Frontispiece: Using the SIPRE auger on Lake Miers.

PROGLACIAL SEDIMENTATION OF LATE WISCONSIN AGE IN MIERS VALLEY, ANTARCTICA

A Thesis

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ABSTRACT

Miers Valley is an ice free valley in the Koettlitz Region of Victoria Land, Antarctica. The eastern basin of this valley has extensive coverage of gypsum and laminated calcite overlying fine silts and sands, while Lake Miers occupies the western basin. Mapping of the surficial sediments has shown that a proglacial lake at least 80m deep extended up the valley (Glacial Lake Trowbridge), dammed to the east by a glacial tongue extending from the coast. C-14 and U/Th dating of the lacustrine carbonates has shown that this lake occupied Miers Valley between 10,000 and 20,000 yrs BP and was at its maximum between 18,000 and 19,000 yrs BP.

Dating of individual lamina has shown that calcite deposition was not continuous, but proceeded in at least three pulses. This is supported by stable isotope analyses which show an enrichment in ^{18}O with time, suggesting the possible concentration through evaporation of ^{18}O during the lake's history. Impressions within the calcite plates indicate that a crystalline material, possibly gypsum, was crystallizing contemporaneously with the calcite. This salt was subsequently flushed from the system on the draining of the eastern basin. At the final stages of occupation, large quantities of gypsum were deposited into the eastern basin from the glacier terminus and draining of the lake resulted in a surface coverage of gypsum throughout the eastern basin.

Shallow mounds of coarse basalt rich drift overlie the carbonate, gypsum and silt beds in places throughout the eastern basin. On the surface ice cover of Lake Miers, similar mounds occur. The ice beneath these mounds is virtually sediment free and the sediment on the surface differs from that on the lake bed. It is suggested that the

mounds on the ice raft and the mounds in the eastern basin share a common origin. Supraglacial and englacial sediments carried into the valley with the Ross Ice Tongue were deposited onto the floating ice cover of Glacial Lake Trowbridge. This material was then rafted across the valley as the lake ice was pushed forward by ice fed from the glacial tongue. The sediments remained supported on the lake until they reached the annual melt out moat or the lake drained. They were then deposited on top of the lacustrine sediments, thereby preserving them (as occurred in the eastern basin), or in the case of the western basin a lake remained so the mounds were preserved on the ice cover which had then become immobile.

Once the proglacial lake drained, most of the lacustrine sediments were then exposed to the environment. Scanning Electron Microscope studies have shown that the exposed lacustrine carbonates have undergone chemical weathering. These carbonates are pitted and weathered compared to much older and less exposed carbonates in neighbouring Marshall Valley.

ACKNOWLEDGEMENTS

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CHAPTER ONE; INTRODUCTION

1:1 PLEISTOCENE GLACIATIONS

The earth's ice sheets have expanded and contracted numerous times over the last few million years, these fluctuations being intimately related to fluctuations of the earth's climate. It has yet to be determined what exactly causes the fluctuations. Many workers (Hays *et al*, 1976; Veeh, H.H. and Chappell, J. 1970; Denton and Hughes, 1983) support Milankovitch's astronomical theory of ice ages, which proposes that the fluctuations in the ice sheets are caused by variations in summertime radiation at high latitudes, which is a result of variations in the earth's orbital geometry.

The last major glacial period the earth experienced occurred between about 75,000 and 11,000 yrs BP and climaxed around 17,000 yrs BP (Epstein *et al*, 1971) or 18,000 yrs BP (Broecker and van Donk, 1970; Shackleton and Opdyke, 1973; CLIMAP, 1976). Studies have attempted to reconstruct the conditions experienced on earth at this time. According to CLIMAP (1976) around 18,000 yrs BP marine based ice sheets and pack ice increased in extent, huge land based ice sheets existed, and vegetation patterns changed, with grasslands, steppes and deserts spreading at the expense of forests. The combined effect of these changes caused an increase in surface albedo and a lowering of sea level by at least 85 metres (Broecker and van Donk, 1970; Veeh and Chappell, 1970; CLIMAP, 1976), although many workers accept a sea level lowering of about 130 metres (Curray, 1965; Bloom, 1971; Shackleton and Opdyke, 1973; Bloom *et al*, 1974).

Investigations into past climatic change has included work such as pollen studies (Hollin, 1980; Turon, 1984), examination of the periglacial record (snowline and sediment studies) (Denton and

Stuiver, 1967; Stuiver *et al*, 1978), stable isotope studies of ice cores (Epstein *et al*, 1971), and biological transfer functions (CLIMAP, 1976) and oxygen isotope stratigraphy of deep sea cores (Broecker and van Donk, 1970; Shackleton and Opdyke, 1973), as well as coastal terrace studies in tropical regions (Bloom *et al*, 1974). Both the on-shore and off-shore studies are important to get a complete view of paleoclimates, although there is sometimes difficulty in correlating what was seen off-shore with what was seen on land. Other work on the ice sheets and their reaction to and influence on global climatic change is being focused on the margins of the polar ice sheets. Sediments in such areas can be examined in an attempt to relate them to a particular glacial environment, thus reconstructing the behaviour of the polar ice sheet in the geologic past.

It has been estimated (Bardin and Suyetova, 1967) that the Antarctic Ice Sheet today covers an area of nearly 14×10^6 km^2, and contains about 24×10^6 km^3 of ice. This ice sheet can be divided into two ice sheets on the basis of differences in basal and surface topography (Stuiver *et al*, 1981), the East Antarctic Ice Sheet and the West Antarctic Ice Sheet. Any change in the global heat balance would be reflected in changes in these ice sheets and in the corresponding change in sea level. Studies of the Antarctic region have shown that the size of the Antarctic ice sheets has fluctuated during the Pleistocene (Scott, 1905; David and Priestly, 1914; Denton *et al*, 1970). The nature of these ice sheet fluctuations has been in debate. Wilson (1964) proposed that the ice sheets have advanced by surging, whilst Hollin (1962) maintains that the glaciations were driven by eustatic sea level fluctuations occurring as a direct result of changes in the Northern Hemisphere ice sheets, causing the ice shelves to ground and expand. Detailed investigations of areas around the ice sheet margins can therefore help to piece together the information on

local ice sheet behaviour and its relationship to climate and sea level changes.

As most of present day Antarctica is covered by ice, sediment studies are restricted to those areas of the continent which are ice free, the dry 'oases'. Three such areas in Antarctica are the Vestfold Hills in Princess Elizabeth Land, the Dry Valley region in Victoria Land, and the Bunger Hills in Wilkes Land.

The Dry Valley region, which is by far the largest of Antarctic oases, is an area of the Transantarctic Mountains along the coast of McMurdo Sound, bound to the west by the East Antarctic Ice Sheet, and bound to the east by McMurdo Sound and its periferal ice sheets. The southern boundary is defined by the Royal Society Range and the Koettlitz Glacier, and to the north, the frequency of occurrences of ice free areas north of the McKay Glacier decreases. The valleys themselves are glacially carved and tend to run from west to east i.e. from the East Antarctic Ice Sheet to the coast. The floors of these glacially carved valleys contain thick deposits of sediment which acts as a partial record of the glacial history of the area. Denton *et al* (1970) outline the three possible sources of glaciation which may have contributed to these valleys; (i) glacial advances from the coast, i.e. expansion of the Ross Ice Shelf up the valleys, (ii) advances of the East Antarctic Ice Sheet which could force ice tongues through the Transantarctic Mountains and down into the valleys (forming glaciers such as Taylor and Upper Wright), and (iii) small alpine glaciers.

Péwé (1960) first recognised the fact that these areas have undergone multiple glaciations, although he proposed that these glacial advances came from the west over the Transantarctic Mountains. Stuiver, *et al* (1981) have expanded this idea and proposed a working hypothesis on the glacial events in the Dry Valleys but including major advances of ice tongues from a grounded Ross Ice Sheet. Their

work has shown that during the Wisconsin Glacial period in the Northern Hemisphere, the West Antarctic Ice Sheet was expanding near the continental shelf margin. The Ross Sea during this period filled with grounded ice which forced its way up into the valleys, blocking them off at the seaward end. The glacial tongues advanced far up these valleys and on retreating left behind large quanties of glacial drift.

It has been proposed (Denton *et al* 1970; Stuiver *et al* 1981) that many such advances have occurred during the Pleistocene and in some areas such as Marshall Valley (see fig. 1.1) a record of a sequence of such glacial advances has been preserved (Dagel *et al*, in press). In valleys such as Taylor, Victoria, Marshall and Miers, lakes have formed in front of ice sheet glacial tongues, the ice front blocking the drainage of the valley. With the removal of the ice front these lakes may disappear, however in some basins smaller lakes may still exist (e.g. Lake Fryxell in Eastern Taylor Valley). In the areas where the lakes have now disappeared sediment and relic geomorphic features are frequently preserved. These can aid the reconstruction of the former proglacial lacustrine environment. Péwé (1960), recognised such features in Miers Valley and named the proglacial lake, Glacial Lake Trowbridge.

1:2 DESCRIPTION OF MIERS VALLEY

Miers Valley occurs between latitudes 78° 06'S and 78° 07'S and longitudes 163° 44'E and 164° 12'E (see fig. 1.1),and is between 80-150 metres above sea level. The valley is bound by the Royal Society Range to the west and slopes down to the Koettlitz Glacier to the east. Glacially carved, it has steep walls and a U shaped cross section. Approximately 11 kilometres in length, the width of Miers

Figure 1.1 Locality map for Miers Valley.

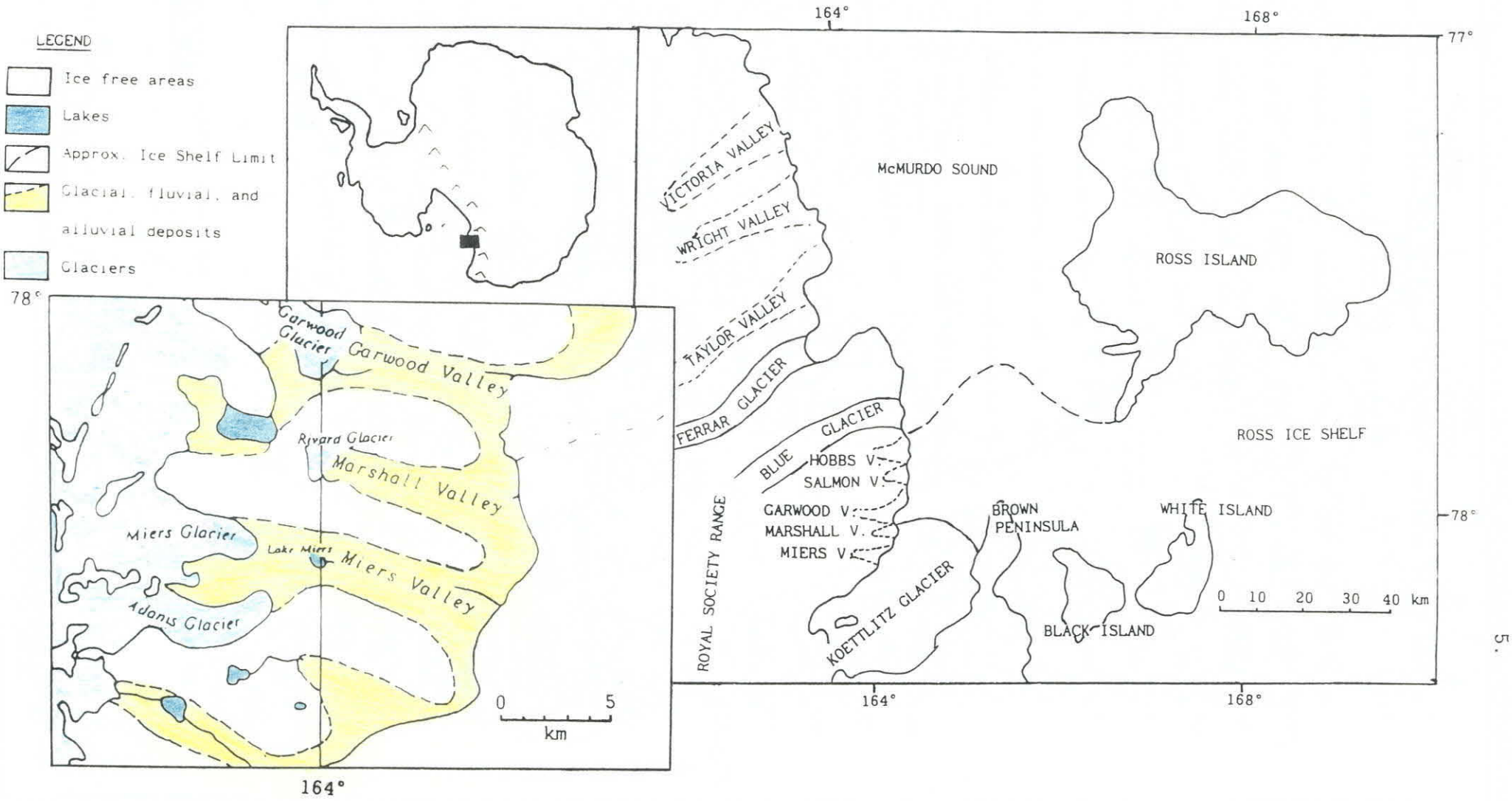
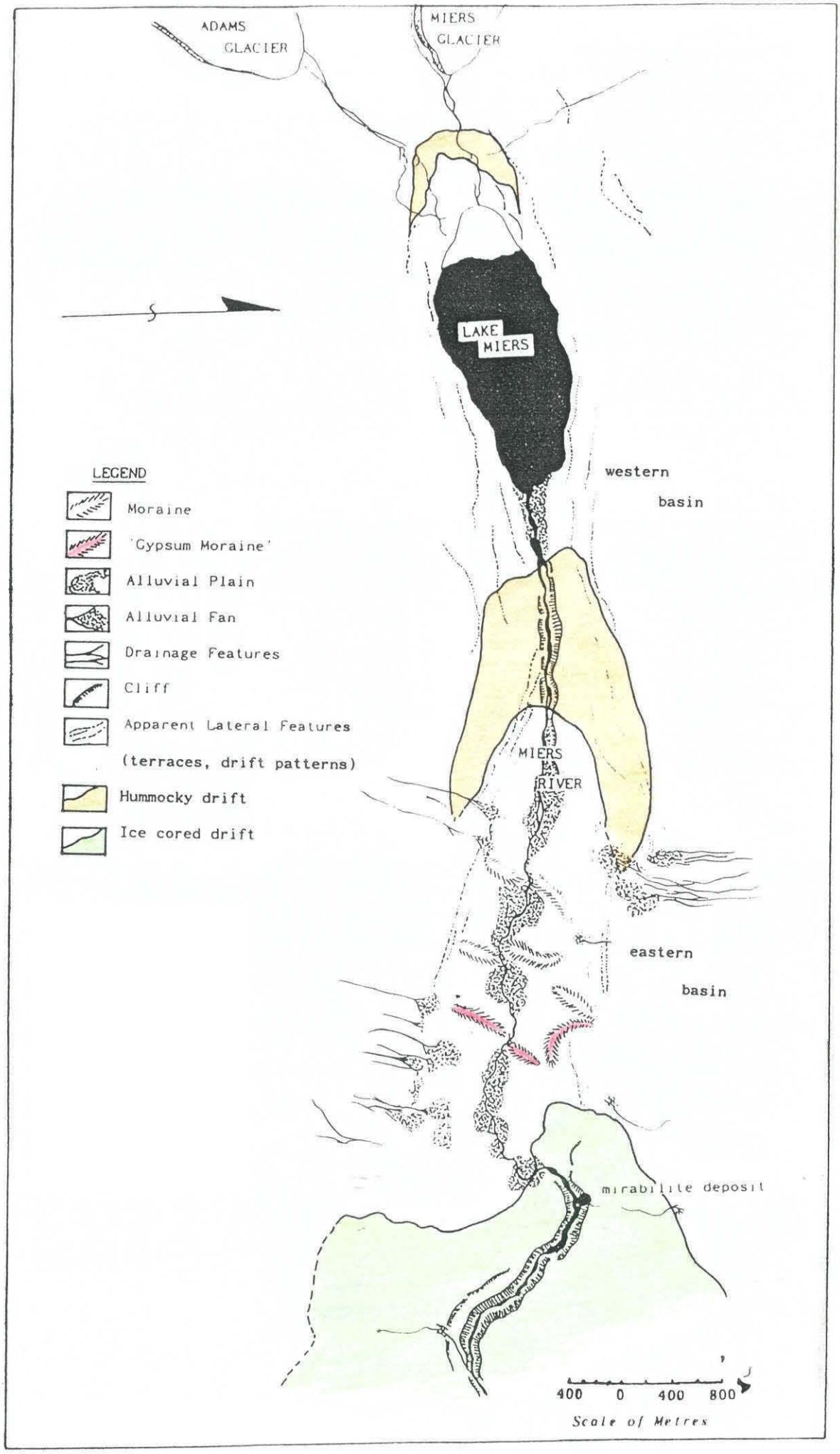


Figure 1.2 Map of Miers Valley showing major geomorphic features.


ADAMS GLACIER
MIERS GLACIER
LAKE MIERS
western basin
LEGEND
Moraine
'Gypsum Moraine'
Alluvial Plain
Alluvial Fan
Drainage Features
Cliff
Apparent Lateral Features
(terraces, drift patterns)
Hummocky drift
Ice cored drift
MIERS RIVER
eastern basin
mirabilite deposit
400 0 400 800
Scale of Metres

Valley ranges from 1.5 to 2.5 kilometres. There are two major basins within the valley separated by hummocky drift.

The western basin is the smaller of the two and contains Lake Miers, a roughly oval shaped lake with a permanent ice cover. The ice cover contains many melt pools and deposits of sands and gravels. Distinctive features on the lake ice surface are groups of ice cored mounds of poorly sorted rock debris. Lake Miers is fed by meltwater from Miers and Adams Glaciers (see fig. 1.2). This water flows through a delta complex at the western end of the lake (see fig. 1.2). Most seasons during summer a moat melts out around the lake and it outflows through Miers river (Bell, 1967; and Torii, *et al*, 1966).

The eastern basin contains at least three moraine complexes (see fig. 1.2), which span the valley floor. Between these moraines are large fans and subdued topography. A large moraine complex to the east of this basin is particularly noticeable due to an extensive coverage of gypsum, giving the moraine a grey to whitish appearance. This moraine is referred to in this work as the 'gypsum moraine'. Ice cored drift occupies the eastern end of the valley, seaward of the basins.

On its route from Lake Miers to McMurdo Sound, Miers river has cut at least two sections through the drift deposits in the valley, one where the river cuts through the 'gypsum moraine' and another where it cuts through the eastern most ice cored material of the valley, exposing a small section containing sands, silts and salts.

1:3 PREVIOUS WORK

Gunn and Warren (1962), mapped the geology of Victoria Land between the Mawson and Mulcock glaciers. Miers Valley occurs within

this area and has been mapped as marble and schist. In 1963, Blank, Cooper, Wheeler, and Willis mapped the area in greater detail separating out the marble and schists and including the granite intrusions that occur through the valley walls. The marbles and schists have been mapped as Skelton Group Metasediments of Cambrian or Precambrian age and the Granitic rocks have been mapped as being younger, although still Cambrian or Precambrian. The valley floor they have mapped as Cenozoic moraines and Quarternary fluvio-glacial lacustrine and deltaic deposits.

Péwé (1960) in his descriptions of the glacial deposits of the McMurdo Sound area, briefly described the lake deposits of Glacial Lake Trowbridge in Miers Valley. He brought attention to the presence of gypsum and calcium carbonate lying on the surface of the valley floor. In a report from the Japanese summer parties on activity in the Dry Valleys (Torii *et al*, 1966), the evaporite salt deposits of Miers Valley were described. Samples of the evaporites from the moraine complexes at the eastern end of the valley were collected and examined. It was concluded that four major 'evaporites' occur within the valley; gypsum, calcite, mirabilite, and thenardite. Although they used the term 'evaporite', they explained that they were not sure of the salt's origin. Several of the samples contained quartz, feldspar and some amorphous minerals.

Bell (1967), published a study on the morphology, physics and chemistry of Lake Miers. Bell's investigations lead him to conclude that the ablation rate from present day Lake Miers is 15-20cm/year. He also showed that although the mean air temperature is -20° C the bottom waters of the lake are +5° C. He attributed this to solar heating beneath the permanent ice cover.

Bradley and Palmer (1967) briefly described Miers Valley including Lake Miers and mentioned the presence of the "dry lake flat" which

occurs in the eastern basin. They discussed the surface of Lake Miers and proposed that the debris cones on the lake surface were ridges formed by compression of the lake ice as it freezes and thaws. They proposed that the debris itself originated from the lake bed, freezing onto the base of the ice around the edges of the lake during winter when the lake edges freeze to the lake bed. Bradley and Palmer (1967), also suggested that the lake ice is doming upwards as a result of the combined action of the compressive forces and upwelling of ice diapairs.

Lyon (1978), looked at the stable isotope geochemistry of the gypsum covering the 'gypsum moraine'. Stable isotope analyses of the $^{34}S/^{32}S$, $^{18}O/^{16}O$ and D/H ratios (see tables 2.1 and 2.2) showed that the mirabilite and gypsum are not closely related as their $\delta^{18}O$ and δD values for their waters of crystallization differ. The mirabilite in the valley is likely to have formed by the freezing of meteoric water containing seawater sulphate, while the gypsum was precipitated from a lake containing water of Koettlitz glacier composition, the sulphate in the gypsum being identical to that expected on the equilibrium crystallization from seawater sulphate.

Miers Valley was surveyed in 1981/82 by P.D. Tinnelly of New Zealand Lands and Survey, and a sketchmap of benchmarks produced.

Very little dating work has been done on the Miers Valley lacustrine deposits. Stuiver *et al*, (1981), give one radiocarbon date (QL-1040; 1300±40 ^{14}C yrBP) of an algal mat from the surface of ice cored Ross Sea Drift found within the mouth of Miers Valley. Denton also submitted several carbonate samples to the Waikato University U/Th dating laboratory, the dates for these samples are included in appendix III. During the summer seasons 1975/76 and 1976/77, the Waikato Antarctic Research Programme included field party visits to Miers Valley. Dr. T.R.Healy collected numerous evaporite samples from

the valley and submitted twelve of these to the Waikato University U/Th dating laboratory, the resulting U/Th dates (University of Waikato Antarctic Research Unit Report No. 7:1978) are summarized in the appendices. During the 1982/83 summer season another Waikato University field party visited Miers Valley and as well as collecting carbonate and gypsum samples, also obtained a core from Lake Miers, the description of this core is in Chapter 2.

CHAPTER TWO; FIELD WORK

2:1 INTRODUCTION

During the 1983/84 summer season, ten days were spent in Miers Valley. During this time the surface sediments in the eastern basin were mapped and sampled. A Department of Lands and Survey, surveyor, Dave Manson, surveyed most of the sample sites, determining their relative locations and their altitude above sea level.

A brief visit was made to Miers Valley during the 1985/86 summer season. On this occasion work concentrated on the western basin. A sediment core was obtained from the eastern end of Lake Miers, and the area around Lake Miers examined for traces of previous lake levels.

2:2 MAPPING AND SAMPLING

Samples were collected from the eastern basin following an arbitrary transect pattern (see fig. 2.1). Where necessary, deviations from the transect line occurred to enable sampling to encompass the evaporite limits (see fig. 2.1). Each sample site was levelled and a sketch map produced which gives the sample site's altitudes above sea level and relationship to other sites (see fig. 2.2).

The surface of the eastern basin is covered with a sequence of lacustrine beds, gypsum overlying carbonate plates, overlying silts and sands. The plates of carbonate vary in thickness from several millimetres to 1-1½ centimetres. In most places the carbonate plates are overlain by a deposit of gypsum and underlain by fine yellow silt or sands.

Figure 2.1 The extent of gypsum and carbonate in Miers Valley. Solifluction on the southern side of the western basin has obscured any trace of carbonate on this side, however obvious terrace marks are seen encircling Lake Miers representing former lake level(s). The arbitrary transects from which samples were collected are shown.

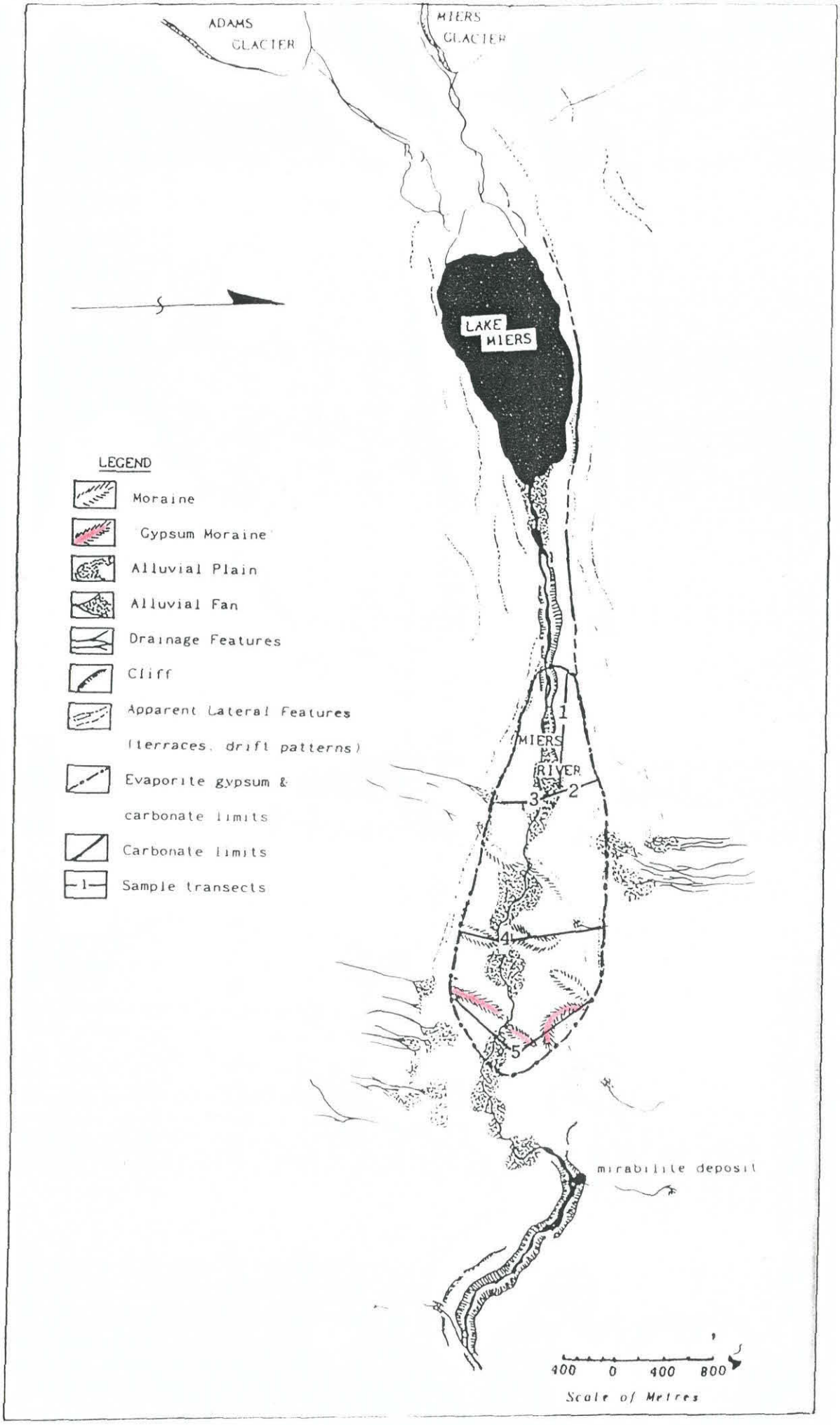
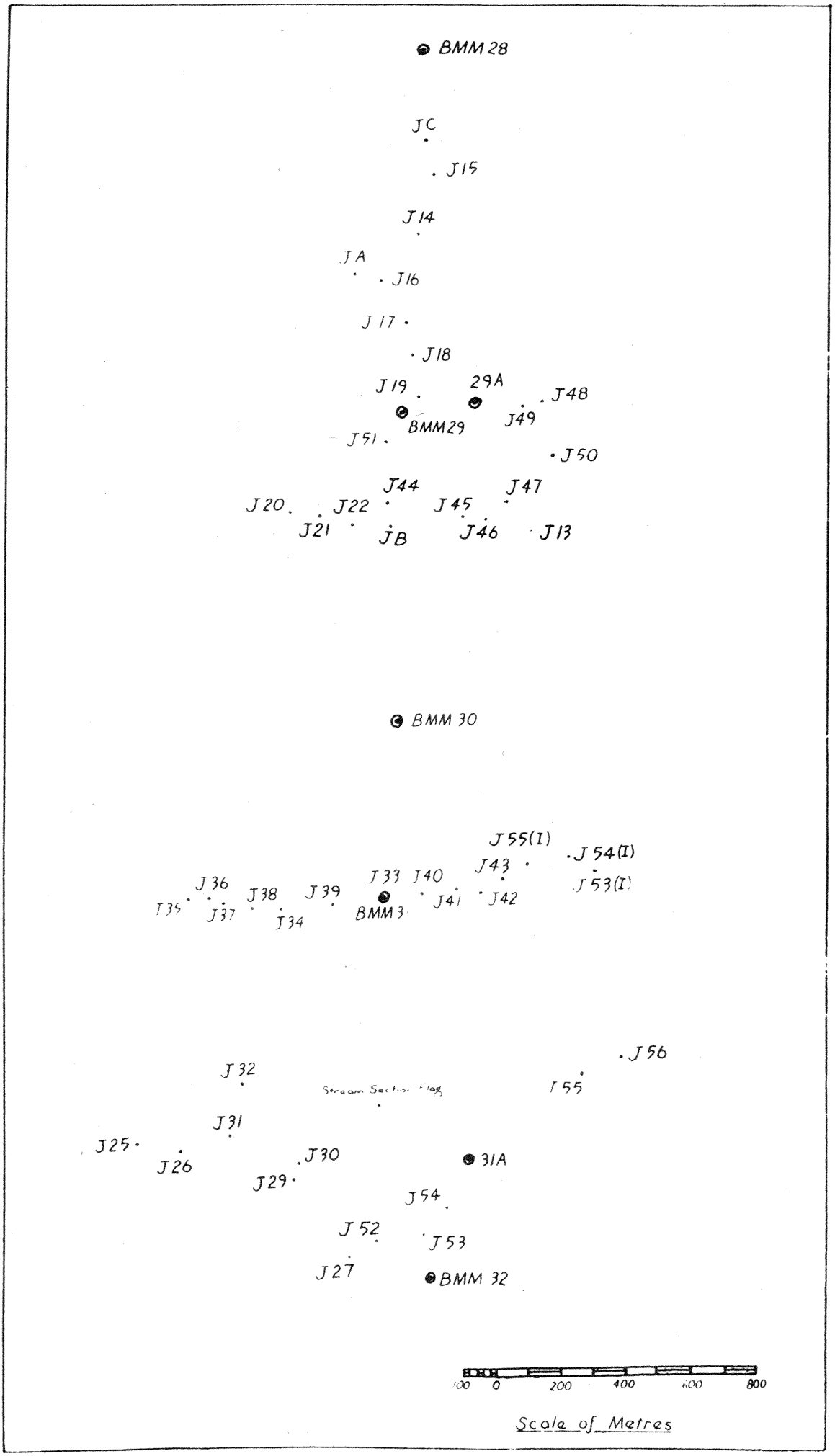

ADAMS GLACIER
MIERS GLACIER
LAKE MIERS
LEGEND
Moraine
Gypsum Moraine
Alluvial Plain
Alluvial Fan
Drainage Features
Cliff
Apparent Lateral Features
(terraces, drift patterns)
Evaporite gypsum &
carbonate limits
Carbonate limits
Sample transects
1
MIERS RIVER
2
3
4
5
mirabilite deposit
400 0 400 800
Scale of Metres

Figure 2.2 The levelled sample collection sites from Miers Valley. The University sample number and altitude of each sample are listed in Appendix I. (from the New Zealand Department of Lands and Survey map produced of the sample sites).


BMM 28
JC
J15
J14
JA
J16
J17
J18
J19
29A
J48
BMM29
J49
J51
J50
J44
J47
J20
J22
J45
J21
JB
J46
J13
BMM 30
J55(I)
J54(I)
J43
J36
J33
J40
J38
J39
J53(I)
J35
J37
J34
BMM3
J41
J42
J56
J32
Stream Section Flag
J55
J31
J25
J26
J30
31A
J29
J54
J52
J53
J27
BMM 32
100
0
200
400
600
800
Scale of Metres

The carbonate and gypsum distributions appear similar (see fig. 2.1) in the eastern basin, although the gypsum appears to be absent from the western basin. The gypsum and carbonate occurs approximately 20 metres higher up the valley walls on the northern side of the valley compared to the deposits on the southern side. The reason for this is probably the greater influence of solifluction on the southern walls which are exposed for longer to the direct sun during the summer months, causing thawing and hence greater downslope movement.

The highest outcrop of carbonate occurs on the northern wall of the valley at 156m above sea level (a.s.l.), (C931:J56 13/12). The lowest point on the valley floor, with carbonate coverage, lies at 77m a.s.l. (C922:J52b). From this it is assumed that Glacial Lake Trowbridge was at least 80m deep. The carbonate deposition would not have occurred directly at the lake edge so it is possible that Lake Trowbridge exceeded 90m in depth. Present Lake Miers is about 154m a.s.l. (Benchmark No 26) and is visible from site C931, even though the hummocky drift lies between the two basins. This implies that the two basins could have been linked. Distinctive lake deposits have also been found in both basins providing further evidence that the basins were in fact linked by one water mass.

The carbonate plates have several discernible features. In most places the carbonate plates occur in layers (see fig. 2.3). Although these layers can be separated by applying pressure, a dividing lithology is absent. In some places only one layer may be preserved, whilst in others up to three or four layers may be present. Needle-like crystal moulds occur in all the carbonate layers. Ranging in length from a few millimetres to about 15 mm and varying in width up to about 5mm, the moulds tend to be square in cross section tapering to points at the ends. Although many of the moulds tend to be roughly

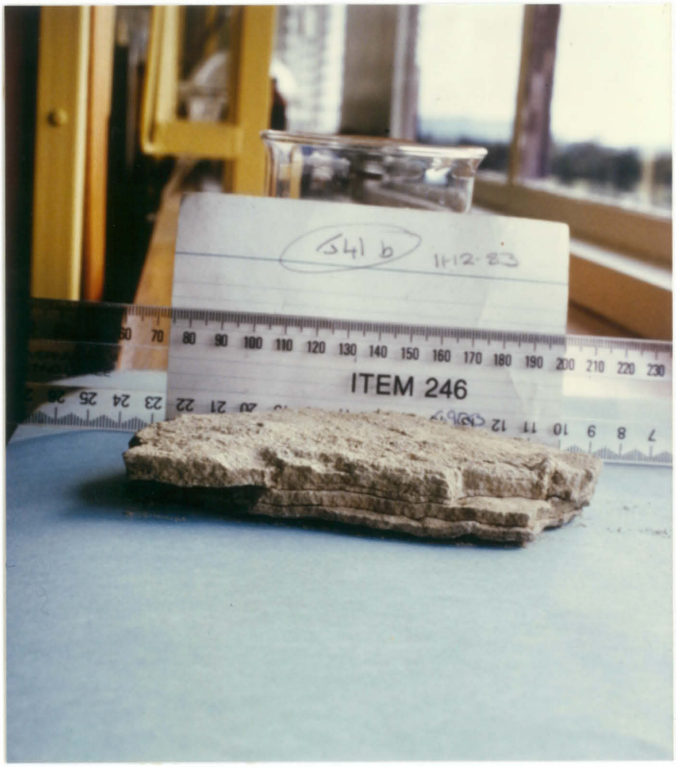


Figure 2.3 A typical carbonate plate showing three possibly four distinct layers. (sample C903).

Figure 2.4 Needle shaped pseudomorphs and thin carbonate plates occurring with the surface lag. A clump of pseudomorphs cemented with carbonate cement also occurs on top of the surface lag.

alligned with the layers of carbonate, some moulds are orientated pointing into the carbonate plate. Impressions on the surface of one plate have corresponding matches on the surfaces of adjacent plates.

At the lower end of the valley, near the site of the 'gypsum moraine' the moulds are usually empty with only the occasional occurrence of calcite regrowth. To the west the extent of regrowth increases, samples such as C941(J58) contain substantial calcite recrystallization within the moulds. Westward again of this site, where the basin begins to ascend the hummocky drift, the carbonate no longer occurs as large distinct calcite plates with infilling pseudomorphs, but as small pieces of fractured plate and with recrystallized calcite as needle shaped pseudomorphs. Further over the hummocky drift, the needle shaped pseudomorphs occur loosely with the surface lag (see fig. 2.4), sometimes being weakly cemented together by a calcite cement (see fig. 2.4). Thus we have a westward trending succession , from carbonate plates containing moulds to carbonate plates containing calcite pseudomorphs, to loose pseudomorphs which are sometimes associated with thin fragments of calcite plates or weakly cemented together by calcite.

Most of the western basin is occupied by Lake Miers. There are higher terraces around the lake representing former higher lake levels. Carbonate deposits are not so evident on the surface of the western basin. On closer inspection it was found that needle shaped calcite pseudomorphs occur amongst the basaltic lag on the flatter areas within the hummocky drift separating the two basins. These pseudomorphs are sometimes weakly cemented to the surrounding silts and sands. They also occur on a terrace approximately a metre above the northern Lake Miers shoreline. This provides a continuity of surface deposits between the two basins.

Figure 2.5 Lacustrine carbonate overlain by a mound of basalt rich drift. (Photo courtesy of Dr. C.H. Hendy)

It was noted during the mapping that in many places in the valley the carbonate and other evaporite deposits were preserved particularly well when they were covered by mounds of basalt rich sands and gravels (see fig. 2.5). It is proposed that these mounds of drift originated from a glacier damming Glacial Lake Trowbridge. The material from the glacier was 'dumped' onto the ice raft of the lake. When the lake disappeared the drift was left overlying the lake bed, thus preserving the evaporites (this is discussed further in Chapter 5).

2:3 STREAM SECTIONS

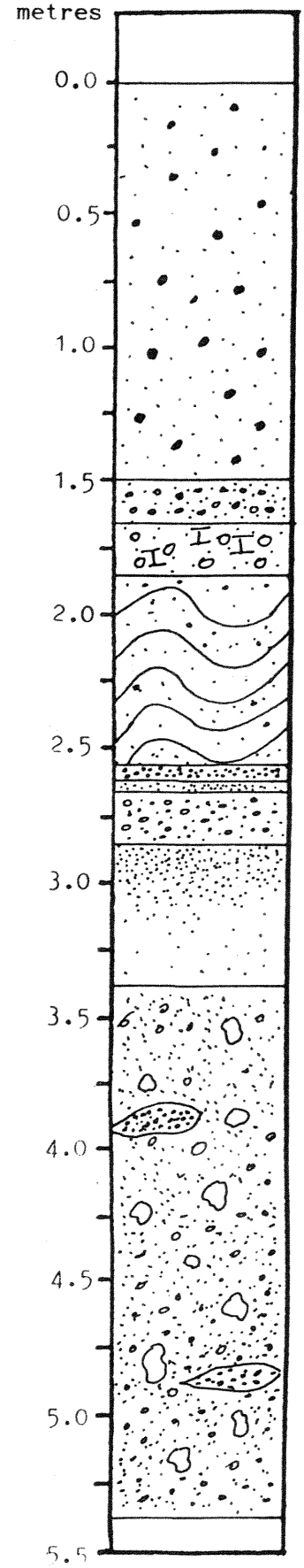
In places where the Miers River has cut through moraines and drift, sections may become exposed (see fig. 2.6). At least two such sections occur in Miers Valley, the first exposes a mirabilite deposit to the east of the two basins and the second exposes a section cut through the 'gypsum moraine'. These sections were cleared, brushed down to the permanently frozen sediment, and described. A statigraphic section of the mirabilite section is shown in figure 2.7.

Figure 2.6 A section cut through the eastern hummocky drift by Miers River, viewed from the river bed. The mirabilite occurs as the white band in the upper half of the section. (Photo courtesy of F.M. Judd)

Figure 2.7

Mirabilite from the section shown in Figure 2.6. The freshly exposed mirabilite is greyish blue, mirabilite exposed for longer, dehydrates to thenardite (white).

Figure 2.8 MIRABILITE SECTION, EASTERN MIERS VALLEY



2:3:1 ORIGIN OF THE MIRABILITE DEPOSIT

The mirabilite section outcrops in a 20m high bank on the northern side of the stream section (see fig. 2.6). Debris had slipped off the mirabilite deposit exposing it about 12m above the stream bed. Instability of the slope restricted the extent of the exposure. The 6 metres of drift overlying the mirabilite appeared to consist of silts and sands and had a hummocky surface topography typical of the Ross Sea Drift deposited in the mouths of valleys in the area.

The poorly sorted sands and gravels of the basal diamicton were presumably laid down by a proxial Ross Ice tongue. Meltout from the ice tongue probably supplied the sands and silts above the diamicton, the fluvial action sorting the sediment. A mixture of sea water and meteoric water flowed from the ice tongue forming a proglacial pond at the ice margin. Sodium sulphate from seawater within the ice tongue crystallized from the pond as mirabilite, possibly forming as salts concentrated by the pond freezing. The basaltic sands and gravels interbedded with the mirabilite crystal beds (see fig. 2.7) are likely to have originated from the ice tongue, either blowing off the ice surface or being washed off. The silts and gravels at the top of the described section are also probably from the ice tongue. The Ross Ice Tongue then presumably advanced over the deposit and on retreating left the overlying drift deposits. Subsequent down cutting by Miers river exposed the section.

Mirabilite deposits occur at several other locations in the Koettlitz Glacier region; near Walcott Glacier (Lyon, 1978), near Hobbs Glacier (Bowser *et al*, 1970) and on the coast next to Koettlitz Glacier between Miers Valley and Walcott Bay (Bowser *et al*, 1970). Stable isotopes of sulphur, oxygen and deuterium have been used in an attempt to relate the mirabilite deposits to a source (see tables 2.1

and 2.2 for a summary of isotope values). These studies attempt to utilise the stability of the oxygen isotopes of the sulphate ion, which will not exchange readily unless they pass through a biological oxidation/reduction cycle. Thus the composition of most of the sulphate samples is identical to that of seawater sulphate. The water of crystallization however is characteristic of the solution from which the mirabilite was precipitated, and in the Antarctic very big differences exist between sea water ($\delta^{18}O$ and $\delta D \simeq 0$) and glacial meltwaters ($\delta^{18}O \simeq -30$ to -40, $\delta D \simeq -250$ to -350). It has been proposed (Bowser *et al*, 1970) that the mirabilite near Hobbs Glacier formed by the freeze concentration of a pond containing sea water sulphate and glacial meltwater. It was suggested that the sulphate concentrated in ponds from aerosols.

Lyon (1978), proposed that the mirabilite in Miers Valley had a similar origin, the anomalously light sulphate oxygen being attributed to microbiological activity. The possibility of the sulphate being directly deposited from a marine source, was dismissed on the grounds that the deposits are too high above sea level (Bowser *et al*, 1970). Lyon (1978), suggested that the gypsum and mirabilite of Miers Valley were not closely related, due to differing $\delta^{18}O$ and δD in their waters of crystallization. The gypsum sulphate had a sea water affinity but appeared to have been precipitated from water of Koettlitz Glacier affinity. From the mapped extent of gypsum, the eastern basin alone contains a gypsum covered area of $2km^2$, with an average thickness of gypsum of about 3cm, it seems unlikely that aerosols, are the source of all the sea water sulphate in Miers Valley.

Mineral	Collection Area	Sulphate		Water of Crystallization	
		$\delta^{34}S$	$\delta^{18}O$	$\delta^{18}O$	δD
Mirabilite[1]	Walcott Glacier	+10.4	−3.6	n.d.	n.d.
Mirabilite[1]	Miers Valley	+19.2	+3.9	-14.6	-151
Mirabilite[2]	Hobbs Glacier	+19.2	+9.3	-9.7	-75.7
Mirabilite[2]	Hobbs Glacier	+21.1	+2.2	-25.3	-198.5
Mirabilite[2]	Taylor Glacier	+18.7	+0.9	-37.9	-305
Mirabilite[2]	Koettlitz Glacier	+19.5	+10.4	-	-24.7
Gypsum[1]	Walcott Glacier	+6.4	-1.6	-3.7	-144
Gypsum[1]	Dromedary Glacier	+15.0	-3.0	-16.4	-209
Gypsum[1]	Miers Valley	+21.6	+7.3	-30.3	-289
	Sea Water	+20	+10		

Table 2.1: Stable isotope analyses of salts from the Koettlitz Glacier Region and Taylor Valley
[1] from Lyon (1978)
[2] from Bowser, Rafter, and Black (1970)

Source of Water	$\delta^{18}O$	δD
Koettlitz Glacier Ice Tongue[3]	-35.8	-273.1
Sea Ice (below Koettlitz Glacier)[3]	+1.7	+13.3
Sea Ice (below Koettlitz Glacier)[3]	+1.8	+14.1
Sea Water (from base of Koettlitz Glacier)[3]	-1.1	-5.6
Fresh Water Ice, derived from meltwater of desalinated sea ice[3]	-0.5	-4.4
Unfrozen fresh water[3]	-3.2	-21.1
Surface water of Lake Miers[1]		-202.
Deep water of Lake Miers[1]		-210.

Table 2.2: Stable isotope analyses of water from the Koettlitz Glacier Region.
[1] from Lyon (1978)
[3] from Gow and Epstein (1972)

Gow and Epstein (1972), from a study of the deuterium and oxygen isotopes of ice cores from the Koettlitz Glacier, concluded that the transition from the Koettlitz Glacier tongue to the sea ice shelf was just to the north of Garwood Valley, which is less than 10km north of Miers Valley. A visible difference in ice was also observed at that line from the eastern end of Garwood Valley in January 1986. Alpine glaciers are believed to have retreated during the last glaciation (Denton *et al*, 1970). A grounding of the sea ice and advance of the ice front up the valleys coupled with a retreat of the Koettlitz Glacier would place a sea water source for the sulphate directly in Miers Valley. Frozen or trapped sea water within the Ross ice tongue could have provided the sea water sulphate which would then have mixed with glacial and meteoric water within lakes or ponds in the area.

2:3:2 THE 'GYPSUM MORAINE' SECTION

This section was extremely variable along its length and depth. To simplify the description for visual display (see figs. 2.9 and 2.10), many of the very small discrete beds are shown together as units.

2:3:3 UNIT DESCRIPTIONS

UNIT A:A deposit of evaporite gypsum, containing black sands and gravels. Towards the base of the deposit the black sands and gravels separate into distinct beds occurring as laminae with gypsum sand.

UNIT B:A lacustrine carbonate bed which appears to be in three layers, each about 3mm thick.

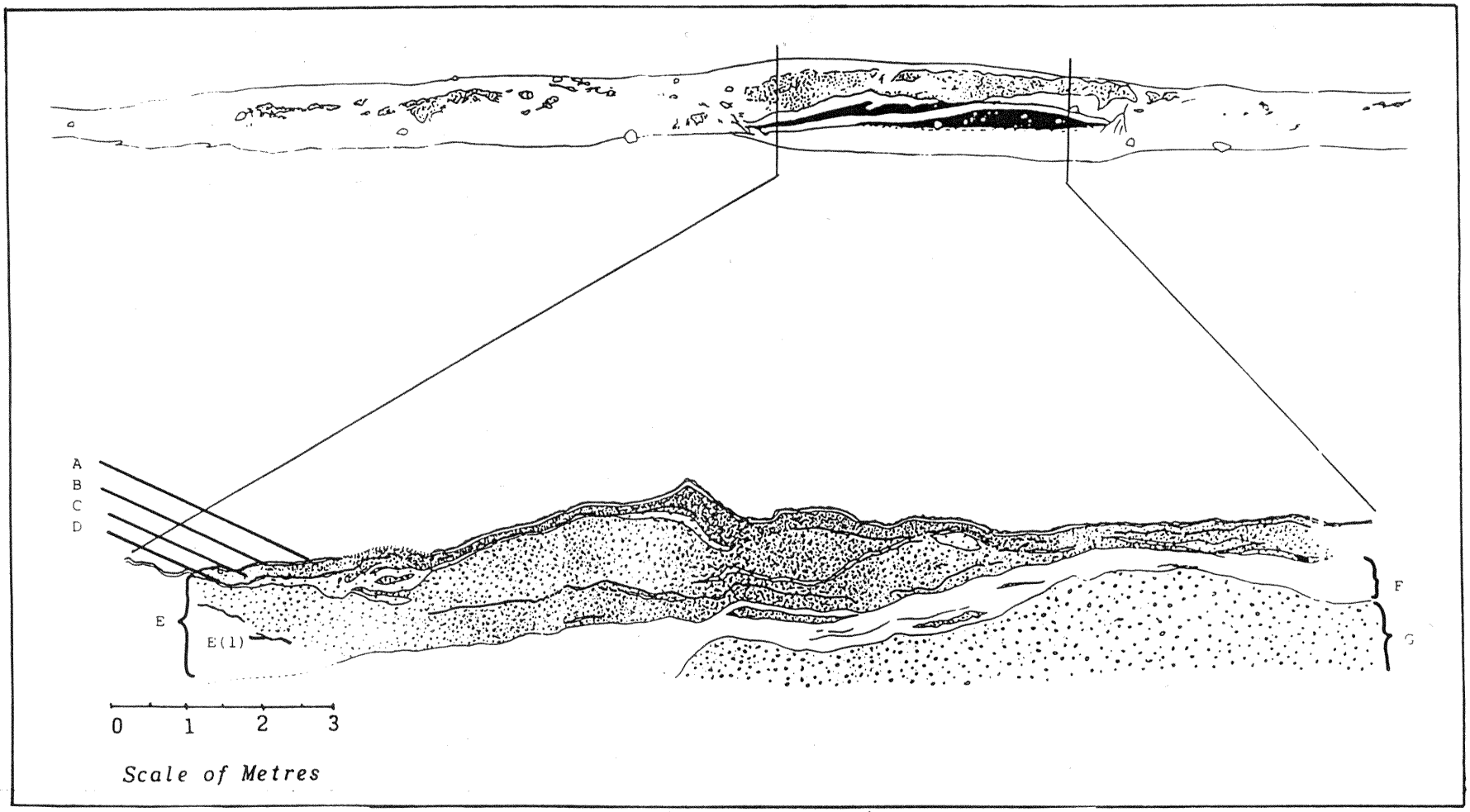


Figure 2.9 Sketch of the cross section of the 'gypsum moraine'. The stratigraphic column of figure 2.10 is a generalized description of this section.

Figure 2.10 Generalized stratigraphic column of the 'gypsum moraine' section.

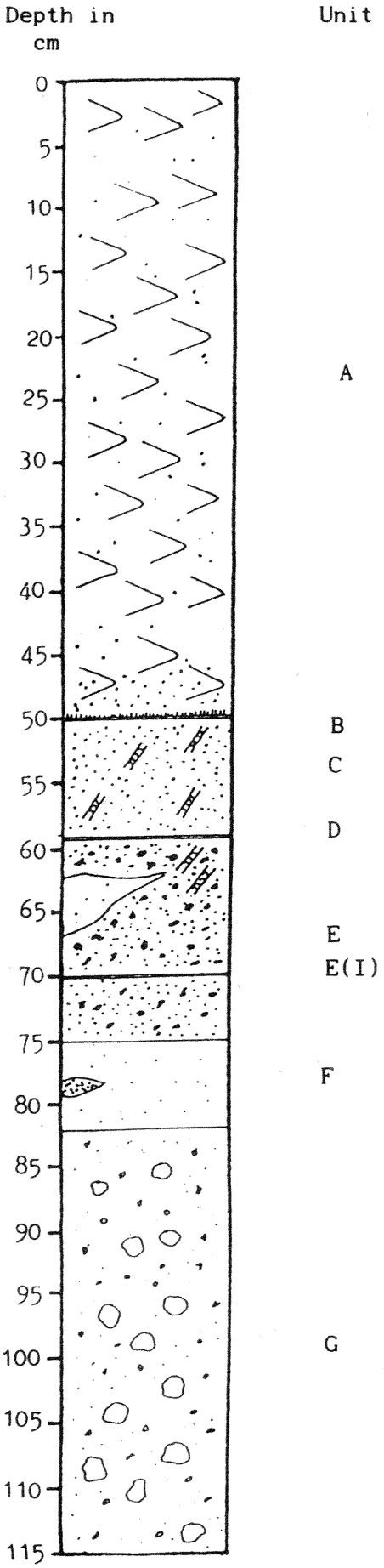
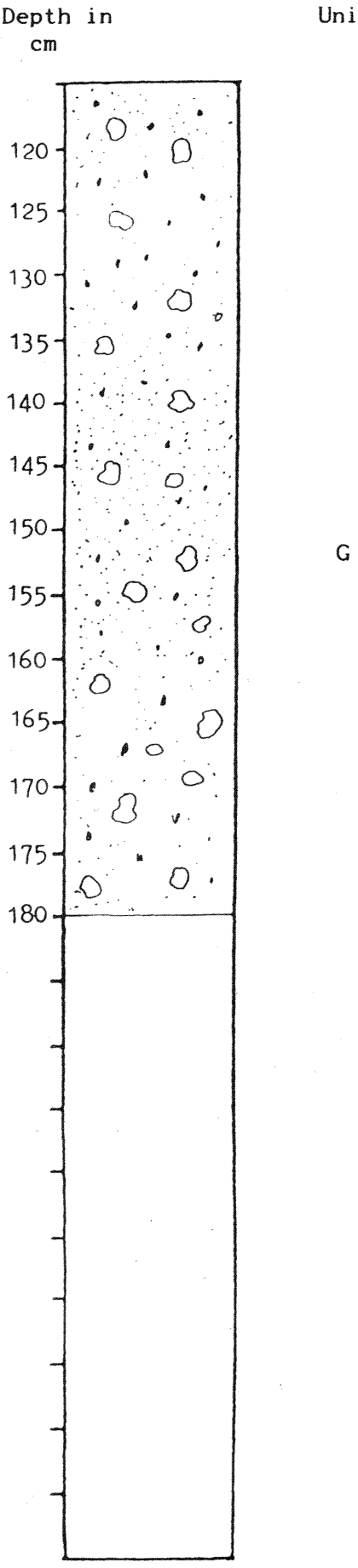


Figure 2.10 cont.



UNIT C:Laminated beds of fine yellow sand and coarse grey sand. The latter grades to black sand and at the base becomes a gravelly black sand. Throughout this unit algae appears as fibrous mats.

UNIT D:A very thin carbonate bed.

UNIT E:A variable (along profile) complex of beds which has been generalized to; 2cm of basalt rich sandy gravel, overlying an orange-red fine sand which appears to have been oxidized. At the eastern end of the profile, algae occurs as fibrous mats amongst the oxidized sands. Beneath this lies 2mm of coarse gravelly sand grading to a sandy gravel, bedded with sand. This overlies a bed of variable thickness (1mm-30mm), of pale orange sand which has been oxidized. Underlying this bed is 1cm of fine grey sand, then 1mm of fine green sand which shows some cementation. In some places this is underlain by an oxidized silt which grades to an oxidized poorly sorted sand, containing cobbles and boulders. An iron oxide skin covers the boulders. This in turn is underlain by a cemented coarse black sand which occurs as laminae due to weak cementation of some beds. Underlying this bed is a very fine (1mm) carbonate bed (UNIT E(I)) with some gypsum occurring on the lower surface of the carbonate. This is underlain by a cemented very coarse sand. Beneath this, are 4cm of laminated fine silts and black sands, the sands sometimes containing pebbles. This overlies 2cm of coarse sandy gravel.

UNIT F:A fine yellow silt with some lenses of sands.

UNIT G:A poorly sorted, basalt rich coarse sandy gravel. Due to permifrost the depth extent of this unit could not be determined.

2:3:4 INTERPRETATION

The basalt rich unit G marks the retreat of the Ross Ice tongue. The ice tongue would still have been damming the valley and it is likely that the fine silt of unit F was preserved by being deposited into a proglacial lake. The capping sands and silts below E(I) are interpreted here as outwash material flowing into the lake. The carbonate unit E(I) formed in a lacustrine environment. Carbonate sedimentation was interrupted by the addition of the basaltic sands, and the poorly sorted material overlying them. This material is likely to be englacial material dumped from the ice margin which would have been in close proximity. After this the lake either drained or evaporated, and the sediments were exposed to the atmosphere, causing gypsum to precipitate below the carbonate E(I), and cement to form within the basaltic sands. The sediment at the surface underwent oxidation, forming the orange skin over the boulders in this layer. The remainder of unit E shows an alternation of sediment textures and oxidation, and it is suggested that these deposits represent a former small stream bed, the algae mats having grown during the brief periods of flow.

At some time after this a proglacial lake either reformed or expanded to cover this area, and the carbonate bed, unit (D) formed. This bed has been dated with the U/Th and C14 radiometric techniques at the University of Waikato, (these techniques are discussed in chapter 3), and yielded a date of 26,000 ^{14}C yr BP (C862).

The algae mats in unit C were sampled and sent to Prof. M. Stuiver of the University of Washington for radiometric dating and its age was determined to be 19,770±110 ^{14}C-yr BP (QL-910).

The three distinctive layers of unit B indicated that it was probably the same carbonate bed observed throughout the eastern basin.

This unit returned a C14 age of 22,950±360 yr BP (WK-609). U/Th dates performed on the same sample however have given much younger ages as discussed in Chapter 3. The final draining and evaporation of the lake concentrated the salts in the waters to produce a brine from which the gypsum was precipitated. Unit A in the 'gypsum moraine' section contains gypsum and volcanic sands which appear to have been deposited directly off an ice margin as a moraine, and this gypsum may also have contributed to that formed in Glacial Lake Trowbridge during its final stages.

2:4 LAKE MIERS SURFACE DESCRIPTION

Bell (1967) and Bradley and Palmer (1967) found the floating permanent ice cover of Lake Miers to be about 6m thick. Annually a moat of between 3 and 6m wide melts out around the lake, passing under a thickening layer of smooth blue ice. The floating ice raft is believed to behave like other Antarctic lake ice rafts, water freezing to the bottom while ablation occurs at the surface of the ice, causing lake ice to move in an upward direction (Bradley and Palmer, 1967; Hendy *et al*, 1972).

The surface of the ice raft is very irregular, and contains numerous melt pools, piles of rock debris and ice ablation tables. The ice surface appears to be ablating away differentially, so that during summer a series of ice sheets occur superimposed over one another (see fig. 2.11). Melt pools lie beneath these sheets and the sheets themselves appeared to be supported by pinnacles of ice. As melting continues the pinnacles or the sheets collapse leaving behind an irregular surface (see fig. 2.12).

Figure 2.11 Ice ablation tables on Lake Miers. Meltwater from these ablation tables accumulates beneath the tables as melting and ablation continues. Eventually the tables collapse producing the irregular surface of figure 2.12. Note also the gravelly sand beneath the ablation tables.

Figure 2.12 The ice surface of Lake Miers with irregular ablation tables and remnent pinnacles of ice. A row of debris cones can be seen in the middle distance.

Figure 2.13 Debris cone on Lake Miers.

Figure 2.1 Closer view of the debris cone of figure 2.13. Ice core is visible.

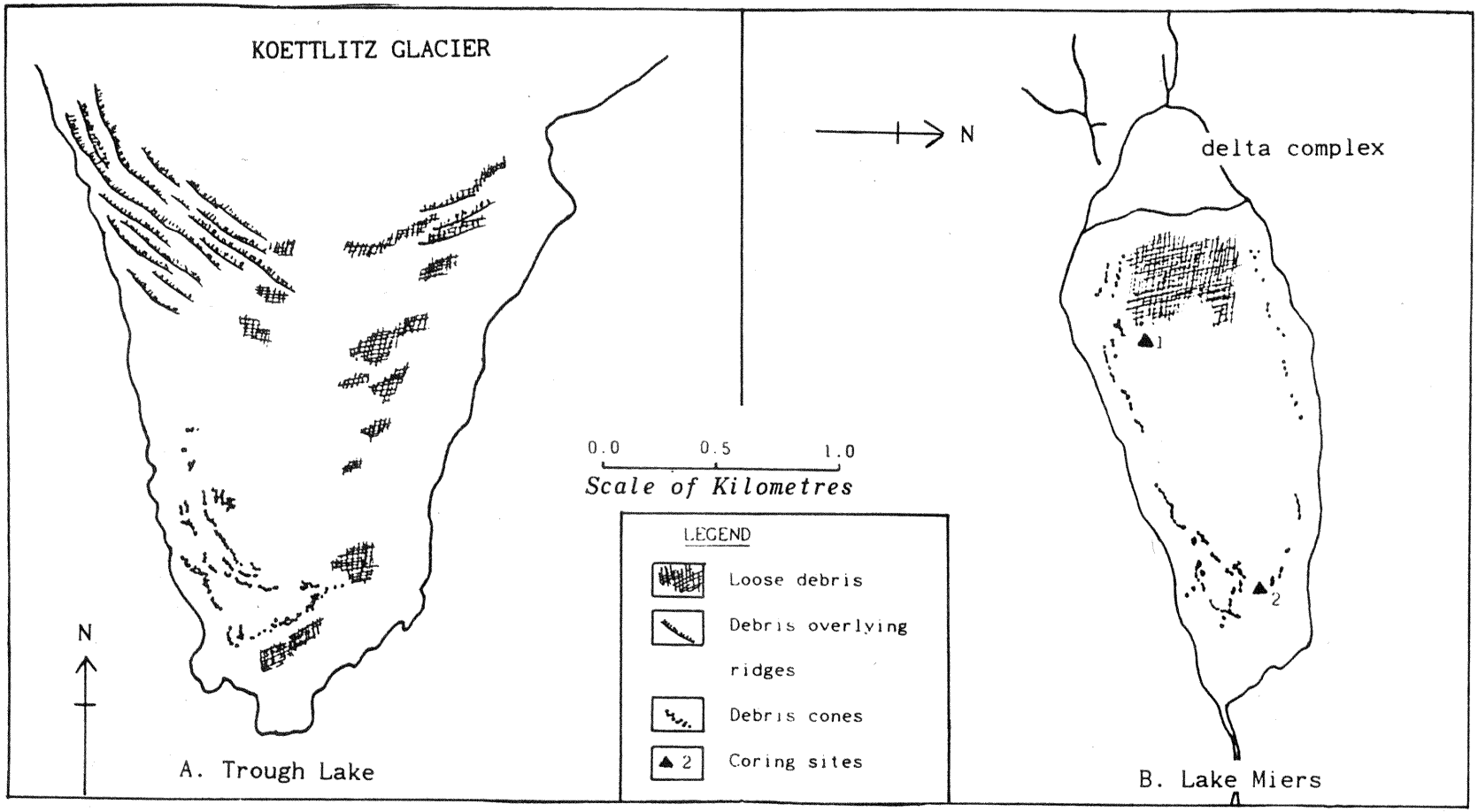


Figure 2.15 Distribution of debris on Lake Miers and Trough Lake. The two coring sites on Lake Miers are marked.

In between the ablation tables smoother blue ice occurs, probably representing former ablation tables which have collapsed and melted, and on refreezing formed hard blue ice. Frequently the surface of the ice is covered by rock debris, either a thin veneer over the smooth ice (see fig. 2.12) or as a prominent debris cone up to 2m high (see figs. 2.13 and 2.14). The debris cones are ice cored and pinnacles of ice can be seen protruding through the debris cover. All the debris on the lake consists of poorly sorted sands and gravels sometimes including large boulders. The distribution of the debris cones is shown in fig. 2.15. Similar such surface debris and debris cones have been observed on other Antarctic lakes such as Deep Lake on Cape Barne (Hendy *et al*, 1972) and Trough Lake in Pyramid Trough (Lyon 1979). A model for the deposition of such debris is proposed in Chapter 5.

2:5 LAKE SEDIMENT CORES

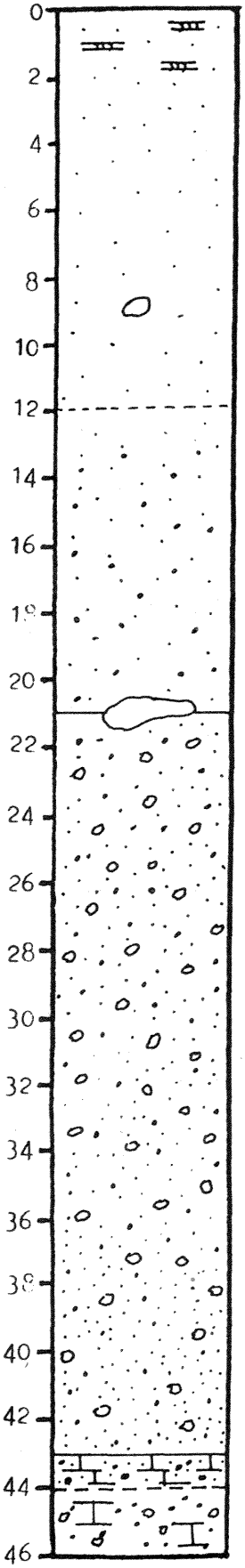
During the 1982/83 summer season, a Waikato University field party led by Dr C.H.Hendy visited Lake Miers and obtained approximately 1 metre of the underlying lake sediment using a small tripod and a percussion piston corer. The core obtained was opened and described at Waikato University (see fig. 2.16). Using the same equipment during the 1985/86 summer season a second core was obtained from Lake Miers (see fig. 2.17).

The first core was taken from the south-western end of the lake (Dr. C.H.Hendy pers. com.). The second core from the eastern end of the lake (see fig. 2.15). The site of the second core was chosen to lie directly beneath a patch of gravel on the lake ice surface.

Figure 2.16 Stratigraphic Log of Core 1

Depth in cm

Lithology



Dark greyish yellow (2.5Y 4/2) silt, massive and uniform, contains occasional sand grains and an erratic pebble (5mm in length). Upper 2cm divided into layers by thin bands of algae (1-3 cells thick)

Greyish yellow brown silt (10YR 4/2) mixed with olive black (5YR 2/2) fine sand producing a sandy silt which grades to a silty sand

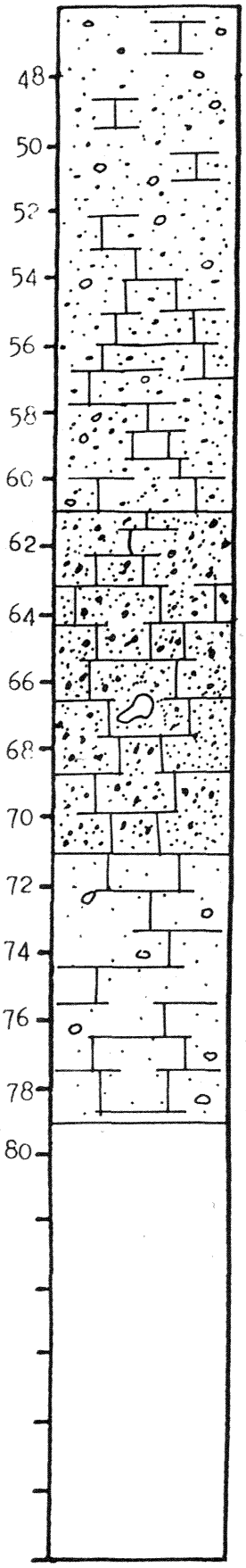
A poorly sorted sandy gravel containing pebbles ranging in size up to 17mm, with one pebble occurring at the boundary being 40mm. A mixture of metasediments and volcanics.

Sandy gravel cemented with calcite which grades to a coarse sand with only a few small pebbles (5-6mm in length).

Figure 2.16 Cont.

Depth in cm

Lithology

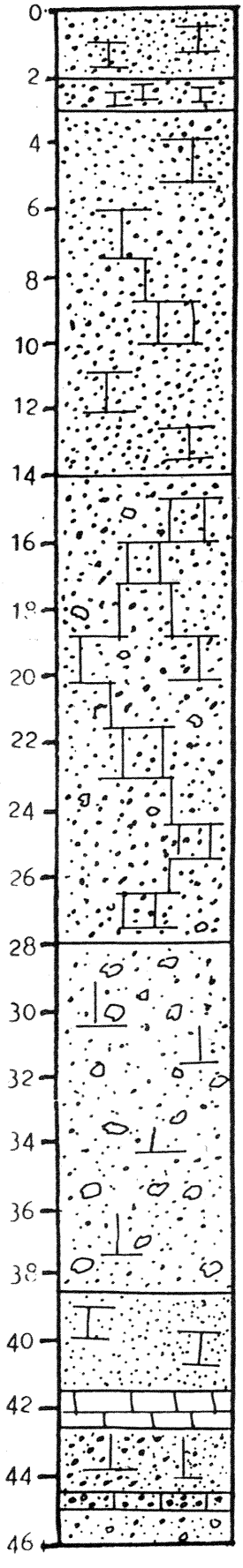


From 44 to 52cm the sand shows weak cementation into aggregates; 52 to 61cm shows slightly stronger cementation into a single unit.

Greyish olive (5Y 4/2) silty sand, well to moderately sorted with pebbles and granules occurring below 67cm. Cemented with calcite

Olive grey (10Y 5/R) fine sandy silt containing small (0.6-8mm) pebbles and granules. Cemented with calcite

Figure 2.17 Stratigraphic Log of Core 2 As the true measurements of this core were distorted on extrusion in the field, the depths given in this description are only approximate.



Lithology

Well sorted medium sand, cemented by $CaCO_3$

$CaCO_3$ cemented silty sand

Well sorted medium to coarse sands, cemented by $CaCO_3$.

Medium to coarse sands with the occasional large basaltic pebble, cemented by $CaCO_3$.

Very poorly sorted coarse sandy gravels containing medium to large pebbles. A mixture of basalt and metasediments, very weakly cemented by $CaCO_3$.

Moderately sorted medium sands, cemented by $CaCO_3$ on exposure to the air.

$CaCO_3$ lens

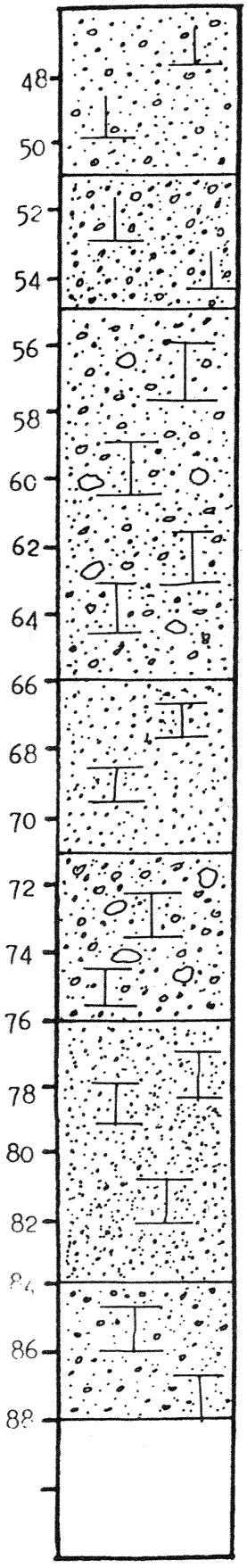
Moderately sorted coarse sand, weakly cemented by $CaCO_3$ on exposure to the air

Poorly sorted sandy gravel, weakly cemented by $CaCO_3$ on exposure to the air

Figure 2.17 Cont.

Approx. Depth in cm

Lithology



Moderately to well sorted medium to coarse sands containing the occasional medium sized pebble, usually basaltic. Weakly cemented by $CaCO_3$ on exposure to air

Poorly sorted very coarse sands and gravels, weakly cemented by $CaCO_3$ on exposure to the air and containing a thin lens of $CaCO_3$

Poorly sorted basalt rich coarse sandy gravel containing large basaltic pebbles, weakly cemented by $CaCO_3$ on exposure to the air

Well sorted basalt rich medium sands, weakly cemented by $CaCO_3$ on exposure to the air

Very poorly sorted coarse sandy gravels containing small to large pebbles which are predominantly basaltic, weakly cemented by $CaCO_3$ on exposure to the air

Moderately to well sorted medium sands, a mixture of basalt and metasediments, weakly cemented by $CaCO_3$ on exposure to the air

Moderately sorted coarse sands, basalt rich but containing some metasediments and subrounded weathered material, weakly cemented by $CaCO_3$ on exposure to the air

Using a SIPRE augre, a hole was drilled through the lake ice. The surface gravel was found to be only a few centimetres thick, below this the ice was virtually sediment free. The floating ice was only 3.4m thick overlying a further 6.1m of water to the lake bed.

Cores were obtained by driving a length of PVC tubing into the underlying sediment. A piston attached to the tripod by stainless steel cable inside the tubing allowed core retention when the tubing was raised from the sediment. The core assembly was raised with the aid of the tripod and winch.

The core described at Waikato University showed varying degrees of calcite cementation in the lower 35cm, so it was decided to extrude a core in the field directly after obtaining it, to determine if the cement was forming in the lake sediments or after exposure to the air. On extrusion of core 2 on site, there was no evidence of cementation, but when the samples arrived back in New Zealand after being allowed to stand after exposure to the air, much of the core had in fact started to harden with calcite cement. It is concluded from this that cements found in relic lake deposits such as in the neighbouring Marshall Valley, are more likely to have formed on draining of the lake rather than whilst the lake still existed.

The large quantity of smaller grain sizes in Core 1 reflects the close proximity to the deltaic complex at the western end of the lake, silts being washed into the lake annually.

The rock debris on the surface of the lake ice at the site of Core 2 was a very poorly sorted coarse sandy gravel, being much coarser grained than the material recovered in core 2. Visually there appears to be a greater proportion of basaltic material on the lake bed compared to on the ice surface. The surface sediments show no sign of cementation, while the subsurface sediments were weakly cemented after exposure to the atmosphere.

CHAPTER THREE: ISOTOPE ANALYSIS

3:1 INTRODUCTION

In this study all the carbonate samples collected from Miers Valley were analysed for $\delta^{18}O$ and $\delta^{13}C$ ratios, to determine the source of waters in Glacial Lake Trowbridge. Fifteen carbonate samples were submitted to the C-14 dating laboratory at the University of Waikato. As mentioned in Chapter 2, one algae sample was sent to The University of Washington for C-14 dating. U/Th dating was attempted on nineteen carbonate samples, however due to difficulties encountered with spike calibration only ten U/Th dates were obtained. This chapter discusses these methods, problems encountered and the results obtained. A summary of C-14 and U/Th dates and δ^{18} and δ^{13} ratios obtained from samples collected in Miers Valley occurs in the appendices.

3:2 STABLE ISOTOPES

The fractionation of oxygen and hydrogen isotopes in the form of different isotopic molecules of water in the hydrosphere and atmosphere means that stable isotope analysis can be used to distinguish between water masses, for example water from the Koettlitz Glacier has a $\delta^{18}O$ value of -35.8 (see table 2.2, Chapter 2), while precipitation falling at the South Pole has $\delta^{18}O$ values of between -55 and -40 (Epstein *et al*, 1963; Epstein *et al*, 1965). The oxygen isotopic composition of waters is determined by CO_2 exchange with water samples at 25°C and analysis of the dissolved CO_2 gas. These values are then compared to similar analyses performed on SMOW (Standard Mean Ocean Water), which is a water standard having a composition closely approximating mean oceanic water, and is currently

supplied and distributed by IAEA (International Atomic Energy Agency) in Vienna. The $\delta^{18}O$ of carbonates is determined by reaction with 100% phosphoric acid at 25°C and analysis of the CO_2 produced. This CO_2 is compared to the standard PDB. The PDB standard is a solid carbonate, a belemnite (*Belemnitella americana*) from the Cretaceous Pee Dee formation, South Carolina, U.S.A. (Friedman and O'Neil, 1977). The relationship between PDB and SMOW is given by:

$$\delta_{SMOW} = 1.030856\delta_{PDB} + 30.86 \qquad \text{(i)}$$

& $$\delta_{PDB} = 0.97006\delta_{SMOW} - 29.94 \qquad \text{(ii)}$$

(Friedman and O'Neil, 1977).

The fractionation of ^{18}O between calcite and water is strongly temperature dependent and forms the basis of isotopic paleothermometry. The relationship between the composition of water and coexisting calcite is given by:

$$10^3 \ln\alpha = 2.78(10^6 T^{-2}) - 2.89 \qquad \text{(iii)}$$

where $10^3 \ln\alpha$ is the per mil fractionation between calcite and water, and T is the temperature in kelvins (O'Neil *et al*, 1969). When, the difference between δ_c and δ_w is small,

$$10^3 \ln\alpha \simeq \delta_c - \delta_w \qquad \text{(iv)}$$

where δ_c is the true δ value of CO_2 gas prepared from the carbonate reaction with %100 phosphoric acid at 25°C and δ_w is the true δ value of CO_2 gas equilibriated with water at 25°C with respect to the same standard. However, analytical problems dictate that carbonates are generally reported with respect to PDB, and waters with respect to SMOW. Equations (i) and (ii) must then be used to convert the δ values

to the appropriate standard. If we assume that the temperature of the lakes in which the calcite precipitated was approximately 3°C then from O'Neil *et al*, 1968 we see that

$$10^3 \ln\alpha = 2.78(10^6 276^{-2}) - 2.98$$

$$= 33.60$$

$$\delta_c - \delta_w = 33.60$$

A lake containing water of Koettlitz Glacier origin (see table 2.2, chapter 2) would have a $\delta^{18}O \simeq -35$ w.r.t. SMOW, calcite precipitating from this water would be expected to have a $\delta^{18}O$ ratio of about -32_{PDB}.

The $\delta^{13}C$ value of calcium carbonate precipitating from aqueous solution is dependent on (i) the $\delta^{13}C$ value of the CO_2 gas in equilibrium with carbonate and bicarbonate ions in solution, (ii) the fractionation of ^{12}C and ^{13}C between the solid calcium carbonate, the CO_2 gas, the carbonate and bicarbonate ions in solution, (iii) the temperature of isotopic equilibrium, and (iv) any other chemical properties which have an effect on the abundance of carbonate and bicarbonate ions in the system (Faure, 1977). Photosynthesis by plants will preferentially remove CO_2 enriched in ^{12}C, leaving the waters enriched in ^{13}C. This effect can be quite pronounced in Antarctic lakes where stratified water columns and permanent ice cover prevent recharge of the CO_2 resevoir from which the algae are removing CO_2 and causing carbonate to precipitate (Lawerence and Hendy, 1985). The isotopic composition of lacustrine carbonates will reflect the nature of the water mass from which they precipitated. In this way they can be used to determine source(s) of waters in the lake.

3:2:1 ISOTOPE ANALYSIS

Dry carbonate samples were prepared by grinding with a mortar and pestle. For oxygen and carbon analysis the samples were reacted with %100 phosphoric acid under vacuum at 50°C. The CO_2 produced was distilled across to the mass spectrometer following the method outlined by Cuthbertson (1985). Samples were analysed on a Micromass 602c spectrometer, comparing samples to a local standard WLS (Waikato Lime Stone). The mass spectrometer compares the ion beams of mass 45/46 and 46/45 and 44, to obtain $^{13}C/^{12}C$ and $^{18}O/^{16}O$ ratios. δ values of these ratios are reported as per mille (‰) deviations from a standard

$$\delta_{std}(SAMP) = \frac{R_{samp} - R_{std}}{R_{std}} \times 1000$$

where R is the isotopic ratio. $\delta^{18}O$ and $\delta^{13}C$ values with respect to PDB are obtained by running an international standard (NBS-19) against WLS and adjusting the sample values with the following equations;

$$\delta^{18}O_{PDB} = 0.999846 + 0.009846 - 1.26$$

$$\delta^{13}C_{PDB} = 1.063845 - 0.034846 - 3.6277$$

These equations were derived by Burns (1980), from the work done by Mook (1968) and Grinstead (1977).

It was found that the NBS-19 standard results changed dramatically after a month of analyses. The $\delta^{18}O$ values changing from -2.55 to -1.37. It is suspected that this change was due to changes in the acid

or sample heterogenity of standard, however the standard values then settled around -1.37 the $\delta^{18}O$ equation was modified to accommodate the change in standard value;

$$\delta^{18}O_{PDB} = 0.999846 + 0.009845 - 0.08$$

The results of these analyses are listed in the appendices.

Where samples occurred in separable layers, each individual layer was analysed separately and these results are listed in table 3.1. Samples containing both calcite pseudomorphs and carbonate plates were also separated and analysed separately. These results are listed in table 3.2.

SAMPLE NOS.		LOWEST		MIDDLE		UPPER	
		$\delta^{18}O_{PDB}$	$\delta^{13}C_{PDB}$	$\delta^{18}O_{PDB}$	$\delta^{13}C_{PDB}$	$\delta^{18}O_{PDB}$	$\delta^{13}C_{PDB}$
C901	(J40b)	-23.76	4.93			-21.66	5.22
C913	(J46b)	-30.44	3.68	-27.20	4.23	-23.57	4.94
C892	(J43b)	-28.09	4.66	-19.30	5.67	-12.56	6.39
C941	(J58a)	-33.20	3.19	-32.55	3.50	-29.29	4.42
C903	(J41b)	-31.86	3.63	-31.20	4.05	-29.38	4.53
C908	(J43b)	-32.15	3.41	-30.90	3.67	-27.66	4.32
C932	(J53)	-30.68	3.81			-28	3.5
C906	(J42b)	-30.42	3.70			-29.51	4.21
C841	(J14)			-29.56	5.31	-30.40	5.03
C843	(J15)			-32.96	4.81	-31.77	5.11
C847	(J17a)			-32.73	4.80	-33.34	4.73
C849	(J18a)	-24.06	6.35	-27.16	5.73	-31.24	5.25
C853	(J19)	-31.97	4.47			-30.57	5.45
C861	(J23)	-33.51	5.29	-30.50	6.54		
C916	(49)			-30.33	4.52	-31.28	5.32

Table 3.1: Stable isotope ratios of the individual layers of the carbonate plates.

SAMPLE NOS.	PSEUDOMORPHS		MATRIX	
	$\delta^{18}O_{PDB}$	$\delta^{13}C_{PDB}$	$\delta^{18}O_{PDB}$	δC_{PDB}
(J13)	-34.01	0.08		
(J12)	-34.06	1.83	-35.85	0.39
(J12)			-26.46	4.36
(J14)	-33.18	1.47	-23.53	-2.52
(J14)			-30.40	2.75
(82/134)	-43.66	-1.98	-33.32	2.99
(J19b)	-37.97	-1.22	-30.35	3.42
(J19b)			-30.56	3.20
(J17a)	-34.67	1.13	-33.34	2.54
(J52)	-33.69	1.42	-30.27	3.06
(J50)	-33.74	1.44	-31.88	2.76
(J50)			-32.01	2.85
(J50)			-32.03	1.98
(82/134)	-34.44	1.30	-33.17	3.11
(82/134)			-32.65	2.95
(J17a)	-34.32	1.29	-33.99	2.94
(J14)	-33.64	1.40	-30.55	3.07

Table 3.2: Stable isotope ratios of the calcite pseudomorphs and their surrounding matrix.

Most of the carbonate $\delta^{18}O_{PDB}$ values are between -26 and -33‰. Values of about -33‰ are what was expected from calcite precipitated from waters similar to Koettlitz Glacier, see table 2,chapter 2, (Gow and Epstein, 1972), and surface ice of the coastal area (Morgan, 1982). In the contour maps of the surface $\delta^{18}O$ values of Antarctica produced by Morgan (1982) the areas to the west of the Transantarctic Mountains have values of around -45‰, while the ice around Ross Island has $\delta^{18}O$ values around -30‰. The carbonates precipitated in Miers Valley closely resemble the local meteoric waters, rather than waters from the interior (the East Antarctic Ice Sheet).

The layered carbonate plates show a progressive enrichment in ^{18}O and ^{13}C from the lowest plate to the uppermost plate (see table 3.1). A plot of $\delta^{18}O$ verses $\delta^{13}C$ (see fig. 3.1) shows that samples enriched in $\delta^{18}O$ are also enriched in $\delta^{13}C$. This might be caused by either a progressive change in the water mass due to a change in the source of water with a concurrent change in $\delta^{13}C$ values due to photosynthesis by algae within the lake preferentially removing CO_2 enriched in ^{12}C, or

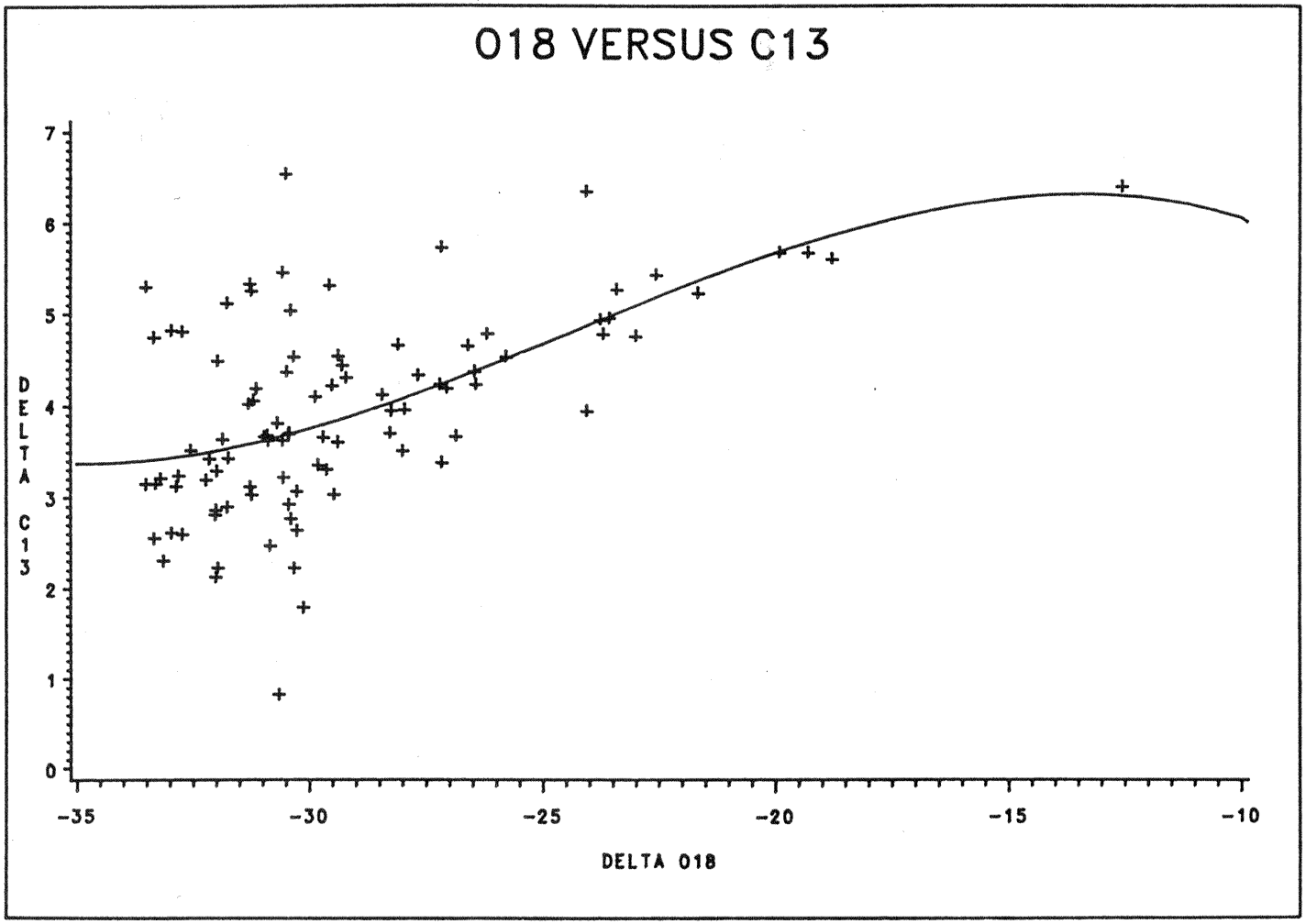


Figure 3.1 Plot of $\delta^{18}O$ versus $\delta^{13}C$, with a computer drawn line of best fit. There is a general trend for ^{13}C to increase with increases in ^{18}O.

by evaporation from the water mass with exchange of CO_2 with the atmosphere. It seems more likely that the enrichment in ^{18}O is due to evaporation from the lake. The water evaporating from the lake would be enriched in ^{16}O relative to the original lake water, causing a concentration of ^{18}O in the lake. In order for oxygen isotopic enrichment to occur however, there must be a direct loss of water vapour from a liquid surface, since $H_2O_{(l)}/H_2O_{(s)}$ fractionation is insignificant (O'Neil, 1968; Stewart, 1974). This would allow CO_2 exchange to occur giving equilibrium values at 0°C of +5.5 or greater at lower temperatures, which may be possible if the lake was slightly saline.

From table 3.2 it can be seen that the calcite pseudomorphs are enriched in ^{16}O and depleted in ^{13}C relative to their surrounding matrix. The difference in $\delta^{13}C$ values indicates that the calcite pseudomorphs formed in different waters to the surrounding matrix. The matrix formed around the original crystals in Glacial Lake Trowbridge, and hardened to preserve their shape, after the original crystals dissolved. The calcite pseudomorphs probably formed after Glacial Lake Trowbridge disappeared, the calcite crystallizing from accumulated meteoric waters, snow banks and drifts accumulating on the carbonate plates.

3:3 RADIOMETRIC DATING

3:3:1 INTRODUCTION TO THE U/TH DATING TECHNIQUE

Uranium has three naturally occurring isotopes, all of which are radioactive: ^{238}U, ^{235}U and ^{234}U. Thorium has six naturally occurring radioactive isotopes: ^{232}Th, which is the primary isotope, and five other isotopes occurring in nature as short lived intermediate

daughters of ^{238}U, ^{235}U and ^{232}Th, (Faure, 1977). Table 3.3 is a summary of the principal naturally occurring isotopes of U and Th.

Isotope	Abundance %	Half-Life years	Reference
^{238}U	99.2739	4.510×10^9	1
		4.468×10^9	2
^{235}U	0.7204	0.7129×10^9	3
		0.7038×10^9	2
^{234}U	0.0057	2.48×10^5	4
^{232}Th	100.	13.890×10^9	5
		14.008×10^9	6
^{230}Th		7.52×10^4	7

Table 3.3: Abundances and half lives of principal naturally occurring isotopes of U and Th (based on Table 12.2 in Faure, 1977).
1. Kovark and Adams (1955)
2. Jaffey *et al* (1971)
3. Fleming *et al* (1952)
4. Strominger *et al* (1958)
5. Picciotto and Wilgain (1956)
6. LeRoux and Glendenin (1963)
7. Attree *et al* (1962)

^{238}U, ^{235}U and ^{235}Th, are parents of a chain of radioactive daughters which ends with the stable isotopes of lead (see fig. 3.2).

The UO_2^{2+} ion is soluble in an oxidizing environment and the Th^{4+} ion is extremely insoluble, thus while thorium is rapidly removed from solution, uranium will remain in solution. The similarity of the UO_2^{2+} ion to the Ca^{2+} ion, means that it is often found in calcium bearing minerals such as calcite. ^{238}U and ^{234}U are usually incorporated in carbonates as they form, but not ^{230}Th. ^{234}U decays to produce ^{230}Th (with a half life, $t_{\frac{1}{2}}$=248,000 yrs). At any point prior to the time when the rate of decay of ^{234}U equals the rate of decay of ^{230}Th i.e. radiometric equilibrium, the age of the sample can be estimated by determination of the $^{230}Th/^{234}U$ and the $^{234}U/^{238}U$ activity ratios (i.e. calculating the amount of ^{230}Th produced since formation of the calcite).

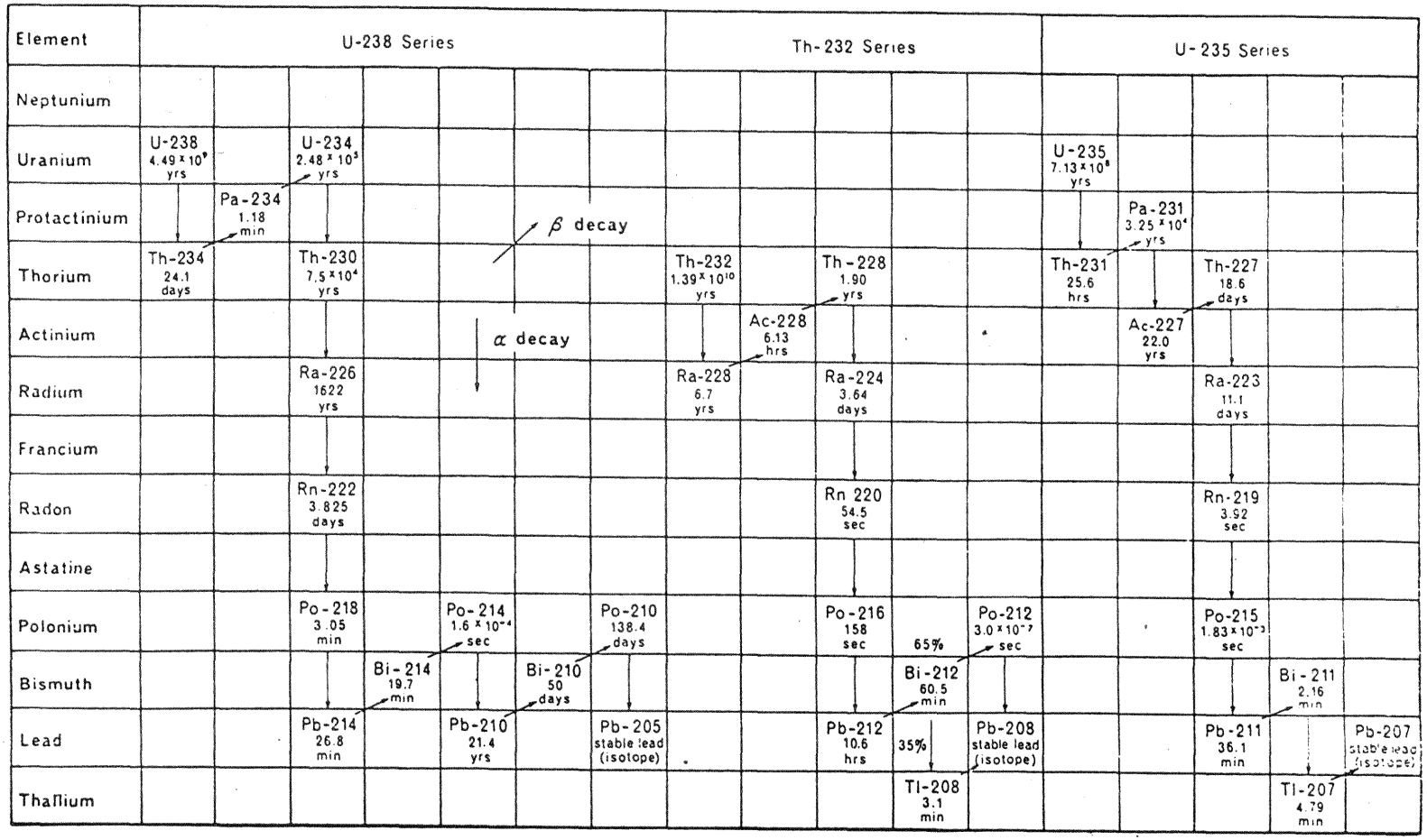


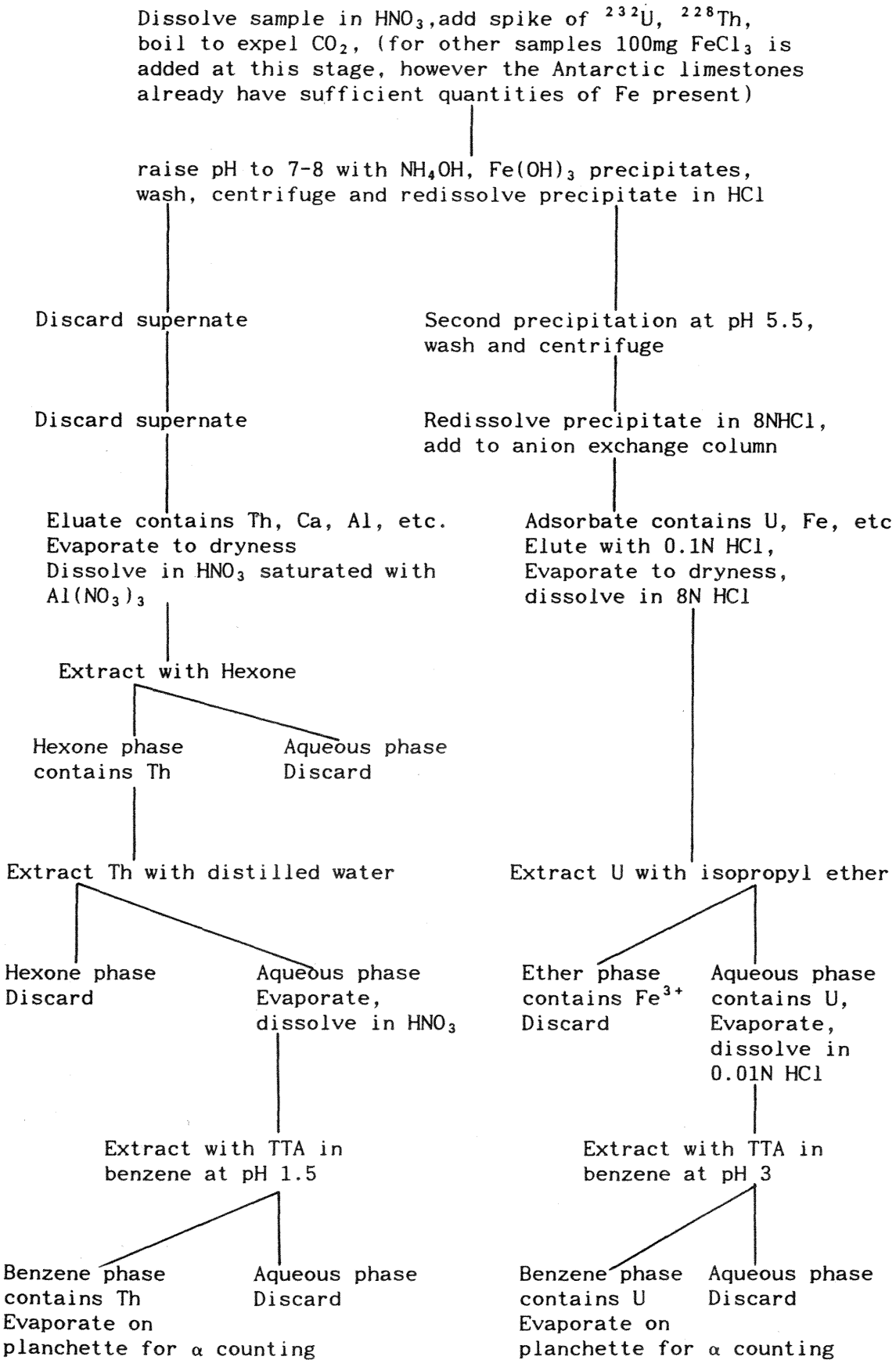
Figure 3.2 The decay chains of primeval uranium and thorium.

3:3:2 METHOD

The Miers Valley carbonate samples range from 68-97% $CaCO_3$, but on average contain 89% $CaCO_3$. The average crustal abundance of U is 1.8 ppm (Taylor, 1964), the Miers Valley carbonates have in comparison high concentations of U (between 1-30 ppm), and low Th concentrations.

The samples were dissolved in nitric acid and equilibriated with a spike of ^{232}U in radiometric equilibrium with ^{228}Th (Hendy *et al*, 1979). Following the techniques outlined by Kaufman (1964) and Goddard (1970), the uranium and thorium fractions were separated on an ion exchange column and recovered from the eluate using solvent extractions. The final extraction was with Thenoyltrifluoroacetone (TTA) in benzene, and this fraction was evaporated onto an aluminium planchette, which was flamed to remove organic matter and bind the uranium and thorium into a thin Al_2O_3 surface. The plated sample was then ready for counting. A schematic outline of the separation and solvent recovery is shown in figure 3.3.

Figure 3.3:Schematic diagram of U/Th separation and recovery



Samples were counted on an α-spectrometer with 450 mm^2 surface detectors (Hendy *et al*, 1979) for at least 24 hours, with the less active samples counting for several days. Examples of typical spectra obtained from the U and Th detectors are shown in figures 3.4 and 3.5.

Isotope concentrations were calculated by determining the area under the appropriate peak and dividing this by the count time. For the U spectra corrections then had to be made for;

1. The background activity of the detector,
2. The tails of one peak which overlap another peak, i.e. the tail of the ^{234}U peak under the ^{238}U peak and the ^{232}U tail under the ^{234}U peak.

Once these corrections had been made the activity ratios $^{234}U/^{238}U$ and $^{234}U/^{232}U$ could be calculated.

For the Th spectra corrections had to be made for;

1. The background activity of the detector,
2. Tail corrections for the tail of ^{230}Th under the ^{232}Th peak and the tail of ^{228}Th under the ^{230}Th peak,
3. The presence of ^{224}Ra in the sample, as the tail of ^{224}Ra is under the ^{228}Th peak and 4.9% of all ^{224}Ra alphas emit in the ^{228}Th region.

After these corrections had been made the activity ratios of $^{232}Th/^{230}Th$ and $^{230}Th/^{234}U$ could be calculated.

The age of the sample corresponds to the combination of $^{234}U/^{238}U$ and $^{230}Th/^{234}U$ ratios and is determined by interpolating between the isochrons shown in figure 3.6 (Kaufman, 1964). These dates can only be interpreted after the following assumptions have been considered;

1. No ^{230}Th was incorporated within the rock at the time of formation. If ^{230}Th was incorporated in a sample at the time of formation it would lead to an overestimation of the age of the sample. This assumption is tested by examining the ^{232}Th ratio of

Figure 3.4 A typical alpha spectra, obtained from the U fraction of sample TX 189. Alphas are emitted from each of the isotopes at, at least two energy levels, the smaller percentage producing the shoulders on the left side of the peaks. (Energy values from the CRC Handbook of Chemistry and Physics.)

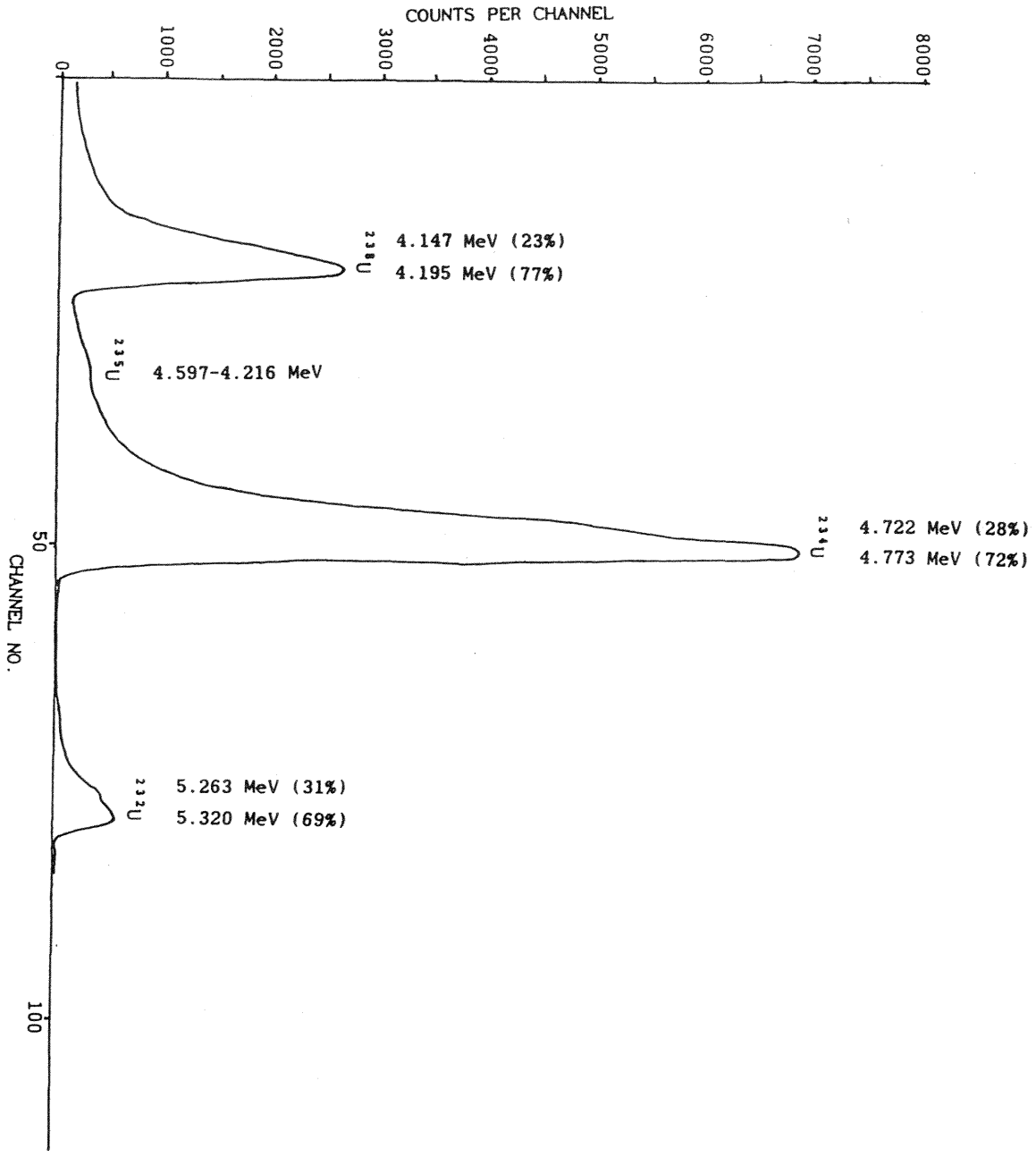

COUNTS PER CHANNEL
0
1000
2000
3000
4000
5000
6000
7000
8000
CHANNEL NO.
50
100
238U
4.147 MeV (23%)
4.195 MeV (77%)
235U
4.597-4.216 MeV
234U
4.722 MeV (28%)
4.773 MeV (72%)
232U
5.263 MeV (31%)
5.320 MeV (69%)

Figure 3.5 A typical alpha spectra, obtained from the Th fraction of sample TX 189. The double peaks showing clearly the two energies at which the majority of alphas are emitted from each isotope. (Energy values from the CRC Handbook of Chemistry and Physics.)

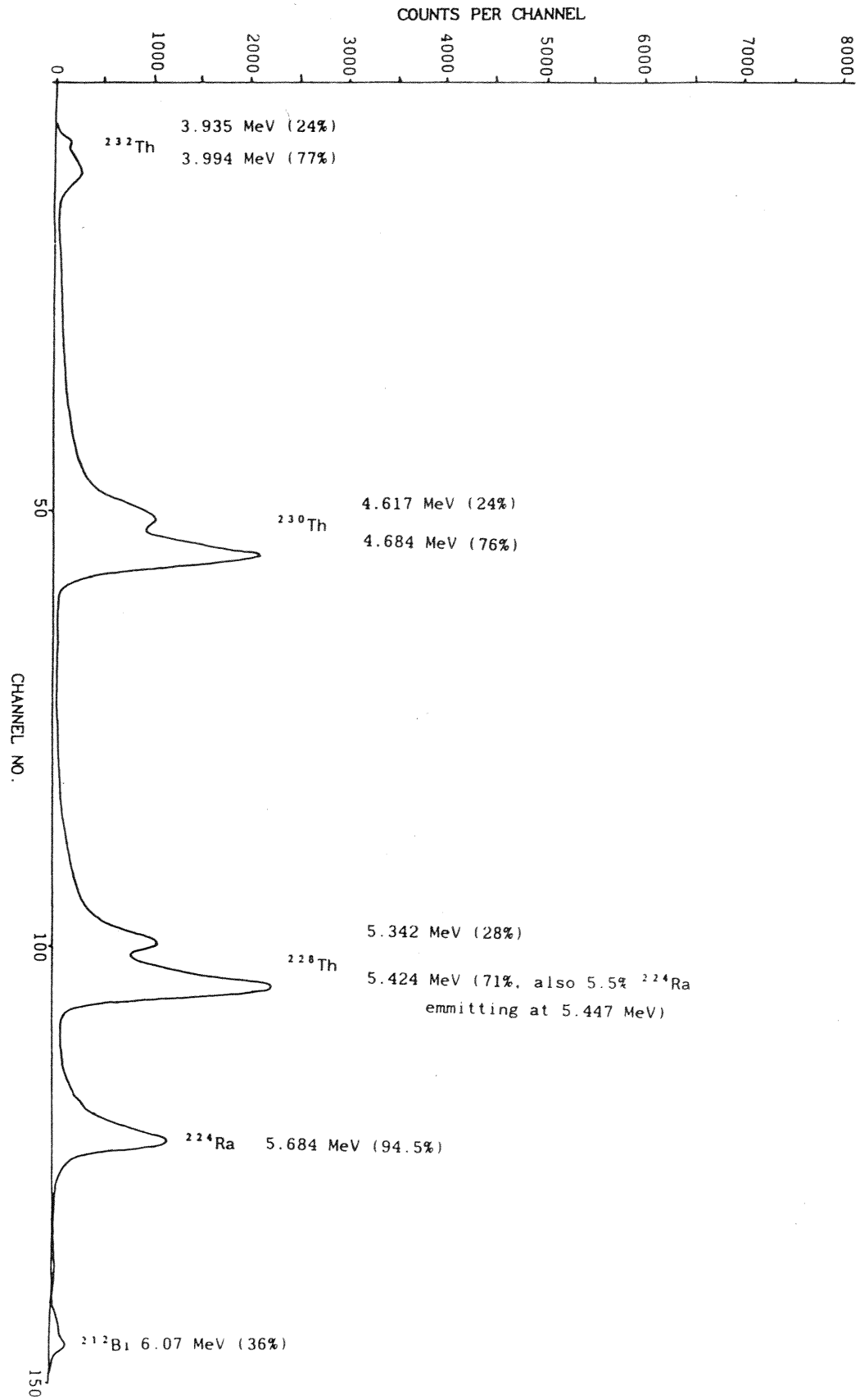
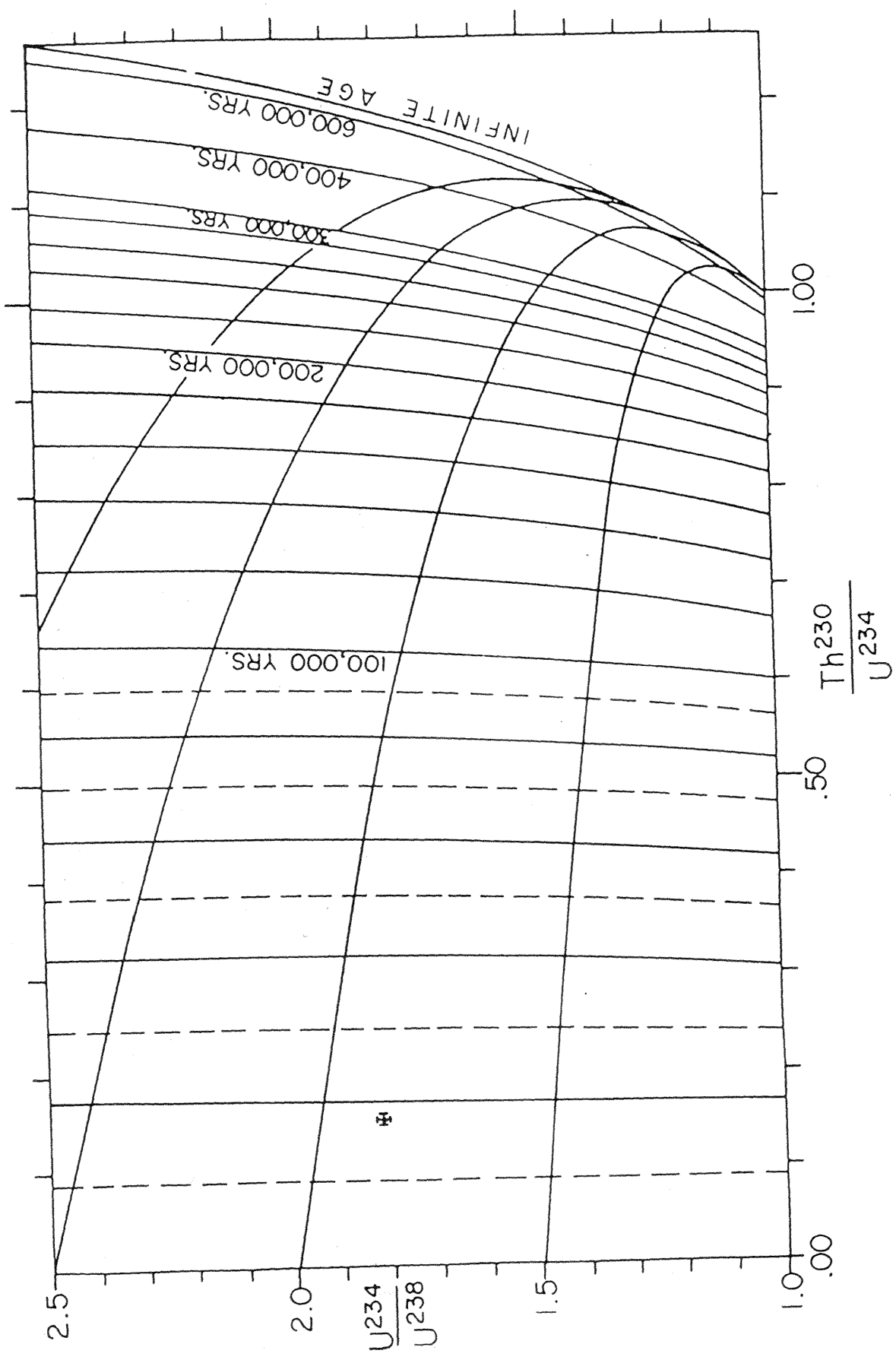

COUNTS PER CHANNEL
0
1000
2000
3000
4000
5000
6000
7000
8000
CHANNEL NO.
50
100
150
^{232}Th
3.935 MeV (24%)
3.994 MeV (77%)
^{230}Th
4.617 MeV (24%)
4.684 MeV (76%)
^{228}Th
5.342 MeV (28%)
5.424 MeV (71%, also 5.5% ^{224}Ra
emmitting at 5.447 MeV)
^{224}Ra 5.684 MeV (94.5%)
^{212}Bi 6.07 MeV (36%)

Figure 3.6 Isochrons produced by $^{230}Th/^{234}U$-$^{234}U/^{238}U$ ratios for various ages (from Kaufman, 1964).
The $^{230}Th/^{234}U$-$^{234}U/^{238}U$ ratios for sample TX 189 are plotted on the isochrons, giving a maximum age of $16.97 \pm 0.36 \times 10^3$ yrs BP. (The calculation sheet for this sample is included in Appendix III.)


INFINITE AGE
600,000 YRS.
400,000 YRS.
300,000 YRS.
200,000 YRS.
100,000 YRS.
1.00
.50
.00
$\frac{Th^{230}}{U^{234}}$
2.5
2.0
1.5
1.0
$\frac{U^{234}}{U^{238}}$

the sample. If ^{232}Th was incorporated in the rock at formation then it is likely that ^{230}Th was also incorporated.

2. The sample remained a closed system after formation, i.e. U and Th has neither left or entered the rock since its formation, except as a result of radioactive decay. To test this assumption samples were dated using both the U/Th and the C-14 methods of dating. The ages obtained with both these methods are compared in section 3:3:5 of this chapter.

3:3:3 PROBLEMS ENCOUNTERED WITH THE U/TH TECHNIQUE

A. Interferring Cations and Anions

It was found that each sample tended to have a unique chemistry, and the presence of trace quantities of some ions could frequently cause problems during the extraction processes and decrease the final recovery of the sample. Two of the commonly encountered ion problems are discussed below.

During the solvent extraction phases of the technique, precipitates, usually white, frequently formed. XRF analysis of the precipitate usually revealed that titanium was present, the pricipitate being TiO_2. The titanium presumably entered solution when the sample was being dissolved in nitric acid, originating from the detrital basaltic sands found within the carbonates. As U and Th may be absorbed onto TiO_2, when large quantities of the precipitate formed an attempt was made to redissolve the precipitate in aqua-regia. This was met with only partial success so for future samples care was taken that the samples were only in contact with the concentrated nitric acid for the minimum time required to dissolve the carbonate, and all extractions and washings were repeated several times to retrieve as much Th and U as possible.

Titanium was not the only ion to interfer in the extraction process. The large quantities of iron in the samples meant that it was possible to overload the anion exchange column. Larger exchange columns were adopted for the Antarctic samples and if overloading did occur the samples were put through a second column. The columns sometimes showed a tendancy to block after some samples. Close inspection of the gel in the column showed a white gelatinous material which proved to be a silica compound. The silica would build up in the column and when the normality of the solution passing through the column changed it would swell and block the column. Between each sample the columns were backwashed with water and 8NHCl to remove as much of the silica as possible from the columns.

B. Spike Disequilibria

The spike used for this technique is a spike of commercially prepared ^{232}U in radiometric equilibrium with ^{228}Th, i.e. the $^{232}U/^{228}Th$ activity ratio equals one. The spike solution is prepared by a thousand fold dilution of a mother liquor, which is in turn a thousand fold dilution of a stock solution of ^{232}U in radiometric equilibrium with ^{228}Th (supplied by the Radiochemical Centre, Amershum, U.K.). 10 ml of the spike solution is normally added to the samples to be dated (a sample normally consisting of 40-60g of lacustrine carbonate).

During this study, a new spike solution was discovered to no longer be in radiometric equilibrium (i.e. the $^{232}U/^{228}Th$ activity ratio no longer equalled one). Another spike solution was prepared from the mother liquor and tested by taking small (5ml) samples of the spike, adjusting to the appropriate pH values (see fig. 3.3), extracting the Th and U into TTA and then plating and counting the activity of the spike samples. This new spike also was not in equilibrium. It was presumed that the mother liquor was no longer in

equilibrium, possibly due colloid formation within the solution. A fresh mother liquor was prepared by diluting 1ml of the stock solution (0.2μCi/10ml) in 100ml of 1N HNO_3. A five hundred fold dilution of the new mother solution made a new working spike which on testing by preparing planchettes proved to be in equilibrium.

Unfortunately before it was realised that the spike was no longer in equilibrium, the disequilibriated spike had already been added to six samples, the dates from which could not be interperated. A summary of dating attempts is listed in Appendix III

3:3:4 CARBON-14 DATING

The sample ages fall within the age range of the Carbon-14 method of radiometric dating. For comparison of the two methods and as a way to test the validity of assumption 2 of the U/Th method, carbonate samples were submitted to the Carbon-14 Laboratory of the University of Waikato. A sample of algae from the 'gypsum moraine' section was sent to the University of Washington for C-14 dating. This particular site allowed the employment of three dating techniques, U/Th and C-14 dating of the overlying and underlying carbonate beds, and C-14 dating of the algae sample which occurs between them.

3:3:5 RADIOMETRIC DATES OBTAINED

Previous U/Th dates (see appendices) on the carbonates from Miers Valley returned ages between 8,000 and 18,000 years BP. Gypsum samples had also been dated with the U/Th technique, these samples returned older less consistent ages (>100,000 yrs BP). The gypsum samples of Miers Valley are weathered and show signs of recrystallization,

therefore the dates from these samples could be incorrect due to the possible post depositional contamination by ^{230}Th or loss of ^{234}U.

In this study only calcium carbonate has been dated and the surface deposits have returned U/Th ages of between 10,000 and 17,000 yrs BP and C-14 ages of between 10,000 and 23,000 yrs BP. A summary of the U/Th and C-14 dates is shown in table 3.4. Although the dates fall within the same age range, between 10,000 and 20,000 yrs BP, frequently the U/Th and C-14 ages of individual samples are not in total agreement. The differences between the two sets of dates are not systematic, five samples give U/Th dates which are younger than their corresponding C-14 dates, and three samples give U/Th dates which are older than the corresponding C-14 dates.

Sample No.	C-14 No.	Altitude a.s.l. (metres)	U/Th age x 1000 yr	σ	C-14 age x 1000 yr	σ
C861	WK:609	81	-	-	22.95	0.36
C862	(J23c)	81	10.5	0.5	26.2	1.15
C878	WK:627	101	13.0	0.4	14.2	0.12
C893	WK:718	121	15.2	0.25	10.3	0.9
C899	WK:717	88	16.97	0.36	14.2	0.12
C908(u)	WK:814	113	15.3	0.5	14.4	0.13
C908(m)	WK:813	113	13.7	0.3	18.1	0.20
C908(l)	WK:812	113	15.7	0.5	19.85	0.22
C916	WK:628	140	-	-	18.6	0.17
C921	WK:630	125	-	-	16.75	0.26
C923	WK:716	78	-	-	19.15	0.18
C931		156	13.25	0.2	-	-
C931	WK:720	156	13.3	1.4	13.7	0.18
C938	WK:719	80	11.2	0.4	14.8	0.17
C941	WK:629	100	-	-	17.95	0.16
C943	WK:582	142	-	-	16.00	0.14

Table 3.4: Summary of radiometric dates from Miers Valley lacustrine carbonates.

Erroneous U/Th dates may be obtained if assumptions of the technique are violated, there was no indication that ^{232}Th was in excess in the samples so it is likely that assumption 1. holds. If assumption 2. is violated and the system has not remained closed to U and Th since the carbonates formed, the dates will be incorrect.

Erroneous C-14 dates may be produced if the carbonates were contaminated by older carbonate in the form of detrital grains or by forming in a water mass which (i) may have exchanged with older carbonate bearing rocks, or (ii) may have been contaminated by dissolved CO_2 which had been trapped in glacial ice which melted and flowed into the water body without exchanging with the atmosphere. All of these possibilities are feasible in Miers Valley, an ice sheet lay in the valley mouth supplying meltwater to the lake, and marble occurs in the valley walls which could also supply detrital carbon. Bell (1967) noted that there were high concentrations of Ca^{2+} in the waters of present Lake Miers and attributed this to the marbles in the area, if the waters are enriched in Ca^{2+} they could also be enriched in carbon. However, Ca^{2+} could also have entered Lake Miers through leaching of gypsum from the Glacial Lake Trowbridge sediments surrounding it. Thin sections of the carbonates were examined to determine if detrital calcite is present within the rocks, the detrital calcite occurring as flakes of crystalline material. Although the occasional flake of detrital calcite was observed, there was not enough to effect the C-14 dates.

The algae sample from the 'gypsum moraine' section (QL-910) returned an age of 19,770±110 yrs BP. This is younger than both the carbonate C-14 dates above and below it, and older than the carbonate U/Th date below it (unfortunately the U/Th date on the upper carbonate was invalid due to the spike disequilibria). The C-14 dates of the carbonates at this site are stratigraphically correct. With the possibility of detrital calcite eliminated, it leaves only the possibility of carbon contamination of the water body from which the carbonates precipitated and in which the algae grew. If this had occurred it would be expected that all the water body would be effected and the errors would be systematic and reflected in the algae

and carbonate alike. However a dynamic disequilibrium between lake waters and the atmosphere such as this is likely to change with time, so that different samples at slightly different ages could show considerable differences in old carbon contamination. There is one factor which casts doubt on the validity of this particular U/Th date, and that is, the sample had stood in solution for a considerable period of time without the spike added to it, during its preparation.

The C-14 dates of the individual layers of sample C908 are also stratigraphically correct, as are the U/Th dates of the uppermost and lowermost layers, however the U/Th date of the middle layer appears out of sequence. This would point to the C-14 dates being more believable, despite the possible contamination of carbon in the water mass.

Whichever set of dates are in fact correct, the differences are not great and they show that Glacial Lake Trowbridge occupied Miers Valley between 10,000 and 23,000 years BP. Proglacial lakes of similar ages have been shown to have occurred in other Dry Valleys, such as Glacial Lake Washburn in Lower Taylor Valley (Stuiver *et al*, 1981), Lake Vida in the Victoria Valley, Lake Vanda in the Wright Valley and possibly lakes in the Salmon and Garwood Valleys (Stuiver *et al*, 1981).

The younger samples tend to occur towards to eastern end of the eastern basin and it is possible that these represent the last stages of Glacial Lake Trowbridge as the Ross Ice Tongue retreated from Miers Valley. The older dates extend up the valley to the west and these probably represent earlier higher levels of the Glacial Lake Trowbridge and the Ross Ice Tongue. The dates from the individual layers of sample C908 show a time difference between the formation of the individual layers, indicating that the calcite deposition possibly was not occurring continuously during the existence of the glacial lake but proceeded in stages.

CHAPTER FOUR; MINERALOGY

4:1 X-RAY DIFFRACTION ANALYSIS

Analysis of lacustrine carbonate beds in other Dry Valleys has shown that the carbonate deposits are frequently a mixture of calcite and aragonite. Lawerence (1982) showed that although calcite is currently precipitating in Lake Fryxell, aragonite was precipitated 10,500 years ago. Hendy and Utting (1984) have shown that in Marshall Valley, calcite and aragonite have been alternately deposited. To determine the mineralogy of Miers Valley carbonates, samples were analysed on an x-ray diffractometer.

A number of the Miers Valley carbonate samples were ground and back-filled into Philips aluminium sample holders and analysed on a Philips P.W. 1050 x-ray diffractometer. The charts produced were then examined to determine the carbonate mineralogy. Any non-carbonate peaks produced from the samples were also noted and identified.

This analysis showed that the Miers Valley carbonates were composed solely of calcite, this included the carbonate plates, the needle-like pseudomorphs, and the carbonate surrounding the psuedomorphs. Some samples showed the presence of detrital material such quartz, feldspars and micas, which occurred as sand grains within the carbonates, originating from the surrounding country rocks.

Hume and Nelson (1982) outlined a method that related the d-spacing of the calcite peak to the magnesium content of the calcite. An internal standard such as silicon powder or halite is added to the samples and intimately mixed. The samples are then scanned at a slow scanning speed ($\frac{1}{2}$°2θmm/min). The exact d-spacing of the major calcite peak is then calculated from the position of the standard peak. Halite

was added to twenty of the Miers Valley carbonates and they were then scanned at the appropriate speed. They all showed a very low magnesium content, less than 7% mol%Mg.

4:2 THIN SECTIONS

Samples of carbonate were prepared by cutting the carbonate plates into small (≈ 1cm x 1cm x 2cm) pieces which could be mounted onto microscope slides. The samples were then thoroughly dried. As the carbonate is not hard enough to be mounted directly onto the glass slides, the samples were set in an epoxy resin. The resin blocks were cut and mounted onto glass slides, then cut and polished. Both horizontal and vertical carbonate cross sections were mounted as thin sections.

General description of the carbonate plates:-

The carbonates are very porous and the pores are not generally connected. All the sections show a calcite matrix containing variable amounts of clastic material. Following the classification of Folk (1968), the carbonates are micrites, that is microcrystalline with a mud-supported texture. Although the crystal shape of the calcite in the matrix is not recognizable, small crystals can be seen to protude into the pore spaces. Large euhedral crystals of recrystallized calcite are also seen to have grown into the pore spaces. Clastic grains occur randomally through the matrix.

The thin sections contain on average, 90% calcite, most of which occurred as a fine grained matrix, 0.5-3% of this calcite was in the form of large euhedral crystals which were recrystallizing in cavities and pores. The samples also contained between 2% and 10% volcanic glass, 0.5-7% rock fragments and 0.5-3% organic material. The percentage of rock fragments tended to increase in samples from lower

altitudes, probably an effect of detritus washing into the valley and accumulating at the base of the valley floor, whilst the higher altitude samples tended to be almost detritus free (0.5-2% rock fragments). Individual section descriptions are shown in Appendix V. A summary of sample compositions is shown in Table 4.1.

Sample No.	Altitude (metres) a.s.l.	%Calcite matrix	%Calcite euhedral	%Glass Shards	%Organic Matter	%Rock Fragments
C843V	143.5	97		2.5	0.5	0.5-1
C843H	143.5	96	0.5-1	2	0.5	0.5-1
C861V	83	80-85	1	10-12	3-5	2-5
C861H	83	85-90		3	1	7
C861H	83	90		5	2	3-4
C861V	83	90		4	1	5
C861H	83	85-90		7	3	4
C861V	83	90-95		4	1	3-5
C899V	88.1	90	1	5	2	2
C899H	88.1	85-89	1	8	2	3-5
C909H	108.7	95	2	1.5	0.5	2
C909V	108.7	90-95	3	2	1	1
C916	140.8	90-95	1-3	4	1	1
C941V	100	88-90	1-2	6	1.5	2-3
C941H	100	95	2	2	0.5	2
C943V	141	90	1	5-6	1.5	3

Table 4.1: Estimated percentage composition of samples examined under a petrographic microscope.

The rock fragments were typically a mixture of quartz, feldspars, biotite, olivine, augite and calcite, with the occasional occurrence of minerals such as epidote and hypersthene. The detrital calcite was visible from the surrounding matrix as it tended to occur as flakes and frequently displayed the cross-hatching effect which is typical of crystalline calcite. The grains varied in size and were subrounded to subangular. The mineralogy of the rock fragments is typical of that expected to enter the glacial lake, the minerals arising from the metasediments of the valley walls or from the volcanics of McMurdo Sound. There are relatively high percentages of volcanic glass in the samples, some of which contains phenocrysts. This volcanic glass would have originated from the McMurdo Sound volcanics and was transported

into the valley by the Ross Ice Tongue or was blown into the valley with other aeolian material.

Several sections were cut through the moulds seen in the carbonate plates. They often appeared to be almost square in cross section and sometimes had calcite recrystallization within the mould. No other noticeable features of the moulds were observed.

4:3 SCANNING ELECTRON MISCROSCOPE STUDIES

Scanning Electron Microscope studies was useful in that they can sometimes provide an insight into the crystallization patterns of precipitating minerals such as calcite, or can indicate the degree of weathering of a sample. Four samples were examined under a Scanning Electron Microscope (SEM) to determine if there was any difference between carbonate plate samples, the needle like psuedomorphs, the carbonate moulds or the carbonate surrounding the pseudomorphs. The upper and lower surfaces of the plates were also examined. The mould surfaces were of particular interest in case any features of the original crystal were present.

The carbonates were all weathered, the calcite crystals showed pitting and had a flaky appearance (see fig. 4.1). The top surface of the carbonate plates showed greater weathering than the lower surfaces (see fig. 4.2), and individual calcite crystals were less discernable. With the exception of sample C774 (collected 1982), both the tops and the bottoms of the plates showed random orientation of the crystals. Diatoms, although weathered, do occur in the carbonates (see fig. 4.3), indicating carbonate formation within a water body. The mould surfaces without magnification appeared relatively smooth compared to the surface of the plates. On examination under a binocular light microscope and the SEM, these surfaces were seen to consist of randomally orientated weathered calcite crystals (see fig. 4.1). The

A. The carbonate surface within a mould, sample C918 (330x)

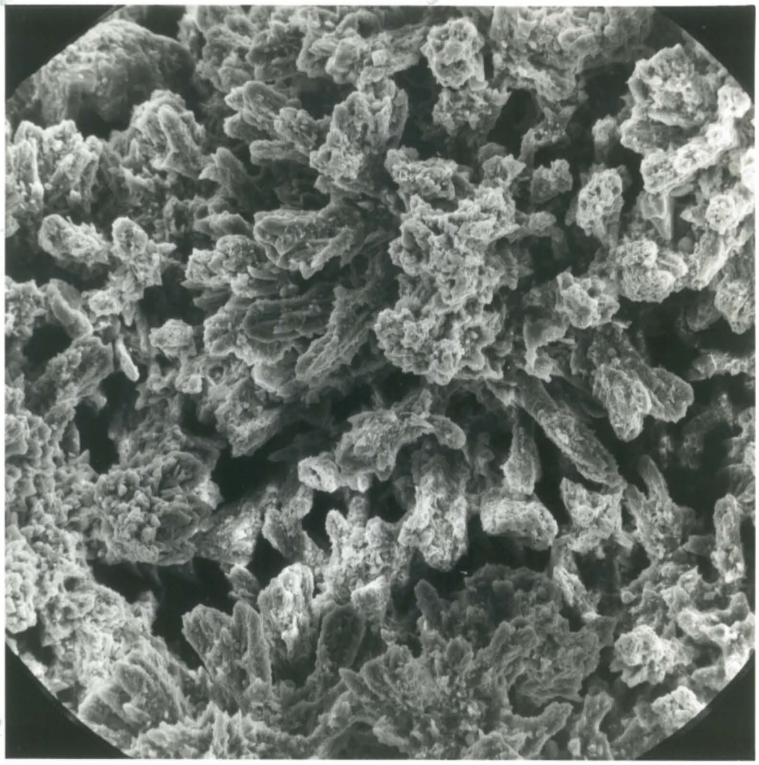
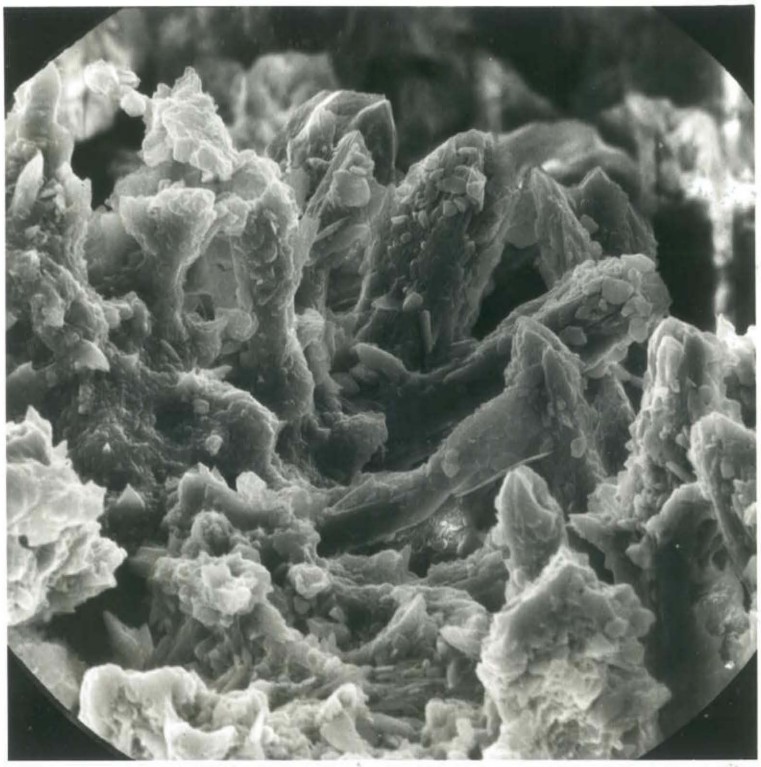
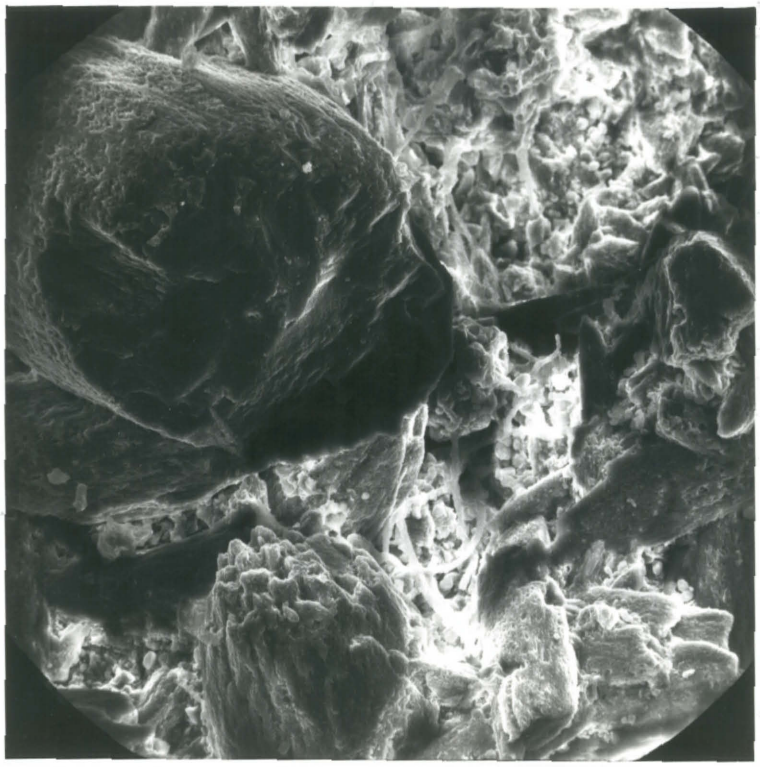


Figure 4.1 Weathered calcite crystals showing pitting and flaking. The crystals are randomly orientated. (by Dr. C. Beltz)



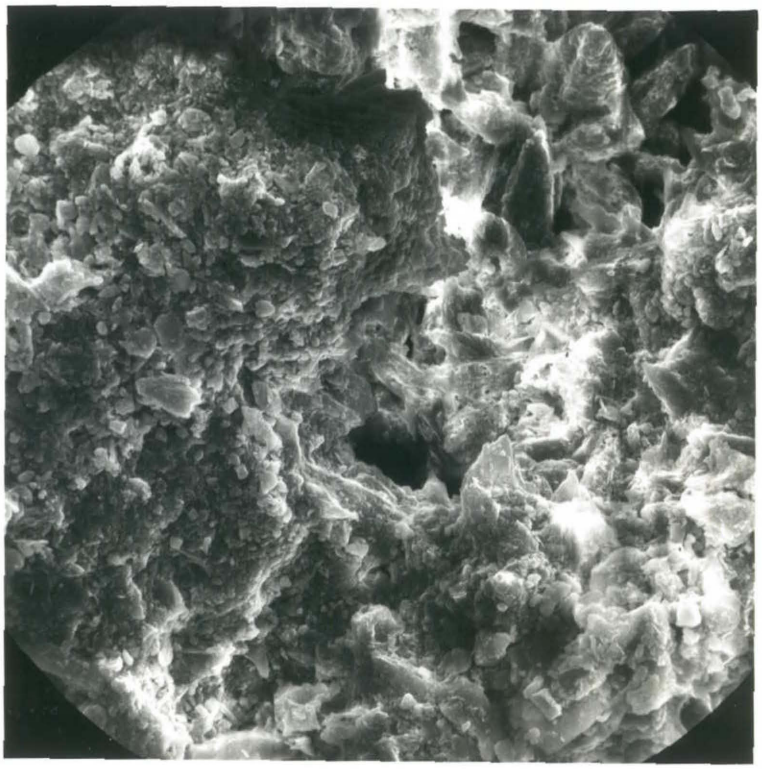
B. The carbonate surface within a mould, sample C918 (1200x)



A. Upper surface (360x). The large calcite crystals are pitted and weathered.

Figure 4.2 Upper and lower surfaces of a carbonate plate (sample C941).
(by Dr. C. Beltz)

B. Lower surface (550x). A 'flat' surface showing no outstanding features, this view is typical of the carbonate plate surfaces.



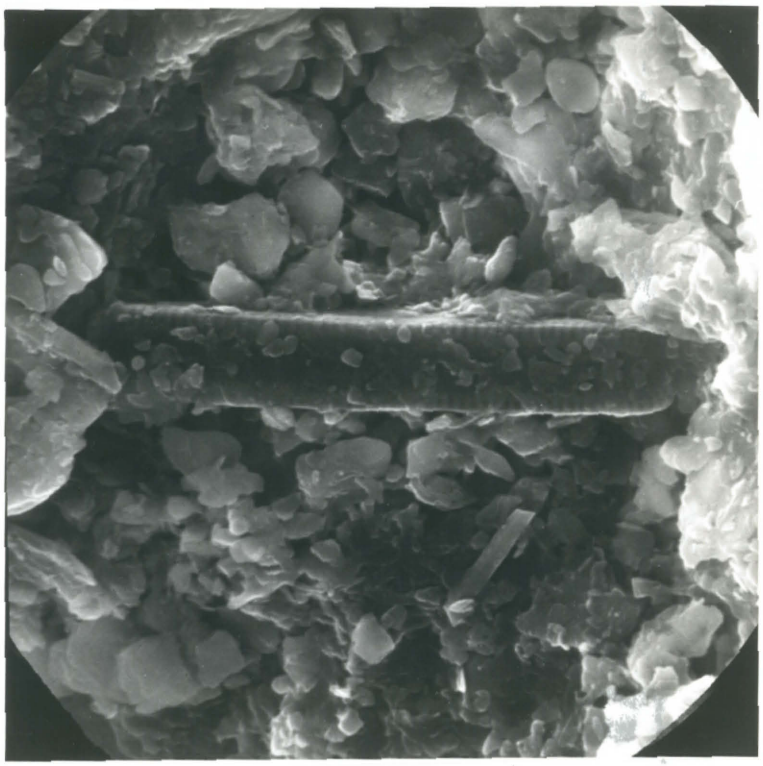


Figure 4.3 Weathered diatom within a carbonate plate, sample C941, (3300x)
(by Dr. C. Beltz)

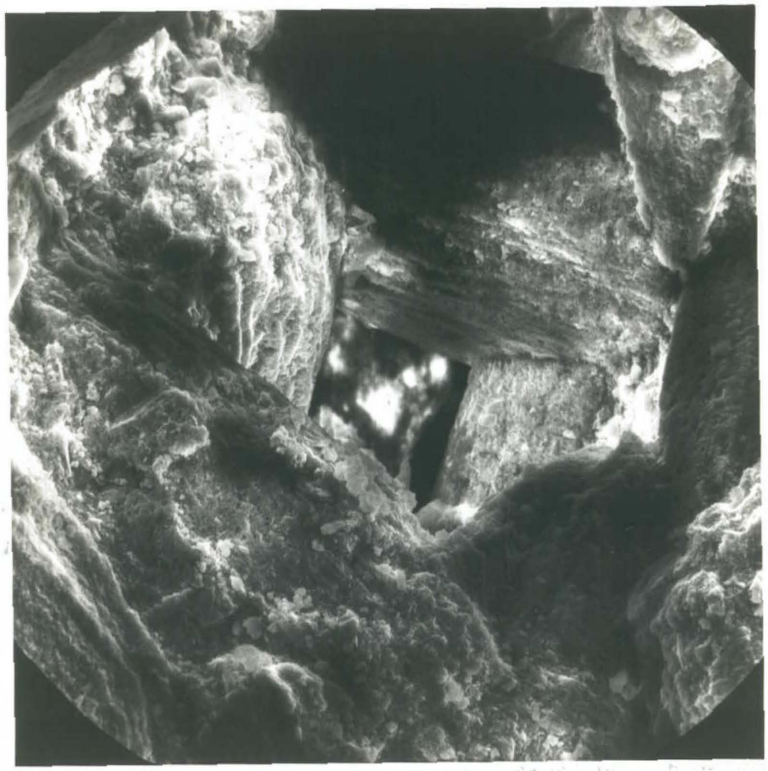
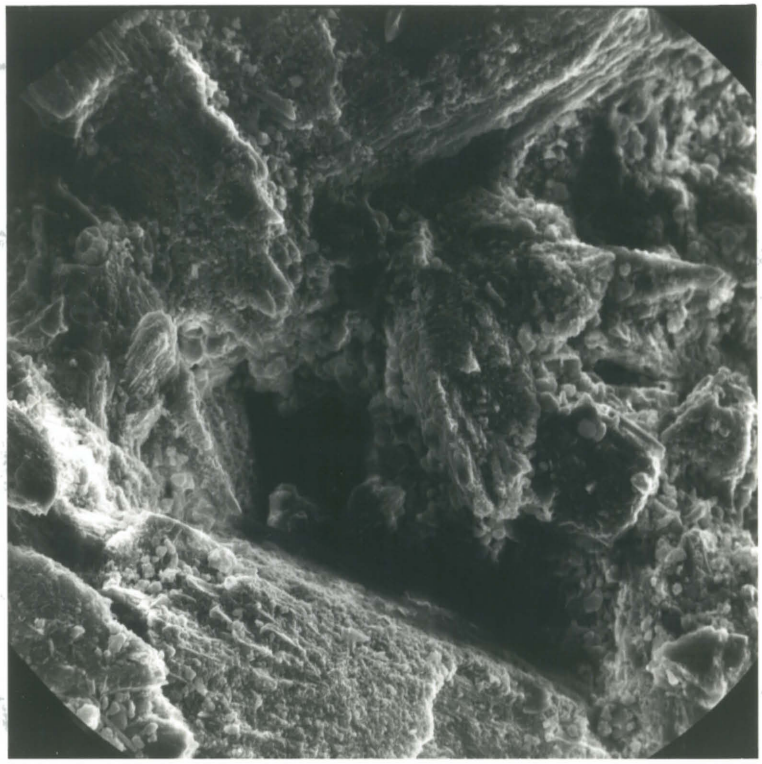


Figure 4.4 Mould within the lower surface of carbonate sample, C857, (390x)
(by Dr. C. Beltz)



A. The lower surface of carbonate sample, C857, showing smooth surfaces within a cavity. This may be a mould or the fortuitous meeting of crystal faces to give mould characteristics. (390x).

B. The upper surface of sample C857. Calcite crystals appear to be infilling a cavity with a square cross section. (750x)

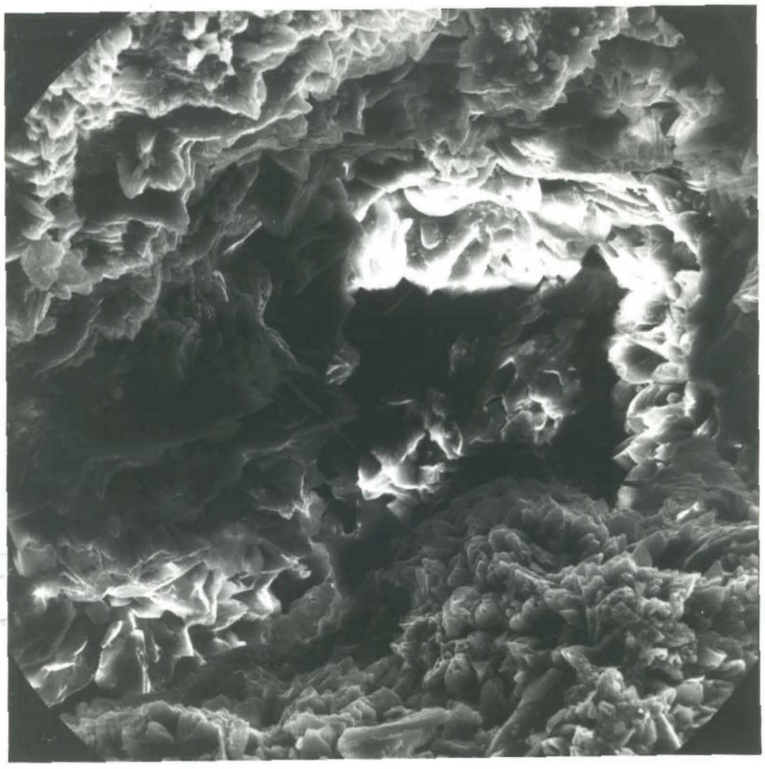
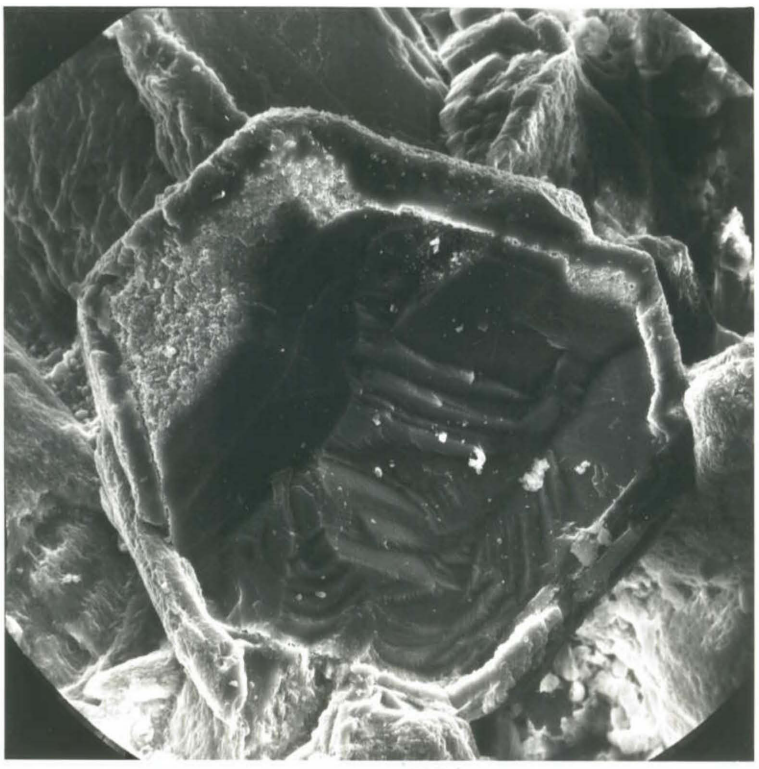


Figure 4.5 Other possible mould surfaces. These features may not be moulds within the carbonate, but they do have a regular appearance characteristic of mould cross sections. (by Dr. C. Beltz)

Figure 4.6 Weathered recrystallized calcite. Both the large calcite rhombs and the smaller recrystallized calcite show pitting and weathering. This is a typical view of the surface of a calcite pseudomorph. Sample C918 (80x)
(by Dr. C. Beltz)

Figure 4.7 Cross section of a single calcite crystal. A weathered crust is visible, contrasting with the fresh, fractured interior. The top left of the crystal interior is slightly more weathered, possibly representing an exposed section of the crystal through a fracture.(Sample C941) (500x).
(by Dr. C. Beltz)



moulds seen without magnification were also observed at 390x magnification (see figs. 4.4, 4.5). Although it is not possible to determine the crystal angles from such viewing one mould appeared to have faces meeting at right angles (see fig. 4.4) while another appeared to have faces meeting at an obtuse angle (see fig. 4.5A).

The calcite pseudomorphs consisted of randomally orientated large weathered calcite crystals with clusters of smaller crystals dispersed amongst them. The clusters of smaller crystals could possibly represent a period of recrystallization, but they are also weathered, thus recrystallization is not currently occurring (see fig. 4.6). One photo (fig. 4.7) shows a cross section of a single calcite crystal which appears to have had its top broken off, exposing a crust around the crystal. This probably represents a weathering crust of the calcite. Most of the internal calcite of this crystal appears fresh and unweathered, displaying a subconchoidial fracture characteristic of calcite. A small area of the cross section is pitted and it is probable that this area had been exposed for longer possibly due to a fracture.

The Dry Valleys are a harsh arid environment where physical weathering tends to dominate over chemical weathering, however the state of Miers Valley carbonates shows that chemical weathering does occur if the carbonates are exposed. Much older carbonates in neighbouring Marshall Valley are comparatively unweathered (Utting, 1983). These carbonates outcrop in a gully cut by a Marshall Valley stream and are generally covered with material which has preserved the samples. The Miers Valley carbonates show that chemical weathering is occurring at the ground surface where the carbonates are exposed. This weathering is seen by the formation of crusts around the individual crystals and pitting and flaking of the crystal faces. The age of the carbonates (10,000 - 20,000yrs B.P.) gives an indication of the rate at which this chemical weathering can proceed.

4:4 MOULDS AND PSEUDOMORPHS

Chapter 2 describes the needle like moulds which occurred throughout the eastern basin. Near the gypsum moraine they occur in all the carbonate plates with some moulds showing calcite recrystallization. Westward the degree of recrystallization increases until at the western edge of the eastern basin there are distinct calcite pseudomorphs held together in a matrix of fine grained calcite. Over the hummocky drift between the basins and in the western basin only calcite pseudomorphs occur within weakly cemented silts. These variations are shown in figures 4.8 and 4.9.

To form these moulds the original crystal would have grown at the same time as calcite was precipitating within the glacial lake, either growing within the soft calcite, growing within the water column and settling onto the bottom, or growing within the ice raft at the surface of the lake. The size of the crystals tends to eliminate the second option, so it seems likely that the crystals were growing in the soft calcite on the lake bottom. The crystal growth appears to have occurred every time calcite was precipitating, or at least has been preserved within the calcite record. At some time after the lithification of the calcite sediment, a change in environment caused the original crystal to dissolve leaving the casts into which calcite precipitated. These moulds and pseudomorphs occupied a considerable area ($\simeq$10 km^2), dissolution of the original mineral would have released considerable quantities of its components. On dissolution these ions, would either reprecipitate or would be flushed from the valley system. The post-dissolution history of these ions should therefore be considered when attempting to assign a mineral to the moulds.

A. Sample C941 showing randomly orientated moulds within a carbonate plate.

Figure 4.8 Types of carbonate found in Miers Valley.

B. Sample C897 showing an infilling of many moulds with recrystallized calcite.

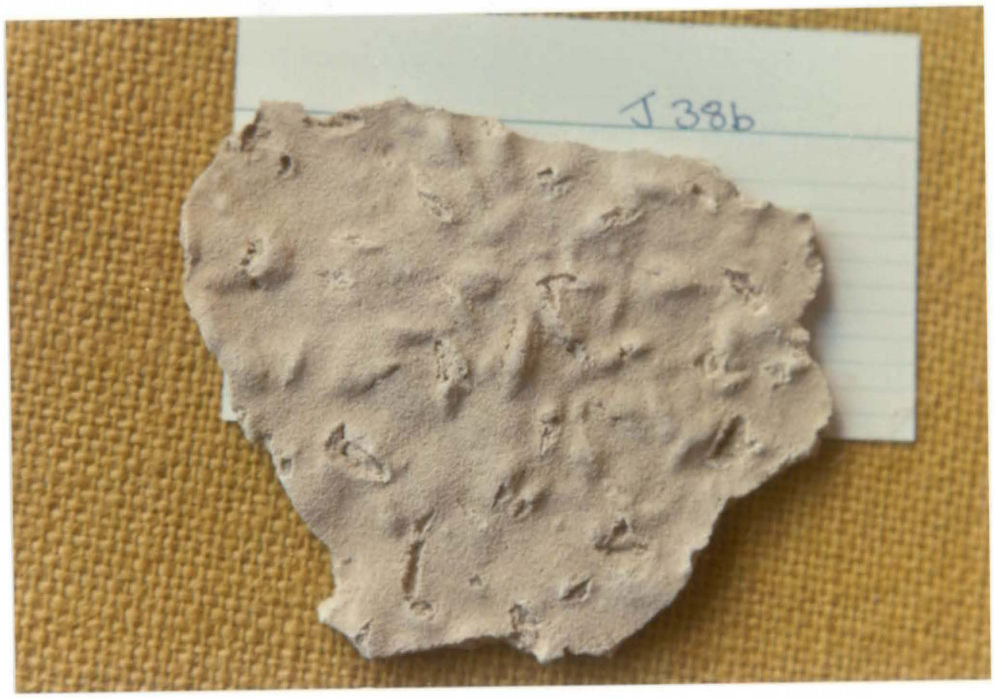


Figure 4.9 Types of carbonate found in Miers Valley, cont.

A. Sample C918, broken carbonate with distinct moulds and pseudomorphs. (Note the carbonate matrix is no longer in distinct plates.)

B. Sample C915, a selection of loose calcite pseudomorphs.

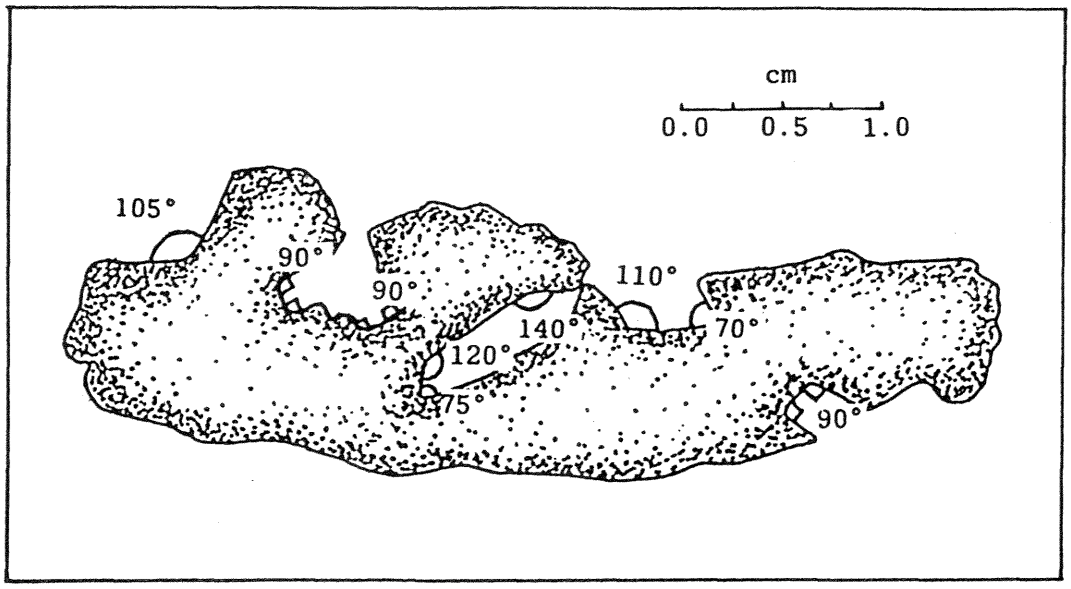


Figure 4.10 Sketch of a typical cross section of a carbonate plate, (Sample C845), showing random orientation of moulds within the carbonate and the angles within the moulds. Note also the twinned mould outline on the upper surface of the plate.

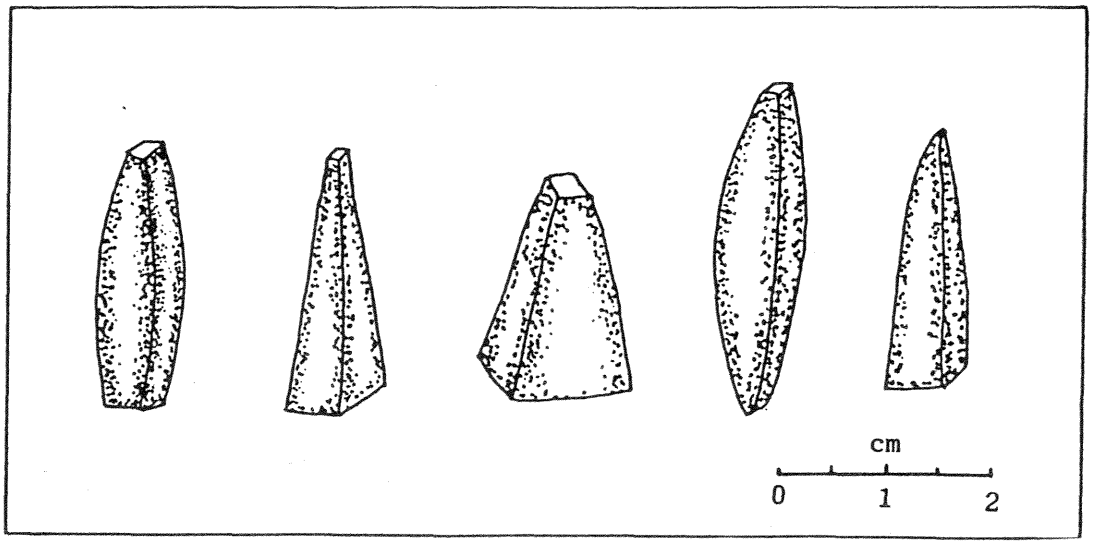


Figure 4.11 Sketches of typical calcite pseudomorphs. (From a number of samples.)

The moulds and pseudomorphs have the same shape and vary in length from several millimetres to approximately 15mm. The crystal faces appear to meet at right angles. The crystals taper out to the ends to give the needle like appearance. A sketch of a typical cross section of a carbonate plate containing these moulds is shown in figure 4.10, and sketches of typical pseudomorphs are shown in figure 4.11.

An attempt was made to obtain resin casts of the carbonate moulds, but the porous carbonates tended to 'soak up' the resin and a complete dissolution of the carbonate in acid was not possible. It was found that a more successful method of obtaining casts was to pack the adhesive "BLU TACK" into the moulds. The "BLU TACK" would hold the impression of the mould on removal from the carbonate, the resulting impression could be used for crystal angle and face measurements. The calcite pseudomorphs were also used to obtain crystal angle and face measurements. It was found that the *a* and *b* crystallographic axes were usually equal and the *c* axis elongated. Although a large proportion of the interfacial angles were 90° the angles could vary from 70° to 120°.

It was presumed that the original mineral was a soluble salt. Although all the possible evaporitic salts were considered it is likely that the salt was one which has already been recorded in the cold, arid region of Antarctica. Of the 30 salt phases recorded in the McMurdo Sound region, 10 were regarded as widespread by Keys (1979) and Keys and Williams (1981). These widespread salts are; thenardite (Na_2SO_4), gypsum ($CaSO_42H_2O$), halite (NaCl), calcite ($CaCO_3$), darapskite ($Na_3NO_3SO_4H_2O$), soda nitre ($NaNO_3$), mirabilite ($Na_2SO_410H_2O$), bloedite ($Na_2Mg(SO_4)_24H_2O$), epsomite ($MgSO_47H_2O$), and hexahydrite ($MgSO_46H_2O$). Thenardite usually occurs as an alteration product of mirabilite, as it is the stable form of sodium sulphate when exposed to the atmosphere. Mirabilite was seen at the eastern end

of the valley and showed a crystal form that was distinct and different from that of the moulds (see fig. 2.7). Soda nitre, darapskite, bloedite, epsomite and hexahydrite all usually occur in the Dry Valleys as accumulations beneath boulders, cobbles or pebbles, or as salt nodules or horizons in the regolith (Keys, 1979). Halite usually forms a cubic crystal which does not resemble the moulds.

Of the polymorphs of calcium carbonate, calcite tends to form rhombic crystals, and aragonite forms orthorhombic 'needles'. Although calcite and aragonite have been reported as occurring together in Marshall Valley (Hendy and Utting, 1984), the aragonite was considerably smaller in size (detectable in SEM studies). It was proposed that the aragonite grew in the water column, raining down onto the sediment, and the calcite subsequently precipitated around it. In the Miers Valley situation the needle-like crystals are considerably larger and probably took longer to grow. The precipitation of calcium carbonate into different polymorphs is thought to be influenced by impurities in the crystal (Zeller and Wray, 1956). It therefore does not seem possible for both calcite and aragonite to precipitate at the same time from the same water body, especially considering the time that was possibly needed for the crystals to obtain their size. If the Marshall Valley situation had occurred, the moulds and pseudomorphs would be expected to occur in particular areas, the needles clustered together and possibly interlocking. In fact, the casts occurred throughout the calcite plates and were dispersed, only rarely interlocked.

Gypsum is the most likely salt to have formed the moulds, although individual crystals are usually not in pronounced needle like shapes, gypsum needles can occur (Palache *et al*,1966) and have been seen growing in clusters in the neighbouring Marshall Valley (see figure 4.12). Individual crystals have been found in sand in the Lake

Figure 4.12 Two views of gypsum needles occurring in Marshall Valley.

Vashka area of Victoria Valley (Gibson, 1962). Gumbley (1975) found that gypsum and calcite occur together on the bottoms of Lakes Vanda and Bonney, however these deposits were fine grained and occurred in varve like bands. Although gypsum appears to be the most likely candidate for the original mould impressions, a further study on the habits of gypsum crystallizing in calcite at cold temperatures would be needed before it could be conclusively identified as the original crystal. It is possible that some other salt phase may have formed within the calcite, a phase which is not stable at present day conditions and hence not seen in the Dry Valleys at present.

If the moulds were in fact from gypsum, the redistribution of the gypsum can also be explained by the extensive gypsum deposits in Miers Valley. The gypsum probably began precipitating soon after the carbonate precipitation commenced, with long needle like crystals growing within or at the surface of the carbonate. Pauses occurred in the calcite sedimentation possibly through a change in temperature, at least three times, forming the layers of calcite plates. It is possible that during the latter stages of the ice tongue occupation when the terminus was in a position near the 'gypsum moraine', large quantities of gypsum were deposited into the lake from the terminus of the ice front. A similar situation is seen at the terminus of Taylor Glacier today, where salts are being discharged onto Lake Bonney. As the lake evaporated or drained, gypsum began to precipitate on the surface of the carbonate plates.

An influx of gypsum into one basin of a two basin system, would explain why the thick, surface gypsum deposits are only seen in the eastern basin. Over the hummocky drift and in the western basin, where thick deposits of gypsum are absent but the calcite pseudomorphs still occur, it is likely that there was only sufficient calcium carbonate in the glacial lake to form the needles in the sediment, either

through a lack of mixing or because the two basins had already separated before the large quantities of gypsum were deposited into the lake.

The gypsum needles in the sediment at some stage after the calcite induration, were dissolved and replaced by calcium carbonate. This may have occurred either as the ice tongue receded, which allowed waters in the sediment to drain seaward and in so doing flushed the calcium sulphate from the carbonate, or after the lake drained, when meltwater from snow drifts mobilised the calcium sulphate and as the water evaporated the calcium sulphate then migrated upwards to finally precipitate with the gypsum at the surface. The former possibility may explain the presence of only calcite pseudomorphs in the western basin. After separation of the basins, meltwaters from the alpine glaciers (Adams and Miers Glaciers) and precipiation, could then have flowed through the western basin flushing out the sulphate. It is possible that these melt waters were enriched in calcium carbonate from contact with marbles which occur in the area. This may have aided the replacement of calcium sulphate by calcium carbonate. After flushing, only the loose pseudomorphs remained from the original deposit.

CHAPTER FIVE; CONCLUSIONS

5:1 ORIGIN OF THE DEBRIS CONES

Previous investigations into the origin of debris cones on lake ice usually concluded that the rock debris forming the cone originated from the lake bed. Bradley and Palmer (1967), proposed that the debris freezes to the base of the ice raft around the margins of the lake during winter. Compressive forces and ice diapairs then cause the raft to dome upwards, the debris eventually being thrust onto the ice surface. Hendy *et al* (1972), proposed that the gravel lenses froze to the bottom of the ice and then moved upwards through the ice by ablation at the top of the raft.

If either of the processes are correct it would be expected to find debris throughout the thickness of the ice raft. In this study the gravel debris was observed to occur only in the top three to four centimetres of the ice, below this the ice was particularly free of detritus, with the exception of the occasional sand grain. This places doubt on the two origins described above. Another observation which contradicts the models is a difference in textures and mineralogy between the lake sediments recovered from the lake bed and those observed at the surface. Those at the surface are much coarser and have a smaller proportion of basaltic material than those from the lake bed.

It is proposed that the rock debris on Lake Miers shares a common origin with the mounds of drift overlying lake sediments observed in eastern basin. These sediments are postulated to have originated from Ross Ice Tongue when it occupied the valley. Carried into the valley as supraglacial or englacial sediments, this material was then

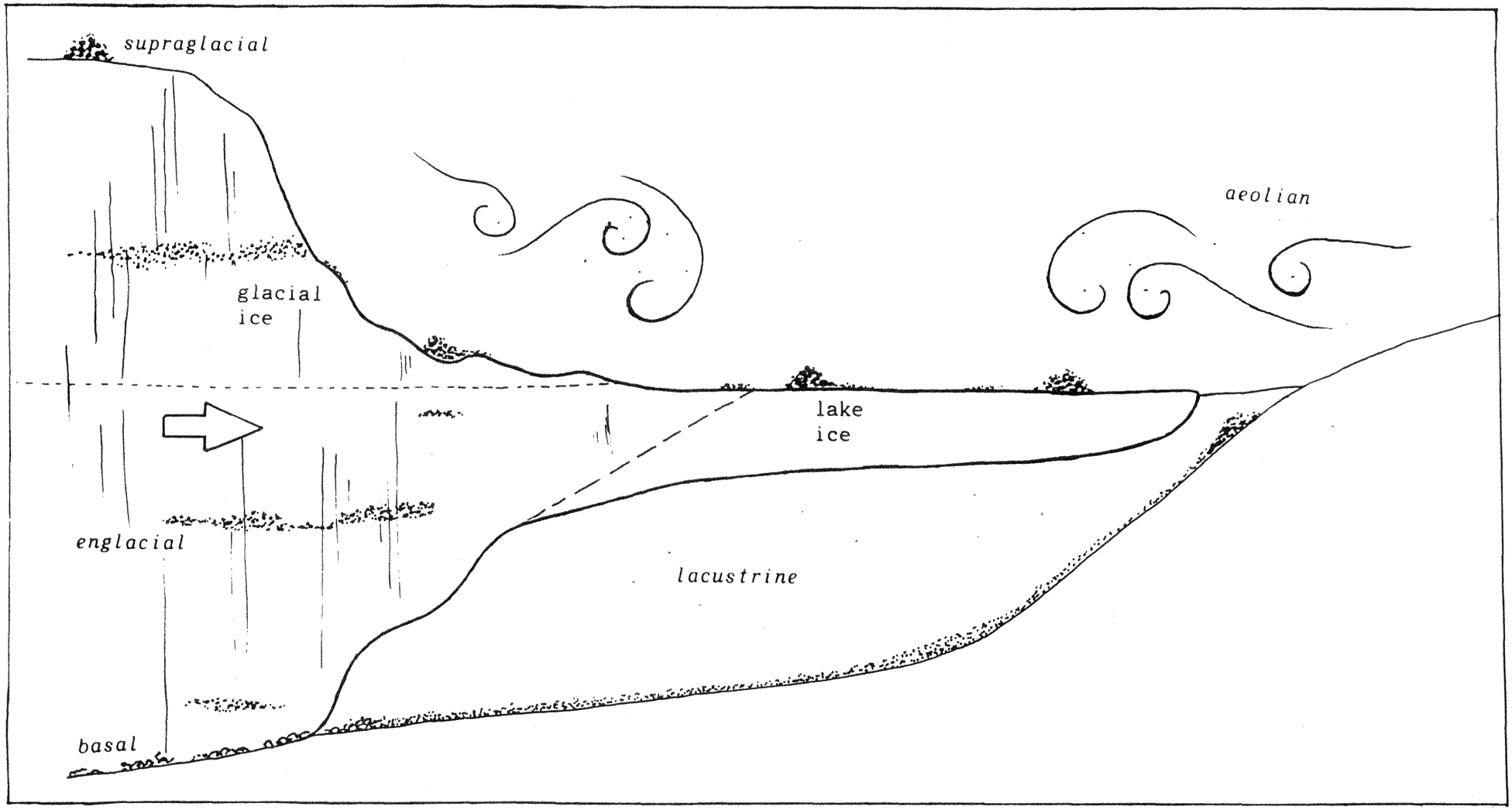


Figure 5.1 Possible sources of sediment into a proglacial lake. Supraglacial, englacial and aeolian material may be deposited onto the lake surface and rafted across the lake as the ice cover is pushed forward by ice fed from a glacier.

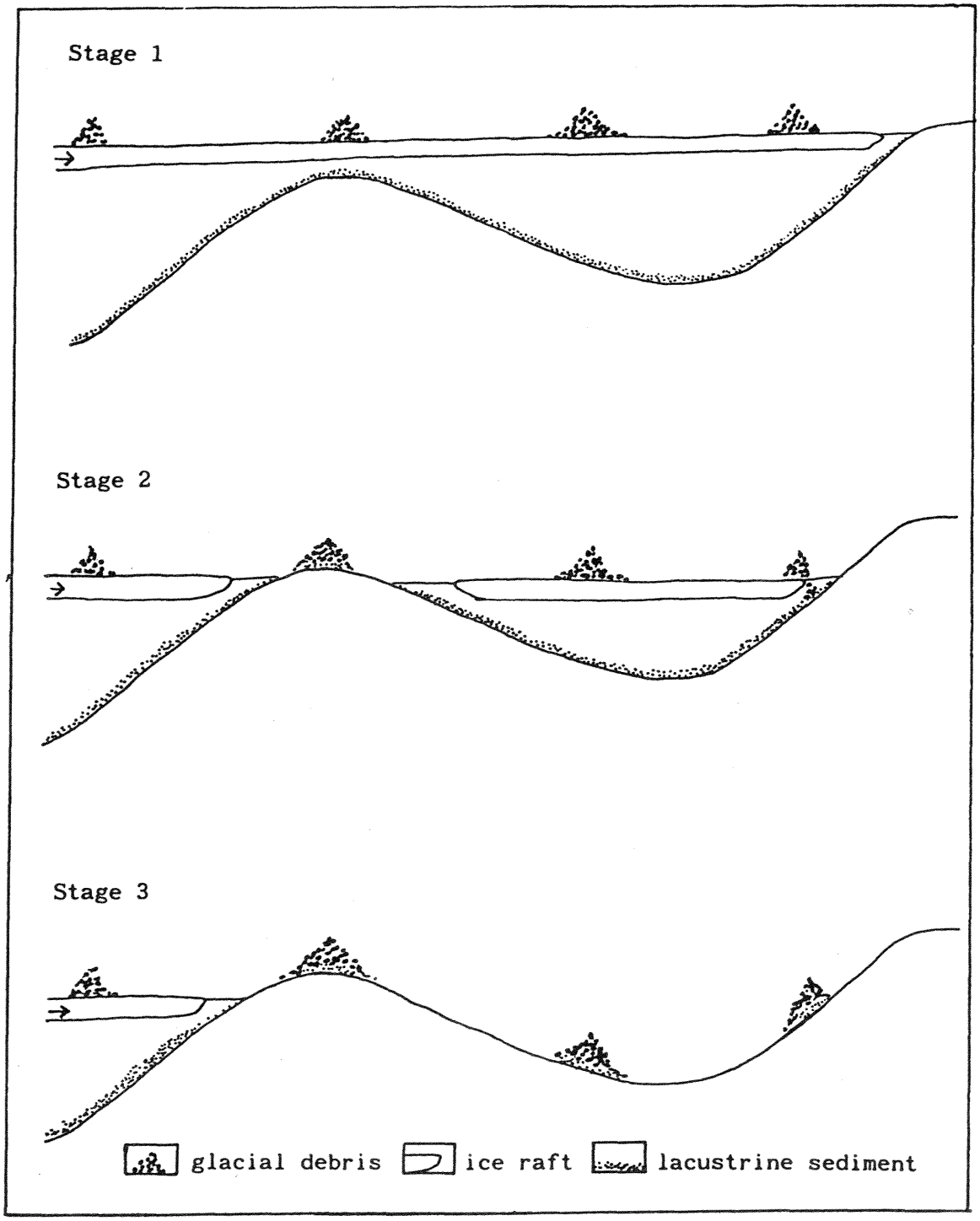


Figure 5.2 Proposed model for the origin of debris cones;
Stage 1: Glacial debris is rafted out across a proglacial lake.
Stage 2: A lowering of lake level causes debris to be isolated on the lake raft, or to be deposited onto the former lake bed.
Stage 3: Drainage of one basin, results in the glacial debris, formerly on the raft, being deposited over the lacustrine sediment, thereby preserving it. If a lake survives in the other basin, debris will be preserved on its surface.

deposited directly on to the floating ice cover of the proglacial lake (see fig 5.1). As the ice raft was pushed away from the ice front the debris was transported across the valley. The sediment remained on the ice surface until it reached the annual melt out moat, or the lake drained and the ice cover ablated away. The mounds in the eastern basin represent the latter case preserving the underlying sediments as the basin drained (see fig 5.2). The cones on Lake Miers were preserved by the ice cover in the western basin remaining intact as the eastern basin drained. The absence of an ice tongue to force the raft forward allowed the cones to be preserved on the ice cover (see fig 5.2).

An active example of such a mechanism exists at Trough Lake. Lyon (1979), observed a series of compressional ridges radiating out from Koettlitz Glacier over the lake surface. He suggested that the gravel on the sides of the ridges, and the large quantity of poorly sorted sands and gravels extending over the ice cover were derived from the glacier. It was noted that large boulders up to 5m in diameter occur amongst the debris. Field observations by Warwick Vincent and Clive Howard-Williams during the summer of 1985/86 (pers. com.) indicate that at least the largest of these boulders still lies on the lake surface in the same position as observed by Lyon (pers. com.). Sketch maps of the debris distribution on Miers and Trough Lakes were shown in figure 2.15). Both lakes show a relative absence of debris towards their centres.

5:2 RECONSTRUCTION OF THE HISTORY OF MIERS VALLEY DURING THE LAST MAJOR GLACIAL PEROIOD

At the peak of the last major glacial period, between 20,000 and 10,000 years BP, the Ross Sea was filled with grounded ice, which expanded and forced ice tongues up into the Dry Valleys. Meltwater from the landward side of these ice tongues ponded in front of the ice fronts forming proglacial lakes. Such an ice tongue extended into Miers Valley at least 23,000 years ago (C-14 age of sample WK:-609). At some stage an ice tongue has advanced far up into the valley, its pathway marked by terminal moraines which expand across the valley floor at intervals along the eastern basin. These moraines could be a feature of this glacial advance or possibly earlier glacial advances, the latter being subsequently covered with younger drift. As we cannot see the subsurface sedimentary record, it is not possible to ascertain which advance was in fact responsible.

The glacial drift in Miers Valley has an obvious basaltic component, which originated from the volcanic area within McMurdo Sound. Much of this drift was probably deposited directly onto the ice cover of the proglacial lake. The material then being rafted away from the glacial terminus as the ice tongue advanced.

At its expansive stages of occupation the primary precipitation within the water column of the glacial lake was calcite. Simultaneously with the calcite precipitation, smaller quantities of a salt, presumed to be gypsum, were crystallizing within the sediment. Changes in the water chemistry, possibly due to temperature, caused the calcite to be precipitated in pulses, of which there were at least three, giving C-14 ages of 20, 18 and 14 kA.

At its peak, the proglacial lake which formed Glacial Lake Trowbridge, occupied both the basins within the valley. Radiometric age of the mapped limits of the carbonate coverage indicate that the major expansion of Glacial Lake Trowbridge probably occurred between 18,000-19,000 years BP. The older carbonates cover a greater extent of the eastern basin compared to the younger carbonates, and it is likely that after about 18,000 years BP Glacial Lake Trowbridge began to recede. This is supported by the oxygen isotope data, which showed an enrichment of ^{18}O with time, suggesting a possible concentration of the heavier isotopes through evaporation. The youngest carbonates occurred in the vicinity of the 'gypsum moraine' and it is likely that this material was deposited at the closing stages of the ice tongue retreat.

Westward, the calcite plates thin and tend to be older. It is likely that the western basin was isolated from the eastern basin by 15,000 years BP. After Glacial Lake Trowbridge separated into the two basin system, a large influx of gypsum entered the eastern basin, possibly when a pocket of sulphate rich brine, from trapped sea water within the ice, was released into the lake. The large concentrations of calcium sulphate within the lake, caused the precipitation of gypsum during the final stages of Glacial Lake Trowbidge when the lake was concentrating through evaporation. Oxygen and sulphur isotopes (Lyon, 1978) showed that the sulphate within the gypsum was from sea water, however the isotopes of the water of hydration showed that the gypsum had precipitated from water which was more akin to that of Koettlitz Glacier, i.e. within the proglacial lake. This precipitation formed the thick layers of gypsum covering the carbonate plates throughout the eastern basin, and particularly, the 'gypsum moraine'.

The separation of the system into two basins also isolated Lake Miers from drift derived from the grounded ice sheet. Glacial debris

still remaining on the surface ice cover of Lake Miers was preserved, as the model for the origin of the debris cones suggests. The glacial debris on the ice cover of the eastern basin was eventually deposited on the lacustrine beds as the glacial lake receded.

At some stage after the calcite plates were sufficiently indurated to retain the shape of the enclosed salt crystals, the salt within the calcite was mobilised. If the salt was calcium sulphate, it was probably flushed from the carbonate by alpine glacial meltwaters flowing down valley. All the sulphate was flushed from the western basin, either draining out to the coast through Miers River, or adding to the sulphate deposits of the eastern basin. Westward, complete replacement of the salt crystals has occurred, with the calcite pseudomorphs being held together only by the remnents of the original plates, or just the calcite pseudomorphs remaining within weakly cemented silts, to the east the plates remain intact and moulds contain less calcite recrystallization.

When the lake evaporated and/or drained, the lacustrine deposits were exposed to the atmosphere. The combination of alpine glacial meltwater and the effect of snow accummulating during winter and melting during summer has caused weathering of the carbonate and gypsum. The surface plates showed pitting and weathering.

The present surface of the eastern basin contains large fans and subdued topography. The occupation by a large lake would have subdued the topography. The fans probably formed by downslope movement of material from the hummocky drift at the head of the basin, as the proglacial lake retreated. Miers River also passes through the drift annually as it drains Lake Miers to the coast. The summer action of the river has transported and redistributed material as it down cut through the loose sediment.

5:3 DRY VALLEY LAKES

The Dry Valley Region is ice free due to the fact that evaporation in these areas is greater than the precipitation. The central Antarctic continent has a very low precipitation rate. Nearer the coast precipitation increases, however this is negated by a drying effect on the coastal side of the Transantarctic Mountains produced by katabatic winds. Cold, dry winds sweep down the valleys from the East Antarctic Ice Sheet to the sea, drying the valleys as they proceed. Therefore snow and ice does not accumulate in the Dry Valleys as it does elsewhere on the Antarctic continent.

When precipitation and accumulation exceeds evaporation in one area (e.g. the East Antarctic Ice Sheet), surplus accumulated ice and water will extend into those areas of lower accumulation (e.g. the Dry Valleys),(Wilson, 1967). The alpine glaciers are believed to reflect the precipitation received by the East Antarctic Ice Sheet from the Ross Sea (Denton *et al*, 1970). When exposed sea is in close proximatey accumulation will occur. Such a situation exists at present with alpine glaciers such as the Miers and Ferrar flowing down into the valleys. If this accumulation slows or stops due to an advance of the West Antartctic Ice Sheet pushing the exposed sea further from the Transantarctic Mountains, these glaciers will retreat. The sedimentary and geomorphic record has shown that the alpine glaciers were in just such minimum positions 10,000 - 20,000 years BP (Stuiver *et al*, 1981).

The presence of lakes within these valleys is the result of the complex interaction within the valley system of the evaporation and precipitation balance and resulting glaciers, meltwater streams, and lakes. Lakes will form in enclosed basins when the input of water into the system is greater than the output, i.e. the addition of water from

precipitation, meltwater off glaciers and the outflow streams is greater than the net loss by evaporation and drainage (Wilson, 1967). Numerous factors contribute to this balance:- the annual precipitation of the area, the volume of meltwater produced in summer, the evaporation rates from the meltwater, and the sublimation rates from the ice. Temperature and humidity are important in this system as the vapour pressure of ice is temperature dependent and low relative humidities will induce evaporation and sublimation. The area of glacial ice exposed to summer heating and the duration of exposure (i.e. the number of cloudless days in summer) will influence the quantity of meltwater produced. The shape of the basin and hence of the lake is also important as lakes with large surface areas will have greater loss due to evaporation and sublimation. Deep narrow basins will contain lakes with smaller exposed surface areas and have reduced water loss due to evaporation and sublimation. The interaction and balance of these factors and others, determines the presence of a lake in a particular valley.

At the same time as Glacial Lake Trowbridge occupied Miers Valley other valleys also had proglacial lakes from in front of ice tongues from the Ross Sea (Péwé, 1960; Denton *et al*, 1970; Stuvier *et al*, 1981). The direction of glaciation has been traced by the deposition of volcanic erratics originating from McMurdo Sound. The largest glacial lake recognized in the Dry Valleys to date, is Glacial Lake Washburn which occupied Taylor Valley between 12,000 ^{14}C-yr BP and 21,000 ^{14}C-yr BP (Stuvier *et al*, 1981). Eskers covering the floor of Lower Taylor Valley indicate that Glacial Lake Washburn was fed by sub glacial flow beneath Ross Sea Ice (Stuvier *et al*, 1981). Fossil deltas occur at former lake levels where streams have flowed into the glacial lake, these deltas usually contain algae which can be radiometrically dated.

As well as Taylor and Miers Valleys, Péwé (1960) suggested that glacial lakes have occupied Garwood Valley and what is now named Salmon Valley. Deltas present in Salmon and Garwood Valleys have also been intepreted as indicating occupation by proglacial lakes (Stuvier *et al*, 1981).

Neighbouring Marshall Valley contains earlier lacustrine deposits and evaporites from previous periods of glaciation (Dagel *et al*, in press) but does not show evidence for major lake formation between 10,000 and 20,000 years BP. The presence of Ross Sea basalt rich drift indicates the advance of an ice tongue but no obvious lake deposits have been recorded. This does not necessarily indicate a lake did not form as lacustrine deposits may not have been preserved. There are fairly extensive gypsum deposits over the surface drift of Lower Marshall Valley which could be where gypsum was deposited from a small pond or lake at the lower portion of the valley. These deposits are weathered and show recrystallization and may have a similar origin to the gypsum deposits of Miers Valley.

Lake occupation in Miers Valley was obviously synchronous with other valleys in the Transantarctic Mountain Region during the late Wisconsin glacial maximum. Although Taylor Valley shows evidence of sub glacial flow into Glacial Lake Washburn, Miers Valley and others do not (Stuvier *et al*, 1981). The presence of these lakes must have therefore been a direct result in the changes in the evaporation/accumulation balance.

A large volume of ice such as the thickened grounded Ross Sea ice filling McMurdo Sound could possibly have effected the local climate. Apart from reducing the precipitation and accumulation in the Transantarctic Mountains, it is possible that the ice sheet at the mouths of the valleys deflected and changed the wind patterns, possibly reducing the katabatic winds. This would have had the effect

of raising the humidity and temperature within the valleys and hence altering the precipitation/evaporation balance in favour of accumulation. The ice sheet itself would present a large surface area to summer melting, increasing the meltwater input into the valleys. Thus, although the last glacial period may have produced overall colder and drier conditions, local factors enabled large glacial lakes to pond in front of ice fronts from grounded Ross Sea ice during the peak of the last glacial period.

APPENDIX I

As samples were collected in the field they were assigned a field reference number. On returning to the University of Waikato, the samples were renumbered with University reference numbers. The University of Waikato C-14 Laboratory and U/Th Laboratory also assign their own references. A summary of samples, their mineralogy, and altitudes (where known), and a check list of laboratory numbers follows. Copies of the surveyor's map of sample sites, can be obtained through the Chemistry Secretary, the University of Waikato, Hamilton.

Uni. No.	Field No.	Mineralogy	Altitude metres a.s.l.	C-14 Lab No.	U/Th Lab No.
C839	J12	carbonate/crystals			
C840	J13	crystals	141.5		
C841	J14	carbonate/crystals	130.4		
C842	J14b	thin carbonate	130.4		
C843	J15	carbonate	143.5		
C844	J15b	carbonate cement	143.5		
C845	J16	carbonate	128.4		
C846	J17	gypsum	122.9		
C847	J17a	crystals/carbonate moulds	122.9		
C848	J17b	thin carbonate	122.9		
C849	J18a	gypsum	119.7		
C850	J18b	carbonate moulds	119.7		
C851	J18c	thin carbonate	119.7		
C852	J19a	gypsum	120.5		
C853	J19b	carbonate/crystals	120.5		
C854	J19c	thin carbonate	120.5		
C855	J20	carbonate	130.9		
C856	J21a	gypsum	120.7		
C857	J21b	carbonate	120.7		
C858	J21c	sand (underlying gypsum)	120.7		
C859	J22	carbonate	111.0		
		MIERS RIVER SECTION			
C860	J23a	gypsum	81.0		
C861	J23b	carbonate	81.0	WK:-609	
C862	J23c	thin carbonate	81.0		
C863	J23d	sandy gravel (red colouration)	81.0		

Uni. No.	Field No.	Mineralogy	Altitude metres a.s.l.	C-14 Lab No.	U/Th Lab No.
C864	J23e	sandy gravel	81.0		
C865	J23f	gravelly sand (red colouration)	81.0		
C866	J23g	green layer	81.0		
C867	J23h	oxidised sand	81.0		
C868	J23i	thin carbonate	81.0		
C869	J23j	yellow silt	81.0		
C870	J23k	coarse sandy gravel	81.0		
		MIRABILITE SECTION			
C871	J24a	silt (above mirabilite)			
C872	J24b	silt (below mirabilite)			
C873	J24c	gravel below mirabilite			
C874	J24d	mirabilite			
C875	J25a	gypsum	124.0		
C876	J25b	carbonate	124.0		
C877	J26a	gypsum	101.1		
C878	J26b	carbonate	101.1	WK:-627	TX168
C879	J27	thenardite	77.7		
C880	J28	carbonate			
C881	J27	gypsum	79.9		
C882	J29b	carbonate	79.9		
C883	J30a	gypsum	79.8		
C884	J30b	carbonate	79.8		
C885	J30c	gypsum	79.8		
C886	J31	gypsum	92.4		
C887	J32a	gypsum	95.0		
C888	J32b	carbonate	95.0		
C889	J33a	gypsum	98.16		
C890	J33b	carbonate	98.16		

Uni. No.	Field No.	Mineralogy	Altitude metres a.s.l.	C-14 Lab No.	U/Th Lab No.
C891	J34a	gypsum	91.2		
C892	J34b	carbonate	91.2		
C893	J35	carbonate	120.8	WK:-718	TX191
C894	J36	carbonate	114.5		
C895	J37	carbonate	102.8		
C896	J38a	gypsum	95.0		
C897	J38b	carbonate	95.0		
C898	J39a	gypsum	88.1		
C899	J39b	carbonate	88.1	WK:-717	TX189
C900	J40a	gypsum	97.3		
C901	J40b	carbonate	97.3		
C902	J41a	gypsum	101.9		
C903	J41b	carbonate	101.9		
C904	J41c	yellow silt	101.9		
C905	J42a	gypsum	106.7		
C906	J42b	carbonate	106.7		
C907	J43a	gypsum	113.0		
C908	J43b	carbonate	113.0	WK:-812 WK:-813 WK:-814	TX220 TX227 TX226
C909	J44	carbonate	108.7		
C910	J45a	gypsum	108.3		
C911	J45b	carbonate	108.3		
C912	J46a	gypsum	114.3		
C913	J46b	carbonate	114.3		
C914	J47	carbonate	124.8		
C915	J48	gypsum/crystals	150.0		
C916	J49a	carbonate	140.8	WK:-628	
C917	J49b	thin carbonate	140.8		
C918	J50	carbonate	150.0		

Uni. No.	Field No.	Mineralogy	Altitude metres a.s.l.	C-14 Lab No.	U/Th Lab No.
C919	J51a	gypsum	124.8		
C920	J51b	carbonate/crystals	124.8		
C921	J51c	thin carbonate	124.8	WK:-630	
C922	J52a	gypsum	77.5		
C923	J52b	carbonate	77.5	WK:-716	
C924	J52c	gypsum/carbonate	77.5		
C925	J53a(13/12)	gypsum	79.3		
C926	J53b(13/12)	carbonate	79.3		
C927	J53c(13/12)	silty sand	79.3		
C928	J54a(13/12)	gypsum	80.5		
C929	J54b(13/12)	carbonate	80.5		
C930	J55(13/12)	carbonate	135.4		
C931	J56(13/12)	carbonate	156.1	WK:-720	TX195
C932	J53(14/12)	carbonate	143.3		
C933	J54a(14/12)	carbonate	141.7		
C934	J54b(14/12)	algae	141.7		
C935	J54c(14/12)	carbonate	141.7		
C936	J55a(14/12)	carbonate	119.8		
C937	J55b(14/12)	thin carbonate	119.8		
C938	J56b(14/12)	carbonate	80.0	WK:719	TX194
C939	J56a(14/12)	gypsum			
C940	J57(14/12)	gypsum			
C941	J58a	carbonate	100.0?	WK:629	
C942	J58b	silt	100.0?		
C943	83/32	carbonate	142.0	WK:-528	

Field No./ Benchmark	Location	Altitude metres a.s.l.
Sites surveyed but not sampled:		
JA	Uppermost appearance of carbonate plate	134.1
JB	Stream bed (transect 3)	104.9
JC	High delta on hummocky drift.	153.7
New Zealand Lands and Survey Dept. Benchmarks in Miers Valley. Surveyed by P.D. Tinnelly 1981/82.		
BMM 21	In large rock on spit just before the junction of the two streams 300 metres from lake.	164.029
BMM 22	In large rock on true right of Adams Stream 100 metres from stream 200 metres from glacier terminal.	242.595
BMM 23	In large rock on moraine between Adams and Miers Glaciers.	287.854
BMM 24	In large rock on slope true left of Melt Stream from Miers Glacier 200 metres from stream and 100 metres from terminal wall.	188.873
BMM 25	In boulder 50 metres above the lake edge and 300 metres down on the true left from the top end of the lake.	159.272
BMM 26	On edge of large boulder 2 metres x1 metre situated halfway along the lake edge and 1 metre from waters edge.	154.879
BMM 27	In rock 200 metres east from lake outlet on slope 50 metres from stream.	158.004
BMM 28	Iron tube on the south east side of rounded hill which is south of highest immediate hill.	175.70
BMM 29	In boulder on true left of stream 6 metres above stream at the bottom of the stream gully.	116.65
BMM 30	In light coloured granite boulder 960 metres downstream from BMM 29 on highest part of hill.	107.33

Benchmark	Location	Altitude metres a.s.l.
BMM 31	In granite boulder on small hill on plains 150 metres from stream bed. An old rock Cairn 0.8 metres high is situated 100 metres to the north of this bench mark.	98.16
BMM 32	In light colour granite boulder on top of black moraine hill in centre of valley where the valley begins to widen out.	87.30
BMM 33	In boulder on highest hill at eastern end of large gully flowing into deep gorge on Miers Stream.	78.450
BMM 34	In granite rock 20 metres from stream bed on a flat ledge 900 metres downstream from BMM 33.	32.031
BMM 35	In large rock on ledge 300 metres downstream on the right of the stream.	

Additional benchmarks surveyed by D. Manson 1983.

BMM 29A	Refer sketchmap (Fig. 2.2)	140.4
BMM 31A	" "	96.53

APPENDIX II

SAMPLE NOS.		$\delta^{18}O_{PDB}$	$\delta^{13}C_{PDB}$	SAMPLE NOS.		$\delta^{18}O_{PDB}$	$\delta^{13}C_{PDB}$
C839	J12	-26.46	4.36	C892	J34b	-30.57	3.61
C841	J14	-30.40	2.75	C893	J35	-27.95	3.96
C842	J14b	-29.47	3.02	C894	J36	-30.13	1.78
C843	J15	-31.77	2.88	C895	J37	-22.56	5.42
C844	J15b	-32.96	2.60	C897	J38b	-19.91	5.67
C845	J16	-32.86	3.11	C899	J39b	-26.19	4.77
C847	J17a	-33.34	2.54	C901	J40b	-23.70	4.77
C848	J17b	-32.73	2.59	C903	J41b	-29.21	4.29
C849	J18a	-31.24	3.02	C906	J42b	-31.14	4.17
C850	J18b	-27.16	3.38	C908	J43b	-28.43	4.11
C851	J18c	-24.05	3.94	C909	J44	-28.24	3.94
C853	J19b	-30.56	3.20	C911	J45b	-27.05	4.19
C854	J19c	-31.97	2.21	C913	J46b	-23.00	4.75
C855	J20	-31.74	3.42	C914	J47	-29.63	3.29
C857	J21b	-18.79	5.60	C915	J48	-26.86	3.67
C859	J22	-29.39	3.59	C916	J49a	-31.28	3.10
C862	J23c	-33.51	3.13	C917	J49b	-30.33	2.21
C861	J23b	-30.48	4.36	C918	J50	-32.01	2.85
C868	J23i	-30.66	0.82	C920	J51b	-33.14	2.29
C876	J25b	-26.43	4.23	C921	J51c	-30.84	2.46
C878	J26b	-23.41	5.26	C923	J52b	-30.27	3.06
C880	J28	-30.44	2.91	C926	J53b	-33.30	3.14
C882	J29b	-29.70	3.65	C929	J54b	-31.99	3.28
C884	J30b	-28.27	3.70	C930	J55	-29.81	3.35
C888	J32b	-31.30	4.01	C931	J56	-30.27	2.63
C890	J33b	-26.60	4.65	C932	J53	-32.02	2.80

SAMPLE NOS.		$\delta^{18}O_{PDB}$	$\delta^{13}C_{PDB}$	SAMPLE NOS.		$\delta^{18}O_{PDB}$	$\delta^{13}C_{PDB}$
C933	J54a	-30.97	3.66	CORE 1; LAKE MIERS			
C935	J54c	-32.81	3.22	43-44	cm	-25.33	2.93
C936	J55a	-32.22	3.18	52-61	cm	-23.23	3.28
C337	J55b	-32.01	2.11	61-65	cm	-23.91	3.53
C938	J56b	-25.78	4.53	65-71	cm	-25.34	3.06
C941	J58a	-29.87	4.09	71-79	cm	-22.77	2.72
C943	83/32	-30.88	3.61				
SAMPLES COLLECTED 1982							
C722	82/82	-31.37	5.78	SAMPLES COLLECTED BY DENTON			
C771	82/131	-30.37	2.02	C403	MC3	-29.99	3.78
C772	82/132	-28.51	2.96	C403	TX23	-31.79	3.55
C773	82/130	-30.77	3.40	C403b	MC3	-30.61	3.80
C774	82/134	-32.65	2.95	C403	MC3	-29.76	3.91
C775	82/135	-24.53	3.43	C403c	MC3	-30.68	3.71
C775	82/135	-28.10	3.51	C403	TX26	-30.67	3.47
C777	82/137	-25.86	4.88				
C778	82/138	-32.44	3.40				
C776	82/136	-25.76	3.21				
C780	82/140	-29.38	4.94				
C782	82/142	-31.63	4.04				

APPENDIX III

A sample record/calculation sheet for U/Th samples. The sample recorded on this sheet is the sample shown in figures 3.4, 3.5, and 3.6.

UNIVERSITY OF WAIKATO

URANIUM/THORIUM SAMPLES

Sample No. TX 189

Description C899 (WK 717)

Weight Taken 61.2

Ml Spike 10

Weight Insoluble Residue ~~5.16~~

Date Plated U ~~18/6/85~~

Net Weight Dissolved 91.6%

Th 18/6/85

Dissolution Method

URANIUM Counter — Started ~~11.09/18/6~~ Stopped ~~16.34/24/6~~ Count Time ~~8965~~

	Count	S.D.	Count Rate	S.D.	-BKGD	-Tail	Net	S.D.
U^{238}	17458	132	1.947	0.015	.002	0.107	1.838	0.016
U^{234}	46598	216	5.198	0.024	0.004	0.013	3.350	0.024
U^{232}	3994	63	0.446	0.007	0.009	0.001	4.36	0.007

$\frac{U^{234}}{U^{238}}$ 1.82 S.D. 0.02 $\frac{U^{234}}{U^{232}}$ 7.726 S.D. 0.13

U concentration 17.8 ppm U efficiency ____ cpm/ml

THORIUM Counter — Started 11.09/18/6 Stopped 16.34/24/6 Count Time 8965

	Count	S.D.	Count Rate	S.D.	-BKGD	-Tail	-Ra	Net	S.D.
Th^{232}	1839	43	0.205	0.005	0.001	0.045		0.159	0.006
Th^{230}	11679	108	1.303	0.012	0.002	0.040		1.261	0.012
Th^{228}	12279	111	1.370	0.012	0.002	0.054	0.036	1.278	0.012
Ra^{224}	6667	82	0.744	0.009	0.089	0.023		0.662	0.009

$\frac{Th^{232}}{Th^{230}}$ ____ S.D. ____ $\frac{Th^{230}}{Th^{228}-Th^{232}}$ 1.127 S.D. 0.016 $\frac{Th^{230}-Th^{232}}{Th^{228}-Th^{232}}$ ____ S.D. ____

Th concentration ____ ppm Th efficiency ____ cpm/ml

$\frac{Th^{230}}{U^{234}}$ 0.146 S.D. 0.003 $\frac{Th^{230}-Th^{232}}{U^{234}}$ ____ S.D. ____

Maximum Age 16.97 S.D. 0.36 years x 10^3 Th^{232} Corrected Age ____ S.D. ____

$\frac{U^{234}}{U^{238}}$ age ____ S.D. ____ (Marine samples only)

Comments:

MIERS VALLEY U/Th DATE LIST (1984-1985)

Sample No.	Collector	Orig. No.	Lab Ref	Location	Type	U ppm	$^{234}U/^{238}U$	σ	$^{230}Th/^{234}Th$	σ	$^{232}Th/^{234}U$	σ	Age x 1000yr	σ
C878	Clayton-Greene	J26	TX168	101m.a.s.l.	95% $CaCO_3$	24.9	2.78	0.02	0.118	0.004	0.004	0.0015	13.0	0.4
C931	"	J56 13/12	TX169	156m.a.s.l.	84% $CaCO_3$	11.2	2.63	0.01	0.119	0.002	0.025	0.0006	13.3	0.2
C943	Hendy	83/82	TX203	142m.a.s.l.	89% $CaCO_3$									
C861	Clayton-Greene	J23b	TX198	81m.a.s.l.	83% $CaCO_3$									
C941	"	J58a	TX197	100m.a.s.l.	94% $CaCO_3$									
C862	"	J23c	TX196	81m.a.s.l.	79% $CaCO_3$	9.5	2.85	0.14	0.092	0.005	0.040	0.023	10.5	0.5
C931	"	J56 13/12	TX195	156.1m.a.s.l.	86% $CaCO_3$	11.0	3.10	0.30	0.116	0.012	0.024	0.02	13.3	1.4
C923	"	J52b	TX202	77.5m.a.s.l.	92% $CaCO_3$									
C899	"	J39	TX189	88m.a.s.l.	92% $CaCO_3$	17.8	1.82	0.02	0.142	0.003	0.018	0.002	17.0	0.4
C893	"	J35	TX191	121m.a.s.l.	84% $CaCO_3$	15.2	1.84	0.03	0.120	0.002	0.017	0.003	15.2	0.25
C938	"	J56 14/12	TX194	≈80ma.s.l.	88% $CaCO_3$	10.8	2.78	0.04	0.096	0.004	0.012	0.001	11.2	0.40
C941	"	J58a Low.	TX207	100m.a.s.l.	94% $CaCO_3$									
"	"	J58a Midd.	TX210	"	97% $CaCO_3$									

Sample No.	Collector	Orig. No.	Lab Ref	Location	Type	U ppm	$^{234}U/^{238}U$	σ	$^{230}Th/^{234}U$	σ	$^{232}Th/^{234}U$	σ	Age x 1000yr	σ
C941	Clayton-Greene	J58a Upp.	TX212	100m.a.s.l.										
C908	"	J43b Upp.	TX220	113m.a.s.l.	93% $CaCO_3$	18.1	2.99	0.03	0.167	0.004	0.014	0.001	15.3	0.5
"	"	J43b Midd.	TX227	"	93% $CaCO_3$	18.5	2.93	0.02	0.121	0.003	0.010	0.001	13.7	0.3
"	"	J43b Low	TX226	"	91% $CaCO_3$	19.2	2.92	0.04	0.137	0.004	0.008	0.003	15.7	0.5

MIERS VALLEY U/Th DATE LIST (pre-1984)

Sample No.	Collector	Orig. No.	Lab Ref	Location	Type	U ppm	234U/238U	σ	230Th/234U	σ	232Th/234U	σ	Age x 1000yr	σ
C408a	Denton		TX23	430'a.s.l.	$CaCO_3$	17.7	2.91	0.04	0.119	0.006	0.011	0.002	13.2	0.7
C403a	"		TX26	"	"	17.7	2.91	0.05	0.095	0.0037	0.007	0.001	10.7	0.4
C403a	"		TX31	"	"	8.1	2.92	0.06	0.096	0.005	0.011	0.002	10.7	0.6
C403b	"		TX36	"	"	14.4	4.20	0.12	0.0702	0.0031	0.010	0.001	7.8	0.4
C403c	"		TX37	"	"	18.9	2.92	0.03	0.0976	0.0032	0.008		10.8	0.3
C294	T.R. Healy	76-26		Koettlitz above L. Miers in front of snout	78% $CaCO_3$	29.9	1.06	0.01	1.166	0.022	0.038	0.003	Excess Th	
C295	"	76-27		191m.a.l.l., N. side	82% $CaCO_3$	18.9	2.85	0.022	0.1074	0.0034	0.007	0.001	11.9	0.38
C296	"	76-28		150m.a.l.l., N. side	83% $CaCO_3$	19.6	2.93	0.022	0.086	0.003			9.5	0.3
C298	"	76-30		174m.a.l.l., S side, W. end of old lake bed	92% $CaCO_3$	24.5	1.80	0.02	1.91	0.04	0.055	0.005	Excess Th	
C299	"	76-31		gypsum from G.G.M.	gypsum	6.0	2.42	0.19	0.627	0.043	0.154	0.020	120.	4
C300	"	76-32		thenardite from moraine complex E. end	99% Na_2CO_3	0.11	0.90	0.16	0.13	0.026	0.177	0.015	15.	4

Sample No.	Collector	Orig. No.	Lab Ref.	Location	Type	U ppm	234U/238U	σ	230Th/238U	σ	232Th/234U	σ	Age x 1000yr	σ
C301	Healy	76-33		lag 40-50m. a.l.l. snout Miers Stream	80%$CaCO_3$	0.03	1.29	0.16	0.672	0.035	0.029	0.012	115	12
C302	"	76-34		from delta E. Miers beneath tuff bed		4.0	1.01	0.03	1.16	0.06	0.0571	0.026	∞	
C303	"	76-35		most easterly gypsum	gypsum	0.25	1.88	0.15	0.900	0.052	0.172	0.008	185	20
C304	"	76-36		near gypsum moraine	68% $CaCO_3$	26.6	1.90	0.017	0.431	0.016	0.021	0.003	58	3.3
305	"	76-37		174m.a.l.l. S. side W. end of old lake bed	gypsum		2.52	0.06	0.131	0.009	0.085	0.004	5-15	

APPENDIX IV

UNIVERSITY OF WAIKATO RADIOCARBON DATING LABORATORY

	SAMPLE	^{14}C Lab. Code & No.	Altitude (m)	Composition of sample	$\delta^{13}C$ °/oo w.r.t. P.D.B.	$D^{14}C$ °/oo w.r.t. P.D.B.	Conventional Age B.P.	Dilution Sample % in Counting vial
	83-32	WK-582	141	Lacustrine Limestone	+3.29	-863.4±2.3	16000±140	91.4
	C861	WK-609	83	" "	+3.88	-942.5±2.5	22950±360	76.2
	C878	WK-627	101	" "	+4.36	-829.4±2.4	14200±120	100.0
	C916	WK-628	141	" "	+3.20	-901.4±2.0	18600±170	100.0
	C941	WK-629	130	" "	+4.36	-893.0±2.0	17950±160	100.0
	C921	WK-630	125	" "	+2.20	-876.0±3.8	16750±260	49.3
*	C862	C862	82	" "	+3.55±2	-961.6±5.1	26200±1150	30.2

* This sample was very seriously diluted with a count rate similar to a 40,000 yr old sample.
There is therefore a considerable possibility that it has an infinite age and the date must therefore not be published.
(Thesis okay though use your sample code number not a WK number)

UNIVERSITY OF WAIKATO RADIOCARBON DATING LABORATORY

Your sample has had its Radiocarbon Age determined. The details and significance are as you supplied them. **Please ammend the significance details if necessary,** and return a copy of this form to the University of Waikato Radiocarbon Dating Laboratory.

WK-: 812 C908 Lower

Conv Age (SE): 19850 ± 220
True Age (SE): 20400 ± 220
D¹³C: +3.2%

Nature of sample: lacustrine carbonate
Collected by: J.Clayton-Green 1983
Submitted by: C.H.Hendy
Chemistry Dept, Univ.of Waikato,Hamilton,NZ
Geographic location: Miers Valley, Antartica; c.156m asl
Lat.& Long.or G.R.: 78°07′S; 164°06′E
Comments:
Sig: (JC) Dating of a former proglacial lake supported by the Ross Sea I ice shelf

WK-: 813 C908 Middle

Conv Age (SE): 18100 ± 200
True Age (SE): 18650 ± 200
D¹³C: +4.0%

Nature of sample: lacustrine carbonate
Collected by: J.Clayton-Green 1983
Submitted by: C.H.Hendy
Chemistry Dept, Univ.of Waikato,Hamilton,NZ
Geographic location: Miers Valley, Antartica; c.156m asl
Lat.& Long.or G.R.: 78°07′S; 164°06′E
Comments:
Sig: (JC) Dating of a former proglacial lake supported by the Ross Sea I ice shelf

WK-: 814 C908 Upper

Conv Age (SE): 14400 ± 130
True Age (SE): 14800 ± 130
D¹³C: +4.2%

Nature of sample: Lacustrine carbonate
Collected by: J.Clayton-Green 1983
Submitted by: C.H.Hendy
Chemistry Dept, Univ.of Waikato,Hamilton,NZ
Geographic location: Miers Valley, Antartica; c.156m asl
Lat.& Long.or G.R.: 78°07′S; 164°06′E
Comments:
Sig: (JC) Dating of a former proglacial lake supported by the Ross Sea I ice shelf

WK-: 716 C923

Conv Age (SE): 19150 (180)
True Age (SE): 19700 (190)
d13C: 2.96

Nature of sample: lacustrine carbonate
Collected by: J.Clayton-Green 1983
Submitted by: C.H.Hendy
Chemistry Dept., Univ.of Waikato,Hamilton,NZ
Geographic location: Miers Valley, Antartica; 77.5m asl
Lat.& Long.or G.R.: 78°07′S; 164°06′E
Sig: (JC) Dating of a former proglacial lake supported by the Ross Sea I ice sheet.

Wk-: 717 C899

Conv Age (SE): 14200 (120)
True Age (SE): 14600 (120)
d13C: 4.37

Nature of sample: lacustrine carbonate
Collected by: J.Clayton-Green 1983
Submitted by: C.H.Hendy
Chemistry Dept., Univ.of Waikato,Hamilton,NZ
Geographic location: Miers Valley, Antartica ; 88m asl
Lat.& Long.or G.R.: 78°07'S; 164°01'E
Sig: (JC) Dating of a former proglacial lake supported by the Ross Sea I ice sheet.

Wk-: 718 C893

Conv Age (SE): 10300 (90)
True Age (SE): 10600 (90)
d13C: 4.66

Nature of sample: lacustrine carbonate
Collected by: J.Clayton-Green 1983
Submitted by: C.H.Hendy
Chemistry Dept., Univ.of Waikato,Hamilton,NZ
Geographic location: Miers Valley, Antartica ; 121m asl
Lat.& Long.or G.R.: 78°07'S; 164°01'E
Sig: (JC) Dating of a former proglacial lake supported by the Ross Sea I ice sheet.

Wk-: 719 C938

Conv Age (SE): 14800 (170)
True Age (SE): 15250 (170)
d13C: 2.87

Nature of sample: lacustrine carbonate
Collected by: J.Clayton-Green 1983
Submitted by: C.H.Hendy
Chemistry Dept, Univ.of Waikato,Hamilton,NZ
Geographic location: Miers Valley, Antartica; c.80m asl
Lat.& Long.or G.R.: 78°07'S; 164°06'E
Comments: Diluted, 63% s.
Sig: (JC) Dating of a former proglacial lake supported by the Ross Sea I ice shelf.

Wk-: 720 C931

Conv Age (SE): 13700 (180)
True Age (SE): 14100 (180)
d13C: 2.90

Nature of sample: lacustrine carbonate
Collected by: J.Clayton-Green 1983
Submitted by: C.H.Hendy
Chemistry Dept, Univ.of Waikato,Hamilton,NZ
Geographic location: Miers Valley, Antartica; c.156m asl
Lat.& Long.or G.R.: 78°07'S; 164°06'E
Comments: Diluted, 56% s.
Sig: (JC) Dating of a former proglacial lake supported by the Ross Sea I ice shelf.

Notes:

1. "Conventional Age" and ^{14}C Depletion are calculated using; the Libby half-life (5568 years); isotope fractionation correction (expressed in ‰ wrt PDB); NBS oxalic acid (*0.95) as "modern" with AD 1950 as reference year; and assumption of constancy in atmospheric radiocarbon levels. This radiocarbon age is normally selected for publication, and is always accompanied by the appropriate error term and Wk number.
2. Quoted errors are based upon counting statistics alone and are equivalent to ± 1 standard deviation.
3. "True Age" (where applicable) is based upon the more accurate half-life of 5730 years.

For further information, please contact the Director (address below).

Dr.A.G.Hogg,
Radiocarbon Dating Laboratory,
University of Waikato,
Hamilton,
NEW ZEALAND.

APPENDIX V

PETROGRAPHIC ANALYSIS OF CARBONATES

SAMPLE NO. C843V (J15) ALTITUDE 143.5m.a.s.l.

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	0.5-1%		large crystals
Calcite	97%		fine grained matrix
Org. Matter	0.5%		
Glass shards	2.5%		smll-med.
Rock fragments Types:	0.5%		
detrital calcite		subrounded	small (brown rims)
augite		angular	small

NOTES
Large euhedral crystals of calcite infilling cavities

SAMPLE NO. C843V (J15) ALTITUDE 143.5m.a.s.l.

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
Calcite	97%		fine grained matrix
Org. Matter	0.5%		
Glass shards	2.5%		
Rock fragments Types:	0.5-1%		
Quartz		angular	smll-med
detrital calcite		subangular	smll-med
hypersthene		"	smll

NOTES

PETROGRAPHIC ANALYSIS OF CARBONATES

SAMPLE NO. C861B V ALTITUDE 83m.a.s.l.

Steam section (J23b)

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	1%		eudedral crysrals
Calcite	80-85%		fine grained matrix
Org. Matter	3-5%		
Glass shards	10-12%	irreg. shapes	v.smll-large
Rock fragments	2-5%		
Types:			
biotite		subangular-	smll-med.
quartz		subrounded	"
felspars		"	"
(plagioclase)		angular	"
augite		subangular-	"
hypersthene		subrounded	"
olivine		prismatic	large
chlorite		subangular	

NOTES
augite shows weathering rim

SAMPLE NO. C861D H ALTITUDE 83 m.a.s.l.

stream section (J23b)

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
Calcite	80-85%		fine grained matrix
Org. Matter	1%		
Glass shards	3%		
Rock fragments	7%		
Types:			
quartz		angular-	medium
felspar		subrounded	"
plag.(labradorite,		"	"
andesine)		"	"
biotite		mainly subang.	"
muscovite		"	"
hypersthene		"	"
olivine		angular-subang	lrge-med
detrital calcite		flakes	

NOTES
V. smll calcite crystals seen in cavities
phenocrysts seen in glass shards

PETROGRAPHIC ANALYSIS OF CARBONATES

SAMPLE NO. C861C H ALTITUDE 83 m.a.s.l.

stream section J23b

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
Calcite	90%		fine grained matrix
Org. Matter	2%		
Glass shards	5%	irregular	smll-med.
Rock fragments	3-4%		
Types:			
detrital calcite		subrounded	med.-large
quartz		subangular	smll.-med
felspar		"	"
biotite		"	"
epidote		"	med.
garnet		"	smll.-med
hypersthene		"	med.
olivene		"	smll-med

NOTES
Bubbly section

SAMPLE NO. C861A V ALTITUDE 83 m.a.s.l.

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
Calcite	90%		fine grained matrix
Org. Matter	1%		
Glass shards	4%	irregular	
Rock fragments	5%		
Types:			
chlorite		subangular	v.smll-med
quartz		"	"
felspar		"	"
detrital calcite		"	"
hypersthene		"	"
augite		"	"

NOTES
Poor section

PETROGRAPHIC ANALYSIS OF CARBONATES

SAMPLE NO. C861 V J23b — ALTITUDE 83 m.a.s.l.

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
Calcite	85-90%		fine grained matrix
Org. Matter	3%		
Glass shards	7%	irregular	v.lrge-v.smll
Rock fragments	4%		
Types:			
augite		angular-	v.smll-smll
quartz		subrounded	"
felspar		"	"
plagioclase		"	"
orthoclase		"	"
biotite		"	"
hypersthene		"	"
garnet		"	"
detrital calcite		"	"
NOTES epidote		subrounded	medium

long viens (unident.) colourless-palebrown mod. relief Upp. 1st order, straight extinction

---

SAMPLE NO. C861 V J23b — ALTITUDE 83m.a.s.l.

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
Calcite	90-95%		fine grained matrix
Org. Matter	1%		
Glass shards	4%	irregular	smll-large
Rock fragments	3-5%		
Types:			
quartz		subangular-	smll-med
felspar(plag.)		subrounded	"
augite		"	"
biotite		"	"
detrital calcite		subangular m	med.

NOTES

PETROGRAPHIC ANALYSIS OF CARBONATES

SAMPLE NO. C899V ALTITUDE 88.1 m.a.s.l.

J39b

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	1%		euhedral crystals
Calcite	90%		fine grained matrix
Org. Matter	2%		
Glass shards	5%	rounded	sml1-med
Rock fragments	2%		
Types:			
quartz		subangular	v.sml1-sml1
felspar		"	"
biotite		"	"
augite		"	"
olivine		"	"
detrital calcite		angular flake	med

NOTES
large euhedral calcite infilling cavitites
augite shows altered rims

SAMPLE NO. C899H ALTITUDE 88.1 m.a.s.l.

J39b

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	1%		lrge euhedral
Calcite	85-89%		fine grained matrix
Org. Matter	2%		
Glass shards	8%	irregular	sml1-med
Rock fragments	3-5%		
Types:			
aegerine augite or aegerine		subangular	sml1-med
plagioclase		"	"
quartz		"	"

NOTES

PETROGRAPHIC ANALYSIS OF CARBONATES

SAMPLE NO. C909H ALTITUDE 108.7 m.a.s.l.

J44

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	2%		lrge euhedral
Calcite	95%		fine grained matrix
Org. Matter	0.5%		
Glass shards	1.5%	irregular	v.smll-smll
Rock fragments	2%		
Types:			
hypersthene		subangular-	v.smll-smll
detrital calcite		subrounded	"
biotite		"	"
quartz		"	"
felspar		"	"
augite		"	"

NOTES
section too thin

SAMPLE NO. C909V ALTITUDE 108.7 m.a.s.l.

J44

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	3%		lrge crystals
Calcite	90-95%		fine grained matrix
Org. Matter	1%		
Glass shards	2%	irregular-rounded	v.smll-smll
Rock fragments	1%		
Types:			
plagioclase		subangular-	v.smll-smll
quartz		subrounded	"
biotite		"	"
detrital calcite		"	"
olivine		"	"

NOTES

PETROGRAPHIC ANALYSIS OF CARBONATES

SAMPLE NO. C916 ALTITUDE 140.8 m.a.s.l.

J49a

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	1-3%		lrge crystals
Calcite	90-95%		fine grained matrix
Org. Matter	1%		
Glass shards	4%	irregular	sml1-med
Rock fragments	1%		
Types:			
biotite		subrounded	v.sml1-sml1
plagioclase		"	"
quartz		"	"
felspars		"	"
detrital calcite		"	"

NOTES

SAMPLE NO. C941V ALTITUDE between 88.1 & 107.3

J58a (summary from 2 slides m.a.s.l.

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	1-2%		lrge crystals
Calcite	88-90%		fine grained matrix
Org. Matter	1.5%		
Glass shards	6%	irregular	sml1-med
Rock fragments	2-3%		
Types:			
quartz		subrounded	sml1-med
felspar		"	"
biotite		"	"
plagioclase		"	"
augite		"	"

NOTES

PETROGRAPHIC ANALYSIS OF CARBONATES

SAMPLE NO. C941H — ALTITUDE between 88.1 & 107.33

J58a — m.a.s.l.

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	2%		lrge crystals*
Calcite	95%		fine grained matrix
Org. Matter	0.5%		
Glass shards	2%	irregular-rounded	sml1-med
Rock fragments	2%		
Types:			
detrital calcite		subang. flakes	v.sml1-sml1
biotite		thin 'needles'	"
quartz		subangular-	"
plagioclase		angular	"
augite		"	"

NOTES
* irregularly infilling cavities

SAMPLE NO. C943V — ALTITUDE 141 m.a.s.l.

83/32

	PERCENTAGE	ROUNDNESS	APPARENT GRAIN SIZE
	1%		lrge crystals
Calcite	90%		fine grained matrix
Org. Matter	1.5%		
Glass shards	5-6%	irregular	sml1-med
Rock fragments	3%		
Types:			
quartz		subangular-	sml1-med
felspar		subrounded	"
olivine		"	"
augite		"	"
biotite		flake	"

NOTES

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