores, photographs of the sea floor, dissolved and particulate carbonate from the waters of the bay, and wind blown sediments collected in aeolian sand traps. The sampling programs conducted by the author account for more than 300 of these samples. Other samples used or referenced include a previous sampling program by the author in 1990, cores collected by ODP Leg 119, a sample collected by HMS Challenger in the 1874, and many others. The location of all samples collected, used or referenced is at Appendix A.

SAMPLE COLLECTION

KROCK

Cores and Grabs

Coring was conducted using an 8 cm diameter gravity corer (Figure 3.1) constructed by the Australian Geological Survey Organisation (AGSO). Six metre core barrels were used for most sites but in areas of hard bottom, 3 m barrels were used. The corer weighed 1100 kg when the 6m barrel was used. It was deployed on 20 mm steel wire rope on the starboard trawl winch. Core cutters were case hardened steel. Cores were stored by removal of the PVC lining which was then cut into ~ 1 m lengths and the segment ends sealed. Cores were stored at 4° C. Thirty-five grab samples were taken using two galvanised steel Van Veen grabs (Figure 3.2) built by AGSO. These grabs weigh ~ 80 kg and have a gape of 0.56 m by 0.45 m. They were deployed on a 6 mm diameter steel wire rope on the hydrological winch in the CTD room of the RSV Aurora Australis.
Figure 3.1 - Offshore gravity corer deployed from the RSV *Aurora Australis* in Prydz Bay.
Materials and Methods

Figure 3.2 - Large Van Veen grab sampler deployed from the RSV Aurora Australis in Prydz Bay

![Eckmann Grab with jaws open and ready for deployment](image)

![Eckmann Grab after release of jaws](image)

**Figure 3.3** - Small Eckmann grab sampler deployed in the shallow marine environment near the Vestfold Hills.
**Bottom Photographs**

Bottom photography was conducted using a deep sea camera with a trigger mechanism activated when a weight on a 5 m line came in contact with the sea floor. Bottom photographs are not oriented. The camera used 35 mm film and a powerful flash unit and was deployed on the bottom of an Armor-Brown CTD rosette frame and the resultant photographs show approximately 4 m² of the sea floor.

Bottom photographs were taken on an opportunity basis throughout Prydz Bay over a number of cruises. Bottom photographs are reproduced at Appendix C. In the case of diamicts and turbates the individual clasts are easily recognised, however, the fine grained sediments were somewhat more difficult to identify, and there is insufficient resolution to differentiate between fine sands and muds in the photos, which were taken from a height of approximately 5 m above the sea floor. In these instances, identification was problematic, however, some grab samples were taken at some of the sites and it was possible to ground truth some of the bottom photos. In some other photos, the trigger weight seems to disturb the fine grained sediments, causing a cloud of fine material to resuspend. There are no photographs which represent the basement or relict till facies.

Without good ground truthing of photographic sites, or the ability to orient the photographs, the photography of sediments on the sea floor is of limited use because it is not possible to determine the sediment characteristics in any but the broadest sense. Nonetheless, it is possible to discern many features such as large clasts, encrustation on clasts, faunal abundance, preservation of shelly material in some instances, and even some information on the current regime.

**Benthic Sled**

Some samples were obtained from an epi-benthic sled designed to sample benthic organisms for another program. The sled provided pebbles and boulders from some sites and fine sediments from others.
Sample numbers

Sample numbering is based on the Antarctic Division marine science Data Logging System Conventions as follows:

Voyage acronym/Station number/ Sample number

The voyage acronym was “KROCK” denoting Krill and Rock indicating krill biology and geoscience as the two voyage determining programs. Station number is a unique integer defined for each station occupied during the voyage and Sample number is a consecutive alpha numeric system with grabs prefixed by GR and cores prefixed by GC, followed by an integer. For this document Grabs and Cores are identified in the text by the sample number only. Appendix A lists all sample numbers, locations, and other relevant data.

Davis Sampling Program

Samples were numbered sequentially from 93001 to 93152. Four samples collected at the end of the program are numbered 94001 to 94004. Two samples collected from Ace Lake are numbered Ace Lake 15 and Ace Lake 25 and were collected at depths of 15 and 25 m respectively.

Offshore

Two sorts of sampling equipment were used, a 1 m gravity core with a 40 kg weight on the top, and a small stainless steel Eckmann grab sampler (Figure 3.3), both constructed by AGSO. A 30 cm hole was drilled in the sea ice with a jiffy ice drill with a 30 cm auger, and the water depth determined with a portable Humminbird™ echo-sounder. If the depth did not exceed the available length of line, a large aluminium tripod with a modified, hand operated boat recovery winch, as used on small boat trailers, was erected over the hole and the corer or grab lowered to the sea floor.
**Sea ice program**

The grab was lowered manually, until it reached the sea floor, whereupon the grab was raised to approximately 2 m from the bottom and then dropped into the sediment. The slack was then taken in and the messenger deployed to fire the spring loaded jaw mechanism. The grab was then recovered manually. After the sea ice had broken out for the summer, the grab was deployed from the side of the Australian Antarctic Division’s 7 m work boat *Southern Comfort.*

**Other Samples**

Sample 93017 was collected from a dirty band in a grounded iceberg. The ice was melted and the water was centrifuged at 2000 rpm for 10 min to ensure that the sediment settled. The water was decanted and the sediment dried and weighed.

**Onshore**

**Mapping and sample collection**

Onshore samples were recovered using a small trowel, and mapping was conducted on foot with outcrop details being inscribed on colour air-photos and later transcribed onto composite maps produced from air photo mosaics. In areas away from those of primary interest to this study, mapping was direct from the air photos (26 January 1979 3050 m height Hughes 500 helicopter using Hasselblad EL500, runs 3-8, 10-13, & 18), with such areas identified in the confidence diagram on each geological map.

**Heidemann Valley Drill Holes**

Heidemann Valley coring was conducted using a portable percussion coring system that provides continuously cored sections with a core diameter of 35 mm. It consists of a Wacker motor driven vibrating unit atop a series of one metre long core barrels and a core cutting device. It was built for, designed, and owned by Nick Poltock of Devonport, Tasmania. Coring proceeded in 50
cm stages until progress was stopped by either a large boulder or basement rock. 50 cm lengths of core were removed from the core barrel and were logged and sampled immediately. The core was cut into 10 cm lengths and the outer 0.5 to 1 cm, which was presumed to have been contaminated by the core barrel during the coring process, was peeled off and discarded. The remaining material was packaged, labelled and returned to Australia for further analysis.

Samples from this program are numbered DHN-nnn where N is the number of the drill hole and nnn is the start depth of the 10 cm core section.

**Aeolian sand traps**

The Heidemann Valley aeolian sand traps (Figure 3.4) were erected at the seaward end of Heidemann Valley. Five evenly spaced sites were selected along Heidemann Beach. The first, third and last sites consisted of single traps at 50 cm height above the ground. The second site was a tower of three traps at 50, 100, and 150 cm. The fourth site was a tower of four traps at 30, 50, 100, and 150 cm height above ground level.

The individual traps were constructed of 15 cm diameter PVC pipe, with two 10 cm diameter holes drilled opposite each other towards the top of the trap. The hole intended to face to windward was left open while the hole towards the rear was covered in 63 μm gauze.

The traps are similar in design to that proposed by Leatherman (1978), whose paper details a number of other sand trap designs. Leatherman’s (1978) design was intended to be buried in the ground and was to have sampled both surface creep and saltation. Its obvious shortcomings relate to it extending only 30 cm above ground level and therefore missing any debris transported above that level. Traps of this design were deployed in the Vestfold Hills but were abandoned because it quickly became obvious that most of the debris transport was happening at heights above 30 cm.
Figure 3.4 - Aeolian sand trap design (A) and tower configuration (B)
Materials and Methods

The trap that was finally used was designed by the author and constructed out of PVC pipe. The limitations of the trap are that it points in one direction only and therefore the amount of sediment collected is greatly reduced when the wind is not blowing directly into the mouth of the trap. There may also be some debris lost from the trap by bouncing out through the open top of the trap, and during high wind events, eddies set up inside the trap may remove fine grained material caught during lower wind events. It is unlikely that the trap would over-sample. The samples are numbered VHN-X where N is the number of the tower and X is the sequential position of the trap from the ground.

ANALYSES

Gravel, Sand, and Mud

Gravel, sand and mud (GSM) determinations were conducted on most samples. The sediments in the study area are of two main types. The most common sediment type is fine grained mud with varying minor amounts of sand and gravel. The second main type of sediments is that with significant amounts of all fractions; gravel, sand and mud.

Sediment was prepared for grain size analysis by disaggregating it in dilute H₂O₂ overnight. The sediment was then sieved to separate gravel (> 2 mm), sand (2 mm - 63 μm) and fines (< 63 μm). The separate fractions were then dried and then weighed.

Grain Size

Sand grain size distribution of the whole sediment was determined in the rapid sediment analyser (settling tube) at AGSO. Fine grain size distributions were determined in a Sedigraph 5100 at AGSO. Subsequent attempts to integrate the mud data with the sand data failed as the algorithms used by each machine presented the results in completely different manners. Consequently, the mud grain size analysis should be viewed qualitatively.
Materials and Methods

Biogenic Silica

Biogenic silica was determined using the method of Mortlock and Froelich (1989) which is a modified molybdate-blue spectrophotometric process of some complexity. Working silica standard solutions were prepared at concentrations of 0, 2.67, 6.23, 8.9, and 12.46 mM SiO₂. Other solutions required are at Table 3.1. The method, using doubly deionised water at all stages, requires the following steps:

Step 1. Sediment is freeze dried and gently disaggregated using a mortar and pestle. A small amount of sample is weighed into a vial such that after all of the following steps, the concentration of Si at step 3 is in the range 700 to 9000 μmol (between 25 mg for high Si samples to 200 mg for low Si samples).

Step 2. This aliquot is then reacted with 5 ml of 10% H₂O₂. After 30 min 5 ml of 1N HCl solution is added and the aliquot sonified for 5 min, capped and left to react for a further 30 min. 30 ml of water is added and the sample centrifuged at 4200g for about 5 min. The supernatant is decanted to remove residual acid and peroxide, and the residue dried overnight.

Step 3. Exactly 40 ml of 2M Na₂CO₃ solution is added to the sample which is then recapped, well mixed and placed in a constant temperature water bath at 85°C and mixed to resuspend solids after 2 and 4 h. After 5 h the sample is removed and immediately centrifuged for 5 min at 4200g. 20 ml of the clear supernatant is then drawn off and transferred to a polyethylene scintillation vial and stored for subsequent analysis.

Step 4. 17.5 ml of molybdate working solution is dispensed into clean, dry polyethylene reaction vessels. At 30 s intervals, 125 ml of sample, standard, or blank, is added to each vessel and immediately mixed by swirling. The vessels are capped and allowed to react for exactly 20 min. At 30 s intervals 7.5 ml of reducing working solution is added and swirled and left to stand for at least 12 h.
Step 5. Absorbances are read in a spectrophotometer at a wavelength of 812 nm. Si concentration is graphically determined from a standard curve produced with standards of known concentration.

<table>
<thead>
<tr>
<th>Solution</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Molybdate Stock Reagent</td>
<td>16g of Ammonium Paramolybdate is dissolved in 1000 ml water. Stable indefinitely.</td>
</tr>
<tr>
<td>Metol Sulfite Stock Reagent</td>
<td>12g anhydrous Sodium Sulfite is dissolved in 1000 ml water then 20g of Metol is dissolved into that solution. The resulting solution is filtered through a No. 1 Whatman filter paper and stored in amber glass bottles with glass stopper. This solution is stable for 1-2 months but should be discarded if a white precipitate forms.</td>
</tr>
<tr>
<td>Oxalic Acid Stock Reagent</td>
<td>60g of Oxalic Acid Dihydrate is dissolved in 1000 ml of water. Stable indefinitely.</td>
</tr>
<tr>
<td>Sulfuric Acid Stock Reagent</td>
<td>300 ml of conc Sulfuric Acid is mixed with 700 ml water. Stable indefinitely.</td>
</tr>
<tr>
<td>Hydrochloric Acid Stock Reagent</td>
<td>48 ml of conc Hydrochloric acid is mixed with 952 ml water. Stable indefinitely.</td>
</tr>
<tr>
<td>Molybdate Working Solution</td>
<td>Molybdate Stock Reagent, HCl Stock Reagent, and water are mixed in the ratio 1:1:5. This solution is stable for up to 6 hours</td>
</tr>
<tr>
<td>Reducing Working Solution</td>
<td>Metol Stock Reagent, Oxalic Acid Stock Reagent, and Sulfuric Acid Stock Reagent are mixed in the Ratio 1:1:1. This solution is stable for up to 4 hours.</td>
</tr>
</tbody>
</table>

Table 3.1 - Working solutions for molybdate-blue spectrophotometry

Biogenic Calcite

The percentage of CaCO₃ in these samples was determined a vacuum gasometric technique as follows. The sample is crushed to a powder finer than 63 µm and dried in an oven at 110 °C. Between 300 and 700 mg of sample is placed in a reaction vessel equipped with a side arm. 5 ml of concentrated phosphoric acid is introduced into the side arm and the chamber is capped. The reaction chamber is then evacuated and the acid is tipped into the sample. The reaction is allowed to continue for 1.25 h and the sample is agitated every 20 min. On completion, the pressure in the reaction chamber is determined.
Materials and Methods

Trace Element Geochemistry

The samples were analysed for trace elements by Ian Snape at the University of Edinburgh. Samples were analysed for trace elements using a Philips PW1480 wavelength-dispersive, automatic, sequential X-ray fluorescence spectrometer fitted with Rh anode side-window X-ray tubes. Samples were analysed twice on different programs for 17 elements on a basalt calibration program and for 19 trace elements on a generalised program.

Sub-samples for trace element geochemistry were taken from the sediment samples collected. These aliquots were weighed wet then dried in an oven at 60°C. The dried sediment was then milled in a rotary tungsten carbide mill until reduced to a fine homogenous powder.

Pellets were prepared for trace-element analysis as follows. Approximately 6 g of rock powder was mixed with 4 drops of binding agent ('mowiol' 2% PVA in distilled water). The mixture was placed in a steel mould, surrounded and backed by boric acid powder, and compressed at 8 tons to form a 40 mm diameter pellet using a hydraulic press.

Water Content

Water content was determined by freeze drying a preweighed aliquot of sediment and noting the weight loss. Freeze drying was conducted using a freeze drier at the Commonwealth Scientific and Industrial Research Organisation (CSIRO) Marine Laboratory in Hobart Tasmania. Sediments were weighed daily until there was no further weight loss. No salt correction has been applied to the dry weight.

Radiogenic Carbon

The samples (Table 3.2) investigated in this study were collected over several years and include the following.
a. Carbonate samples from core tops and surface sediments (when the sediment/water interface in the grab sample was collected undisturbed). These samples were stained (Rose Bengal) to identify living organisms and were then wet sieved. The > 63 μm fraction was retained and oven dried. Carbonate fragments were manually picked using a microscope and a fine brush and then cleaned in 15% H2O2 overnight, and dried. Samples were analysed by either the Accelerator Mass Spectrometer (AMS) facility at the Australian Nuclear Science and Technology Organisation (ANSTO) for analyses prefixed by “OZA” or at the AMS facility of the New Zealand Institute of Geological and Nuclear Sciences Limited (IGNS) for analyses prefixed “NZA” (*Neogloboquadrina pachyderma*).

b. Total Organic Carbon (TOC) from core samples was analysed by the Institute of Geological and Nuclear Science (IGNS) in New Zealand. The samples were ground and sieved and then treated with hot solutions of HCl and alkaline pyrophosphate to remove CaCO3.

c. Particulate Organic Carbon (POC) from surface water samples was filtered from a continuous sampling facility which drew water from beneath the RSV *Aurora Australis*. POC analyses were conducted at the ANSTO AMS facility.

d. Dissolved Inorganic Carbon (DIC) was collected from a coastal sampling site (depth of 15 m) in the vicinity of O’Gorman Rocks approximately 1 km offshore Davis Station, over a 12 month period. DIC samples were analysed at the ANSTO AMS facility.
<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Analysis number</th>
<th>Latitude °S</th>
<th>Longitude °E</th>
<th>Water Depth</th>
<th>Core Depth</th>
<th>Material</th>
<th>Max Age</th>
<th>Age</th>
<th>Min Age</th>
<th>Error</th>
<th>δ¹³C</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>GR31</td>
<td>OZA159</td>
<td>67.27</td>
<td>65.42</td>
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**Particular Organic Carbon Samples**

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**Dissolved Inorganic Carbon Samples**

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**Total Organic Carbon Samples**

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Notes

#1 $\delta^{13}$C figures from Rao et al. (1996)

#2 $\delta^{13}$C figures from John Gibson (Personal Communication)

#3 Actual determination of $\delta^{13}$C

Table 3.2 - Results of radiometric carbon dating.
CHAPTER 4 - CROSS COASTAL PROCESSES

Introduction

The major processes involved in the movement of sediment from the Antarctic continent to the marine environment are the erosion and incorporation of continental material into glacial ice and the subsequent movement of that ice offshore. These particular processes are discussed in the following chapters. Deposition of diamicton as moraines fringing continental ice cliffs provides source material for the processes discussed in this chapter. Similarly, meltout at the base of grounded coastal ice cliffs deposits debris in the nearshore region which can then be affected by processes discussed in Chapter 5. The aeolian transport of granular material directly across the coast, or via melting snow drifts accounts for some cross-coastal transport. Fluvial transport of debris by catastrophic failures of ice dams may be important and, locally, water movement after such events appears to incise glacially deposited valley fills.

GLACIAL

Ice flows from the interior of the continent towards the coastline and incorporates terrigenous material into a basal layer which can be up to tens of metres thick (Anderson et al. 1980, Hambrey 1991). Glacial sediment load is discussed in depth in Chapter 6. In the study area, glacial ice debouches through glaciers and/or ice shelves, forms coastal ice cliffs, or terminates on exposed rock such as the Vestfold Hills. These three glacial modes result in very different sedimentary patterns. The sedimentary processes associated with glacial and ice shelf formation are discussed in Chapter 6.

Ice cliffs are maintained by thermal erosion or wave action (Chinn 1990, Fitzsimmons and Colhoun 1995). Cliffs form when basal ice velocities are significantly lower than surface velocities and where ice is removed by processes that generate or maintain a cliff form, including erosion of the ice
face by wave action (Fitzsimmons and Colhoun 1995). Chinn (1986, 1991) argued that when the semi-rigid zone of a glacier (a brittle layer ~ 20 m in thickness) is grounded, it obstructs ice flow and determines the formation and position of a cliff, or ramped terminus.

Coastlines with grounded ice cliffs are present to the north of the Vestfold Hills. Measured water depths suggest that the water at the face is about 50 m deep based on the slope between localities 93001 and 93002, and their distance from the ice cliffs.

Small sections of glacial ice can be seen having fallen away from the cliff face, probably due to wave action which maintains the cliffs in a state of equilibrium. Summer wave action undercuts the cliff face and sections fracture or slump away from the face. In the study area the ice cliffs were in the order of 50 m high, with a total thickness of ice of up to 100 m. With up to 50 m of ice below the water line it is likely that tidal action may also weaken the relatively buoyant ice in the cliff and assist in weakening the structure. No samples were collected at sites 93001 or 93002 but the recovery of sessile marine organisms with holdfasts indicates either crystalline basement or boulders in this area.

Continental glacial ice which terminates onshore is found in the eastern part of the Vestfold Hills. In this area, the ice edge is in relative equilibrium with deposits of basal meltout fringing the ice sheet (Fitzsimmons and Colhoun 1995). The rate of sediment transport in glacial ice is discussed in Chapter 6. Sample number 94004 was collected from one such basal meltout deposit near Platcha Hut (Vestfold Hills 1:100000 2nd edition GR 982979), to which the ice sheet abutted, where the glacial ice in contact with the hill melts and deposits the debris load.

Two samples were collected from locations which represent the supply of terrigenous material to Prydz Bay. Sample number 93017 was a sample of debris rich glacial ice from a grounded iceberg just offshore the Vestfold Hills (Figure 4.1). 32.71 g of sediment were collected from 3.12 l of melted glacial ice or ~ 1% by mass which is lower than other values of between about 5%
and 8% reported by Anderson et al. (1980) and Fitzsimmons and Colhoun (1995), and significantly lower than the 40-50% reported by Shaw (1977) for dry based glaciers.

The grain size distribution of these samples (Figure 4.2) is difficult to interpret. While it may be expected that these sediments should be completely unsorted, the deficiency in coarse silt is similar to grain size distributions from icebergs reported by Anderson et al. (1980). Anderson et al. (1980) remarked that bimodality seen in debris rich layers of ice in several Antarctic icebergs was similar to that produced by mechanical abrasion in the laboratory (Rogers et al. 1963) and that this attests to the effectiveness of comminution through glacial abrasion.

In summary, a sample of basal glacial ice collected in the vicinity of the Vestfold Hills was depauperate in englacial debris when compared to other reported values but it and another sample of basal meltout debris adjacent to the ice sheet displayed grain size distributions consistent with others reported from Antarctica for similar material. While they were not sampled, basal meltout deposits at the base of grounded coastal ice cliffs are probably similar to the basal meltout sample collected, notwithstanding they are probably modified by current action in the shallow marine environment.

AEOLIAN

The aeolian transport of terrigenous material is important only in the vicinity of ice free areas where there are surface rocks or sediment able to be eroded, entrained, and transported. In Antarctica, such places are few, and in Prydz Bay, the most important of these is the Vestfold Hills.
Figure 4.1 - Grounded iceberg (68 25 S, 78 05 E) showing debris rich layer "a", a large visible clast "b", and some evidence of deformation of ice "c".
Figure 4.2 - Gravel, sand, and mud proportions and grain size distributions of samples 93017 and 94004
The results of aeolian transport can be readily viewed in the Vestfold Hills, especially after high winds. Material is entrained from the existing deposits of glacial and glaciomarine sediments which occur throughout the Vestfold Hills. This material is transported until deposited in the lee of hills (Figure 4.3), in snowdrifts (Figure 4.4), in the lee of smaller features such as boulders (Figure 4.5), onto the sea ice (figure 4.6) or directly into the sea when it is ice free. In other areas, wind may also deposit material onto glaciers or ice sheets, but as the prevailing winds around the Vestfold Hills blow away from the ice sheet, this is not important in the area.

To quantify the aeolian transport rate in the Vestfold Hills, a series of aeolian sediment traps was deployed along Heidemann Beach (Figure 4.7) for 11 months. Traps were placed at 5 sites, evenly spaced along the beach. Three sites (Sites 1, 3, & 5) contained single traps at 50 cm height above ground level, Site 2 had three traps in a tower at 50, 100, and 150 cm height above ground level, and Site 4 had four traps at 30, 50, 100, and 150 cm height above ground level. The traps were emptied approximately monthly on the dates referenced. All periods are identified by the date the traps were emptied.

**Aeolian transport rate**

The monthly mass of sediments (Figure 4.8) collected shows three periods of high debris transport and, at those sites with traps at more than one height, the vertical decrease in sediment collected is clear. Monthly debris yields at Site 4 (Figure 4.9) demonstrate that the selection of the heights at which the sediment traps were placed did not identify the level at which maximum transport of material occurred, except in one particular month in which unusually large amounts of sediment were collected. In that month, a maximum yield of 118.91 g was collected at 50 cm (trap number 4-2).
Figure 4.3 - Snow and debris (a) accumulating in the lee of Anchorage Island (GR 750915 Vestfold Hills 1:100000 2nd edition. The highest point on the island is 70 m.)

Figure 4.4 - Debris (a) rich snow drift in the lee of a hill at GR 788909 - Vestfold Hills 1:100000 2nd edition. Debris is deposited in the meltpool in the foreground (b).
Figure 4.5 - Small deposit of aeolian sand in the lee of a large boulder in Heidemann Valley, Vestfold Hills.

Figure 4.6 - Small piece of stranded sea ice with significant amounts of aeolian sand and grit on the top. Item photographed on Heidemann Beach.
Figure 4.7 - Sites of aeolian sand traps on Heidemann Beach.
Figure 4.8 - Area plots of monthly aeolian debris transport at the 30, 50, 100, and 150 cm levels. Figures represent the mass (g) of debris collected in each trap during the period. The key explains the features of each graph.
Figure 4.9 - Monthly aeolian debris yields for Site 4.
All of the plots show a roughly logarithmic decrease in sediment yield with increasing distance from ground level (Figure 4.8). In September 1994 (period ending 6 October 1994), when the maximum yield was at 50 cm, the sediment yield decreased below 50 cm, probably due to the lower wind velocities in the vicinity of the ground in the boundary layer.

While the curves of sediment yields rise and fall in an apparently regular fashion the relationships are complex and for the purpose of this study, an empirical method of estimation and a set of assumptions are applied.

Assumptions

The rate at which sediments are normally transported is low for most months of the year. More than half of the sediment transported moved during the period 4 September to 6 October 1994. More than half the remainder was transported from 6 June to 5 July 1994. Wind data collected at Davis (Figure 4.10) indicate that during these two months, a number of particularly high wind events occurred. These events probably resulted in the movement of most of the sediments. It is assumed that such high wind events are the norm and that this data set represents a typical year in respect of wind events and sediment transport.

The prevailing wind in the Vestfold Hills is from the north-east. The sediment traps were placed with their collection holes directed to the north east. Directional variation of the wind is therefore not taken into account by the sampling method, however, the wind strengths reported in Figure 4.10 represent the vector component in the direction of the trap mouth. Onshore winds blowing from any direction between west and true north will not transport sediment offshore.

Aeolian transport can only occur where there is material available to be transported. Loose material is common throughout the Vestfold Hills and this study assumes that aeolian transport occurs uniformly across the study area.
Figure 4.10 - Three hourly wind velocity (component in direction of sand trap mouth) for the twelve collection periods (each approx 30 days). The three months of high debris transport are in blue. Individual high wind events during those months are in red and marked A, B, C, & D. Wind speed measured at 10 m above ground level.
### Daily sediment transport

| Height (cm) | 30 cm | 50 cm | Trap | 1 | 1 | 2 | 1 | 3 | 1 | 4 | 2 | 1 | 5 | 1 | Av | sd |
|------------|-------|-------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|
| Period 1   | 0.216 | 0.005 | 0.051 | 0.037 | 0.048 | 0.039 | 0.034 | 0.018 |
| Period 2   | 0.051 | 0.011 | 0.008 | 0.011 | 0.009 | 0.025 | 0.013 | 0.007 |
| Period 3   | 0.640 | 1.619 | 0.964 | 0.530 | 0.206 | 0.715 | 0.819 | 0.515 |
| Period 4   | 0.004 | 0.006 | 0.003 | 0.005 | 0.007 | 0.004 | 0.005 | 0.001 |
| Period 5   | 1.227 | 2.431 | 1.506 | 2.196 | 0.445 | 1.160 | 1.547 | 0.801 |
| Period 6   | 0.072 | 0.399 | 0.186 | 0.373 | 0.017 | 0.165 | 0.228 | 0.158 |
| Period 7   | 0.009 | 0.079 | 0.008 | 0.006 | 0.023 | 0.029 | 0.029 | 0.030 |
| Period 8   | 2.629 | 3.618 | 1.440 | 1.170 | 3.716 | 1.615 | 2.312 | 1.248 |
| Period 9   | 0.012 | 0.012 | 0.006 | 0.017 | 0.004 | 0.004 | 0.008 | 0.006 |
| Period 10  | 0.077 | 0.041 | 0.169 | 0.191 | 0.012 | 0.065 | 0.095 | 0.080 |
| Period 11  | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| **ANNUAL** | **1828.64** | **3002.12** | **1587.57** | **1658.16** | **1666.13** | **1431.90** | **1160.18** | **640.28** |

SD as %

| Height (cm) | 100 cm | 150 cm | Trap | 4 | 3 | 2 | 3 | 4 | 4 | Av | sd |
|------------|-------|-------|------|------|------|------|------|------|------|------|------|------|------|------|------|------|
| Period 1   | 0.005 | 0.015 | 0.10 | 0.010 | 0.011 | 0.009 | 0.002 | 0.001 |
| Period 2   | 0.001 | 0.001 | 0.10 | 0.001 | 0.001 | 0.001 | 0.001 | 0.000 |
| Period 3   | 0.132 | 0.024 | 0.078 | 0.077 | 0.046 | 0.022 | 0.034 | 0.017 |
| Period 4   | 0.003 | 0.003 | 0.003 | 0.000 | 0.002 | 0.004 | 0.002 | 0.001 |
| Period 5   | 0.291 | 0.090 | 0.10 | 0.012 | 0.012 | 0.002 | 0.070 | 0.060 |
| Period 6   | 0.017 | 0.004 | 0.10 | 0.009 | 0.020 | 0.009 | 0.015 | 0.008 |
| Period 7   | 0.010 | 0.024 | 0.017 | 0.010 | 0.012 | 0.012 | 0.012 | 0.001 |
| Period 8   | 1.825 | 1.295 | 1.560 | 0.374 | 0.883 | 0.983 | 0.733 | 0.212 |
| Period 9   | 0.001 | 0.003 | 0.002 | 0.000 | 0.005 | 0.004 | 0.004 | 0.001 |
| Period 10  | 0.013 | 0.001 | 0.007 | 0.008 | 0.003 | 0.001 | 0.002 | 0.002 |
| Period 11  | 0.003 | 0.003 | 0.003 | 0.000 | 0.003 | 0.001 | 0.002 | 0.001 |
| **ANNUAL** | **839.87** | **535.23** | 687.55 | 215.41 | 397.57 | 243.72 | **520.65** | **108.79** |

SD as %

|  | 31.33 | 33.93 |

**Table 4.1 - Average daily aeolian debris transport (g)**
### Cross Coastal Processes

#### Daily sediment transport

<table>
<thead>
<tr>
<th>Height (cm)</th>
<th>30 cm</th>
<th>50 cm</th>
<th>100 cm</th>
<th>150 cm</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Trap</strong></td>
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<td>4</td>
<td>1</td>
<td>2</td>
</tr>
<tr>
<td><strong>Period 1</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.216</td>
<td>0.005</td>
<td>0.051</td>
<td>0.037</td>
</tr>
<tr>
<td><strong>Period 2</strong></td>
<td>0.051</td>
<td>0.056</td>
<td>0.008</td>
<td>0.011</td>
</tr>
<tr>
<td><strong>Period 3</strong></td>
<td>0.640</td>
<td>1.619</td>
<td>0.964</td>
<td>0.530</td>
</tr>
<tr>
<td><strong>Period 4</strong></td>
<td>0.004</td>
<td>0.906</td>
<td>0.003</td>
<td>0.005</td>
</tr>
<tr>
<td><strong>Period 5</strong></td>
<td>1.227</td>
<td>2.431</td>
<td>1.506</td>
<td>2.196</td>
</tr>
<tr>
<td><strong>Period 6</strong></td>
<td>0.072</td>
<td>0.399</td>
<td>0.186</td>
<td>0.373</td>
</tr>
<tr>
<td><strong>Period 7</strong></td>
<td>0.009</td>
<td>0.079</td>
<td>0.008</td>
<td>0.006</td>
</tr>
<tr>
<td><strong>Period 8</strong></td>
<td>2.629</td>
<td>3.618</td>
<td>1.440</td>
<td>1.170</td>
</tr>
<tr>
<td><strong>Period 9</strong></td>
<td>0.012</td>
<td>0.062</td>
<td>0.006</td>
<td>0.017</td>
</tr>
<tr>
<td><strong>Period 10</strong></td>
<td>0.077</td>
<td>0.041</td>
<td>0.169</td>
<td>0.191</td>
</tr>
<tr>
<td><strong>Period 11</strong></td>
<td>0.000</td>
<td>0.000</td>
<td>0.000</td>
<td>0.000</td>
</tr>
<tr>
<td><strong>ANNUAL</strong></td>
<td>1828.64</td>
<td>3002.12</td>
<td>1587.57</td>
<td>1658.16</td>
</tr>
<tr>
<td><strong>SD as %</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th><strong>Height (cm)</strong></th>
<th><strong>100 cm</strong></th>
<th><strong>150 cm</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Trap</strong></td>
<td><strong>Av</strong></td>
<td><strong>sd</strong></td>
</tr>
<tr>
<td><strong>Period 1</strong></td>
<td>0.005</td>
<td>0.014</td>
</tr>
<tr>
<td><strong>Period 2</strong></td>
<td>0.001</td>
<td>0.001</td>
</tr>
<tr>
<td><strong>Period 3</strong></td>
<td>0.132</td>
<td>0.024</td>
</tr>
<tr>
<td><strong>Period 4</strong></td>
<td>0.003</td>
<td>0.003</td>
</tr>
<tr>
<td><strong>Period 5</strong></td>
<td>0.291</td>
<td>0.093</td>
</tr>
<tr>
<td><strong>Period 6</strong></td>
<td>0.017</td>
<td>0.004</td>
</tr>
<tr>
<td><strong>Period 7</strong></td>
<td>0.010</td>
<td>0.024</td>
</tr>
<tr>
<td><strong>Period 8</strong></td>
<td>1.825</td>
<td>1.292</td>
</tr>
<tr>
<td><strong>Period 9</strong></td>
<td>0.001</td>
<td>0.003</td>
</tr>
<tr>
<td><strong>Period 10</strong></td>
<td>0.013</td>
<td>0.001</td>
</tr>
<tr>
<td><strong>Period 11</strong></td>
<td>0.003</td>
<td>0.003</td>
</tr>
<tr>
<td><strong>ANNUAL</strong></td>
<td>839.87</td>
<td>535.23</td>
</tr>
<tr>
<td><strong>SD as %</strong></td>
<td>31.33</td>
<td>33.93</td>
</tr>
</tbody>
</table>

Table 4.1 - Average daily aeolian debris transport (g)
Figure 4.11 - Distribution curve for sediment transport measuring annual sediment transport (g) against height (cm). Figures represent a unit width of 8.86 cm. Figure for saltation/ground creep is from Leatherman (1978).
Figure 4.12 - Grain size distributions for aeolian sands collected from Site 4 over the sampling period. For the high transport months of May and October (Two columns on left), the samples from the four traps (30, 50, 100, and 150 cm) were analysed. For the rest of the period only the 50 cm level was analysed. The key at the bottom right indicates the features of each graph.
Grain size distributions show that the aeolian sediments from the Vestfold Hills are generally coarse to medium grained sands (between 0.5 φ and 2.2 φ), moderately to moderately well sorted (between 0.5 φ to 1 φ), and strongly positively skewed to strongly negatively skewed (between -0.5 φ to 0.5 φ). Aeolian deposits are commonly considered to be fine, well-sorted sands and these results indicate that the aeolian transport of sand in this environment is atypical showing grain size distributions significantly broader in nature than expected for aeolian sands. Most work in aeolian systems is done in areas of dune sands, beaches (Dingler et al. 1992, Bagnold 1954, Hsu 1973), or deserts (Fryberger and Dean 1979, Glennie 1970). In these environments, the sands are already well sorted and the processes can take place over very long distances. In the Vestfold Hills the source material is glaciomarine or glacial and therefore contains a wide variation in grain sizes.

Transport processes in the Vestfold Hills can occur only over the width of the ice free area, or no more than approximately 20 km. Additionally, wind strengths are widely variable throughout the year, with wind speeds between 0 and 27.8 m s⁻¹. The threshold wind velocity for the transport of sediment is given by the equation

\[ U_{*t} = A \sqrt{\frac{(\rho_s - \rho)gD_m}{\rho}} \]

Equation 4.1

where \( \rho_s \) is the sand density (2.34), \( \rho \) is the density of air (0.002), \( g \) is the acceleration due to gravity, \( D_m \) is the grain diameter (in cm) and \( A \) is a constant equal to 0.1 in air (Bagnold, 1941). Using this equation, the threshold value for the transport of a particle of 4 mm diameter is 21 m s⁻¹, well below the maximum wind speed value for the Vestfold Hills of 27.8 m s⁻¹.

This poor sorting of aeolian material is probably due to a number of factors. The sorting of the source sediments is generally poor and a wide range of grain sizes is available for transport. When wind velocity is high a wide range of grain sizes is transported. In addition, the fact that the traps were emptied
monthly means that analyses represent an “averaged” sample. This particular problem could be overcome by collecting the trap contents daily and should be considered for a later study. Nonetheless, this “averaged” material is that which will be reflected in the resultant sediment.

Sediment transport is directly related to the average wind speed and maximum wind gust (Figure 4.13). The relationship is best developed in the months of high sediment transport and appears to break down in times of low sediment transport. The low sediment transport rates in the two groups marked “A” and “B” are problematic. In particular, time period ending 8 Jan 94 has an average wind speed towards the high end of the range at 5.64 m s\(^{-1}\) with only ~ 8 g of sediment collected. This compares to an average wind speed of 5.42 m s\(^{-1}\) with a corresponding mass of ~ 275 g of sediment in period ending 5 Jul 93. This period had a maximum wind gust of only 2 m s\(^{-1}\) more.

The difference in sediment transport is probably due to a combination of factors. Firstly, the period ending 5 Jul 93 seems to have had a high wind event of long duration while the period ending 8 Jan 94 had a less variable distribution of wind speeds. Secondly, it is possible that because the period ending 8 Jan 94 is in the middle of summer, the surface sediments are wet and grain adhesion results in a higher threshold value of wind required to entrain sediments. Other periods of low sediment movement but high average or maximum wind gust are the periods ending 1 Mar 94 (with only one, unsustained high wind event), 4 Apr 94 and 6 Jun 94 (with peak events of low maximum wind gust), 4 Sep 94 and 5 Nov 94 (with mostly low wind speeds). These periods are in the winter and the low transport rate may be explained by the presence of snowdrifts covering the source material or even by ice cementing the grains on the surface. As snowdrifts form in low wind areas, it is unlikely that the snowdrifts actually affect the areas from which most debris is sourced.
Figure 4.13 - Bivariate plot of sediment transport against average wind speed and maximum wind gust for aeolian sand traps in Heidemann Valley.
Because of the high bed roughness in the area, a relatively high threshold wind speed is required for debris to be entrained. Regressions on sediment transport and wind speed data (Figure 4.13) indicate that the threshold average wind speed is about $5 \text{ ms}^{-1}$ while the threshold maximum wind gust is about $17 \text{ ms}^{-1}$.

In summary, aeolian processes can be locally important. In the Vestfold Hills up to $6500 \text{ t a}^{-1}$ of terrigenous material is transported across the coast by wind. The material transported includes all fractions up to $-2.0 \phi$. The material is poorly sorted, of various skewnesses, and atypical of aeolian sediments. Even samples collected over a single month are poorly sorted and coarse, reflecting the highly variable wind regime in the Vestfold Hills.

Sediment transport is proportional to average wind speed and maximum wind gust but the relationship is ambiguous due to additional factors including: duration and intensity of high wind events; maximum wind gust; and time of the year. Threshold values for wind speed needed to transport debris in the Vestfold Hills are in the vicinity of $10$ to $12 \text{ ms}^{-1}$.

The sediments resulting from aeolian transport look very similar to the finer fractions of the source sediment as wind speeds are sufficiently variable to move all debris below $-2 \phi$ and display strong positive skewness.

**FLUVIAL**

The fluvial transport of sediments in the Vestfold Hills is minimal. Gore *et al.* (1996) described annual small scale snow drift damming and dam failure in an area adjacent to the Sørøsdal Glacier in the south east of the Vestfold Hills, in which ephemeral lakes of up to $1 \times 10^6 \text{ m}^3$ are released on a regular basis and form fluvial channels in glacial valley fills. Apart from these individual, large scale events such transport is restricted to the very small scale transport of sediment by meltwater from melting snow and ice in the summer. Even so, free water can be seen to flow in small amounts in the Vestfold Hills over $\sim 2$ to 4 weeks at the height of summer. One result of this free flowing water is the formation of snowdrift meltout fans (Figure 4.14).
Figure 4.14 - Photograph of fan developing below a melting snow drift at GR 830823 (Vestfold Hills 1:100000 2nd edition). Hammer for scale. View is from below looking up to the edge of a snow drift.
The melt of snowdrifts in the Vestfold Hills mobilises material transported by aeolian processes and deposited concurrently with snow in the lee of hills. The material forms fans and splays which become active when the snow melts, moving the sand downslope and leaving surface features such as small scale (10's of cm) splays, anastomosing channels, and braided streams. In one such snowmelt fan (Figure 4.15) the sediments become coarser grained downfan and display increasing sorting, indicating that the action of meltwater removes the fine grained material in the downslope locations.

When grain size parameters of the snowdrift samples are compared with those from the aeolian sand traps (Figure 4.16), they plot within a coherent group, albeit at the coarser end of the distribution. A trend towards increasing grain size from 93091 to 93092, followed by an more effective sorting from 93092 to 93093 is evident.

These results may indicate that coarse sediments which are transported during high wind events are deposited with the snow, move to ground level during the nival melt, then are transported downslope by meltwater to a point where the slope is insufficient for the meltwater to continue transporting it. At this stage, water winnows out the fine material and moves it away resulting in the better sorting.

In summary, snowmelt fan material is remarkably similar to the aeolian debris from which it is composed, except that it is coarser grained and less strongly fine skewed, and in the distal parts of the fan, better sorted due to the water borne transport. In practice however, these sediments can only be differentiated from other aeolian deposits by the large scale architecture of the deposit and the surficial fluvial features of the fan deposit.
Figure 4.15 - Cross-sectional representation and grain size distributions for samples from the snowmelt fan at GR 823833 (Vestfold Hills 1:100000 2nd edition)
Figure 4.16 - Plot of mean grain size against standard deviation (sorting) in three samples from the snowmelt fan at GR 823833 - Vestfold Hills 1:100000 2nd Edition

Figure 4.17 - Bivariate plot of standard deviation against skewness shows the trend to improved sorting by aeolian processes acting over distance from the primitive (englacial) material.

Figure 4.18 - Bivariate plot of mean grain size against skewness showing winnowing of fines in beach sands at Lake Dingle and the similarity between "primitive sands" from englacial debris, aeolian debris, and beach sands at Heidemann Beach
Wave mobilisation of sediments in Antarctica is not an important process. The results of wave action are, however, apparent in beach formation along the coastline and on the shores of lakes. Some discussion of beach sediments at Heidemann Bay is in Chapter 5. This study does not deal with wave action at any depth but three samples were collected from the western edge of Lake Dingle (Vestfold Hills 1:100000 2nd edition. GR. 800910).

These samples (93082, 83, & 84) are well sorted, coarse sands, deposited in a small but well developed beach. The samples form a tight group (Figures 4.17, 4.18) with coarse mean grain size (-0.2 to 0.25 \( \phi \)), well developed sorting (standard deviation between 0.35 to 0.7 \( \phi \)), and strong positive skewness (0.6 to 1.0 \( \phi \)). These sediments are the end product of a number of processes. The poorly sorted, coarse grained source sediments are entrained by the wind and deposited in the lake or in snowdrifts. The process preferentially mobilises the finer fractions of the source material and results in a strong fine skewing of the resultant deposit. That debris deposited in snowdrifts is further mobilised in the summer when the snow melts and again, the fines are preferentially mobilised and deposited in the lake basin. Both of these processes cause strong fine skewing.

In the summer, the prevailing north easterly wind generates waves on the lake surface, which, in storm conditions can be substantial. Wave action pushes debris against the western shore and a beach is formed. Continuing wave action winnows fines, ameliorating the strong fine skewing transport and depositional processes described above. The coarse nature of the sediment has two probable causes. Firstly, the material on the beach may be sourced from the lake floor which is probably composed of the same glacial valley sediments as are found throughout the rest of the Death Valley System (which includes Lake Dingle). Secondly, coarse material up to -2 \( \phi \) is demonstrably transported by high wind events as discussed previously.
THE EVOLUTION OF SAND CHARACTERISTICS WITH TRANSPORT

The transport of sediment from the point of origin through the processes discussed previously results in observable changes in the statistical parameters of the sand fraction of those sediments. Two trends in the evolution of sand parameters can be seen (Figure 4.17, 4.18). The first trend represents the evolution of the primitive sediment (represented by the englacial material from 94004 and 93017) to a much better sorted sediment with a coarser mean grain size due to aeolian winnowing of the fine material. Material labelled as "aeolian sands" in Figures 4.17 and 4.18, is the material collected in aeolian sand traps at the seaward end of Heidemann Valley. These are the sediments transported away from the primitive material. They are assumed to represent the signature of sands being transported throughout the Vestfold Hills. The sediments labelled "L. Dingle sands" were collected from the saline Lake Dingle in the Vestfold Hills, from a beach at the western edge of the lake. The second grouping, labelled "Heidemann Beach sands" are samples from the beach at the western end of Heidemann Valley. The marine waters of Heidemann Bay maintain the level of sorting of the sand but appear to remix finer grained sand into the sampled beach sands.

CHAPTER SUMMARY

This chapter has identified the major processes (Figure 4.19) which transport sediment across the coastline of the Vestfold Hills as a guide to similar areas of Antarctica. The pathways for sediment transport across the coastline include the erosion, entrainment, rafting and meltout of sediment in basal glacial ice (to be further discussed in chapters 5 and 6), the entrainment and deposition of sediments into the marine environment by aeolian processes either directly or via snowdrift or the sea ice, and the transport and deposition of sediments via fluvial processes and snowdrift meltout onto sea ice or directly into the sea.

Sediments cross the coastline by being entrained in basal ice in glaciers. The amount of sediment so transported is discussed in Chapter 7.
Cross Coastal Processes

Debris Source - Exposed sediments, crystalline basement

Aeolian erosion and transport

Glacial entrainment

Meltout at terrestrial ice margins

Ice rafting and meltout

Sea level fluctuations

Deposition in snow drifts

Fluvial transport in summer

Sea ice entrainment and transport

Sea-ice meltout

Debris destination - glaciomarine sediments of shallow marine environment

Fig 4.19 - Cross coastal sediment pathways
Wind transports over 6500 t of sediments annually from the 400 km$^2$ of the Vestfold Hills. This sediment is deposited onto sea-ice in the winter, later to be deposited in the sea when the sea ice melts in the summer. Some of the aeolian sediment is deposited onshore in snowdrifts in the lee of hills, where it is transported downslope during the summer nival melt. This fluvial transport sometimes deposits sediment into the sea in the summer or onto or beneath sea ice in the winter. Large scale fluvial transport of sediments is restricted to uncommon, events, mostly related to ice dam failure, although annual small scale ice dam failure has been reported.