Frontispiece; Drilling in the Marshall Valley 1986.

THE CHRONOLOGY OF THE ROSS SEA II GLACIATION, AN ANTARCTIC GLACIATION OF ILLINOIAN AGE.

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Abstract

A meltwater stream in the Marshall Valley, southern Victoria Land, Antarctica has revealed a stratigraphy of glacial drift and lacustrine lake beds.

The sequence has been differentiated by U/Th dating into the Ross Sea I Glaciation (15-30,000 years B.P.), represented by a gypsum bed and the Ross Sea II(?) Glaciation (130-180,000 years B.P.), represented by three calcium carbonate and gypsum beds.

The three Ross Sea II(?) lake beds may be differentiated from one another by an increasing calcite content with age , and a systematic change in the crystal type and morphology of the carbonates.

Chemical analysis of the lake beds reveals that their components have differing sources. The freezing on of seawater and/or frozen out seawater precipitates can quantitatively provide the salts required. However U isotope studies shows that a weathered crustal component is present in the lake beds. Stable isotope analysis indicates that the meltwaters were formed at high altitudes of 1,900 m (Ross Sea I) and 2,500 m (Ross Sea II(?)), and originated as snow, precipitated approximately 5,000 km to the south.

Ice sheet incursion into the valley created a proglacial lake. Calcium carbonate and gypsum were precipitated due to increased lake saturation conditions. Stable isotope studies show that the calcium carbonate deposition was triggered biologically by phytoplankton. The morphology of the calcium carbonate and the presence of gypsum indicates that evaporitic conditions were also necessary for lake bed deposition, with the distribution of gypsum and calcium carbonate within the lake being controlled by depth of water.

Correlation of lake bed geochronology and geochemistry, together with the stratigraphy of the stream sections provides a glacial history of Marshall Valley for the last 180,000 years B.P. The last two Ross Sea glacial sequences rest upon an unconformably eroded glacial surface. The Ross Sea II(?) Glaciation commenced at 180,000 years with a glacial ice incursion into the valley, a stillstand, proglacial lake occupation and deposition of a lake bed. The ice front then advance farther up the valley, depositing large terminal loop moraines between Sections III and IV. The ice then retreated at least 2 km.

A second glacial advance commenced prior to 160,000 years, stillstanded and produced the 160,000 year lake bed, before retreating. A minor fluctuation and readvance, deposited another lake bed (160,000 years) or reworked the previous lake bed into the overlying stratigraphy, before the ice retreat.

The deposition of the 130,000 year lake bed was also a the result of a stillstand, before the retreat and cessation of the Ross Sea II(?) glaciation.

The Ross Sea I Glaciation involved the advance of a glacier front from an ice sheet occupying the M^{c}Murdo Sound and the deposition of a gypsum lake bed (15-30,000 years B.P.). With the retreat of the ice front, downwasting of ice-cored moraines in the mouth of the valley began, and currently continues.

The geochronology of the Ross Sea II(?) Glaciation correlates with eustatic sea level lowering in the Illinoian (Isotope Stage 6). Subsurface drilling indicates that the calcium carbonate deposition was a unique event but at least 8 earlier glacial incursions had occurred in the Marshall Valley.

The geochronology of the Ross Sea I Glaciation correlates with the eustatic sea level lowering in the Wisconsin (Isotope Stage 2).

The chronology and isotopic composition of the proglacial lake deposits can be matched to those in other ice-free valleys exposed to M^{c}Murdo Sound.

This thesis is

dedicated

to my parents

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CHAPTER 1

INTRODUCTION

Chapter 1.

Introduction.

Studies of the Antarctic have revealed that a substantial ice cover has existed at least throughout the Quaternary, and possibly since the early Miocene (Hughes *et al*,1974, Kennett *et al*,1974).

Large outflowings of the outlet glaciers draining the inland ice sheets of East Antarctica have occurred throughout the Transantarctic Mountains, resulting in the carving of the Dry Valleys of the M^{c}Murdo Sound region. These occurred at a time when there had been less uplift of the Transantarctic Mountain range and at least the eastern ends of the valleys were developed as fjords, and still retain Miocene and Pliocene glacial marine sediments (Brady,1979). Evidence exists suggesting that regular cycles of glaciation have continued, but with substantially less erosion, during the last 4M yrs (Denton *et al*,1970, Hendy *et al*,1979), accompanied by the gradual uplift of the Transantarctic Mountains. These have been termed the Taylor Glaciations, and result from the draining of the East Antarctic Ice Sheet which has been demonstrated to be a stable ice sheet and subject to change only after prolonged alteration of climatic conditions (Hollin,1962).

From glacial evidence found in the Dry Valleys, it is also suggested (Denton *et al*,1970) that there has been glaciations in which the floating ice shelf of the Ross Sea has enlarged and grounded on the sea floor to form a large ice sheet. Fed by the increased outflow of the south-west outlet glaciers of the Ross Sea region this ice has pushed westwards towards the M^{c}Murdo Sound Dry Valleys, infilling and damming the mouths of the valleys and depositing erratics and glacial moraines (Stuiver *et al*,1981).

The distribution of drift sheets in the Taylor Valley, and the limited erosional exposure shows that the two major forms of glaciation have not occurred in phase with one another, so that differing triggering mechanisms are required to explain their growth and decline.

The Taylor Glaciations have produced drift which has been the subject of research by a number of investigators (Denton, 1977, Robinson, 1979, Kovach and Faure, 1977). This research has shown that the Taylor drift sheets have a composition of continently derived sediment. The drift sheets are characterised by units of yellow sands and silts and diamictons. Units of lacustrine origin are present and some lacustrine clasts are found *in situ*. The occurrence of volcanically derived material and the presence of marine microfossils are rare within the Taylor drift sheets.

The drift sheets of Ross Sea Glaciations have also been investigated (Denton, 1970). The drift sheets have a composition of both continental and volcanic origin. However the largest percentage of the sediments are basaltic and therefore volcanic. The drift sheets are characterised by units of grey-black sands and diamictons. Units of lacustrine origin are common (frequently *in situ*) together with deltaic beds and strandlines. There is also an abundance of marine microfossils found in the sediments. These are commonly reworked. This drift also contains clasts of kenyite, a rare basalt, whose only known local source is on the western slopes of Mt. Erebus, 60 km to the east of the Tranantarctic Mountains.

The Ross Sea Glaciations are thought to result from the grounding of the Ross Ice Shelf, which is in turn fed by the West Antarctic Ice Sheet. This ice sheet rests on topography that is largely below sea level and therefore has an inherent instability (Weertman,1974, Whillans,1976), which might lead to rapid changes in the extent of the ice shelf.

Hollin (1962) suggested that the glacial response of an ice shelf is related to eustatic sea level. Hughes *et al* (1985) have developed a mechanism for the advance and retreat of ice shelves, dependent on the position of the "grounding line" of the ice shelf in relation to sea levels and therefore topographical highs on the sea floor. Since the sea floor rises to within 150m of the base of the Ross Ice Shelf close to the northern limit of the ice shelf, a glacial eustatic sea level drop, such as that experienced during a Wisconsin glaciation would have the effect of advancing the grounding line 1000 km, (Stuiver *et al*, 1981). If this happened, the West Antarctic Ice Sheet would effectively extend to a position north of M^{c}Murdo Sound.

Therefore if Ross Sea glaciations are triggered by eustatic sea level, then there should be a correlation in ages between glaciations and the known history of eustatic sea levels.

Glacial deposits attributed to Ross Sea I Glaciations have been dated by C^{14} techniques on freshwater algae found in the frozen moraines. This has suggested that the recession of the Ross Sea I Glaciation has occurred from 10,000 years B.P. to the present. Coinciding with an increase in eustatic sea level. The peak of the glacial eustatic sea level lowering during the Wisconsin glaciation occurred between 18,000 and 22,000 years B.P. This closely matches the earliest dates of the Ross Sea I advance in the Taylor and Miers Valleys (Clayton-Greene,1986).

From the known history of eustatic sea level change, periods of low sea level can be defined. If there is a correlation between Ross Sea glaciations and eustatic sea level, we should be able to predict the ages of the older Ross Sea Glaciation deposits.

The penultimate low sea level period occurred 180,000 years B.P.- Isotope Stage 6. Therefore the ages of the penultimate Ross Sea Glaciation (Ross Sea II (?) Glaciation) should correlate with this date. A chronology of this glaciation has been difficult to obtain due to the lack of datable material until the discovery of datable glacial deposits in Marshall Valley, although Porter (1977, 1979) has dated Hawaiian glacial drift of this age range.

In the Marshall Valley, the stream that drains the Rivard Glacier has cut down into the glacial sediments that form the valley floor. It has revealed a series of lake beds (suspected to be older than Ross Sea I) layered between glacial drift.

As there are no topographical highs in the valley capable of impounding lake waters, it is assumed that these proglacial lakes are the result of the impeded drainage of the valley created by the damming of the valley mouth by the grounded Ross Ice Shelf. Salts accumulated in the basin, which deposited beds of calcite and gypsum. These can be reliably dated by U/Th methods to an age limit of 400,000 years B.P. Therefore the collection and dating of the Marshall Valley glacial lake deposits would test the correlation between Ross Sea glaciations and eustatic sea level change.

If it can be shown from U/Th dating of proglacial deposits that there is a correlation between the last two Ross Sea glaciations and the glacioeustatic sea level lowering, it is likely that earlier Ross Sea drifts would also coincide with sea level low stands. As Isotope Stages 8 and 10, lie within the limit of the U/Th dating technique it would greatly strengthen the argument for a causal connection between sea level and Ross Sea glaciations if drifts of these ages could also be dated. As the Marshall Valley reveals in its upper subsurface the deposits of the last two Ross Sea glaciations, then it may be possible, that at depth in it's subsurface can be found the sedimentary remains of earlier proglacial lakes.

A diamond drilling programme was undertaken to investigate the subsurface of Marshall Valley and to retrieve subsurface samples.

The glacial deposits collected in Marshall Valley were analysed by X-Ray Diffraction, $\delta O^{16}/O^{18}$ methods and Scanning Electron Microscopy, to determine the origins of the lake waters and the salts now present in the deposits and the composition and crystal structure of the minerals in the lake evaporite beds.

The near continuous cropping out of the lake deposits provided a detailed chronology of the position of the ice front within the valley through time. Information on the mode of formation and deposition of the lacustrine beds will provide a model for the environment created by a grounded ice sheet incursion.

Field Work Objectives.

The first season in the Antarctic was involved in the collection of samples for analysis and commenced in mid November 1983. The field work involved 6½ weeks of work in the Taylor, Marshall and Miers Valleys examining proglacial sedimentation derived from Ross Sea ice incursions. The objective of this part of this study was to produce, by radiometric methods of dating, a dated sequence of beds that could be correlated laterally and thereby produce a glacial history for Marshall Valley for the last 180,000 years. The subsequent use of other geochemical techniques (Mass Spectroscopy, X-Ray Diffraction, Scanning Electron Microscopy) was initiated to develop an understanding of the geochemical events that occurred concurrent to the glaciation history. Major questions studied included the probable source of water for the lakes and the calcium and sulphates for the evaporites, the trigger for the deposition of the lake beds, and the mechanisms by which the surficial features seen in the valley were laid down.

The second season in Antarctica was to study this possibility. In January 1986 , Waikato University in a joint programme with the New Zealand Geological Survey, undertook the drilling of three holes in Marshall Valley using the Winkie drill rig. This produced a frozen core from which older glacial events could be studied, to determine whether evidence for older Ross Sea incursions exist in the subsurface and whether these include proglacial sediments containing potentially dateable carbonates.

Marshall Valley.

Location.

Marshall Valley (see Fig.1.1) is situated at 77°26'S, 164°45'E. A largely ice free valley, Marshall Valley is one of the Dry Valleys trending West-East in the Koettlitz Glacier area of southern Victoria Land. It is a small valley situated in the valleys between the Joyce and Garwood Glaciers in the north and the Miers and Adams Glaciers in the south.

It is bounded at each end by glaciers. In the western head of the valley is the small cirque glacier, the Rivard, which is drained by the ephemeral Marshall Valley Stream. The Marshall Valley Stream flows eastwards to the coast which is bounded by the floating Koettlitz Glacier as it flows into M^{c}Murdo Sound. The Koettlitz Glacier currently terminates close to the mouth of the Marshall Valley and since regular observations were initiated (1958) has been steadily retreating.

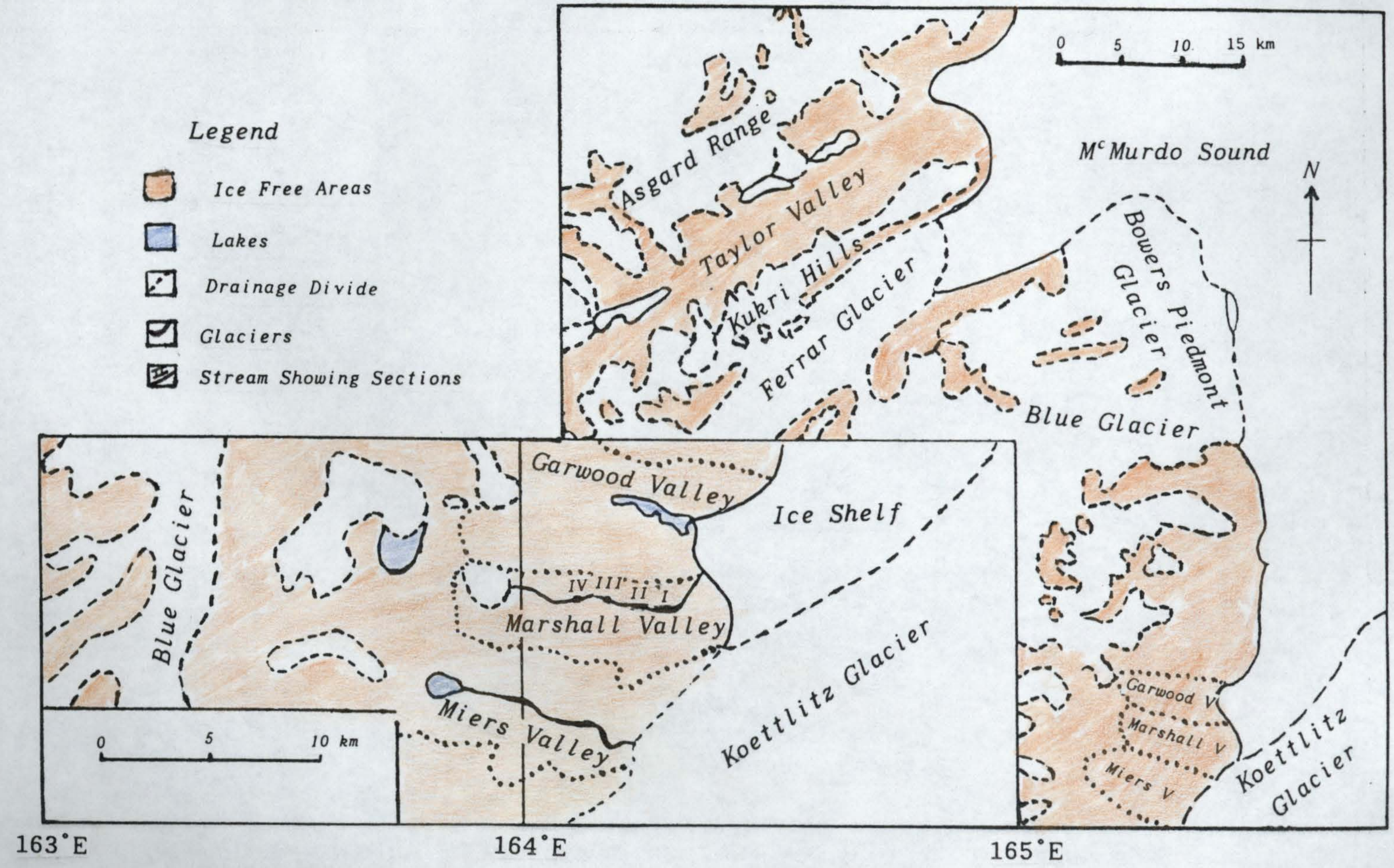


Fig. 1.1 Locality map of Marshall Valley.

Topography.

Marshall Valley as part of the south Victoria Land area, lies near the centre of the vast chain of the Transantarctic Mountains. This large mountain arc extends for 4000 km from the Admiralty Ranges in the north to a few hundred kilometres from the South Pole. The Transantarctic Mountains form the western boundary of the Ross Sea and the Ross Ice Shelf (Gunn and Warren,1962) and the eastern boundary of the massive East Antarctic Ice Shelf.

The Royal Society Range forms the highlands of the Koettlitz Glacier region. These are the eastern mountains of what has been termed the Antarctic Horst (David and Priestly,1914), the north-south series of mountain blocks which show varying degrees of elevation, tilting and dissection.

The walls of Marshall Valley reach an altitude of up to 1000 m and exhibit a steepness of slope comparable to that found in other glaciated areas of the world.(Gunn and Warren, 1962).

The floor of the valley ranges in elevation from sea level to 400 m and contains several topographical highs as the result of moraines left behind by glacier action.

Climate.

The existence of the Dry Valleys in southern Victoria Land is the result of several factors. There is presently, a greatly reduced inflow of glacier ice from the plateau than in the past. The topography of the high altitude polar plateau to the west of the region reduces the precipitation of ice and snow in the valley and provides an extremely dry katabatic wind. This combined with the increased ablation of the large expanses of bare rock reduces the net accumulation in the valleys.

Therefore the combination of the low albedo of bare rock and the comparative warm dry katabatic winds results in a localised climate which is appreciably warmer and drier then the surrounding glaciated regions (Gunn and Warren,1962).

Marshall Valley exhibits a shorter summer season then some of the more northerly valleys eg. the Taylor Valley. Temperatures rise above zero as a emphemeral stream flows in the valley.

Several weeks were spent in field investigations of Marshall Valley in two separate summer seasons. The first period of study was in early December (1983). At this time the valley was still occupied by small snow drifts and much of the southern bank of the stream was under an extensive snowbank. This was assumed to be permanent. The stream was flowing freely at this time.

The second period of study was for a month in January (1986). On arrival the valley was completely free of snow drifts and the stream flowed freely. It was discovered that most of the previously assumed, permanent south snow bank had disappeared. However in the last week of our occupation of the valley (23rd-30th January) the climate began to steadily cool. There were two days of snow which later ablated from the valley floor but remained on the higher ridges. The flow capacity of the stream was greatly diminished and there was partial freezing of the stream (Jan Clayton-Greene, pers.comm.). This marked the beginning of the oncoming winter season in Marshall Valley.

The range of temperatures recorded over our two summer seasons in Marshall Valley were $^{+}2$°C to $^{-}7$°C . This represents a warmer climate than the surrounding glaciated regions or the Ross Island region, but colder then the much larger Wright and Taylor Dry Valleys to the north.

Bedrock Geology.

The timespan represented in the bedrock of the Marshall Valley is from the Pre Cambrian through to the Silurian (1000 million to 400 million years B.P.). This timespan can be characterised as a period of sedimentary deposition followed by metamorphism and then intrusion. This was followed by an orogeny and subsequent folding, metamorphism and further intrusion. After the final Silurian episode there was a long period of erosion that resulted in the creation of the Kukri Peneplain (400 million years B.P. - late Devonian).

Subsequently in Devonian through to late Triassic there was a period of slow continental sedimentation, however these rocks have been eroded away from the valley ridges of the Koettlitz area and so no rock stratigraphy exists beyond the peneplanation episode.

The oldest rocks in the bedrock geology of Marshall Valley are those belonging to the Skelton Group. These consist of marbles, shists and gneisses which were originally laid down as impure limestones, greywackes and mudstones. This was a part of a geosyncline which lay off the coast of present day Victoria Land in the Pre Cambrian. These sediments were subsequently folded and metamorphosed during the Ross Orogeny and subsequently intruded (Gunn and Warren,1962).

The Ross Orogeny is thought to have occurred 500 million years B.P. (late Cambrian) and field evidence suggests that it can be divided into two phases of folding and 3 phases of plutonic intrusion. The period of orogeny resulted in the invasion of a complex series of intrusions ranging in composition from diorite through granodiorite to alkaline granite (Blank *et al*,1963). These intrusions consequently also underwent metamorphism to varying degrees.

The name Granite Harbour Intrusives is given to all plutonic and hyperbyssal intrusions that invade the Skelton Group and are older then the Kukri Peneplain (Blank *et al*,1963).

After this period of orogenic activity there was a period of erosion and the creation of the aforementioned Kukri Peneplain.

Therefore the bedrock of Marshall Valley consists of folded and metamorphosed marbles, shists and gneisses (termed the Marshall Valley Formation),intruded by diorites and granites (of various metamorphosed grades) then eroded to a peneplain. This rock stratigraphy has been block faulted, tilted and eroded to the present topography.

Surficial Deposits.

The surface of Marshall Valley is covered by glacial debris as the result of glaciation in the valley (see Fig.1.2).

These deposits can be divided into two types:

The first group of sediments are those representative of glaciations that have resulted from the expansion of outlet glaciers draining the inland ice plateau. This resulted in the movement of glacier ice in the eastward direction.

The deposits of these glaciations consist of the light coloured rocks of the basement- shists, gneisses and marbles with metamorphosed intrusions. This material is found only on the valley walls at high altitudes and at the head of the valley, although it appears that this material extends the length of the valley, but is overlain by the glacial deposits of the second group.

The sediments of the second group are representative of glaciation that has resulted in the expansion of the Ross Ice Shelf. This expanded ice shelf grounded and subsequently flowed up the valley

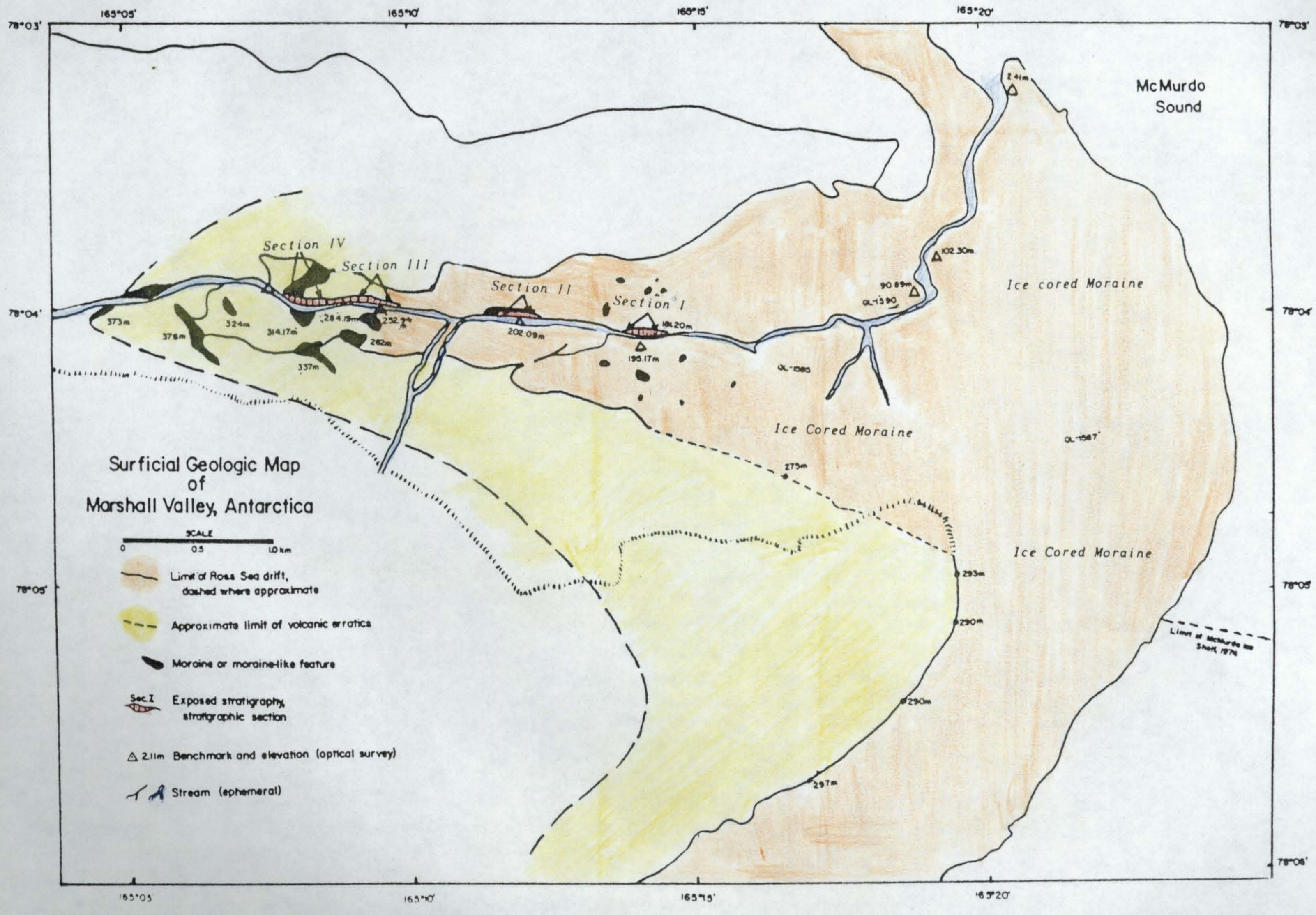


Fig. 1.2 Surficial geology map of Marshall Valley.

in a westerly direction. As a consequence of this incursion, glacial moraines have been deposited in the valley.

The deposits of this glaciation have a dark basaltic rich lithology. Present in the glacial drift is kenyite, a volcanic rock whose only known origin is the Ross Volcanic Complex. This implies that the movement of the Ross Ice Shelf has been in a circulatory direction flowing westwards around and over Ross Island and the $\mathrm{M^cMurdo}$ Sound to the Dry Valleys (Stuiver *et al*, 1981).

The expansion and damming of the mouth of Marshall Valley has created lateral moraines on the coastal ridges of the valley. These are present to an altitude of 270 m. Extrapolation from the lateral sides of this glacial incursion into the valley, includes a lobe of dark basaltic drift extending a distance of 7 km up the valley. The drift encompassed to this point is patchy and subdued on the valley floor and at the valley mouth rises in a series of ice cored moraines (Dagel *et al*, in press)

The ice cored moraines have a relief of up to 12 m and have a rugged, hummocky topography containing numerous kettle lakes which sustain algal mats (Dagel *et al*, in press) (Plate 1.I and Plate 1.II).

This glacial drift is thought to represent Ross Sea I Glaciations and has been dated in several locations. Radiocarbon dates from fossil algae document the recession of ice from the lower Marshall Valley- 13,300 years ± 150 years B.P. at sample site QL-1584; by 12,530 years ± 40 years B.P. at sample site QL-1590; and by 8,710 years ± 50 years B.P.at sample site QL-1587; (Fig.1.2) (Stuiver *et al*, 1981).

Dark glacial drift extends past the presumed extent of Ross Sea I and is therefore thought to predate this glaciation. Tentatively thought to be Ross Sea II, this drift extends 6.5 km up the valley. It is revealed in section, through the downcutting of the ephemeral

Plate 1.II Meltout of a Marshall Valley ice-cored moraine
(photo C Hendy

Plate 1.I Ice-cored moraines at the mouth of Marshall Valley
(photo C Hendy)

Marshall Valley Stream, in a series of lake deposits layered between dark glacial drift including tills and outwash sediments. These sections are known as the Marshall Valley Stream Sections.

The lake deposits are from proglacial lakes resulting from the impeded drainage of the valley when occupied by a Ross Sea glaciation and include deposits of gypsum and calcium carbonate, layered between the sands and gravels of the glacial drift, to form a near continuous feature along the 3 km of stream sections. Deposits of calcium carbonate are also present at the surface of the valley.

CHAPTER 2

PREVIOUS WORK.

Chapter 2.

Literature Review.

Research into Antarctic glaciations: their origins and sequence, commenced in 1841 when Sir James Ross mapped the edge of the Great Icy Barrier, which we know today as the Ross Ice Shelf.

This area was remapped by the National Antarctic Expedition of Scott in 1902. From this expedition, scientists determined that the Ross Ice Shelf was floating and that the edge of the "barrier front" appeared 24 to 32 km south of Ross's positioning. This raised the question of whether the Ross Ice Shelf was in a state of retreat, and implied further by this question that the present Ross Ice Shelf was the remanent of a far larger feature in the past. The 1902-1903 expedition scientists also gathered evidence as to the past extent of the ice shelf.

In publication of the expeditions' findings, Scott (1905) stated that at the southern glacial maximum, the Ross Ice Shelf was far more enlarged than today. At the time of the glacial maximum, the Ross Ice Shelf grew too large to float and grounded, filling the area with an ice sheet that extended north to Cape Adare. In discussing the recession of this glaciation, Scott suggested that the ice sheet became buoyant once more and retreated slowly, by iceberg calving to the remanent that remains today.

In his following expedition Scott (1914) carried out further surveys on the Ross Ice Shelf and suggested that the present movement of the ice shelf is fed almost wholly by the deposition of snow on its surface. Having calculated the annual movement of the ice shelf (by remapping the positions of old food depots from the previous expedition) he stated that the calculated increment of snow on the ice shelf was sufficient to account for any movement.

Scott did not completely discount the contribution of the Victoria Land glaciers to shelf movement but, based on his observations (Scott,1914) that their movement was exceedingly slow, he concluded and that their occurrence at infrequent intervals along the coastline would greatly diminish their contribution.

Concerning the larger inland draining glaciers of the Beardmore region, Scott,(1905) stated " that movement must mainly be a measure of the pressure behind, that is the Great Inland Ice Sheet or other sources of supply". Scott having no evidence of the rapid movement of these glaciers felt " that the supposition of a rapidly moving glacier is purely the effect of imagination and is without doubt invoked to account for the barrier movement". Far slower movement he felt could be assumed " if the Antarctic glaciation is receding, the internal ice sheet must be suffering denudation ...... glacier movement everywhere on the coast is mainly due to the local neve fields and therefore never rapid" (Scott,1914).

Shackleton's British Expedition 1907-1909 also focused on the Ross Ice Barrier. Further evidence for the northward advance of the ice shelf was found in the distribution of ice-cored moraines and foreign erratics at Cape Royds on Ross Island. From the mapping of this evidence (David and Priestly,1914) postulated that at its greatest extent the Ross Ice Shelf had grounded and that ice had built up to an altitude of 305 m above sea level and that this ice sheet had extended 320 km farther north then the present extent of the ice shelf.

The presence of continental erratics on Ross Island (David and Priestly,1914) was attributed to the northwards thrust of outlet glaciers from Skelton, Mulock, Barne and Shackleton Inlets and possibly the Beardmore Glacier. David and Priestly (1914) stressed the importance of the outlet flow of the thickened glaciers that drained into the Ross Ice Shelf. It was their presence that provided the thrust from the south that deflected the west coast outlet glaciers in a northward direction.

However in publication of their work the two geologists had conflicting ideas as to the flow of ice in M^{c}Murdo Sound. In their combined paper, David and Priestly,(1914) suggested that during the maximum glaciation two west coast outlet glaciers, the Taylor and the Ferrar, in greatly expanded form, flowed to the Ross Sea where they were deflected northward by the flow of the expanded ice shelf.

However Priestly,(1915) suggests that the flow of ice within the M^{c}Murdo Sound was in a more circulatory direction and that the ice flowed around and over Ross Island (to the east of M^{c}Murdo Sound) before it flowed over the sound and onto the coastland of the western valleys. Evidence for this pattern of flow was the presence of kenyite-rich debris in the Strand Moraines of the M^{c}Murdo west coast. The only origin of such a rock is from the Ross Island Volcanics. Kenyite was also found by Priestly in the erratics on the morainal debris of the lower Taylor Valley.

The glacial deposits found in the M^{c}Murdo Sound region by subsequent Antarctic expeditions were given an alternative interpretation. Debenham (1921) felt that the foreign erratics and the ice-cored moraines of Ross Island were deposited by an expanded lobe of the Koettlitz Glacier from the west of M^{c}Murdo Sound. Debenham suggested that although the Ross Ice Shelf subsequently thickened by this advance it did not ground but remained afloat. The distribution of the erratics and the ice-cored moraine on Ross Island was achieved by the subsequent expansion of the Erebus Ice Tongue on Ross Island. Debenham's (1921) interpretations of the events during the maximum glaciation was a localised situation compared to that of David and Priestly (1914). Although Debenham (1921) felt that the west coast outlet glaciers were contributors to glacial advance, he felt that their former sizes were greatly exaggerated.

Very little evidence was added to either interpretation of glacial phenomena until the 1950's commencing with the renewed interest in Antarctica in the years of Operation Highjump and the International Geophysical Year 1957-58. The continued exploration of Antarctica and the use of man's increasing technology enabled the use of aerial photography, seismic and gravity surveys to increase knowledge of the internal structure of the Antarctic beneath the vast ice sheet. Because of these remote sensing techniques, the structure and topography of Antarctica provided new evidence as to the origins and sequence of glaciations.

The division of Antarctica into two ice sheets: East Antarctica and West Antarctica divided by the Transantarctic Mountains, the increased knowledge of the topography underneath these ice sheets and the direction of flow of the ice sheets themselves, (Bentley, 1961, Bentley *et al*, 1965, Giovinetti *et al*,1966) has provided a new perspective for the study of Antarctic glaciation (Weertman,1976).

Further exploration of the M^{c}Murdo Sound region and the study of the glacial deposits found in the region has continued since IGY.

Péwé (1962) in his investigations of the Koettlitz Glacier and Taylor Valley regions first recognised the presence of multiple glaciations within the western valleys of M^{c}Murdo Sound. He differentiated between four glacial deposits on the basis of their separation in altitude and degree of weathering and moraine topography. He named each successively smaller glaciation and described the characteristics of their glacial drift and their origin. The oldest and most extensive glaciation , the M^{c}Murdo, was characterised by highly weathered glacial deposits found on flat topped ridges at altitudes of 700 m above the valley floor. The next two succeeding glaciations, the Taylor and the Fryxell were represented by well distributed, preserved but greatly subdued glacial moraines of the outlet and alpine glaciers. The youngest glaciation was the Koettlitz. These glacial deposits were found deposited close to the coastline and their moraines were still ice-cored.

In discussing the origins of these glaciations Péwé (1962) stated that in times of glaciation, the outlet glaciers, the Taylor and Ferrar, draining the expanded ice cap, filled their respective valleys and then coalesced with the expanded Koettlitz Glacier in the M^{c}Murdo Sound.

The M^{c}Murdo Glaciation represented the largest of these glaciations and the M^{c}Murdo Sound region became filled with ice to 700 m altitude. The successive glaciation were less extensive and the deposits were also formed by expanded alpine glaciers occupying the western valleys. With each successive glaciation the Koettlitz expansion became smaller, culminating in the Koettlitz Glaciation in which the coalesced Ferrar-Koettlitz ice tongue deposited the ice-cored moraines found on the western coast.

Péwé (1962) also discussed the presence of the glacial lakes in the interglacial periods between successive glaciations. He noted the successive strandlines exhibited in Taylor, Miers and Garwood Valleys and associated these with stillstands.

In attempting an age correlation of the glacial sequence, Péwé was able to radiocarbon date dried algae from the ice-cored moraines of the Garwood Valley. He therefore suggested that the Koettlitz Glaciation had a minimum age of 6,000 years B.P. In dating the earlier glacial events Pewe was unable to assign absolute ages. On the basis of morphology he placed a longer time interval between the first M^{c}Murdo Glaciation and the successive Taylor-Fryxell Glaciations. The timespan of the glaciation events he could only tentatively place in the Pleistocene (2.0 million years B.P. to the present).

On the basis of evidence collected during the the International Geophysical Year, Péwé (1962) suggested that " climatic warming and glacier retreat have been contemporaneous in Antarctica during the last few thousand years."

Following the work of Péwé (1962) the extensive mapping of the ice-free Wright and Victoria Valleys by (Bull *et al*,1962) resulted in the recognition of four glacial advances in the valleys, all successively smaller with age. The first two glaciations originated with the increased westward flow of the outlet glaciers draining the ice plateau. These glaciations excavated the valleys and their remaining deposits were thin and highly weathered. The third glaciation consisted of a smaller advance of the plateau glaciers and the fourth glaciation marked the smallest advance of plateau glaciers.

The occurrence of the successive strandlines surrounding the saline lakes, found in these valleys, was suggested as evidence of previous glacial events.

As with Péwé, (1962) Bull *et al*,(1962) suggested that the impetus of glaciation was from the expanded ice plateau that was drained by the outlet glacier and the sequence of glaciation was thought to be initiated by the response of the inland ice sheet to climatic change.

It was the work of Hollin (1962) that placed renewed emphasis on the response of the Ross Ice Shelf. From a consideration of glaciological theory, Hollin calculated that whilst changes in the response of inland ice caps to climatic changes would be small, the responses of ice sheets, at or below sea level, would be large enough to account for the glaciation seen in Antarctica. Hollin (1962) suggested that a sufficient drop in sea level, resulting in the lowering of the grounding line of an ice shelf, produced the response of a glacial advance into an embayment. " The glacial history of most ice-free areas is governed, not so much by climatic as by sea level changes. Therefore Antarctic glacial fluctuations are dependant on and in phase with those of the Northern Hemisphere" (Hollin,1962).

The glacial chronology of Taylor Valley was further investigated by Denton *et al* (1968,1970,1974). Denton recognised 3 major glacier systems: outlet glaciers that drained the dammed East Antarctic Ice Sheet, the Ross Ice Shelf that floats on the surface of the Ross Sea and independent alpine glaciers that occur throughout the Transantarctic Mountains.

By the detailed mapping of the glacial deposits left behind by these glacier systems, Denton (1970) suggested, " fluctuations of the three major glacier systems of the M^{c}Murdo Sound region are not synchronous."

In considering each glacier system independently Denton, (1970) assigned the name Taylor Glaciations to all eastward flowing glaciations attributed to the overflow of outlet glaciers from the inland plateau. Denton (1970) recognized at least five Taylor glacial advances, (see Fig.2.1) Taylor V, the oldest glaciation carving out the major features of Taylor Valley and reaching the M^{c}Murdo Sound. The successive Taylor glaciations were all smaller with age. Taylor I was assumed to be current.

Denton (1970) was able to date these older glaciations by K/Ar geochronology of lava flows separated by layers of glacial drift. This extended the oldest dated period of Taylor Glaciation to greater than 3.5 M yrs B.P. and the onset of Taylor IV Glaciation. By this method the Taylor III Glaciation is estimated to have occurred at least 1.6 - 2.1 M yrs B.P.

Denton (1970) also recognised at least four separate expansions of the Ross Ice Shelf into a grounded ice sheet. These he termed Ross Glaciations and suggested " During these Ross Glaciations , ice sheets in the Ross Sea and the M^{c}Murdo Sound reached high on the flanks of

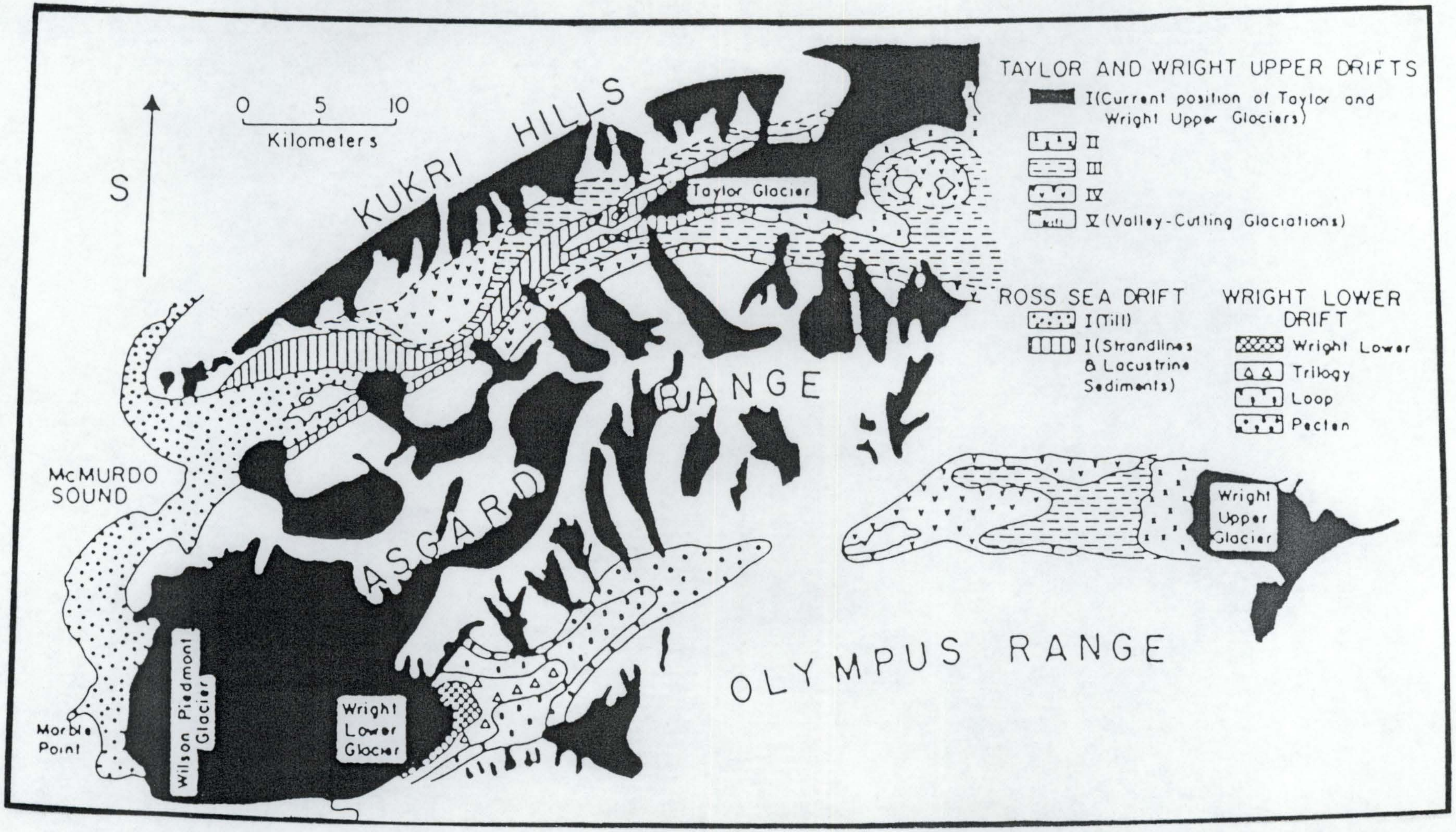


Fig. 2.1 Surficial glacial unit map of Taylor and Wright Valleys (Map from Denton, Armstrong and Stuiver,1971).

Mount Discovery, Brown Peninsula, and Black, White, and Ross Island. Glacier tongues from these ice sheets pushed westward up Taylor Valley and the valleys fronting the Royal Society Range, leaving well preserved moraines on the valley floors and along the coast."

Because of the possible confusion with the New Zealand Ross Glaciation, Denton subsequently renamed the Antarctic Ross Glaciations, "Ross Sea Glaciations".

The use of K/Ar and C^{14} dating techniques have placed ages on these glaciations (Denton 1970). K/Ar dates on a lava flow suggested that the Ross Sea IV Glaciation was older than 1.2 - 3.1 M yrs B.P. The dating of shells from raised marine beaches between glacial tills provided dates for the interval between Ross Sea I and II at 34,000-49,000 yrs B.P. (Hendy *et al*, 1969) (see Fig.2.2).

Denton (1970) also dated the recession of the Ross Sea I glaciation by the use of C^{14} dates on buried freshwater algae found in the ice cored moraines. He suggested that the recession of the Ross Sea I ice from the west coast Dry Valleys had occurred from 10,000 yrs B.P. to the present and " coincided closely with the rapid rise in sea level during the late Wisconsin (Würm) time."

In summary Denton (1970) stated that the Ross Sea Glaciation was unlikely to have been climatically controlled as the glacial chronology of the Dry Valleys showed that alpine glaciations did not fluctuate in phase with the Ross Sea Glaciations.

Denton (1970) suggested that the explanation of the correlation between Ross Sea Glaciations and sea level could be explained by Hollin's (1962) model of the alternate grounding and floating of the Ross Ice Shelf. Another alternative suggested but not considered likely by Denton was that surging of ice from West Antarctica caused Ross Sea Glaciations.

Taylor Glaciations (Ice sheet in East Antarctica West of Taylor and Wright Valleys)	Ross Sea Galaciations (Ross Ice Shelf)	Alpine Glaciations
	4550 yrs. B.P.(L-627; Marble Point)[1]	
	5900 yrs. B.P.(L-462; Hobbs Glacier)[2]	
	6110 yrs. B.P.(Y-2401; Hobbs Glacier)	
Taylor I	9490 yrs. B.P.(Y-2399; Hobbs Glacier)	**Alpine I**
		12,200 yrs. B.P.(I-3019; Hobbs Glacier.)[3]
	Ross Sea I	
	34,800 yrs. B.P. (no laboratory no. given; Cape Barne)[4]	**Alpine II**
	>47,000 yrs. B.P. (Y-2641; Cape Barne; same locality as sample dated 34,800 yrs. B.P.)	K/Ar dates; 2.1 to 0.4 m.y. (Walcott Glacier area.)
	>49,000 yrs. B.P. (Y-2642, Cape Barne)	
	Ross Sea II	
Taylor II		
	Ross Sea III **Ross Sea IV**	
Taylor III	K/ Ar dates; 3.1 to 1.2 m.y. (Walcott Glacier area)	
K/Ar dates; 1.6 to 2.1 m.y. (Taylor Valley)		K/Ar dates; 2.1 m.y. (Taylor Valley)
Taylor IV		**Alpine III**
K/Ar dates; 2.7 to 3.5 m.y. (Taylor Valley) and 3.7 m.y. (Wright Valley)		K/Ar dates; 3.5 m.y. (Taylor Valley)
Taylor (s) IV		

[1] Nichols (1968, p.471) ; Olson amd Broecker (1961, p. 150). The C_{14} date given in the chart and text is corrected. The uncorrected date is 5650 ± 150 yrs. B.P. (L-627).

[2] Pewe (1961).

[3] Black and Bowser (1969).

[4] Hendy et al (1969)

Fig.2.2 Schematic correlation chart and chronology of glacial events from the M$_c$Murdo Sound Region. (from Denton, Armstrong and Stuiver, 1970) ammended to include renaming of glaciations.

Continued remote sensing surveys of the West Antarctic Ice Sheet which the Ross Ice Shelf drains, revealed that the topography of the West Antarctic land mass was an island archipelago and that the majority of the landmass was 0.5 - 1.0 km below sea level and would remain below sea level if all the ice was to be removed (Bentley, 1961).

Workers have investigated the stability of the West Antarctic Ice Sheet (Whillans,1976) and it has been postulated that the West Antarctic Ice Sheet is capable of surging into its ice shelf which provides the inducing mechanism for ice ages, through the effect of increased albedo (Wilson,1968). In assessing this possible instability (Hughes,1973) studied the theoretical dynamics of the West Antarctic Ice Sheet to deduce if the ice sheet is presently in a regime of retreat. On theoretical grounds, a polar ice sheet grounded below sea level was found to be unstable and therefore capable of disintegration. Direct evidence of disintegration was limited however and the possible surging capacity of the ice sheet remains unresolved.

With the knowledge of the possible instability of marine ice sheets, mechanisms of the movement of a grounded ice sheet have been investigated.

Hughes *et al* (1985) developed a model for grounded ice shelf advance and retreat. The ice sheet advances along a trough until the grounding line rests over the sill. Isostatic adjustment of the sill lowers the sill to the grounding line creating instability by the thinning of the ice sheet. The grounding line then retreats by the process of iceberg calving. This period of retreat is balanced by the calving rate of the sheet versus the supply of ice to the grounded ice shelf by the downdraw of outlet glaciers.

The glacial history of several outlet glaciers draining the East and West Antarctic Ice Sheets have been studied. Mercer (1968) in his investigations of the Reedy Glacier area notes three stranded lateral moraines and ice marginal lake sediments. He suggested this was evidence of the varying thickness of the Reedy Glacier through time. Mercer (1968) states " The varying thickness of the Reedy Glacier after it has become an outlet of the East Antarctic Ice Sheet are thought to have been determined mainly by the position of the grounding line of the Ross Ice Shelf, which was controlled by the eustatic changes in sea level, so that the glacier thickened during the Northern Hemisphere glaciation".

Oxygen isotope and palaeomagnetic stratigraphy (Shackleton and Opdyke, 1973) has been used to produce a sea level curve, that is capable of extending to the Brunhes-Matayama boundary (700,000 years B.P.). Based on the equivalance that a 0.1‰ change in the oxygen isotope ratios found in deep sea cores is equivalent to a 10 m sea level change, a sea level curve is generated (see Fig.2.3). Therefore the sea level curve represents the eustatic adjustment of sea level as a response to the glacial-interglacial cycle of the world's climate. The curve is divided into Isotope Stages, the even numbered stages representing glacials and the odd numbered stages representing interglacials. The curve, first produced by Shackleton and Opdyke (1973) represents 10 cycles of glaciations within the late Pleistocene Epoch. As a response to glaciations during this timespan they calculated fluctuations in the eustatic sea level of 140 m.

Further work has continued on the glacial deposits found on the M^{c}Murdo Sound region. There has been continued work into the chronology of the Ross Sea glaciations to test the correlation between the glaciations and the eustatic sea level record. The work of Stuiver *et al* (1981) contains the investigations of Ross Sea I glacial drift. The extent and characteristics of the drift have been mapped and described. Correlated with the extensive C^{14} dating of raised

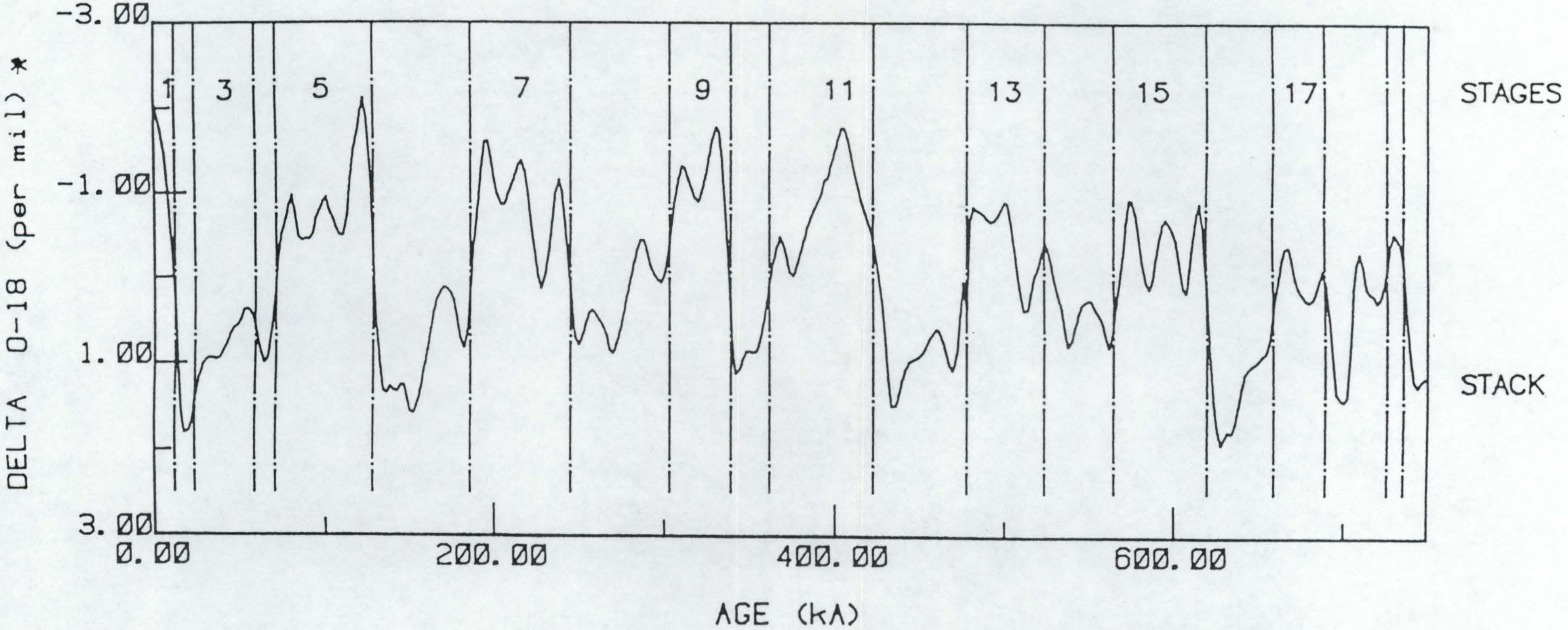


Fig. 2.3 Late Quaternary isotope stage 1 to 19 defined for age_based "SPECMAP" δO^{18} isotope stack (Imbrie *et al*,1984) in(Cuthbertson,1985).

marine terraces, dead seals and penguins and fresh water algae samples, this work provides a detailed account of the chronology of the Ross Sea I glacial advance and its continuing recession. The correlation of the numerous dates produced a chronology of the fluctuations of glacial lake Washburn in the Taylor Valley. The presence and size of this proglacial lake is considered indicative of the accompanying phase of Ross Sea Glaciation and is determined by the extent of incursion of glacial ice into Taylor Valley. (see Fig.2.4)

Investigations into the Ross Sea II Glaciations (tentatively dated Isotope Stage 6, 180,000 yrs B.P.) has continued in the Taylor Valley and in the Marshall Valley where possible Ross Sea II lake beds have been discovered in the downcut sections of the Marshall Valley Stream. The deposits have been U/Th series dated and a geochronology of Marshall Valley established (Dagel *et al*,in press). The results of this work in Marshall Valley forms the basis of this thesis.

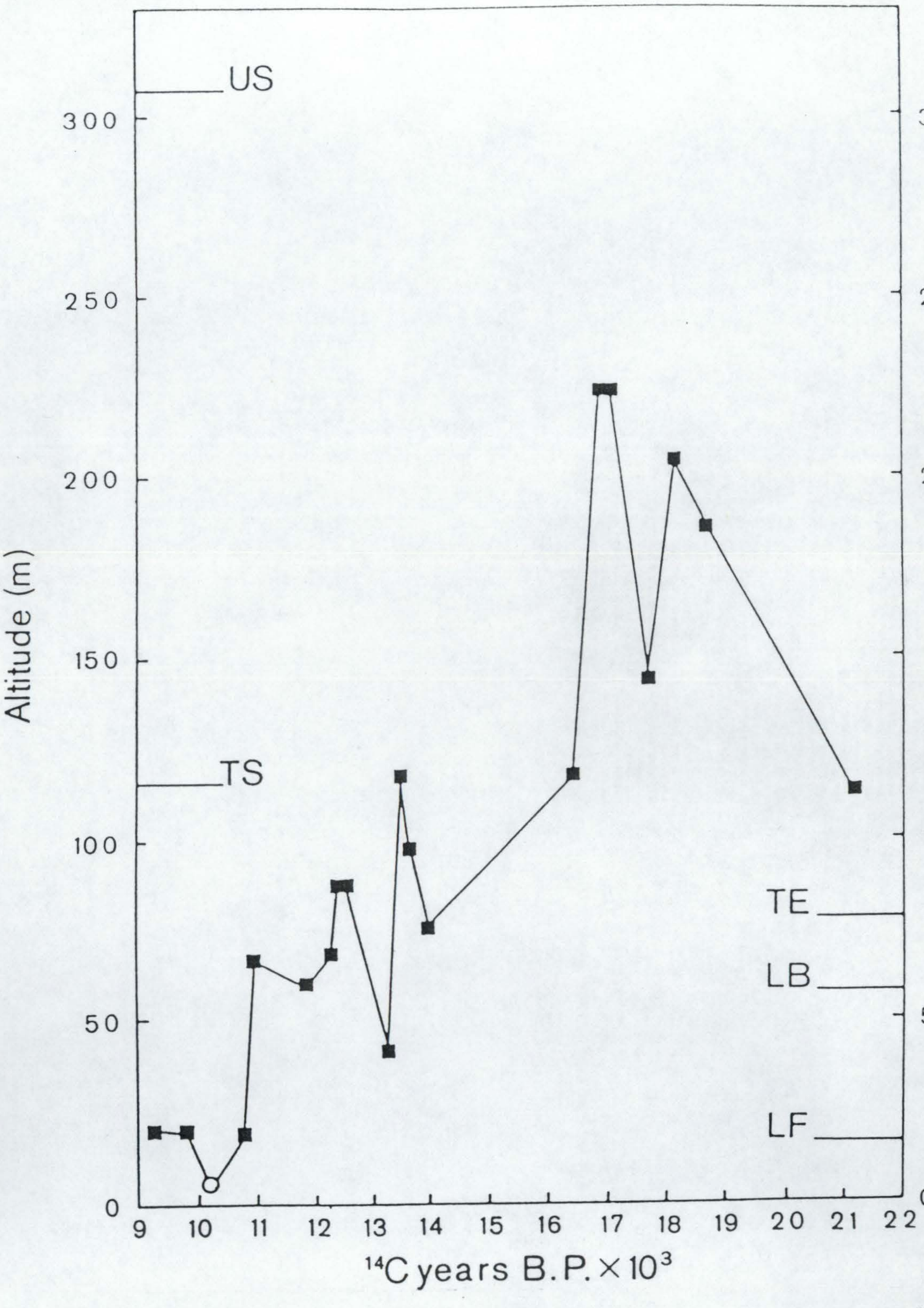


Fig. 2.4 Lake level curve for glacial lake Washburn, Taylor Valley (Lawrence,1981)

CHAPTER 3

U/TH GEOCHRONOLOGY

Chapter 3

Introduction

This chapter explains the general theory of the radiometric dating and then focuses on its application to Antarctic lake beds and its advantages and disadvantages. The results of the U/Th dating are then presented and their implications discussed.

Discussion of the radioactive decay theory.

The theory of radioactive decay explains the transformation of a unstable nuclide to a stable nuclide (often through a series of unstable nuclides). The rate of decay of the parent nuclide is proportional to the number of atoms present at the time and so occurs at a definite rate or decay constant.

$$\frac{-\ dN}{dT} = \lambda N$$

where -dN/dT is the rate of change of the number of parent atoms as it decreases with time.

The proportionality is transformed to an equality by the introduction of the term λ which is the decay constant (the probability that an atom will decay in a set unit of time). If the radioactive decay of an isotope occurs at various stages towards a stable atom then there is a relationship among the abundance of the isotopes in the series. It can therefore be hypothesised that if the rate of decay of the isotopes in the series is known and if the proportions of the isotopes to one another in the series can be measured, then this relationship together with a close estimate of the proportions of the isotopes at the initial time of synthesis will enable the age of the material to be calculated.

Often it is the uncertainty in the original proportion of isotopes in a sample, that hinders the accuracy of a date. The reliability of dating can be improved if the initial abundance of the daughter isotopes was zero and therefore all the measured abundance of the isotope is due to the subsequent radioactive decay of a parent nuclide.

U/Th dating theory

The U decay series is a radioactive decay series (see Fig.3.1) involving a sequence of alpha and beta decay along a large series of unstable isotopes before a stable isotope of lead is achieved.

Of main interest is the decay of isotopes of U through to isotopes of Th. In particular is the production of Th^{230} from U^{238} and U^{234} that enables this series to be used for dating purposes. As discussed later, it is the assumption that our dating material contained little or no Th^{230} originally, that allows a high degree of certainty to be achieved in dating.

U/Th dating and it's applicability to carbonates

$CaCO_3$ deposited in the oceans and lakes contains appreciable amounts of uranium (up to 100 ppm) however virtually no Th is incorporated in the sediments. Thus due to the chemical nature of uranium and thorium; the 4^+ oxidation states of both uranium and thorium form very insoluble salts and ions which strongly hydrolise water and strongly absorb onto particulate matter. Thus their ions are rarely found in solution in natural waters. The 6 + oxidation of

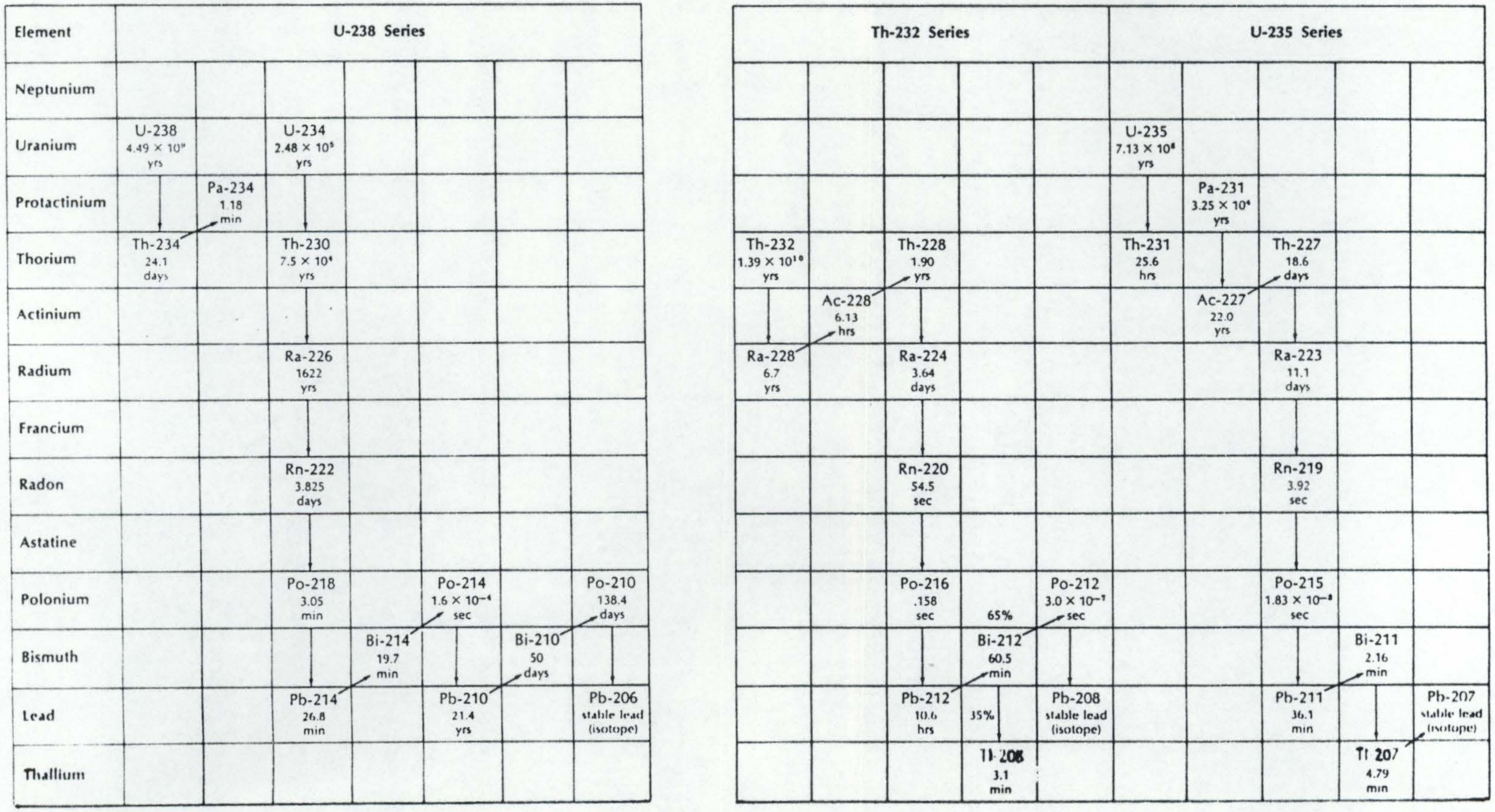


Fig. 3.1 Radioactive decay series of U^{238}, Th^{232} and U^{235} (Broecker, 1974)

uranium however is stable at moderate oxidation potentials and forms the UO_2^{2+} ion which produces soluble salts. It complexes strongly with CO_2 and mimics Ca^{2+} for which it readily substitutes into minerals such as calcite.

If we assume that no Th^{230} is present at the deposition of the carbonate then the amount of the Th^{230} generated since the formation is a measure of the age of the sample. As stated Th^{230} is generated by the decay of U^{234}, itself a daughter of U^{238}. First attempts at dating carbonates measured the ratio Th^{230}/U^{238} as it was assumed that the uranium isotopes U^{234} and U^{238} were in equilibrium. Thus since U^{234}/U^{238} was assumed =1;

$$\frac{Th^{230}}{U^{238}} = 1-{}^{e-\lambda 230t}$$

Thurber (1962) demonstrated that for natural waters this was not the case and that in seawaters the U^{234}/U^{238} = 1.15 and so the simple equation Th^{230}/U^{234} activity ratio does not apply as it does not allow for an excess of U^{234}. In smaller and more confined water bodies the ratio of U^{234}/U^{238} shows considerable variation, apparently responding to the relative importance of alpha recoil (by which U^{234} is preferentially removed from the crystalline matrix) to leaching as a mechanism for mobilising the uranium.

In the Antarctic, ratios range from greater then 3.7:1 (Lake Vanda) to 1.1:1 (Lake Fryxell) (Hendy *et al* 1979). The ratio of U^{234}/U^{238} measured in our samples = 1.5:1 and has a large effect on the date .

It is therefore necessary to allow for the decay of U^{234} , ie

$$dU^{234}{}_{dt} = {}_{-}\lambda_{234}U^{234} + \lambda_{238}U^{238}$$

and

$$dTh^{230}{}_{dt} = -\lambda_{230}Th^{230} + \lambda_{230}U^{230}$$

since solution of these simultaneous differential equations is not possible by direct means, a graphical approach is utilised (see Fig.3.2) in which isochrons are plotted on a graph of U^{234}/U^{238} versus Th^{230}/U^{234} ratios. The age of any combination of these ratios can be found by interpolating between the isochrons.

Therefore to calculate an age by U/Th dating ,the ratio of Th^{230}/U^{234} (the amount of Th^{230} generated since deposition) and the ratio of U^{234}/U^{238} (the measure of the equilibrium between U isotopes) must be known.

Several assumptions are made:

1) The initial Th^{230}/U^{234} activity ratio is close to zero.

2) That the sample is from a closed system with respect to uranium and the intermediate nuclides between U^{238} and Th^{230} .

The validity of these assumptions

1) Because of the low solubility of Th^{4+} salts, natural waters have low concentrations of Th^{230}/U^{234}, resulting in little Th being incorporated into Ca minerals. If however Th^{230} was to be incorporated into sediments on fine particulate matter it would have associated with it Th^{232}. An approximate 1:1 association between Th^{230} and Th^{232} is assumed as this is the activity ratio of Th^{232} to Th^{230} in average crustal rocks. Thus a measure of Th^{232} will allow an approximation of initial Th, the radiogenic Th^{230} being the total Th^{230} in the sample minus the initial Th^{230} (measured from the Th^{232}).

2)Incorporation or removal of uranium as a secondary migration has been discussed by Broecker (1963). He suggested that there was a postdepositional process that does add uranium to marine mollusks in the initial period of the deposition. However due to the nature of our deposits (to be discussed later) the incorporation and enrichment

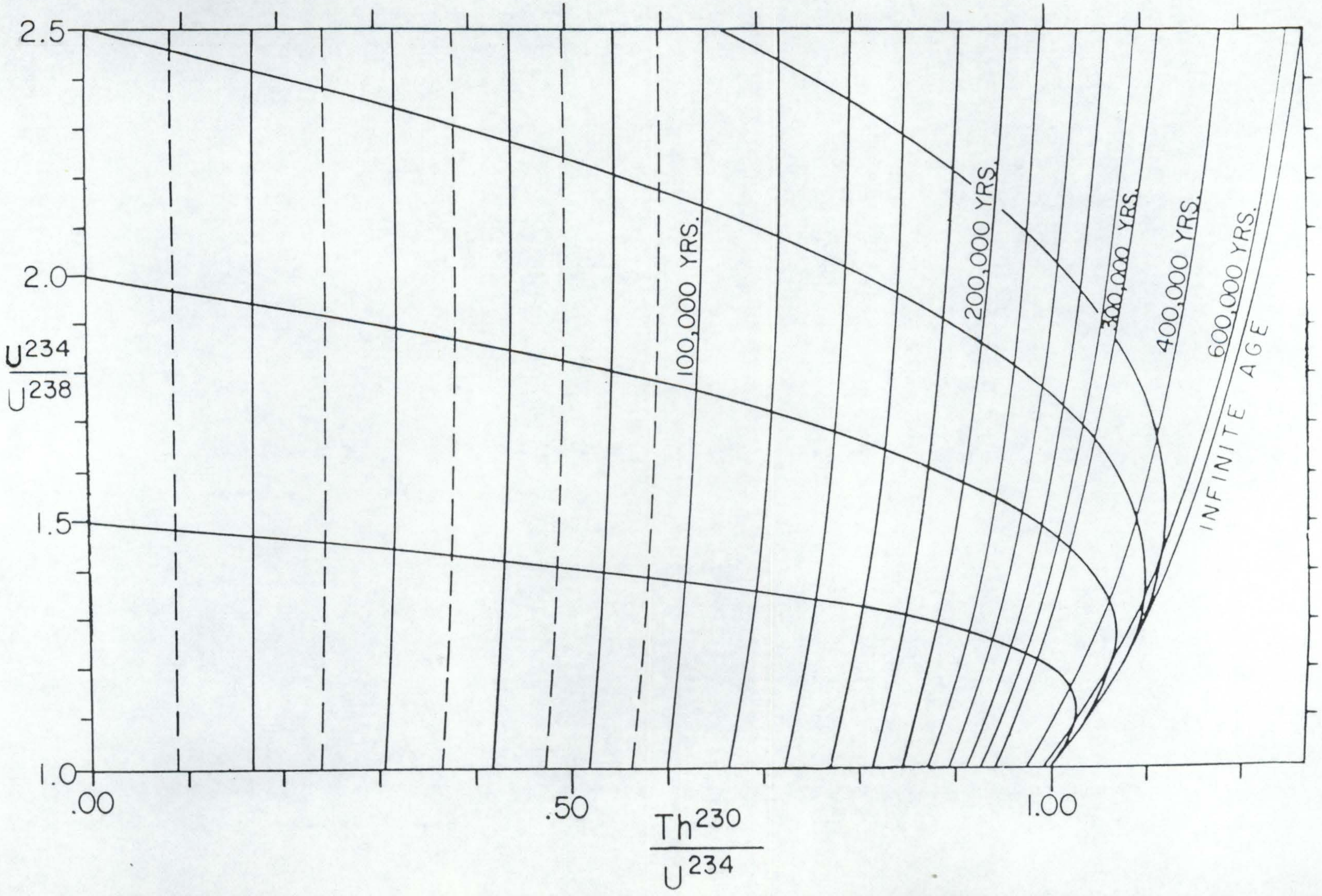


Fig. 3.2 U/Th isochron plot used in age determinations(from Kaufman,1964).

or depletion of our sediments is not significant and therefore the system can be considered closed.

Assumptions of Method with respect to Antarctic Lacustrine Carbonates

The applicability of U/Th dating methods to particular carbonate environments requires special considerations:

1)Are the expected age of these carbonates within the optimal age range of the method?

2) Is there sufficient uranium present in the samples that can be accurately measured?

3)Is there a stratigraphic correlation between samples that permits a correlation between dates of the samples?

4) Are the samples preserved in such a way that the general assumptions of dating are applicable, ie. is the sample closed for the system?

5)Is there a separation between the samples that has a low or predictable Th^{230}/U^{234} ratio? (Kaufman 1964).

Validity of Assumptions.

Antarctic lacustrine carbonates are very good samples for U/Th dating as they validate the assumptions inherent in the dating method.

1) The expected age range of the lake beds in the Marshall Valley extends back to 180,000 years B.P. (Isotope Stage 6). This is within the optimum range of dates obtained by U/Th dating methods (0-400,000 years B.P).

2) The lake beds contain an average of 6 ppm U. This is considered a high concentration for carbonates and the sample is therefore capable of accurate measurement.

3) The Marshall Valley carbonates, present as they are in a stream section, provides a good stratigraphic correlation between carbonate deposits. The carbonate deposits are found interbedded between tills of the glacial sequence.The beds themselves form near continuous layers down the length of the stream section and appear to have been deposited in place and to have remained intact.

4) The deposits appear in general to have been well preserved. One bed does show slight oxidation (evidence of weathering). However it can be generally assumed that the carbonates were covered by drift a relatively short time after deposition which has preserved them in a permanently frozen state. The preservation is important as it confirms that the carbonate, as a system, was closed with respect to time and that there has been no post depositional enrichment or depletion of elements. This is due to the nature of the environment of deposition of the carbonate, cold temperatures and an inactive leaching regime that allows little weathering of deposits.

Summary of the Suitability of Antarctic Carbonates for Dating.

Therefore in general it can be hypothesised that Marshall Valley carbonates should be good subjects for U/Th dating as firstly they contain a high concentration of uranium and are deposited in an environment that suggests little and probably no postdepositional contamination. Secondly they were deposited in a glacial sequence that has a stratigraphy that enables an age correlation up the sequence together with a lateral depositional model along the valley.

Discussion of the Chemical Method used in U/Th dating.

The following is a brief resume of the chemical techniques used. A detailed account of the U/Th separation techniques used in dating is given in Appendix I.

For the dating of Marshall Valley samples an average of 35g of sample was used . The yield of $CaCO_3$ from the sample was > 85% and this together with the " high" concentrations of uranium is the reason for the size of sample. The sample was then digested in concH NO_3 and then a known spike of U^{232}/Th^{228} was added to the sample which was then boiled for ½ hour to ensure that the isotopes in the sample and in the spike had equilibrated . A spike is added because the addition of a known amount of isotope allows for an estimate of the % recovery of U and Th isotopes and therefore a more accurate quantification of the concentrations.

Next the insoluble residues were removed by centrifuging. The U and Th are precipitated by the addition of $FeCl_3$ followed by NH_4OH and the precipitate washed and collected. The addition of Fe forms a scavaging complex of $Fe(OH)_3$ and removes the multivalent anions and cations from solution. The precipitate is then dissolved in 8M HCL and passed through an anion exchange column where by washing with differing molarities of HCl the thorium and then the uranium fractions are stripped from the column. These solutions are then evaporated to dryness and by methods of solvent extraction the Th is separated from other cations and the U from the Fe.

These ions are then complexed with 2-Thenoyltrifluoroacetone (TTA) dissolved in a benzene compound and evaporated onto an aluminium disk before their decay is measured by an Alpha Spectrometer consisting of a surface barrier detecter, biased amplifier and a multichannel analyser. Figures 3.3 and 3.4 show the typical spectra produced by the U/Th method. The isotope ratios are calculated from the printout of the multichannel analyser, and the ages calculated from these as set out below.

Chemical Problems Encountered With U/Th Separation.

Several problems were encountered with this method. The trace element chemistry of the Marshall Valley carbonates appears to vary widely and in general they had a particularly high Ti content so that a white precipitate often formed in the evaporation of the anion solution stage . This Ti precipitate (presumably TiO_2) holds Th and its presence often resulted in poor Th recovery.

The carbonate samples also contained high concentrations of Fe and frequently no $FeCl_3$ was needed to be added at the precipitation stage. This was a problem because high concentrations of Fe clogged the ion exchange columns, requiring the volumes of washings to be increased to a litre, resulting in increased work time at the evaporation stage.

Problems were also encountered at the U purification stage with solvent extraction techniques needing to be repeated to remove Fe . Failure to remove Fe resulted in a bright red TTA complex which when plated onto the disk for counting revealed a "thick source" - The Fe absorbing the energy of the alpha particles being measured and resulting in a spectra that lacked the sharp peaks needed for ratio determination.

The two common problems encountered therefore were the incomplete removal of trace elements leading to energy absorption during counting, and the scavaging of U or Th by undesirable precipitates.

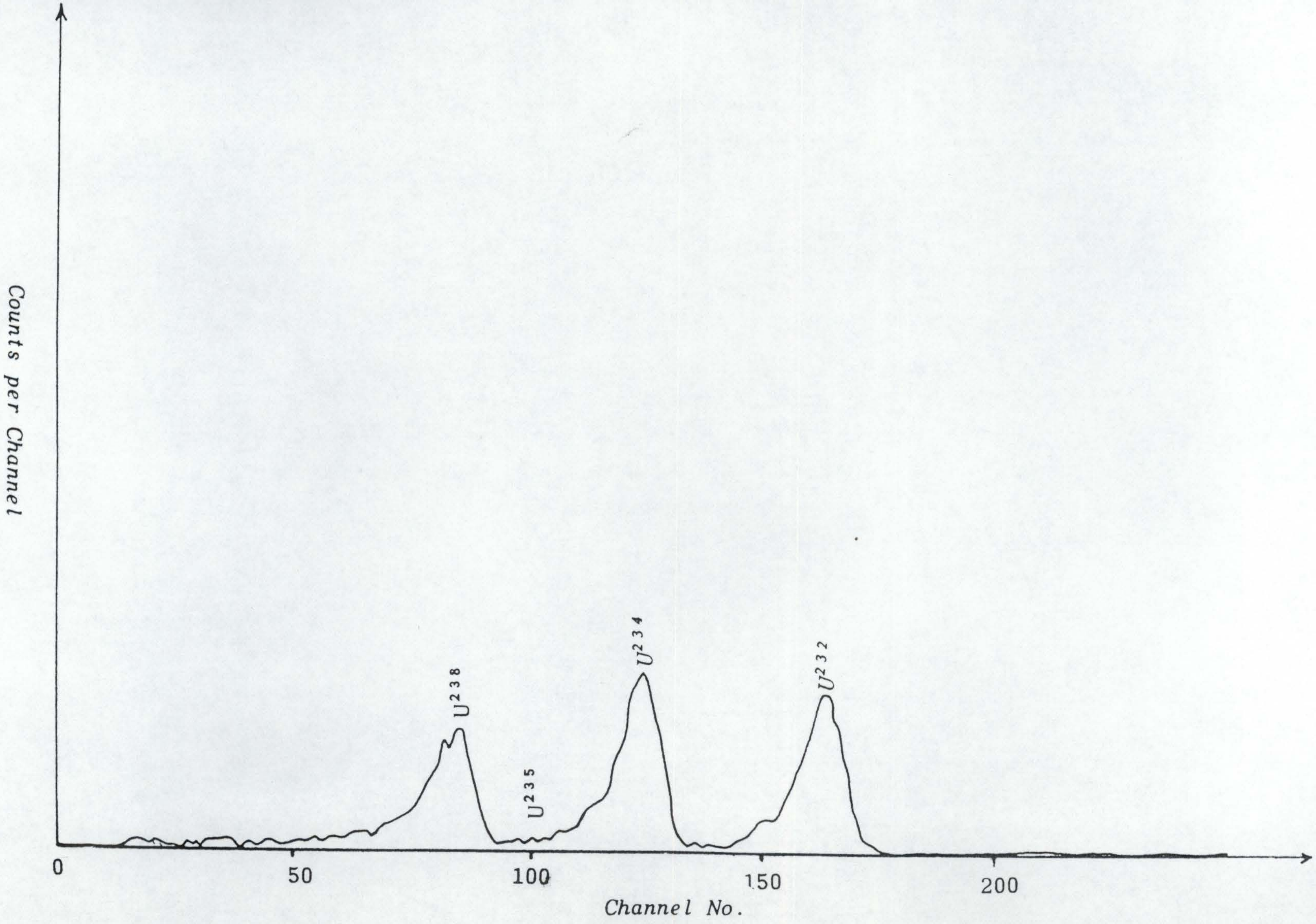


Fig. 3.3 Graph of typical U spectra.

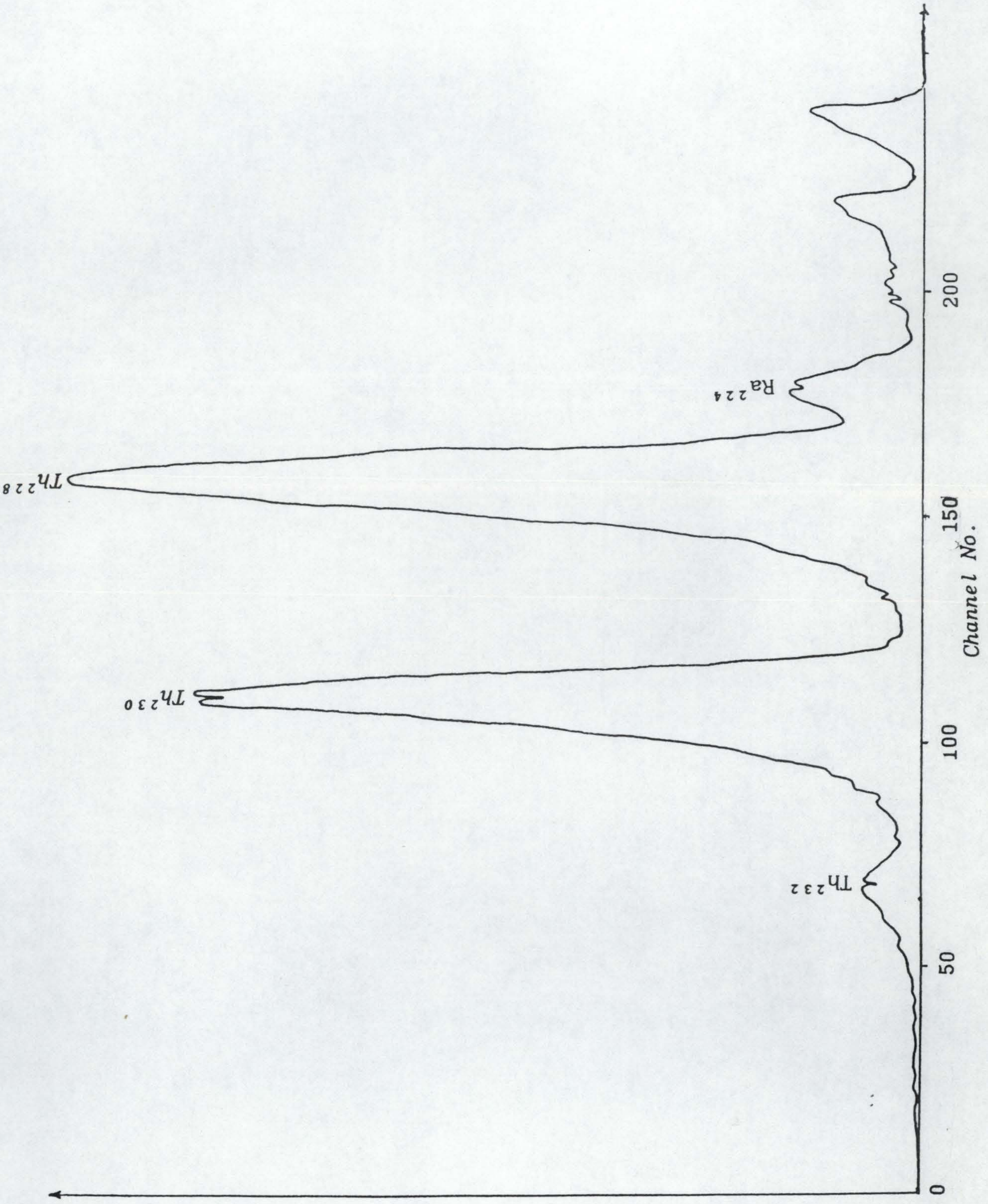


Fig. 3.4 Graph of typical Th spectra.

Interpretation of Radiometric Spectra and Calculation of U/Th dates

Together with spectra produced from the carbonate sample, a background spectra is necessary for each detector. This should be taken every month and ensures that the spectra being analysed has been produced from the radioactive decay from the sample, and not from a faulty detector or from a contaminated detector.

The age of the samples was calculated in the following way (see Fig 3.5);

1) A region of interest was chosen to span each peak produced on the spectra. This should span all the peak above half the maximum value.

2) Datum points within the region of interest, were summed and divided by time. This produces the count rate.

3) The square root of the count rate divided by time gives the standard deviation.

4) The background count rate from the summed region of interest of the sample peak was then subtracted from the sample count rate.

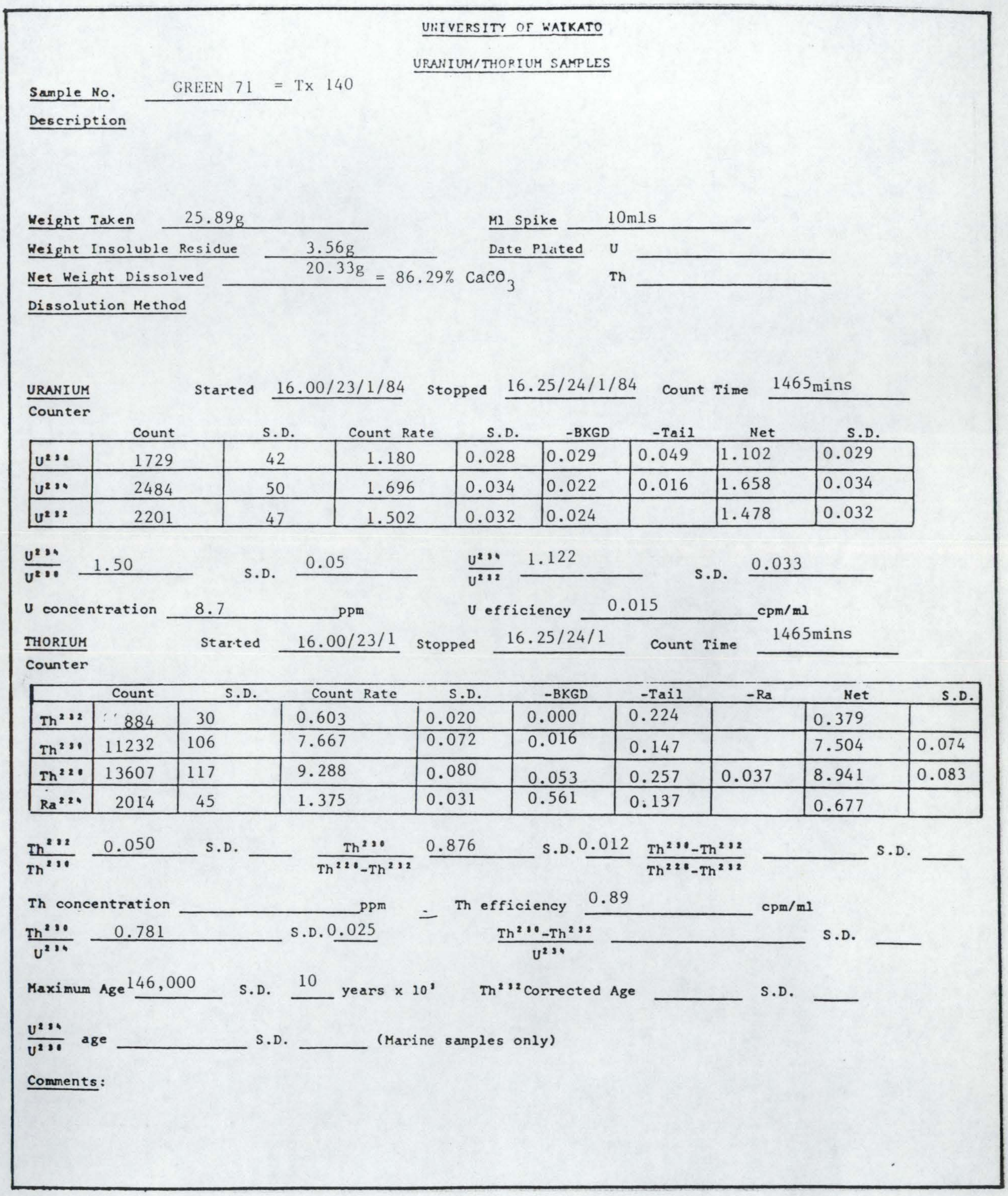
5) The spectra were plotted and extrapolated from each peak to extend under the next (lowest in energy) peak, to estimate the significance of peak tailing.

6) The sum of the tail under the region of interest of the next peak, divided by time, was subtracted from that of the "next peak".

7) The process was repeated for the isotope spectral peaks of U^{238},U^{234},U^{232},Th^{232},Th^{230},Th^{228} and Ra^{224}.

8) 0.054 x Ra^{224} was subtracted from Th^{228} to correct for the low energy α produced by Ra^{224}.

9) The activity of Th^{232} was subtracted from the activity of Th^{228}. This corrects for any Th^{228} already present in the sample before the introduction of the spike as Th^{232} decays to produce Th^{228} as one of its daughters.



UNIVERSITY OF WAIKATO

URANIUM/THORIUM SAMPLES

Sample No. GREEN 71 = Tx 140

Description

Weight Taken 25.89g

Ml Spike 10mls

Weight Insoluble Residue 3.56g

Date Plated U

Net Weight Dissolved 20.33g = 86.29% $CaCO_3$

Th

Dissolution Method

URANIUM Started 16.00/23/1/84 Stopped 16.25/24/1/84 Count Time 1465mins

Counter

	Count	S.D.	Count Rate	S.D.	-BKGD	-Tail	Net	S.D.
U^{238}	1729	42	1.180	0.028	0.029	0.049	1.102	0.029
U^{234}	2484	50	1.696	0.034	0.022	0.016	1.658	0.034
U^{232}	2201	47	1.502	0.032	0.024		1.478	0.032

$\frac{U^{234}}{U^{238}}$ 1.50 S.D. 0.05 $\frac{U^{234}}{U^{232}}$ 1.122 S.D. 0.033

U concentration 8.7 ppm U efficiency 0.015 cpm/ml

THORIUM Started 16.00/23/1 Stopped 16.25/24/1 Count Time 1465mins

Counter

	Count	S.D.	Count Rate	S.D.	-BKGD	-Tail	-Ra	Net	S.D.
Th^{232}	884	30	0.603	0.020	0.000	0.224		0.379	
Th^{230}	11232	106	7.667	0.072	0.016	0.147		7.504	0.074
Th^{228}	13607	117	9.288	0.080	0.053	0.257	0.037	8.941	0.083
Ra^{224}	2014	45	1.375	0.031	0.561	0.137		0.677	

$\frac{Th^{232}}{Th^{230}}$ 0.050 S.D. $\frac{Th^{230}}{Th^{228}-Th^{232}}$ 0.876 S.D. 0.012 $\frac{Th^{230}-Th^{232}}{Th^{228}-Th^{232}}$ S.D.

Th concentration ppm Th efficiency 0.89 cpm/ml

$\frac{Th^{230}}{U^{234}}$ 0.781 S.D. 0.025 $\frac{Th^{230}-Th^{232}}{U^{234}}$ S.D.

Maximum Age 146,000 S.D. 10 years x 10^3 Th^{232} Corrected Age S.D.

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

Comments:

Fig. 3.5 Data calculation sheet used in U/Th dating.

10) The net activity of Th^{230} is divided by that of Th^{228}, U^{234} is divided by U^{232} and U^{234} is divided by U^{238}.

11) The Th^{230}/Th^{228} ratio was divided by U^{234}/U^{232} ratio. This produces the ratio of Th^{230}/U^{234} since the activity of U^{238} and U^{232} in the spike was 1:1.

12) The Th^{230}/U^{234} ratio (the daughter isotope produced since deposition) was plotted against the U^{234}/U^{238} ratio (a measure of the U isotope disequilibria at the time of deposition) on the Kaufman plot (Kaufman, 1964) and the age was obtained including the standard deviation of the ratios which were obtained by;

$$\sigma \text{ ratio} = \sqrt{\sum\left(\frac{\text{Standard Deviation of CR}}{\text{net CR}}\right)^2} \text{ x ratio}$$

$$\text{eg} \qquad \sigma\frac{U^{234}}{U^{238}} = \frac{U^{234}}{U^{238}}\sqrt{\left(\frac{C.\sigma.U^{234}}{CR.U^{238}}\right) + \left(\frac{\sigma.U^{234}}{CR.U^{234}}\right)^2}$$

Problems of Spike Disequilibria.

The spike utilised was purchased from the Radiochemical Centre, Amersham in 1973 as a synthetic mixture of U^{232} and Th^{228} with a 1:1 activity ratio. The spike is stored in 1M HCl and has been diluted 1000 fold to produce a stock solution which in turn has been diluted 1000 fold to produce a working spike. For most samples 10 mls of spike was employed for 35 - 50gms of $CaCO_3$. The isotopic ratio of spike was checked periodically by running 500gms samples of marble.

For the first 34 dates of the 37 dates of this thesis work the same spike U^{232}/Th^{228} was used.However when this volume was exhausted it was necessary to make another volume of spike and so 1 ml of stock solution was diluted to the appropriate volume. The dating process was continued. However on viewing the spectra produced there appeared to be an excess of Th^{230} present. This invalidates the assumption that the initial abundance of the daughter isotope Th^{230} was zero.

Throughout the period that the U/Th process was being used , a marble standard was run at regular intervals. The marble age was outside the dating range of U/Th methods and so gives an infinite age. Because of this we can assume the ratios of U^{234}/Th^{232} were 1:1 and so if the spike was still in equilibrium, then the ratio of U^{232}/Th^{228} should be 1:1. On running the marble standard it was discovered that the ratio U^{232}/Th^{228} was 0.438 $\sigma = 0.014$.

It was hypothesised that the disequilibria was due to absorption of the Th^{228} in the spike onto colloids and so the ratio Th^{230}/Th^{228} was giving an inflated result.

In order to test this hypothesis a quantity of spike was extracted, centrifuged, and the U and Th isotopes separated from the supernatent and solution separately. Each fraction was plated. The spectra of each of the isotopes supernatent and solution were then compared.

The U fraction showed no differences between phases and so it was assumed that the U had remained in solution. However the Th from the centrifuged fraction spectra showed an enrichment relative to the Th supernatent spectra and so it was assumed that the Th had been removed from solution forming colloids. This would be possible if "dust particles" had entered the solution providing particulate matter for the absorption of Th^{232}.

In tracing the source of contamination of the spike , the stock solution that provided the concentrated solution of isotopes was plated. This also showed disequilibria between U^{232} and Th^{228} and therefore it was necessary to remake the stock solution from the original spike and then prepare a new working spike solution.

This spike was also plated and found to be in equilibrium. Another marble standard was run and the results confirmed the equilibria between the spike isotopes.

The problems encountered with spike disequilibria highlights the need for care in the handling of isotope spikes as even minute contamination can have a large effect on isotope equilibria. It also emphasizes the usefulness of running a standard at regular intervals to check spike equilibria.

Results

The results of the U/Th dating of the Marshall Valley Stream Section Lake Beds are presented in Tables 3a and 3b.

The dating of the Marshall Valley lacustrine carbonates was begun in 1977 after the first reconnaissance of the Marshall Valley Stream by Dr. Terry Healy and Prof. George Denton. The dating process was carried out by Mr. Des Dass. Work continued through until 1982 with Mr Theo Vanderlee processing samples. Further trips to the valley resulted in detailed stratigraphic mapping and sampling.

The remaining dates were generated as the result of the 1983-1984 trip to the valley when the fieldwork for this thesis commenced. As the result of this work 37 dates have been obtained from the stream section carbonates.

A previous paper (Dagel *et al*, in press) contains a number of the Marshall Valley Stream Section dates, however this thesis contains the now completed U/Th chronology.

Table 3a presents the set of 37 dates and includes U concentrations and isotopic ratios. The table of results uses the stratigraphy of lake beds assigned by Dagel *et al* (in press) and each separate lake bed has been statistically analysed. A mean age

Table 3a Results from U/Th Dating Marshall Valley Carbonates.

Lake Bed	Unit	Sample	Analysis	U_{ppm}	U_{234}/U_{238}	Th_{232}/U_{234}	Th_{230}/U_{234}	Date	s.d.
A	2 IV*	G82	Tx 93	5.5	1.45	0.08	0.862	180	9
A	6 III	G33	Tx	9.1	1.44	0.00	0.968	240	80
A	6 III	G27	Tx 95	9.6	1.38	0.03	0.862	182	14
A	6 III	G33	Tx 113	8.4	1.43	0.02	0.857	179	5
A	6 III	G47	Tx 114	7.0	1.43	0.03	0.918	217	13
A	6 III	G36	Tx 115	7.9	1.37	0.04	0.901	205	16
A	2 IV	G71	Tx 140	8.7	1.50	0.04	0.781	146	10
A	2 IV	G71	Tx 141	9.4	1.39	0.03	0.778	144	12
A	6 III	G27	Tx 187	6.8	1.51	0.00	0.985	250	20
B	7 I	R181a	Tx 80	14.9	1.44	0.13	0.864	180	10
B	3 III	G30	Tx 97	15.8	1.42	0.03	0.858	180	12
B	3 III	G37	Tx 98	7.7	1.42	0.05	0.857	172	20
B	8 II	G12	Tx 126	6.8	1.76	0.02	0.823	164	15
B	7 I	G23	Tx 180	8.3	1.65	0.15	0.928	204	14
B	7 I	G23	Tx 181	9.2	1.55	0.11	0.814	160	6
B	8 II	G12	Tx 184	6.9	1.62	0.01	0.783	140	11
B	8 II	R171	Qx 13	11.0	1.46	0.05	0.879	190	30
B	8 II	R171	Qx 18	7.0	1.62	0.05	0.930	210	15
C	6 II	C552	Tx 46	8.4	1.39	0.04	0.822	170	10
C	6 II	C552	Tx 76	7.1	1.55	0.03	0.809	157	6
C	6 II	G8	Tx 164	6.7	1.56	0.18	0.707	121	4
C	6 II	G4	Tx 165	8.4	1.56	0.02	0.704	120	5
C	6 II	G4	Tx 182	---	1.53	0.02	0.831	162	11
C	6 II	G8	Tx 183	---	1.59	0.02	0.769	138	7
D	5bII	G6	Tx 99	11.4	1.51	0.03	0.782	145	10
D	5bII	G13	Tx 100	8.9	1.42	0.05	0.796	145	16
D	5bII	G14	Tx 101	7.2	1.44	0.04	0.827	164	11
D	Note I	82/120	Tx 222	1.7	1.50	0.03	0.841	170	8

Table 3a (cont) Results from U/Th dating Of Marshall Valley Carbonates.

Lake Bed	Unit	Sample	Analysis	U ppm	U_{234}/U_{238}	Th_{232}/U_{234}	Th_{230}/U_{234}	Date	s.d
E	1 III	G45	Tx 102	33.0	1.37	0.02	0.730	130	---
E	1 III	G40	Tx 160	21.9	1.37	0.02	0.730	130	3
E	4 II	G18	Tx 176	11.9	1.51	0.03	0.760	137	10
E	4 II	G18	Tx 177	13.6	1.41	0.04	0.689	120	10
E	Note 2	G69	Tx 223	20.4	1.45	0.03	0.797	150	12
E	Note 2	G69	Tx 225	28.7	1.44	0.02	0.713	127	6
F	2 I	183A	Tx 81	2.7	1.66	0.19	0.262	32	2
F	2 II	R174	Qx 12	4.8	2.80	0.25	0.400	<15	-
F	2 II	R178	Qx 22	3.0	2.19	0.10	0.140	11	3

Note 1 Samples taken from south side of stream adjacent to Section III, and underlying a lake bed equilvalent to 1III

Note 2 Surficial bed between Sections III and IV, laterally traced to Unit 1III.

* Equilvalent exposure to Section IV but on south side of stream.

Table 3b.Comparisons of statisically meaned dated for Marshall Valley Carbonate Lake Beds

Dagel's Lake Beds - U/Th dating.

Bed	$\bar{x}\frac{U_{234}}{U_{238}}$	$\sigma\frac{U_{234}}{U_{238}}$	$\bar{x}\frac{Th_{230}}{U_{234}}$	$\sigma\frac{Th_{230}}{U_{234}}$	Date	σ
A	1.43	0.05	0.879	0.073	190	36
B	1.56	0.12	0.860	0.049	174	22
C	1.53	0.07	0.774	0.057	142	19
D	1.46	0.05	0.802	0.023	154	8
E	1.44	0.05	0.732	0.040	129	12
F	2.22	0.57	0.267	0.130	33	18

Judd's Lake Beds - U/Th Dating.

Bed	$\bar{x}\frac{U_{234}}{U_{238}}$	$\sigma\frac{U_{234}}{U_{238}}$	$\bar{x}\frac{Th_{230}}{U_{234}}$	$\sigma\frac{Th_{230}}{U_{234}}$	Date	σ
A-B	1.49	0.12	0.869	0.061	182	28
C-D	1.51	0.07	0.809	0.095	155	38
E	1.44	0.05	0.732	0.040	129	12
F	2.22	0.06	0.256	0.132	33	18

and standard deviation is given for each bed. The comparison of averaged dates is listed in Table 3b.

Dagel et al (in press) has recognized 6 separate lake beds in the Marshall Valley Stream Section, the first lake bed being present during a glaciation that occurred 190,000 σ 36,000 years B.P. The next 4 lake beds are suggested as evidence of another 4 proglacial lakes before 130,000 years B.P. These lake beds are dated at 174,000 years σ 24,000 years B.P., 142,000 years σ 19,000 years B.P.,154,000 σ 8,000 years B.P., and 129,000 σ 12,000 B.P. respectively.

The dating of the final lake bed at 33,000 years σ 19,000 years B.P. suggests that this lake bed is representative of a later glaciation. The U^{234}/U^{238} isotope ratio of the gypsum of this bed is radically different to all other Marshall Valley lake beds, but falls within the range found in the Ross Sea I Lake Bed occupying the Miers Valley (Clayton-Greene, 1986) and dated 10,000-17,000 years B.P.

Dagel *et al* (in press) therefore suggests that on the basis of U/Th dating 2 major periods of glaciations can be recognised. These have been assigned to the Ross Sea I and II Glaciations as their chronologies date them at Isotope Stage 2 and 6 which correlates with eustatic low sea levels (see Fig. 2.3,Chapter 2).

We agree with the hypothesis that the Marshall Valley Stream Section represents Ross Sea I and II Glaciations, however we suggest that there is evidence (chronological, mineralogical and stratigraphical) that the glacial history within these glaciations can be simplified.

We suggest that rather than 6 separate lake bed layers representing 6 lake occupations of the valley, it is possible to show that some of these lake bed layers are continuous and therefore represent the same lake bed and lake occupations.

On the basis of statistical analysis, some of Dagel's lake beds, show little variance in age between each lake bed and the meaned data have large standard deviations. It is suggested on this evidence that lake beds A plus B and lake beds C plus D if considered as composite lake beds produce results that have a higher statistical validity. These results are presented in Table 3b and the beds are labelled to represent their new combinations, eg lake bed A-B is the composite of Dagel's lake bed A and lake bed B. These are referred in the text as Judd's lake beds to differentiate them from the previous paper.

Therefore we suggest that the Marshall Valley Stream Section contains lake beds that represent the last two Ross Sea Glaciations.

The penultimate glaciation, Ross Sea II, occupied the valley for a period of time from 130,000 to 180,000 years B.P. Within this timespan we find 3 separate lake beds that represent 3 proglacial occupations of the valley. These lake beds are dated at 182,000 years σ 28,000 years B.P. , 155,000 years σ 38,000 years B.P., 129,000 years σ 12,000 years B.P. respectively. The lake beds are separated by glacial drift and the accuracy of dating allows valid statistical separation of their ages.

The simplification of the stratigraphy of the sections can be explained by lake beds with similar age but stratigraphic separation representing minor fluctuations of the advance and retreat of the glacier ice within the valley during a glacial episode.

An example of this can be seen at Section II (see Appendix II) where lake bed D overlies lake bed C. On the basis of U/Th dates there is little statistical variation between the ages of the lake beds (lake bed C = 142,000 years σ 19,000 years B.P. , lake bed D = 154,000 years σ 8,000 years B.P.).

However by considering that the two layers represent a fluctuation of the proglacial lake it is plausible to combine the two layers to represent one lake bed.

Similarly in Section III the stratigraphy of the two closely dated lake beds A and B (190,000 years σ 36,000 years B.P. and 174,000 years σ 22,000 years B.P. respectively) have been explained as representing a single lake bed deformed by a complex glaciotectonic thrust fault (Dagel *et al*, in press).

We agree with Dagel's interpretation of a single stratigraphic unit, however we suggest that the complex stratigraphy of Section III can be explained by post-depositional deformation of the lake bed.

Further evidence for a simplified stratigraphy of lake beds is discussed in later chapters on the mineral composition of the lake beds and the fieldwork from the 1985-1986 season.

The final glaciation, Ross Sea I, is represented in the Marshall Valley Stream Sections by lake bed F. This bed has a mean age of 33,000 years σ 19,000 years B.P. The U/Th dating was carried out on samples of gypsum (no carbonates are found in the upper stratigraphy). This age correlates with radiocarbon dates on fossil algae from the ice-cored moraines (see Fig.2.1, Chapter 2). The U/Th dates represent the presence of a proglacial lake, the radiocarbon dates document the downwasting of the ice during the recession of the glaciation.

In summary: U/Th dating has provided age control on the glacial stratigraphy of the Marshall Valley Stream Section. The identification and dating of proglacial lake beds documents the history of the glacial ice occupation within the valley. On a statistical basis it is possible to interpret the sequence of lake beds in terms of the advance and retreat of glacial episodes. As a result a glacial geochronology of Marshall Valley for the last 180,000 years is possible.

CHAPTER 4

ANALYSIS BY X.R.D. AND S.E.M.

Chapter 4

Introduction.

In this chapter the methodology of X-Ray Diffraction and the results obtained by this technique are discussed. X-Ray Diffraction has been used to estimate the percentage by weight of the two polymorphs, calcite and aragonite present in our carbonate samples.

X_Ray Diffraction Analysis and Standardization.

The method used was developed by Hume and Nelson (1982) and is applicable to the analysis of carbonates. The initial study of XRD analysis of Marshall Valley carbonates was undertaken by Mark Lawrence as part of his M.Sc thesis. Further work was completed by Alison Utting for an undergraduate research exercise.

The analysis was carried out on a Philips P.W. 1050 X-Ray Diffractometer using copper radiation ($\lambda = 1.54$ Å) at machine settings given below.

kV	36
mA	16
Time Constant	4
Chart Speed	20 mm min $^{-1}$
Scanning Speed	2° 20mm min $^{-1}$
Ratemeter Settings	4(400 cps)

Machine Settings for XRD Analysis

Samples were prepared for analysis by grinding in a ringmill until a silty texture was obtained, followed by grinding by hand with a mortar and pestle until the silty texture had been lost. The powder sample obtained was back-packed into a Philips aluminium sample holder, thereby avoiding direct pressure or handling of the sample face that was to be exposed to the X-rays. By this method a randomly orientated mount of crystallites is achieved, which is necessary for quantitative analysis.

Calibration of the machine is achieved by the use of a permanent quartz standard. A sample of pure quartz was ground to less then 45 μm particle size. The sample was mixed with an Araldite resin which was set in a Philips aluminium sample holder and ground flush with the sample holder.

Calibration data for the standard is provided below and after a warmup period of 30 mins, the machine settings are adjusted so that the operating conditions closely matched the calibration values. The standard if then run between the 2θ angles of 20.00°-21.50° should produce the required quartz peak at 20.8°2θ with a peak intensity of 37 ± 2.

Peak Position		Ratemeter Setting	Peak Height above Background	Background Height
°2Θ	dÅ			
20.8	4.26	RMS4	37	17
26.6	3.34	RMS10	88	4

Calibration Data For Quartz Std. used in XRD Analysis.

The carbonate samples were then run and their diffractograms produced.

The standard curves which relate mineral abundance to peak intensity were constructed by Nelson and Cochrane (1970) from powdered aragonite and calcite admixtures in known proportions. The curves are reproduced in Figs. 4.1 and 4.2.

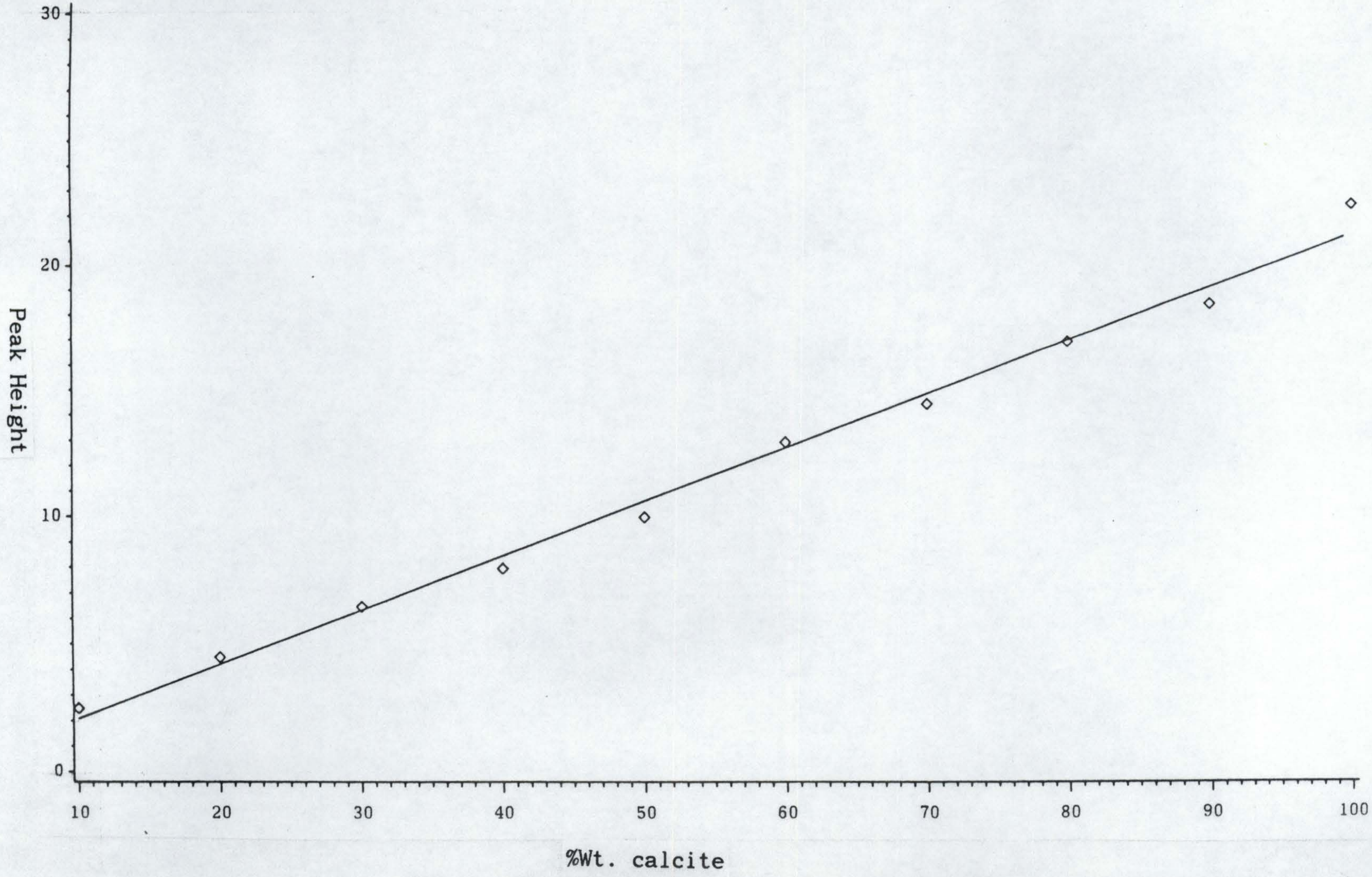


Fig. 4.1 Standard curve for calcite in XRD analysis

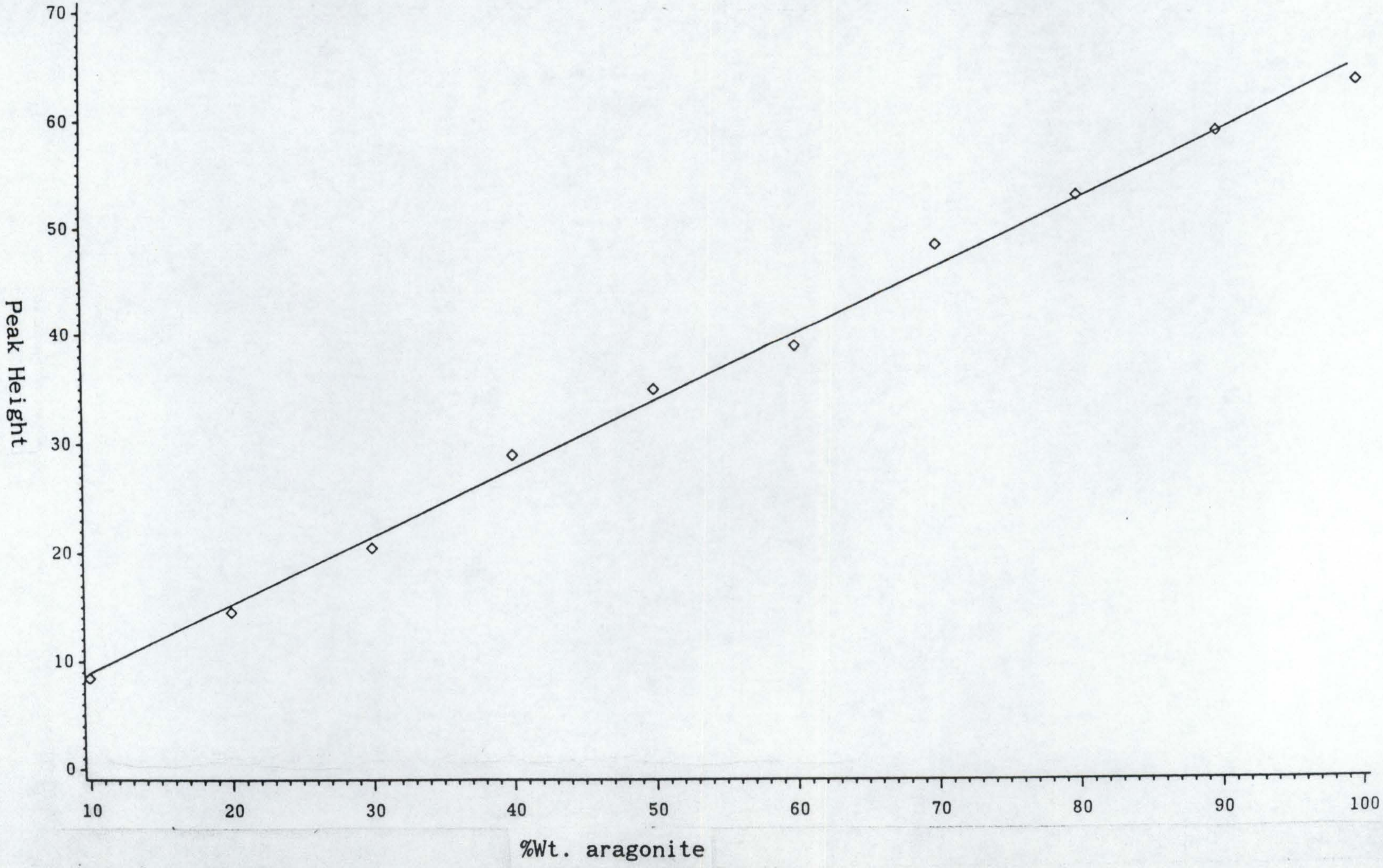


Fig. 4.2 Standard curve for aragonite in XRD analysis

The aragonite shows a linear relationship between peak height and mineral abundance. However the calcite plots show a curved relationship instead.

Diffractogram Interpretation.

For the analysis of calcite and aragonite it was only necessary to run the X-ray diffractometer between the 2θ angles of 25.50°-30.50° at the chart speed of 20mm/min. This is because the major aragonite peak is present at 26.2° and the major calcite peak at 29.3°.Each sample was run three times for statistical purposes to determine average peak height intensity corrected for background.

On each diffractogram a straight background line was drawn. The peak height intensity is defined as the vertical distance between peak apex and the background line and is expressed as the number of small grid units on the paper chart. The peak height intensity is then read off the standard curve to determine the mineral concentration of the samples.

Results

The accuracy of the technique varies with the mineral type and the machine settings for the standard (Hume and Nelson, 1982) and this trend was noticeable in the analysis of our samples. The error involved in this technique is estimated to be in the region of ± 10%.

The results of X-Ray diffraction are tabulated in Table 4a and show that the majority of carbonate samples contain both calcite and aragonite.

The samples analysed appear to yield more than 100% carbonate, possibly because they have retained a greater degree of crystallinity then the standards.

Conclusions.

The weight percentages of calcite and aragonite by X-ray diffraction shows distinct differences between units of the proposed same age along the sections of Marshall Valley.

The lake beds A and B (Dagel *et al*,in press) are dated at 190,000 σ 36,000 years B.P. and 174,000 σ 22,000 years B.P. respectively. In Chapter 3 it was suggested that these layers could collectively be considered one bed (Lake Bed A-B).

Comparisons of the wt.% calcite vs. aragonite ratios show that the beds have a similar composition. Lake bed A has a mean wt.% composition of 75% σ 21% calcite: 39% σ 24% aragonite. Lake bed B has a mean wt.% composition of 75% σ 24% calcite: 25% σ 9% aragonite. Thus lake bed A-B has a mean composition of 73% calcite : 32% aragonite.

The lake beds C and D (Dagel *et al*,in press) are dated at 142,000 years σ 19,000 years B.P. and 154,000 σ 8,000 years B.P. respectively. It is suggested that these lake beds should collectively be considered lake bed C-D.

These two lake beds also exhibit a similarity in composition. Lake bed C has a mean wt.% composition of 52% σ 19% calcite: 42% σ 18% aragonite. Lake bed D has a mean wt.% composition of 53% σ 22%: 45% σ 26% aragonite. Lake bed C-D therefore has a mean composition of 53% calcite : 44% aragonite, which is distinctly different from the composition of lake bed A-B. Thus this bed which appears in outcrop as a fragmented and disturbed bed, cannot have been derived from the underlying lake bed A-B.

These results confirm the conclusions reached with U/Th dating, that lake beds A and B appear to have a common origin, as do lake beds C and D. Their present superposition in Section III (for beds A and B) and in Section II (for beds C and D) must therefore result from

Table 4.a XRD results for Marshall Valley carbonates.

Sample	Cement?	Wt% Aragonite	Wt% Calcite
G 1	N	28	100
G 2	N	24	100
G 3	N	58	73
G 4	N	27	61
G 5	N	60	29
G 6	N	25	67
G 7	N	46	34
G 8	N	24	65
G 9	N	39	52
G10	N	20	64
G11	N	29	74
G12	N	11	85
G13	N	82	29
G14	N	27	65
G16	N	60	30
G17	N	83	15
G23	Y	84	6
G24	Y	100	15
G27	N	46	100
G30	N	33	54
G33	N	27	56
G37	N	29	55
G41	N	22	47
G42	N	25	46
G44	Y	26	18
G45	N	78	2
G46	Y	100	16
G47	Y	36	52
G48	Y	74	4
G49	Y	40	7
G50	N	35	57

Table 4a.(cont) XRD results for Marshall Valley carbonates.

Sample	Cement?	Wt% Aragonite	Wt% Calcite
G51	Y	100	5
G52	Y	100	5
G53	N	99	5
G54	N	79	4
G55	N	49	92
G56	Y	68	25
G57	Y	23	48
G60	N	39	100
G61	N	11	52
G63	N	21	100
G64	Y	6	63
G65	Y	20	46
G67	N	9	58
G68	Y	4	61
G69	N	100	8
G70	N	100	5
G75	Y	100	7
G76	Y	89	1
G78	N	100	1
G80	Y	19	64
G82	N	21	49
G85	N	100	83
G87	N	26	100
G88	Y	18	42
G91	N	36	100
G92	N	49	100
G96	N	23	100
G97	N	31	100
G98	N	100	58

postdepositional events within the valley. The possibilities for these events and the pertinent field evidence are presented in subsequent chapters.

The lake bed E (Dagel *et al*, in press) has been dated at 129,000 σ 12,000 years B.P. It is suggested that because of this date, lake bed E is a separate lake bed. The mean wt.% composition of 5% σ 2% calcite: 94% σ 11% aragonite supports this suggestion.

Furthermore, distinct differences in the calcite to aragonite ratios in these carbonates allows clastic carbonates contained within drift or cropping out on the valley floor to be identified. Therefore the XRD analysis (Tables 4b and 4c) provides further evidence for the proposed simplification.

Lake bed F has been dated at 33,000 years σ 18,000 years B.P. This also suggests that lake bed F is a separate lake bed. The bed however only consists of gypsum, and no carbonate has been found to enable the calcite/aragonite ratios to be examined.

With the assumption that the lake bed stratigraphy can be simplified, there is a stratigraphic trend in the wt.% composition of the revised lake beds (see Table 4b and 4c). The wt.% composition of the lake beds changes from a calcite dominant mineralogy to an aragonite dominant mineralogy as the stratigraphy becomes younger.

There are three possible reasons for this trend.

1) The trend may represent a change in the mode of deposition of the lake beds in relation to the depth of the lake in which they formed. Lake bed A-B may have formed in the deepest lake where calcite was able to form slowly at the expense of the thermodynamically less stable aragonite. This may also relate to a deeper lake allowing a long period of deposition before evaporation of the lake. Lake bed C-D may have formed in a lake of shallower water where the time and quantity of calcite formation may have been decreased. Lake bed E may

Table 4b. Dagel's Lake Beds.

Lake Bed	Wt% Aragonite	Wt% Calcite
A	39% σ 24%	71% σ 21%
B	25% σ 9%	75% σ 24%
C	42% σ 18%	52% σ 19%
D	45% σ 26%	53% σ 22%
E	94% σ 11%	5% σ 2%

Table 4c. Judd's Lake Beds.

Lake Beds	Wt% Aragonite	Wt% Calcite
A-B	35% σ 24%	73% σ 27%
C-D	44% σ 21%	53% σ 19%
E	94% σ 11%	5% σ 2%

have formed in a very shallow lake, where it formed not by the previous mode of deposition of the other lake beds but in more evaporitive conditions that favoured aragonite deposition.

2) The trend in composition may be related to the stratigraphy of the sections. Lake beds A-B and C-D appear to have been buried a short time after deposition. Samples of these lake beds show little evidence of oxidation caused by weathering. Lake bed E however appears near the surface of the section and therefore may have been exposed to some weathering for 130,000 years (Plate 4.0). The change in composition may be related to post-depostional conditions that the lake beds have experienced.

3) All three lakes may have originally deposited aragonite, but with increasing age a large percentage of this may have recrystallised to form the more stable calcite.

A further discussion of this trend in mineralogy is continued in the section of work on Scanning Electron Microscopy.

It is also apparent in Table 4a that the carbonate cements have a distinct mineralogy consisting largely of aragonite.

In summary:There is a broad trend between the lake beds and there is sufficient differences between lake beds as to allow a check of stratigraphic correlation between dated samples of lake beds.

The stratigraphy of the dated lake beds agrees with the stratigraphy of the sections derived from XRD analysis with one exception.

The results of XRD analysis creates uncertainty as to the reliability of U/Th dates obtained from lake bed unit 7I. As already discussed in Chapter 3 on U/Th dating, the dates and stratigraphy of Section 1 suggests lake bed unit 7I should give an approximate age of between 130-160,000 years and correlate with lake beds of these ages present further up the stream at Section II.

Plate 4.0 Signs of oxidation in lake bed E (photo C Hendy)

If we consider the XRD results we find further evidence to substantiate this hypothesis. If lake bed unit 7I was to be representative of a 180,000 year glaciation and correlate with lake beds A and B then we would expect it's composition to show an enrichment of calcite over aragonite.

However the analysis of Unit 7I shows that its composition is largely aragonite and so this further suggests that it should represent lake bed E with its similar composition and therefore younger age.

Further discussion of this stratigraphic problem is introduced in Chapter 5 containing 1986 field work.

Scanning Electron Microscopy Analysis.

Introduction.

Examination of the Marshall Valley lake beds by Scanning Electron Microscopy was undertaken to further determine the mineralogy of the deposits and to provide evidence for the mechanism which has produced lake beds with a mixed calcite/aragonite mineralogy. S.E.M. provides information on the shape and orientation of the calcite and aragonite crystals within the lake bed. This information can be used to understand the mechanisms and the sequence of deposition of the lake beds. S.E.M. can also be used, as a further check, that separate samples from different sections represent the same lake bed unit. This can also be used to test the correlation between lake beds in a section and the geochronology of the entire stratigraphy.

The initial Scanning Electron Microscopy study was undertaken by Miss Alison Utting as part of an undergraduate study. As part of the experimental work for this thesis , several previously unstudied samples were analysed. These samples were representative of each of the lacustrine lake beds occurring within the Ross Sea II Glaciation sequence. Several of the samples were taken from the south side of the Marshall Valley Stream. This was to check the assumed correlation to the lake beds revealed in the studied sections of the north side of the stream.

Previous Work.

The previous study (Hendy and Utting,1983) has revealed that the lacustrine lake beds have a mixed mineralogy of calcite and aragonite in varying proportions. The broad trend in mineral composition suggested by XRD analysis can be identified in the S.E.M. analysis. Therefore it is possible to check the correlation between samples of one unit and between samples of differing units.

Hendy and Utting (1983) suggest that the mode of deposition of the lake beds can be deduced from the S.E.M analyses. The lacustrine lake beds of mixed mineralogy contain long needlelike crystals of aragonite and rhombic crystals of calcite. The wt.% composition and size of these crystals varies between the lake beds. The aragonite needles have a random orientation and exhibit no common centre of growth, the calcite crystals appear to grow to an ordered orientation and to often appear as overgrowths enclosing the aragonite needles. The degree of overgrowth also varies between lake beds.

Hendy and Utting (1983) suggest that on the basis of the morphology of the crystals, the lake beds were formed firstly by the deposition of aragonite and then by the deposition of calcite. Because of the random orientation of the aragonite crystals it is suggested that they did not form *in situ* , but formed as a precipitate in the water column that produced a "rain" of needles onto the lake floor.

Later under less saturated conditions the calcite crystals grew in place and enclosed the aragonite needles.

The examination of the aragonite needles reveals few cases of pitting and therefore Hendy and Utting (1983) suggest that the calcite crystals do not form as the result of aragonite recrystallisation.

S.E.M. analysis of the cements found in the Marshall Valley Stream Sections (Hendy and Utting,1983) suggests that they have a different mode of deposition. XRD results show that these deposits consist almost entirely of aragonite. This is confirmed by S.E.M. which further reveals that unlike the lacustrine carbonates, the aragonite crystals found in the cements have a radiating acicular habit. This orientation of crystals suggests that the mineral aragonite grew *in situ*. Therefore it is suggested that the cement was deposited in an evaporitive environment and grew subaerially.

Results From The Experimental Work Of The Thesis.

In this experimental work of this thesis samples from each Ross Sea II lake bed were analysed.

Lake Bed A-B

Sample Green 37 previously unstudied by S.E.M. was selected. The sample was taken from Section III and was a hard tabular lacustrine calcium carbonate. The sample used had a thickness of approximately 1 cm. The sample was representative of lake bed A-B, an extensive layer found in all sections. The lake bed has a mean age of 182,000 years σ 28,000 Years B.P. and XRD analysis shows that it has a mean wt.% composition of 55% calcite: 29% aragonite. In preparation for mounting, the sample was cut into 1 cm cubes before being randomly sliced horizontally and vertically at varying proportions. By this method it was proposed to study any differences in deposition within the lake bed layers.

Green 37 appears to exhibit the morphology of crystals characteristic of the unit. At low magnification (300x) the random orientation of aragonite crystals is clearly shown (Plate 4.I). Plate 4.I shows a vertical section through the sample. At increased magnification (550x) another vertical section (Plate 4.II) shows aragonite crystals with large overgrowths of calcite. The calcite rhombs are well formed and much larger in comparison to the aragonite needles. At high magnification (900x) the overgrowths of calcite can be clearly seen in Plate 4.III. Plate 4.III shows clearly aragonite needles enclosed in calcite rhombs. The aragonite needles show a random orientation.

Plate 4.IV shows the top surface of the sample. The random array of broken aragonite needles can be seen and the surface also has a layer of subhedral calcite crystals. The shape of the crystals suggests some degree of weathering present at the surface with recrystallisation to form the small imperfect crystals found at the surface.

The composition of the lake beds are not purely calcium carbonate. S.E.M analysis showed the presence of detrital grains and micro-organisms. Plates 4.V and 4.VI have detrital grains present in them surrounded by the calcium carbonate matrix. The detrital grain in Plate 4.V is identified as probably being basaltic in origin. This due to the circular vesicles that are present in its surface. The detrital grain in Plate 4.VI cannot be so easily identified but due to its sharply cut angular surfaces it is suggested that it is a quartz sand grain.

Plate 4.I Randomly orientated aragonite needles (300x).

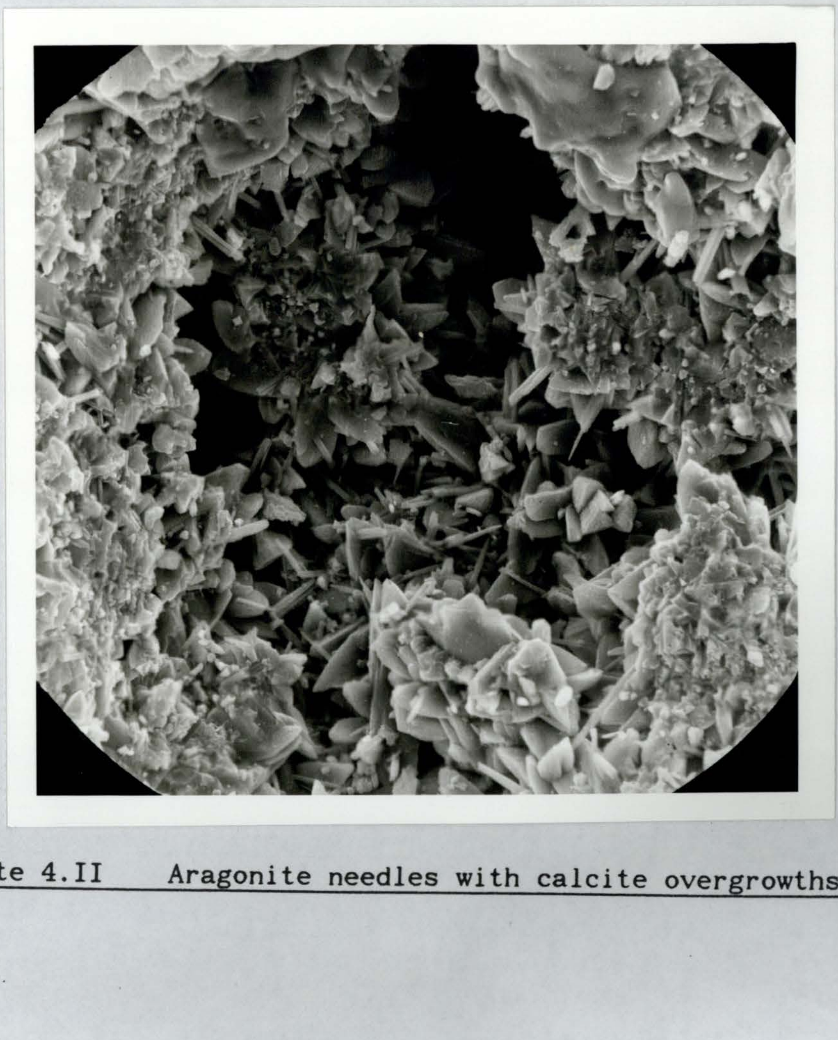


Plate 4.II Aragonite needles with calcite overgrowths (550x).

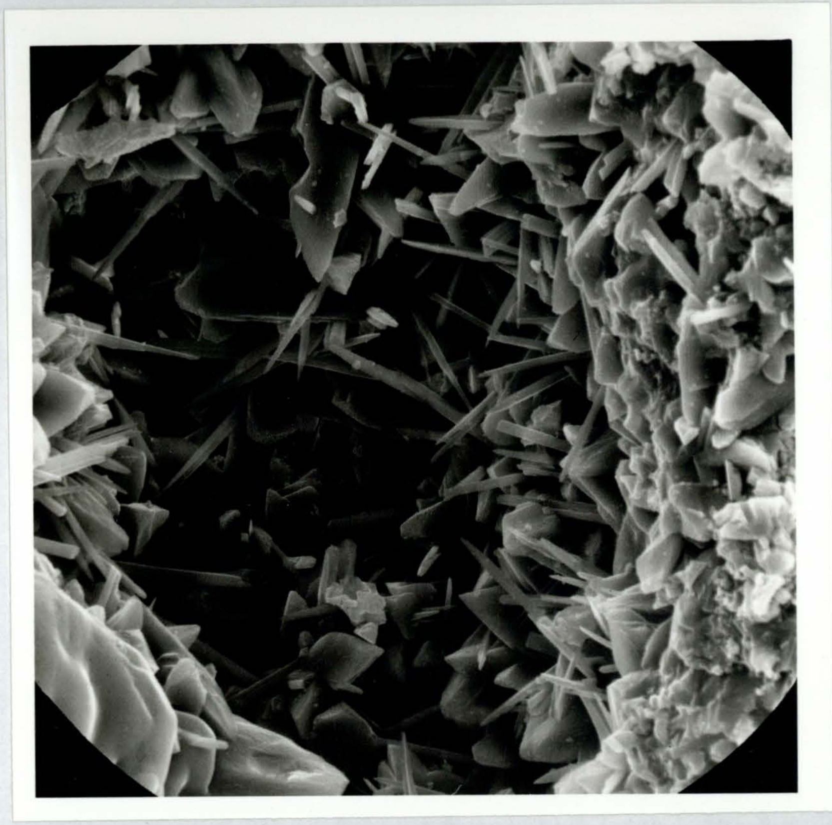


Plate 4.III Overgrowths of calcite around aragonite needles.(900x).

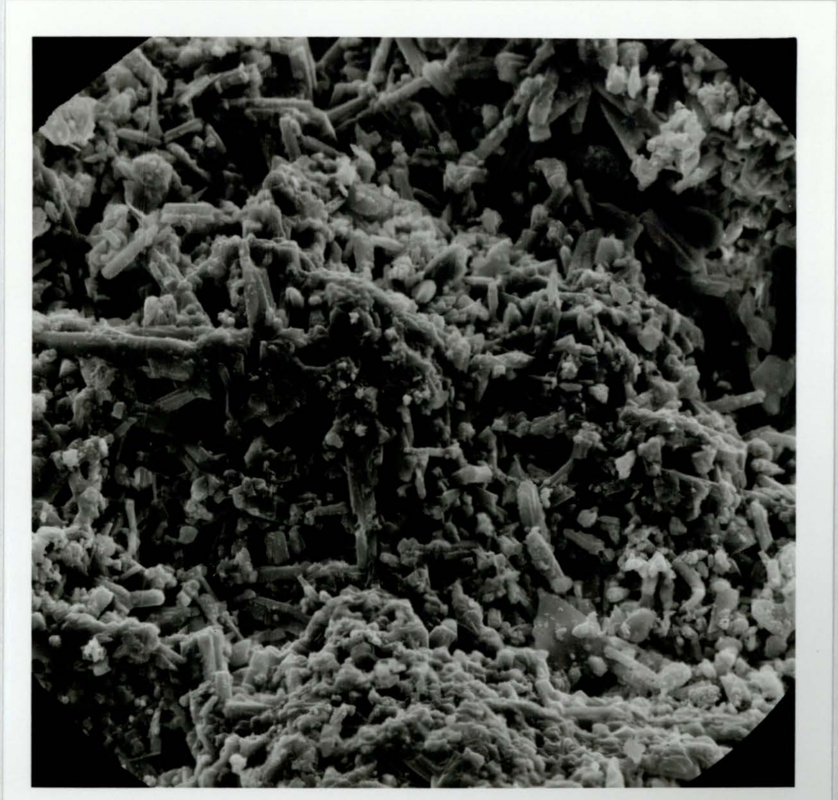


Plate 4.IV Broken array of aragonite needles and subhedral calcite rhombs (600x).

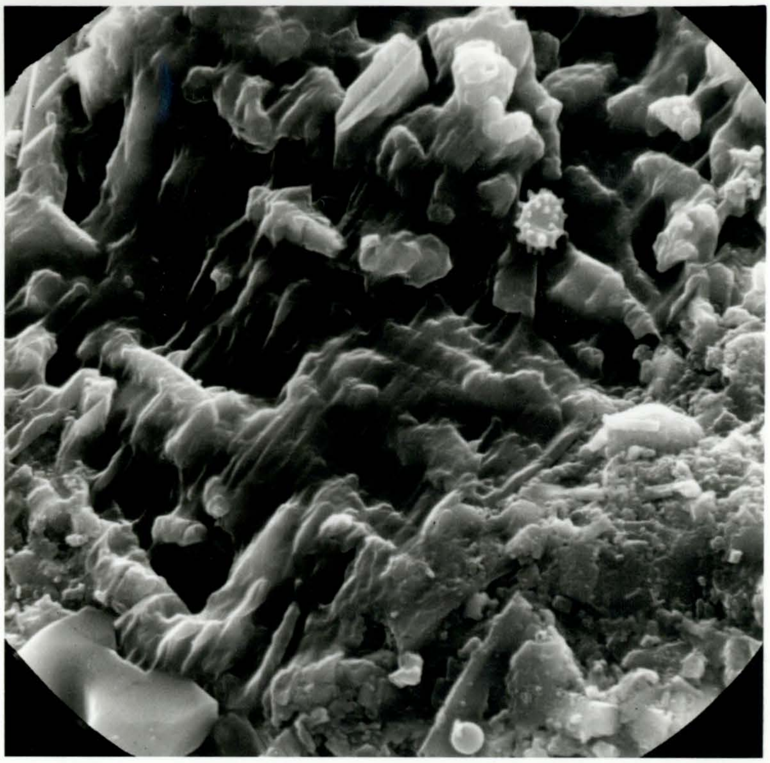
The presence of detrital grains in lake sediments is not unlikely. We suggest that the detrital grains are part of the wind blown "loess like" component of the glacial system. This would explain the angular abrasion of the particles. The present of basalt rich detritus is also to be expected. This material forms the bulk of the Ross Sea glacial sediments and is the result of the circulatory flow of glacier ice over the Ross Island Volcanic system.

It is possible that a proportion of the wind blown detritus will become incorporated in the lake ice and then melt out to become incorporated in the lake sediments.

Plates 4.VII and 4.VIII show the remains of biological life that existed in the photic zone beneath the ice. These also have fallen through the water column to be incorporated in the lake beds.

In summary sample Green 37 shows a morphology similar to that suggested by Hendy and Utting (1983) for samples of the same lake bed. In general lake bed A-B has a morphology of long thin needles of aragonite enclosed in large rhombs of calcite. The overgrowth of the crystals is widespread and suggests that the deposition of the calcite around the aragonite crystals occurred over a long period of time. This implies that a large lake occupied the valley during this period.

There is very little weathering present in the sample with the exception of the upper surface, where only slight weathering can be seen. This suggests that the lake was not subaerially exposed for a long period following deposition. It appears that the glacial sediments topping the lake beds were deposited in a period contemporary to lake bed deposition.



0·6µm

Plate 4.VIII Remains of biological life found in the lake bed. (2,000x).

0·6µm

Plate 4.VII Remains of biological life found in the lake bed. (2,000x).

Lake Bed C-D.

The samples analysed representing this lake bed were 82/120 and 82/103. Sample 82/120 was from a section uncovered on the south side of the stream and sample 82/103 was from a newly uncovered section on the north side of the stream. This was as a result of field work in the 1982/1983 season.

XRD results show that this lake bed also has a mixed mineralogy but that a slightly higher concentration of aragonite is expected then in Lake Bed A-B. The mean wt% composition of lake bed C-D is 53% calcite: 44% aragonite.

Plate 4.IX at 1,100x magnification shows the characteristic random array of aragonite needles, however there is a smaller proportion of calcite crystals present than in the previous lake bed. Many of the blunt square shaped crystals in the photograph appear to be sheared aragonite crystals with square faces.

Plate 4.XI in low magnification (650x) shows the radiating habit of the calcite crystals. These photos show that the calcite has grown *in situ*.

Plate 4.XI at the high magnification (1,500x) shows both crystals together. The calcite crystals are not as large and well formed as the previous lake bed.

Plate 4.XII at very high magnification (12,000x) shows the faces of an aragonite needle. There is no sign of any pitting or other signs of weathering.

In summary: Samples of lake bed C-D show a similarity in unweathered composition to lake bed A-B however the calcite crystals appear smaller and have less well defined shapes than lake bed A-B. The overgrowth of aragonite crystals by calcite crystals is not as common as in the previous lake bed.

Plate 4.IX Random array of aragonite needles. Many are sheared (1,100x).

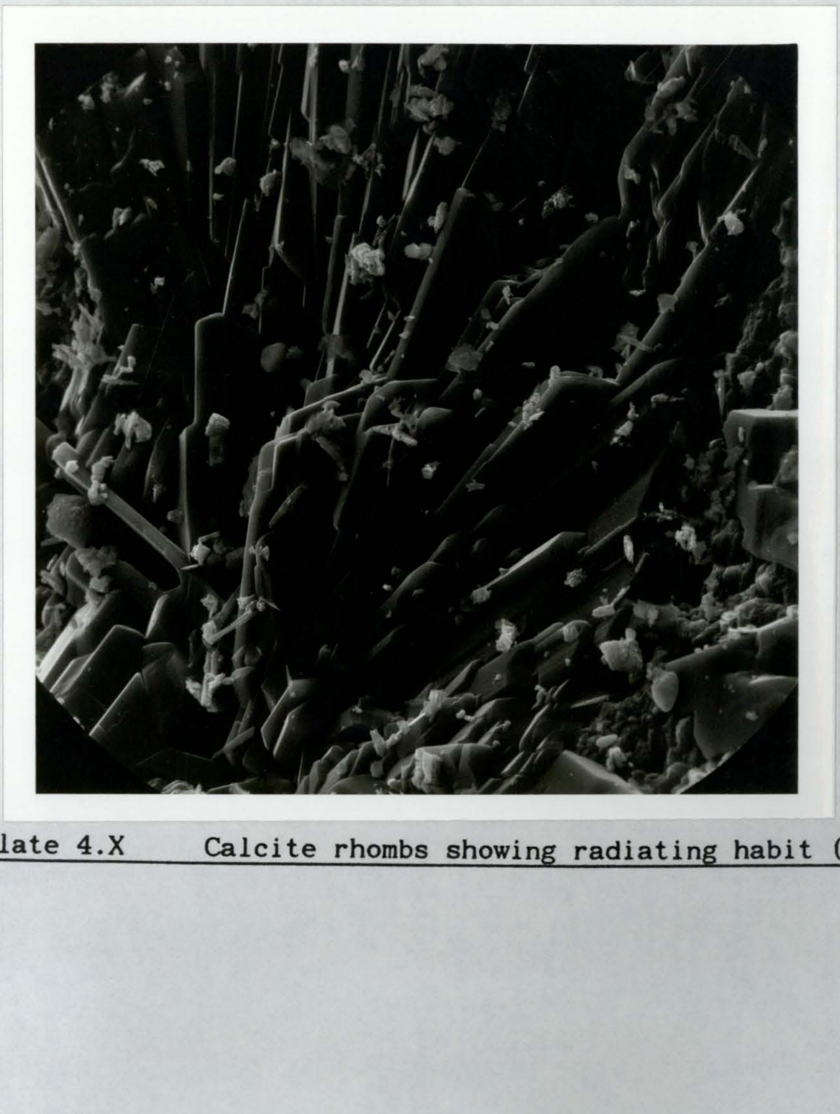


Plate 4.X Calcite rhombs showing radiating habit (650x).

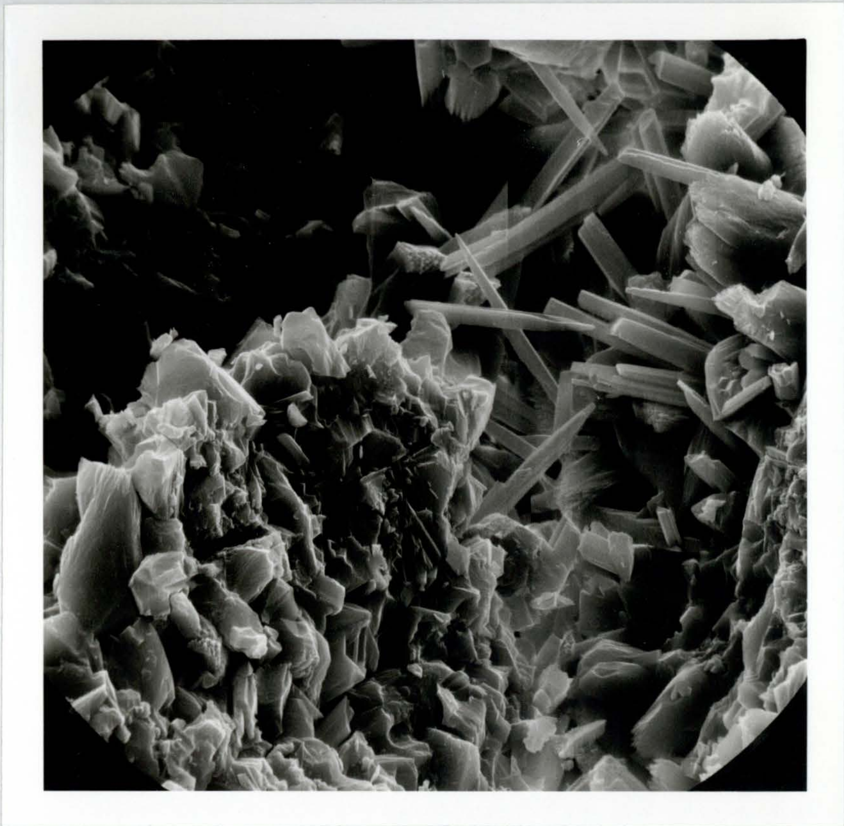


Plate 4.XI High magnification micrograph of calcite and aragonite crystals (1,500x).

Plate 4.XII Unweathered aragonite needles (12,000x).

Lake Bed E.

The samples analysed that represent lake bed E are Green 69 and 82/124. XRD analysis shows that based on composition lake bed E is very different from the previous beds. Lake bed E has a mean wt.% composition of 5% calcite : 94% aragonite.Green 69 is a lacustrine calcium carbonate with a crumbly cement component. Sample 82/124 is a hard lacustrine carbonate.

Green 69 was analysed by XRD and has a mean wt% composition of 8% calcite: 100% aragonite.

Plate 4.XIII shows aragonite crystals that show a common origin of growth. At 1,500x magnification there appears to be clusters of small crystals that could be a cement component. Plate 4.XIV also shows radiating aragonite needles. The smaller crystals may be the smaller calcite crystals or may be crystals of a cement. Plate 4.XV (550x magnification) shows large calcite crystals grown *in situ*. Combined with these are the very much smaller crystals of calcite.

In summary: lake bed E represents a different environment than that which deposited lake beds A-B or C-D. Both minerals are still present in the lake bed however aragonite crystals dominate and the same crystals now show a radiating acicular habit. Calcite crystals can be found in the same large form as previously, however they are now present with aragonite as a smaller crystalled component, and this is presumed to be a cement.

The lake bed appears to have been formed in a much shallower water column, where the aragonite crystals have grown *in situ*, rather then fallen down upon deposition from higher up in the water column. This shallow more evaporitive environment is further substantiated by the presence of a cement component in some of the samples.

0·6μm

Plate 4.XIII Aragonite needles with common origin of growth (1,500x).

0·9μm

Plate 4.XIV Radiating acicular aragonite needles (1,000x).

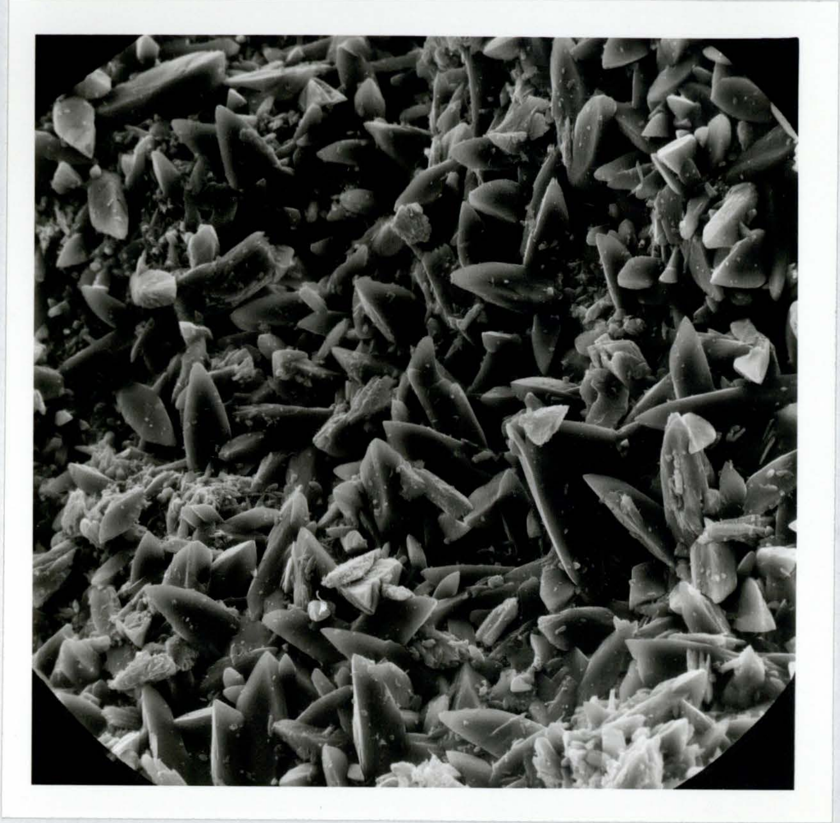


Plate 4.XV Large calcite crystals grown *in situ* (550x).

Plate 4.XVII Radiating habit of calcite crystals inside moulds (750x).

S.E.M. Conclusions on Morphology and Deposition Of Ross Sea II Lake Beds.

S.E.M. further confirms the XRD conclusions on the mixed mineralogy of the lake beds. The trend of a calcite dominant mineralogy to an aragonite dominant mineralogy, as the lake beds youngs, can be seen in the S.E.M. photos. However, differences are not as great or as representative as to be diagnostic. The S.E.M. photos gives further information on the morphology of these minerals and shows that the lake beds formed in different environments.

The calcite dominant lake beds were deposited in deep lakes where the aragonite crystals formed in the water column and then settled to the lake bottom. The calcite crystals then formed around the aragonite needles. The depth of the lake and the length of deposition time, determined the relative composition and morphology of the lake beds.

In the aragonite dominant lake bed E, the crystals appear to have grown *in situ* and suggests evaporitic growth. The presence of finer grained cements supports this theory.

The lake beds A-B, C-D show little evidence of weathering. The crystal faces are sharp and well defined with no pitting. This is in contrast to the lake beds of Miers Valley (one valley to the south). Here the 10-20,000 yrs old lake bed (s) are characterised by extremely weathered faces (Clayton-Greene,1986). Yet Marshall Valley contains unweathered 180,000 year old lake beds.

This suggests that unlike the Ross Sea I Miers Valley lake bed, the Ross Sea II lake beds A-B, C-D of the Marshall Valley, have never been subaerially exposed. The fluctuations in the advance and retreat of the glacier front during this Isotope Stage 6 glaciation, resulted in the freshly deposited lake beds being covered by advancing and retreating glacial drift. This sequence was repeated twice in the Ross Sea II Glaciation.

Lake bed E however does show oxidation of Fe^{II} to Fe^{III}, and a more friable texture and has probably been exposed to weathering for a considerable period of time.

S.E.M. Analysis of Calcium Carbonate Cemented Moulds.

Present within the stratigraphy of the Marshall Valley Stream Section are units of calcium carbonate cemented moulds. These units are found in Sections II,III, and IV and are closely associated with the calcium carbonate lake beds. Generally the cemented moulds units form localised deposits within the sections, in contrast to the continuous units of calcium carbonate lake beds (for a detailed stratigraphy - see Appendix II).

The moulds probably represent the past presence of gypsum crystals enclosed in a micrite cement, and there is field evidence and laboratory results from investigations of the crystal moulds, to support this proposition.

An examination of the morphology of the crystal moulds suggests that the size, shape and orientation of the crystal faces is similar to that of selenite crystals (a polymorph of gypsum). Selenite crystals are present in all the sections of the Marshall Valley Stream and vary in size from 1 mm to 4 cm. The selenite crystal layer also has a close association with the calcium carbonate lake beds and can be present in a sequence together with the cemented moulds, eg. units 3III and 6III in Section III.

Strong evidence that the cemented moulds are casts of gypsum is the occurrence of layers of selenite crystals encased in a micrite cement (see Plate 4.XVI).These layers represent the cemented moulds prior to the dissolution of the crystals.

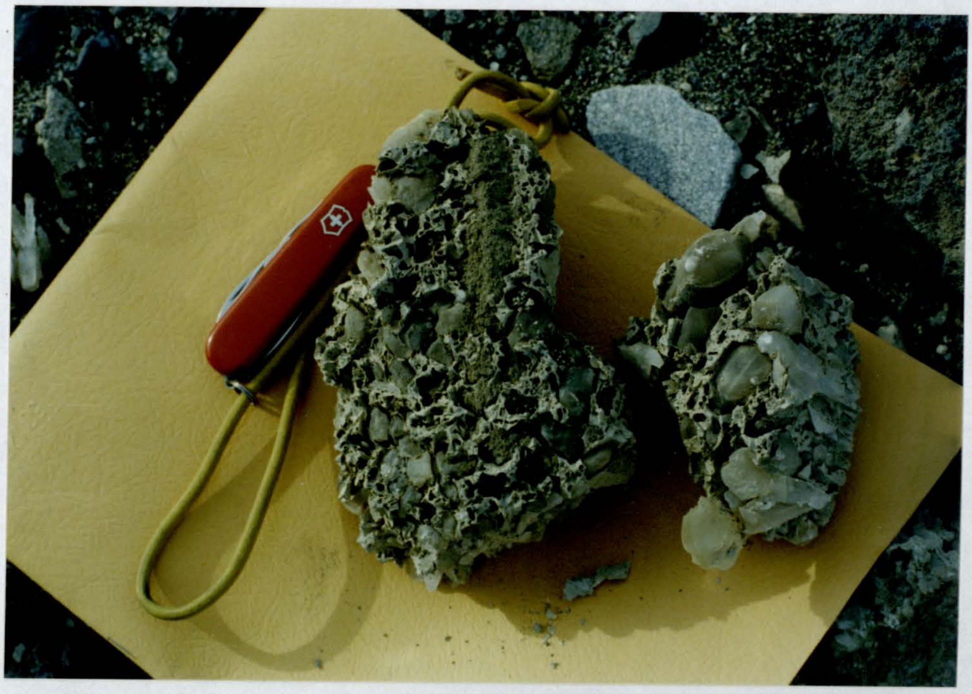


Plate 4.XVI Selenite crystals in gypsum cement.

In Section IV the selenite crystals in a micrite cement form a continuous layer that undergoes a transition to micrite cemented moulds. In the extreme west of the section, the unit consists of large selenite crystals, which forms a continuous unit that trends into a micrite cemented unit, and finally a micrite cemented crystal mould unit in the extreme east of the section.

Results of S.E.M analysis.

Samples of the large moulds were too large for S.E.M. analysis. However smaller crystals were analysed.

Plate 4.XVII (1,600 x magnification) was taken inside the mould. Large calcite crystals are present and show a radiating habit. This is the micrite cement. The crystals show evidence of weathering and lack the sharply defined crystal faces seen in the lacustrine carbonates. No evidence of any gypsum crystals is present.

Plate 4.XVIII (600 x magnification) was taken of the outer surface of one of the moulds. This photograph clearly shows the weathering of the surface. The crystals of calcite show pitting from weathering.

Plate 4.XIX (750 x magnification) shows the variety of crystals present in the sample. This photograph is also taken of the outer surface of the moulds. The weathered surface of the calcite crystals can be seen, partially obscured by the radiating acicular growth of the aragonite needles. We suggest that the aragonite crystals, are a secondary growth of crystals, and formed after the dissolution of the crystals that formed the moulds. This is because there is little evidence of weathering on the aragonite needles. Also present in the centre of photograph is a large abraided detrital grain tentatively identified as a sand grain.

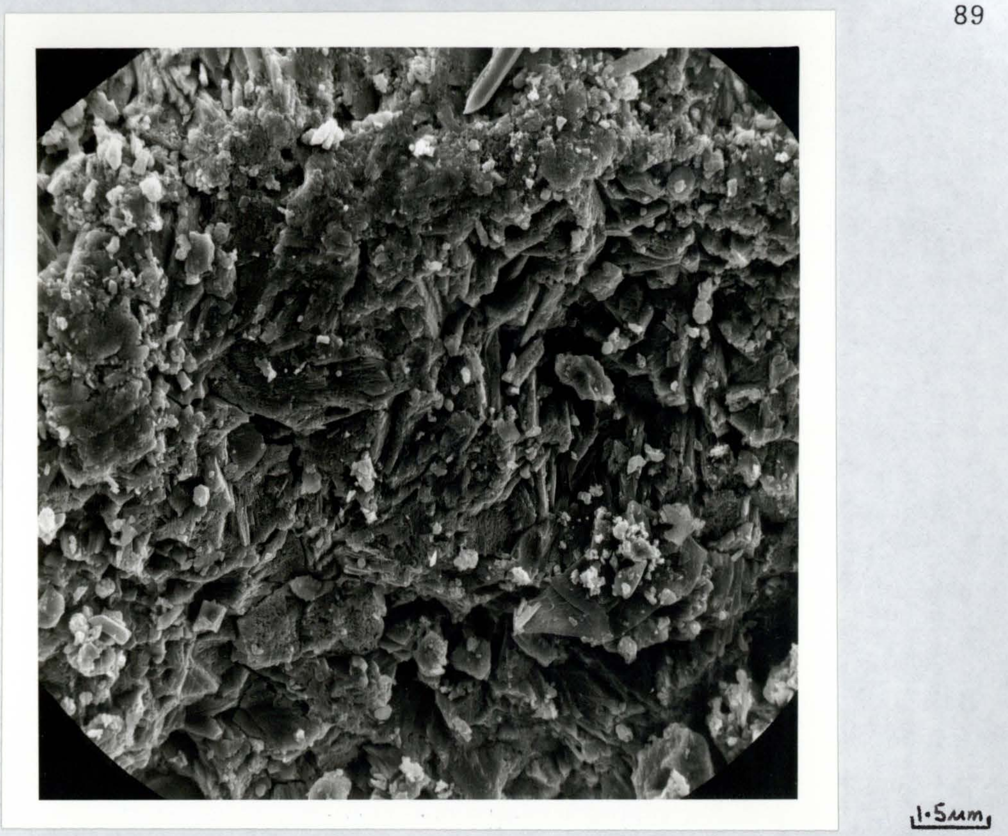


Plate 4.XVIII Weathering on outer surface of crystal moulds (600x).

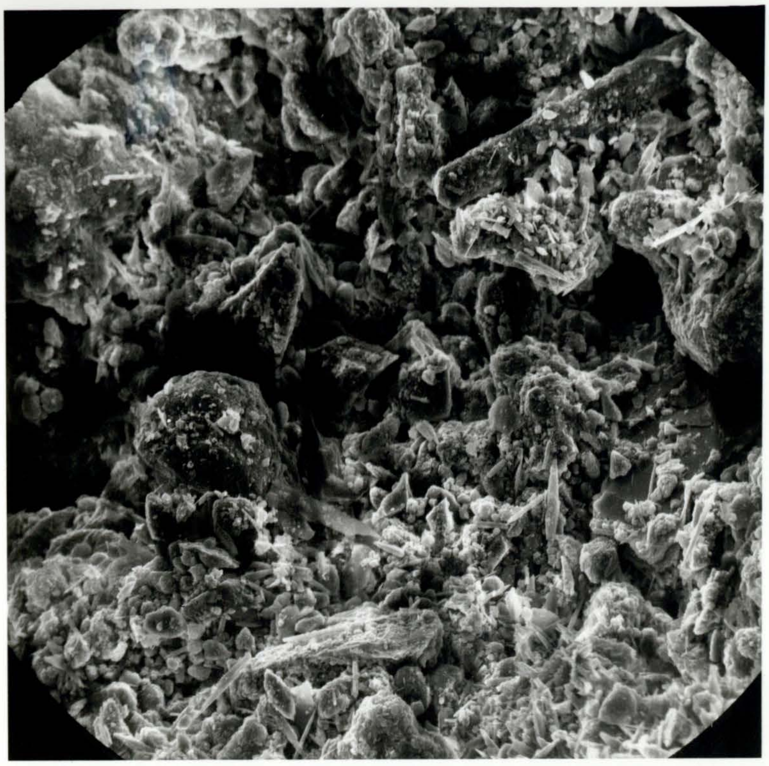
Plate 4.XIX Secondary growth of crystals on the outer surface of crystal moulds (750x).

Plate 4.XX (470 x magnification) was taken just inside the lip of the mould. The calcite crystals show definite signs of weathering, with much pitting. A diatom is present in the upper right quadrant of the photograph. A small quantity of aragonite is also present. The crystals have the appearance of growing on the surface and show little signs of weathering.

Plate 4.XXI (1100 x magnification) was taken on the outer surface of the moulds. The surface is largely aragonite crystals and these appear to have grown in place. The crystals appear to have grown through one another and once again show little sign of weathering.

In summary: Analysis inside the moulds shows no evidence of the past occupation the crystal. The crystal dissolution appears to have been complete. As a result of this dissolution the minerals of the cement show signs of weathering with much pitting. As well as the calcite crystals present in the cement, aragonite crystals are present. These crystals appear to have grown in place and because of a lack of weathering it is suggested that these crystals are a secondary deposit. These crystals have therefore grown after the dissolution of the presumed selenite crystals.

Because of the radiating habit of the crystals and the high degree of weathering, the cementing of the crystals and the following dissolution of the cemented crystals, appear to have occurred subaerially. As the units are only localised deposits in the stratigraphy, we suggest that the units deposition and dissolution occurred in very shallow ponds. Therefore this probably occurred in the final stages of proglacial lake recession from the valley, and dissolution may also have occurred as the Marshall Valley Stream cut down to the effected beds providing a flushing of undersaturated water.



1·9um

Plate 4.XX Weathering inside lip of crystal mould (470x).

0·8um

Plate 4.XXI Weathering on outer surface of crystal moulds (1,100x).

CHAPTER 5

INTERPRETATIONS OF 1986 FIELDWORK.

Chapter 5

Introduction

Due to an unusually warm summer season in 1985/86, snowbanks which normally obscured much of the southern side of the Marshall Valley Stream, receded revealing sections not previously described, and allowing new interpretations to be made of those previously described. This additional field evidence is discussed in this chapter.

1986 Season Field Observations.

Work has been carried out on the descriptions and interpretations of the stream sections of Marshall Valley for a number of years.

In the 1979/1980 season Mark Dagel began his masterate thesis on the stratigraphy of the stream sections. The stream sections were described in detail and their sedimentology and mineralogy studied .

A detailed discussion of this work is outside the range of my thesis, however the section descriptions are necessary to correlate the stratigraphy of the carbonate lake beds. The unit descriptions and stratigraphic sections are therefore reproduced in appendix II.

Prior to 1986, work in Marshall Valley had been carried out in the early part of the Antarctic field season, in November and December.

Our 1986 Marshall Valley drilling project was present in the valley in January, just after the peak of the Antarctic summer. In January, the stream was flowing freely and the large ice banks previously found on the south side of the stream had disappeared.

This revealed two stream sections that had previously not been described. These were described, photographed and sampled.

Newly Revealed Section A.

The section revealed was on the south side of the Marshall Valley Stream and was 100 m E of the north side Section III. In previous seasons, this section was covered by a snow bank. In January 1986, it was free of snow and the section was slumping away from the bank. The section rested on an ice core.

Fig.5.1 is the detailed description of the section and Plate 5.I is the photographed section.

The section is cut down through glacial drift units of basaltic medium sands with basalt pebbles and cobbles and basaltic gravels. Layered between these units are cemented beds,(at approx. 1.0 m, 1.3 m and 1.6 m) and calcium carbonate beds which contain crystal cavities (at approximately 1.1 m - 1.2 m, 1.3 m - 1.5 m, and 1.6 m - 1.8 m) and are presumed to have been gypsum.

Also present are two layers of hard calcium carbonate that forms sharp contacts with surrounding beds. Both layers, we suggest are lacustrine lake beds and correlate with the lake beds found in the section on the north side of the stream.

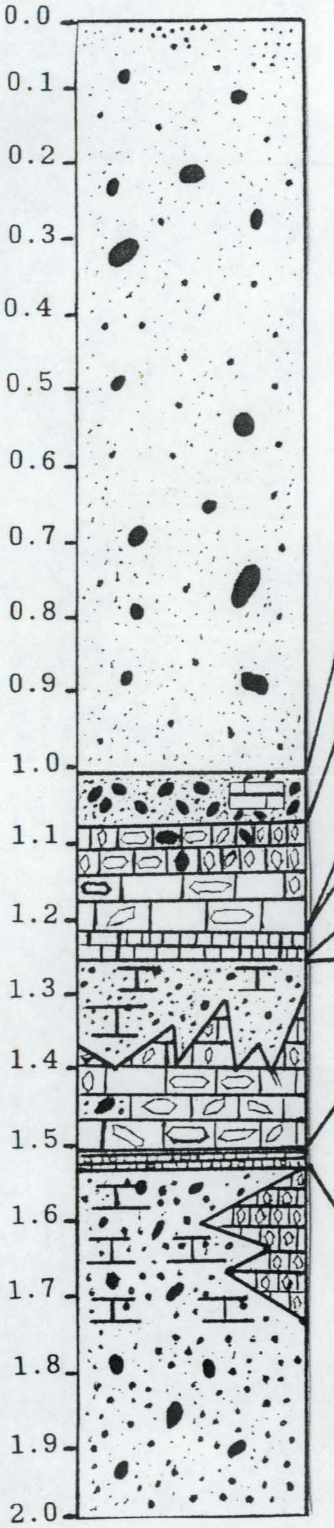
The lower of the two lake beds,(1.50 m) and therefore the older, we suggest correlates with lake bed A-B and is therefore approximately 180,000 years old. We base this suggestion on the stratigraphy of the section and on the section opposite on the north side of the stream. The morphology of the lake bed is a hard tabular calcium carbonate of approximately 1 cm thickness. This lake bed can be clearly seen in Plate 5.I. This morphology is characteristic of lake bed A-B as it is revealed in all sections of the Marshall Valley Stream.

The upper lake bed (at approx. 1.2 m - 1.3 m) is harder to correlate to a known lake bed. Although it is possible that this lake bed represents either lake bed C-D or lake bed E, with ages of

Plate 5.I Newly revealed Section A (south side of stream)

(photo C.Hendy)

Fig. 5.1 New Southside Section 100m E of Section III.



1m of wellsorted medium variegated sands with occasional pebbles and cobbles. Possible aoelian deposit.

5-10cm medium sands with pebbles and cobbles.Small pieces of $CaCO_3$ but these are not in place. Evidence of down slope movement.

10-20cm of bright white $CaCO_3$ with crystal cavities. Basaltic cobbles also present. The white $CaCO_3$ becomes more massive towards the base.

2cm thin grey $CaCO_3$ coating on underlying beds. Forms sharp contact.

10-15cm of laminated dark grey $CaCO_3$ cemented coarse sand to fine gravels merging downwards into 10cm of light grey $CaCO_3$ with crystal moulds of gypsum. Bed forms sharp contact at base of bed.

1-2cm of finely laminated pale yellow brown $CaCO_3$ into which the underlying clasts project.The $CaCO_3$ bed is not connected to other beds and can be seen growing.

50cm thick bed consisting of coarse variegated sands with occasional gravels. The bed is heavily cemented with large crystal moulds of gypsum to 20cm depth beneath the $CaCO_3$ bed. Locally the 5-10cm of white crystal moulds are heavily overgrown. These occur at the top of the coarse sand.

The section is supported on relatively clear laminated? ice.

approximately 160,00 years and 130,000 years respectively, it is most likely that the lake bed in the section is lake bed C-D. Both lake beds extend further westwards in the sections on the north side of the stream.

The samples from these sections have not been analysed and it is probable that by U/Th dating, XRD and SEM, it would be possible to define the chronology of the section.

However the presence of the two lake beds and their possible correlation increases knowledge as to the lateral extent of the lake beds.

Newly Revealed Section B.

This section was also revealed on the south side of the Marshall Valley Stream and was 250 m west of the campsite between Sections III and IV on the opposing north side of the stream. This section was also revealed, due to the removal of the large snow bank, that had previously obscured the south side of the stream.

Fig. 5.2 is the detailed description of the section and Plate 5.II is the photographed section.

This longer section cuts down through several metres of glacial drift units, of basaltic medium sands with basaltic pebbles and cobbles, and units of basaltic medium sands with basaltic gravels. Interbedded between the glacial drift are two layers of cemented glacial units; a crumbly calcium carbonate containing basalt pebbles and cobbles (at a 4.4 m) and a cemented sandy gravel (at a 4.9 m). One layer of calcium carbonate cemented gypsum moulds is present underlying the crumbly calcium carbonate (at a 4.7 m).

Three laminated calcium carbonate units are present in the section, layered between the glacial till units (at 1.5 m, 3.3 m and 3.8 m).

Plate 5.II Newly revealed Section B (south side of stream)
(photo C Hendy)

Fig. 5.2 New Southside Section 250 m West of Campsite.

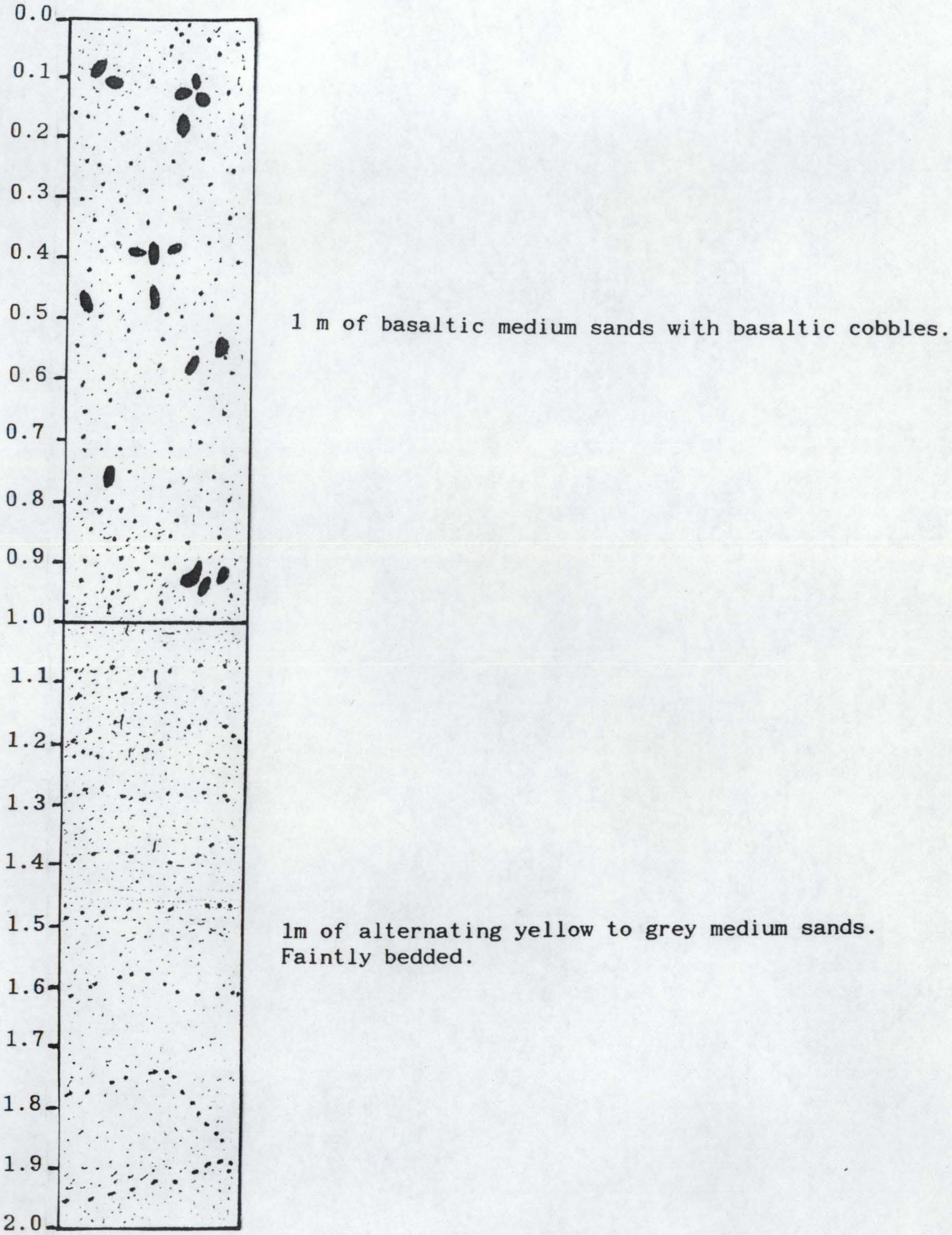
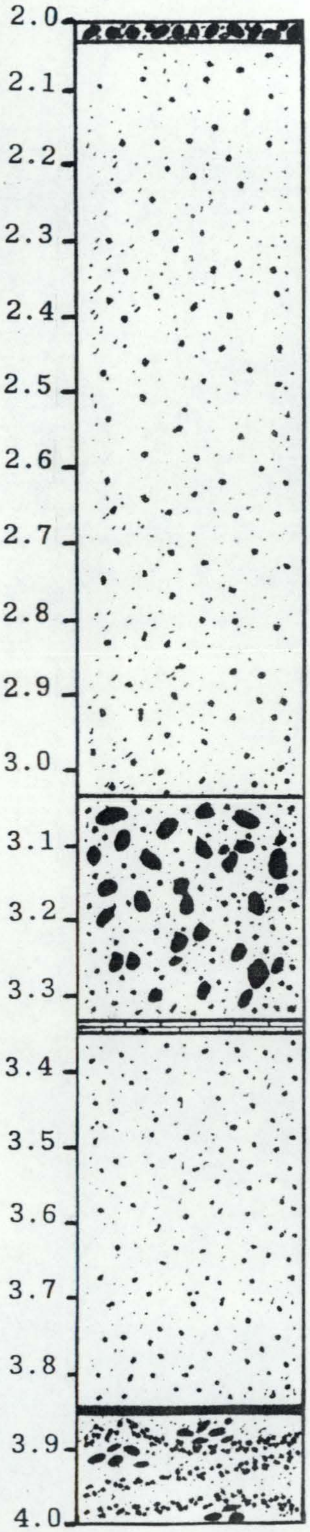


Fig. 5.2(cont) Stratigraphic column of newly revealed Section B.



1-2cm of rounded basaltic gravels - 1 pebble thick.

1m of medium sand.

30cm of moderately rounded coarse sandy gravels. Composition >95% basalts.

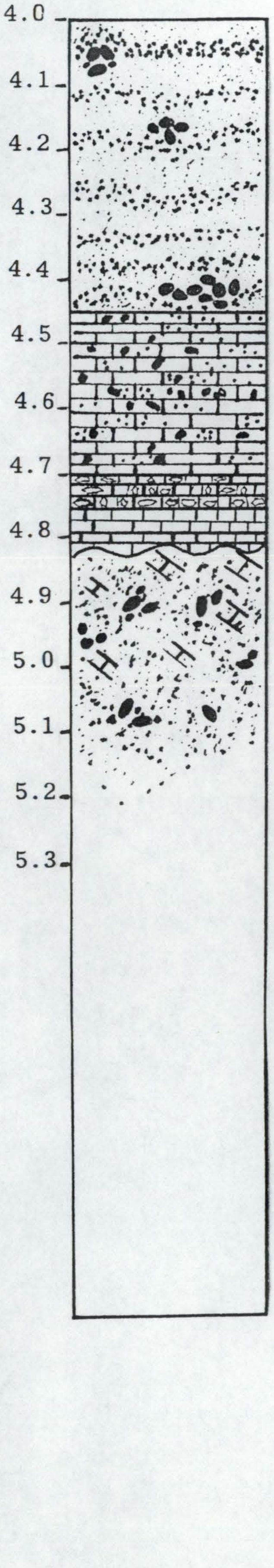
1cm finely laminated cream $CaCO_3$, firmly cemented to pale grey medium sands with 1cm basaltic clasts.

50cm pale grey basaltic sands. A sharp contact at base of bed with laminated $CaCO_3$ cement.

2-3 mm laminated $CaCO_3$ cemented to overlying bed.

unit discription over page.

Fig. 5.2(cont.) Stratigraphic column of newly revealed Section B.



60cm of well bedded medium sands with gravel lenses. Has a higher basalt concentration then unit above. Bottom of unit is cemented to the underlying unit.

25cm of bright white crumbly $CaCO_3$ containing basalt pebbles and cobbles.

5cm of grey brown $CaCO_3$ with crystal moulds of gypsum.

2-3cm of laminated $CaCO_3$ which drapes over underlying clasts and unconformably rests on the next unit.

$CaCO_3$ cemented sandy gravel or ice. The base of this unit cannot be seen. The sandy gravel beds dip to the west.

It is suggested on the basis of the observed stratigraphy, on the opposing sections of the north side of the section, that the laminated calcium carbonate layers correlate to the 3 Ross Sea II lake beds with ages of approximately, 130,000 years , 160,000 years and 180,000 years B.P.

Once again this hypothesis is based only on field evidence as no experimental work has been undertaken on the samples retrieved from this section. However the presence 3 discernible lake beds in the stratigraphy of the south side of the stream, is evidence of the lateral correlation of stratigraphy between both sides of the Marshall Valley Stream.

A Review of Section III Stratigraphy.

As part of the field work of the 1986 field season. The previously mapped sections (Dagel *et al*, in press) were reviewed. New field evidence leads to the suggestion of a new interpretation of the stratigraphy of Section III.

Section III is the most complex of the Marshall Valley Sections (Plate 5.III). All Ross Sea II lake beds are revealed in the section. The detailed description of the section is contained in Appendix II (Dagel *et al*, in press). Based on this work, a brief resume of the 1986 field description is :

From the base: A basement rich muddy sandy gravel underlies a basement rich gravelly sand. Unit 6III overlies these two units. It has three layers: a tabular calcium carbonate, overlain by 1-4 cm selenite crystals, overlain by a fine gypsum crystal layer with sands and pebbles. Unit 5III overlying the previous bed is composed of basaltic sandy gravel lenses and this underlies Unit 4III which is a muddy sandy gravel. Both the last two layers are not extensive. Unit

Plate 5.III Downslope slumping of lake beds at Section III
(photo C Hendy)

3III contains three layers identical to Unit 6III, however the layers are reversed being; a fine gypsum crystal layer, overlain by 1 - 4 cm selenite crystals overlain by a tabular calcium carbonate. This reversed unit is overlain by a basalt rich muddy sandy gravel identical to Unit 4III. Unit 1III contains a layered sequence of micrite with crystal moulds, calcium carbonate layer varying laterally from a hard tabular to a soft chalky morphology, a medium sand unit and a finely laminated tabular calcium carbonate.

New Interpretation.

As a result of the completed geochronology and field investigation of the Section III stratigraphy, it is suggested that the interpretation can be simplified. The similarity of Units 6III and 3III and 4III and 8III have been explained (Dagel *et al*,in press) by an isoclinal fold with an axis perpendicular to the stream. This folding has been attributed to the action of the advancing and retreating glacial front.

On reinspection, after more exposure, it is suggested that the complex stratigraphy is the result of slumping down slope of the mineral beds and glacial drift .This slumping may be envisaged as a sheet deposit that has slumped downslope, forming folds due to the non moving toe at the base of the slope. This produced the overturn of beds. The axis of the fold is parallel to the stream and the identical 3 layers of the Units 6III and 3III can be traced through the slumped folds as the sequence is overturned. The selenite crystal layer very clearly shows the overturn of the beds on slumping.

It is proposed therefore that the two lake beds A-B represent a continuous layer contorted by the slumping event. This is further substantiated by the similar geochronology, morphology and mineral composition. This lake bed was deposited 180,000 years B.P. as the result of glacier advance and retreat. At some time after deposition (possibly as the result of instability due to the retreat of glacier ice) slumping occurred.

The slumping may also have occurred as the result of the formation of a proglacial lake in the valley, reducing the strength of the unconsolidated beds through melting the previous ice cement and lubricating any movement down the steep north valley wall. A pronounced hummocky terraine (Plate 5.IV) extends up slope from Section III and appears to originate from a horizontal bench extending up, into the valley at approximately 310 m a.s.l. Since the slumped material contains a basalt rich drift overlying lake bed A-B, but does not contain lake bed C-D carbonates, which elsewhere immediately overlies this drift,it is postulated that the slumping may have occurred during the initial formation of the 160,000 year old lake.

Plate 5.IV Hummocky ground on slopes above Section III

CHAPTER 6

1986 WINKIE DRILLING SEASON

Introduction

The 1986 drilling project in the Marshall Valley is discussed in this chapter. A comparison is made between the upper portions of the cores and the glacial sequences already represented in the stream sections, as a discussion of the deeper subsurface sediments is outside the range of this thesis.

Aims and Objectives of Drilling in Marshall Valley.

As previously stated in this thesis, Marshall Valley contains in its stream section, a well dated glacial sequence representing glaciations at Isotope Stage 2 and 6 (30,000 and 180,000 years B.P.). This is related to the incursions of the Ross Sea into the valley and the ponding of drainage water to create lakes in which lacaustrine carbonate lake beds were deposited.

Times of glaciation in Marshall Valley are characterised by incursions of an ice sheet from the Ross Sea and the creation of lakes. The valley sediments are an accumulation of these deposits, representating a separate glaciation. Such older glacial sequences are likely to contain lake beds capable of dating by U/Th method, and if so it could therefore be possible to date these glaciations. As the last two glaciations are present in the stream sections, it is probable that earlier glacial sequences will be present in the subsurface.

In addition it is possible that, in the past glacial history of the valley, there had been a reaction of the interior glaciers to climatic change,with the Rivard Glacier or perhaps the Blue Glacier flowing down valley and depositing glacial drift in a seaward direction. This would have been in contrast to the upvalley direction of movement of the Ross Sea glaciations.

It was hypothesised that the direction of flow in the valley and therefore the origin of the glaciation could be determined by the mineralogy of the sediments represented in the glacial drift. Ross Sea I and II Glaciations are characterised by glacial drift rich in basalts derived from the Ross Island Volcanics, whilst the glacial drifts whose origins are from interior glaciers should have compositions dominated by granitic and plutonic rocks of continental origins.

Therefore investigation of the mineralogy of the subsurface glacial sequences could give information as to the origins of previous Marshall Valley glaciations.

A drilling project in Marshall Valley would determine if prior glacial sequences had been accumulated in the valley. Sedimentalogical evidence found could determine the origins of these glaciations and a possible geochronology of glacial events could be determined . If the hypothesis, that Ross Sea Glaciations are iniated by low sea levels, is true, then any dates obtained on subsequent Ross Sea Glaciations should correlate to the eustatic sea level curve.

The 1986 Drilling Season.

In January of 1986 we commenced a drilling project in Marshall Valley using a Winkie drill. The Winkie drill has the capacity to drill to a depth of 80 m and produced a 28.5 mm diameter core.

Drilling was through permafrosted silts and sands and used circulating refrigerated D.F.A. (drilling fuel arctic) to maintain the core in its frozen form, whilst the drill was in operation. Once extracted from the core barrel, the core was given a preliminary log on site and then boxed and stored in a refrigerated box (refrigerator units were run off a diesel generator to maintain the core in its frozen form). At the end of drilling of each hole the drill camp and rig, was relocated by helicopter and the core from that hole was returned to Scott Base where it was described in detail, photographed and sampled . Still in its frozen form it was shipped to New Zealand and stored in a refrigerated room at the University of Waikato.

Three drill sites were chosen for drilling. The locations were surveyed by N.Z. Lands and Survey surveyors and altitudes for each site are given on the map provided. (Fig.6.1). The drill sites form a transect down the valley and are roughly parallel to the stream sections but offset approximately 200 m to the south of the stream.

Their distances from the stream sections were positioned so that they were in the middle of the valley and therefore in the area of expected deepest sediment. Care was taken that drill sites were not too close to the stream as these were likely to be affected by subsurface drainage due to meltout of the sediments. However the drillsites were still close enough to allow possible correlations between the upper portions of the core and the glacial sequences found in the stream sections.

It was proposed to drill to basement and from a visual appraisal of the valley morphology it was estimated this depth would not exceed 50 m.

The Drilling Of Site M.V.1

Site M.V.1 was chosen as it was adjacent to Section III and therefore this section could be used for correlation purposes. This site was also beyond the proposed extent of Ross Sea I glacial drift deposits (Dagel *et al*, in press)(Fig.1.2) and so any glacial sequences drilled in the subsurface would be representative of glaciations older then Ross Sea I.

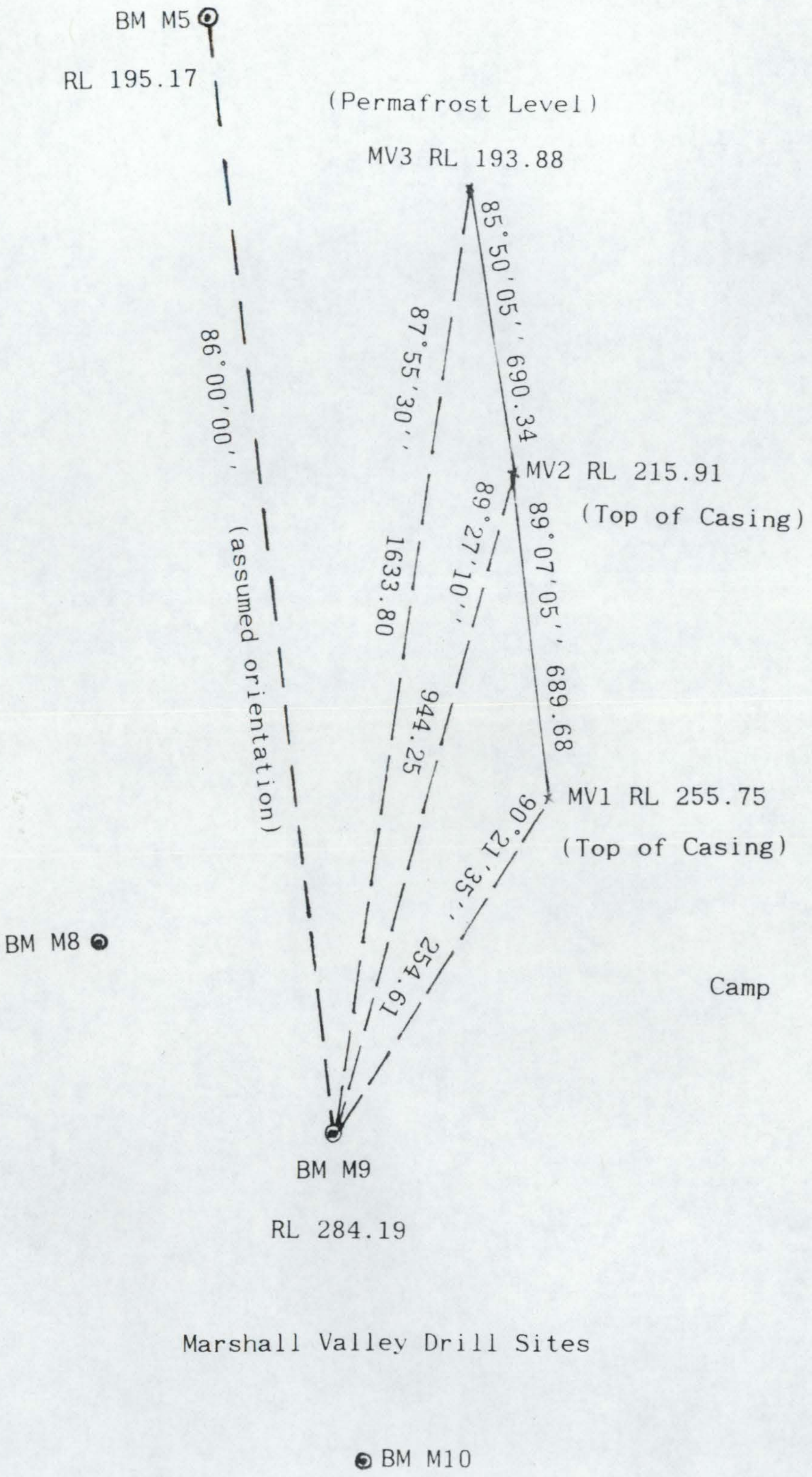


Fig. 6.1 Locations and altitudes of 1986 Marshall Valley drillsites (Department of Lands and Survey)

The drilling of M.V.1 commenced on the 11th of January and the hole was completed on the 16th of January. The depth to basement was 25.63 m and of this depth 22.89 m of core was recovered (89.3 % recovery).

In general the drilling of this hole was easily accomplished and the core recovery was very good. Core was lost at the top of the hole due to layers of coarse sands and pebbles which caught in the core catcher and prevented drilling and recovery of core. Further problems were encountered at a depth of approximately 22 m. Drilling was through a very coarse and porous layer and a large volume (150 litres) of D.F.A. was lost down the hole. An attempt was made to seal the hole bottom by the use of a permafrost cement. Reaming was required to drill down through the cement before drilling could recommence. Drilling continued through coarse sediments and another 200 litres of D.F.A. were lost down the hole. The core recovered showed signs of oxidation and it was suspected that we were close to basement. Another 70 litres of D.F.A. was lost down the hole in the last run that produced 1.5 m of plutonic rock which we assumed to be basement .

The Drilling of Site M.V.2.

Site M.V.2. was chosen 689.68 m down the Valley from site M.V.1. The hole site sat on Ross Sea I drift and was presumably ice rafted. Drilling commenced on the 20th of January and was completed on the 24th of January. 35.11 m was drilled and 33.02 m of the core was recovered (96% recovery).

Some loss of core was sustained at the top of the hole due to coarse sediments .The recovery rate of the carbonate cemented core was lower then the overall recovery due to the grinding and washing away of the finer particles. However the drilling of the medium sands and gravels was easily achieved and recovery rates were excellent.

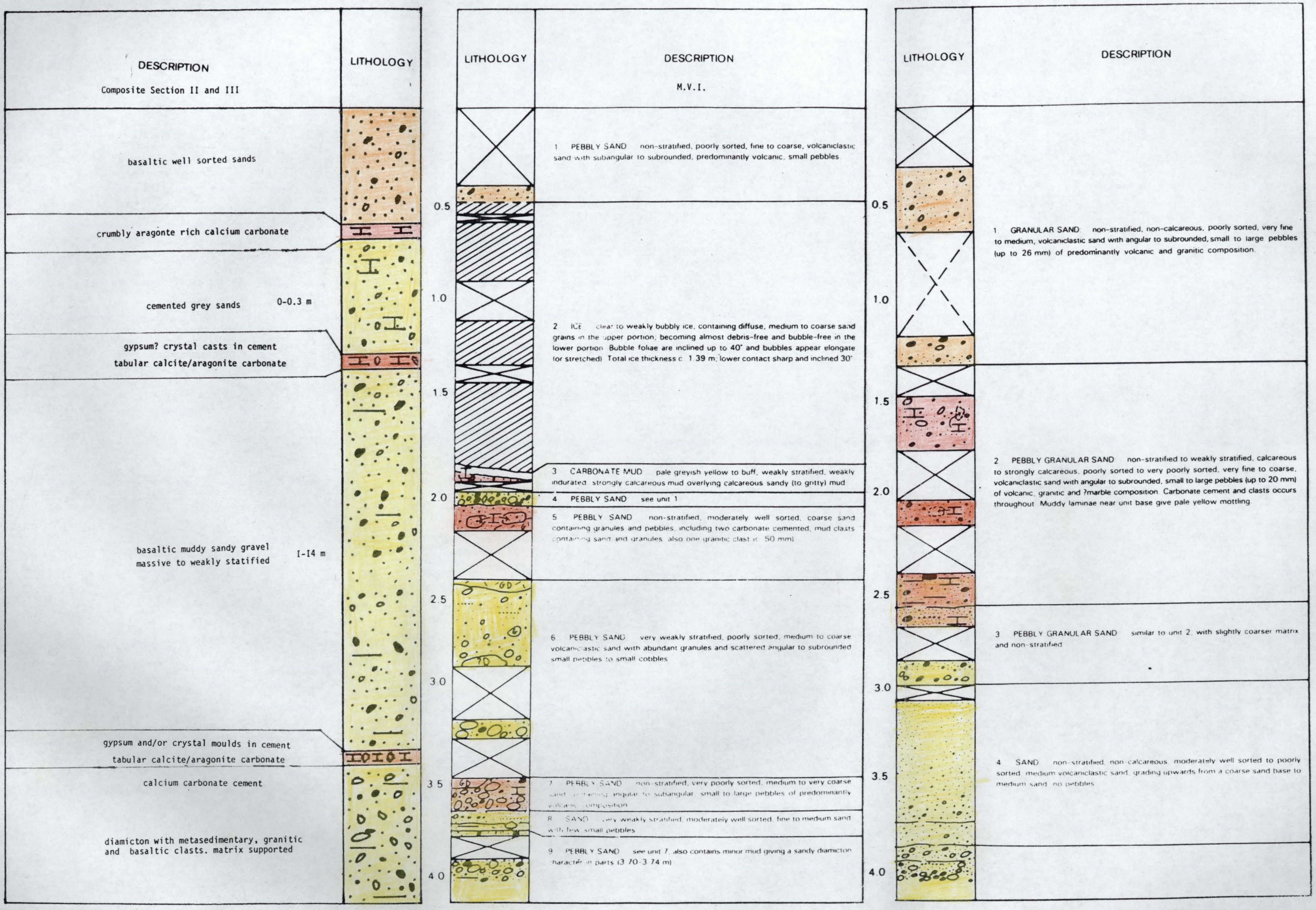
At approximately 31 m oxidation rinds on pebbles were recovered in the core and it was felt that the hole was close to basement. At this point we had a mechanical failure with the drill that necessitated its return to Scott Base for repairs. On its return drilling recommenced and on the first run evidence of basement was encountered . A further drill run confirmed that basement had been reached.

Correlations of M.V.1. and M.V.2. to Marshall Valley Stream Sections

Sites M.V.1. and M.V.2. (Fig. 6.2) both encountered laminated carbonates and carbonate cements within the top 4 m of the core. Because of the narrow diameter of the core, and the poor recovery of the upper 3 m, it was not possible to demonstrate the presence of continuous carbonate beds. However, at each site three carbonate zones were observed, which visually could be correlated with lake beds A-B, C-D and E in Sections II and III.

Between the lower two carbonate zones, both sites exhibited approximately 1 metre of pebbly-pebbly granular volcaniclastic sands (units 5,6 M.V.1., units 2,3 M.V.2). This bed presumably represents Units 7II, 2III (5III) in Dagels discription of the exposed stream sections (Appendix II), where the thickness ranged from 1 m at Section II to 14 m at Section III, at which point the basaltic sands thicken to form a series of loop moraines.

In Section III,lake beds C-D and E are separated by 0-30 cm of granular volcaniclastic sands, which match unit 4 in M.V.1. In Section II, lake beds C-D and E are separated by a diamicton which varies in thickness from 0 to 0.3 m. A similar bed was not observed in M.V.2., however core recovery was only 50% over the expected interval 1.3 - 2.3 m depth, so that it is not possible to determine whether this variable unit extended across the valley between Section II and M.V.2.



1 PEBBLY SAND non-stratified, poorly sorted, fine to coarse, volcaniclastic sand with subangular to subrounded, predominantly volcanic, small pebbles.

2 ICE clear to weakly bubbly ice, containing diffuse, medium to coarse sand grains in the upper portion; becoming almost debris-free and bubble-free in the lower portion. Bubble foliae are inclined up to 40° and bubbles appear elongate (or stretched). Total ice thickness c. 1.39 m; lower contact sharp and inclined 30°.

3 CARBONATE MUD pale greyish yellow to buff, weakly stratified, weakly indurated, strongly calcareous mud overlying calcareous sandy (to gritty) mud

4 PEBBLY SAND see unit 1

5 PEBBLY SAND non-stratified, moderately well sorted, coarse sand containing granules and pebbles, including two carbonate cemented, mud clasts containing sand and granules, also one granitic clast (c. 50 mm)

6 PEBBLY SAND very weakly stratified, poorly sorted, medium to coarse volcaniclastic sand with abundant granules and scattered angular to subrounded small pebbles to small cobbles

7 PEBBLY SAND non-stratified, very poorly sorted, medium to very coarse sand containing angular to subangular, small to large pebbles of predominantly volcanic composition

8 SAND very weakly stratified, moderately well sorted, fine to medium sand with few small pebbles

9 PEBBLY SAND see unit 7, also contains minor mud giving a sandy diamicton character in parts (3.70-3.74 m)

1. GRANULAR SAND: non-stratified, non-calcareous, poorly sorted, very fine to medium, volcaniclastic sand with angular to subrounded, small to large pebbles (up to 26 mm) of predominantly volcanic and granitic composition.

2 PEBBLY GRANULAR SAND: non-stratified to weakly stratified, calcareous to strongly calcareous, poorly sorted to very poorly sorted, very fine to coarse, volcaniclastic sand with angular to subrounded, small to large pebbles (up to 20 mm) of volcanic, granitic and ?marble composition. Carbonate cement and clasts occurs throughout. Muddy laminae near unit base give pale yellow mottling.

3 PEBBLY GRANULAR SAND similar to unit 2, with slightly coarser matrix and non-stratified

4 SAND non-stratified, non-calcareous, moderately well sorted to poorly sorted, medium volcaniclastic sand, grading upwards from a coarse sand base to medium sand, no pebbles

SCALE 1:20

Fig. 6.2 Stratigraphic correlation of M.V.1. and M.V.2. to Marshall Valley Stream Sections II and III

Mineralogical and isotope studies in progress should confirm the indentification of these beds.

It is apparent from the two cores that carbonate beds are confined to the upper 4 m of sediment at both sites, so that carbonate sedimentation has been a rare event in the Marshall Valley, and is apparently confined to the Ross Sea II (?) Glaciation, 180,000-130,000 years B.P. It is also apparent that the underlying sediment contains an irregular succession of sand-mud/ pebbly sand/ diamicton which probably represents successions of ice distal, ice proximal and ice contact deposition (Robinson,1986). Six to eight such successions appear below the Ross Sea II beds. These contain both basaltic and basement clasts but detailed measurements of provenance have yet to be made. It is likely therefore that the valley has been subjected to many incursions of ice from M^{c}Murdo Sound.

M.V.1 revealed 1.39 m of clear to weakly bubbly ice lying approximately 0.5 m below ground level. Similar ice was also encountered in the newly exposed stream section A approximately 50 m east of M.V.1. The origin of this ice remains uncertain. The lack of bubbles suggest that it is not a buried snowbank or glacial ice. However it could be refrozen meltwaters buried under aeolian sands, or perhaps a remanant of the Ross Sea I lake ice buried under ice rafted debris. A distinction between these two will be attempted using O^{18}/O^{16} and D/H. Units 1 in M.V.1 and M.V.2 represent Ross Sea I sediments, possibly augmented by recent aeolian reworking at M.V.1.

The Drilling of Site M.V.3.

Site M.V.3. was chosen 690.34 m down the valley from site M.V.2. The site was chosen opposite Section I and could therefore allow correlation between section I glacial sequences and glacial sequences found at the top of the drill core. To therefore ensure the recovery of the core containing possible evidence of Ross Sea II glaciations, the drill site was positioned on the top of the most westerly Ross Sea Drift moraine . The extra 12 m of sediment contained in the moraine allowed for " spudding in " and the maximum core recovery to be achieved before the presumed Ross Sea II glacial sequences were drilled. Also drilling vertically through a moraine would give more information on its internal structure and its possible mode of deposition then could be achieved by the digging of vertical trenches over its surface.

The drilling of M.V.3. commenced on the 28th January and ended on the 29th of January. The depth of the hole was 13.79 m and 7.22 m of core was recovered. (52% recovery rate).

The drilling of this hole was difficult and the recovery rate was much lower than in the two previous holes. The material being cored was very coarse and problems were encountered with gravels jamming in the core barrel and preventing core recovery. There was evidence that the sands were being flushed away from the core sediments.

At approximately 3 m recovery rates improved and sands, carbonate layers and a gypsum layer were recovered.

However at approximately 12 m very coarse sediments were encountered . Drilling was stopped and the hole abandoned due to the damage being sustained to the diamond bits and our lack of replacements. The material being recovered was also very weakly cemented coarse sands and was unable to be retained in the core barrel.

Possible Correlation between M.V.3 and Marshall Valley Stream SectionI

M.V.3 revealed a contrasting sediment to M.V.1 and M.V.2 (Fig.6.3). It was dominated by basaltic pebbly sands and contained neglible carbonates. It did however penetrate a well crystallised selenite horizon (3.0 - 3.2 m depth). Similar beds crop out in Sections I and II where they are identified as lake bed F. The selenite overlies a sandy muddy bed and is consistant with a lacustrine origin. The 3 m of basaltic pebbly sands (unit 1) overlying the gypsum represent a late Ross Sea I moraine. Below the gypsum, all of the sediments appear to be Ross Sea I in origin, making at least 11 m thickness. In the stream section 180 m north of M.V.3., a total of 8 m of basaltic coarse pebbly sands separate the gypsum bed (F) from the lake bed E.

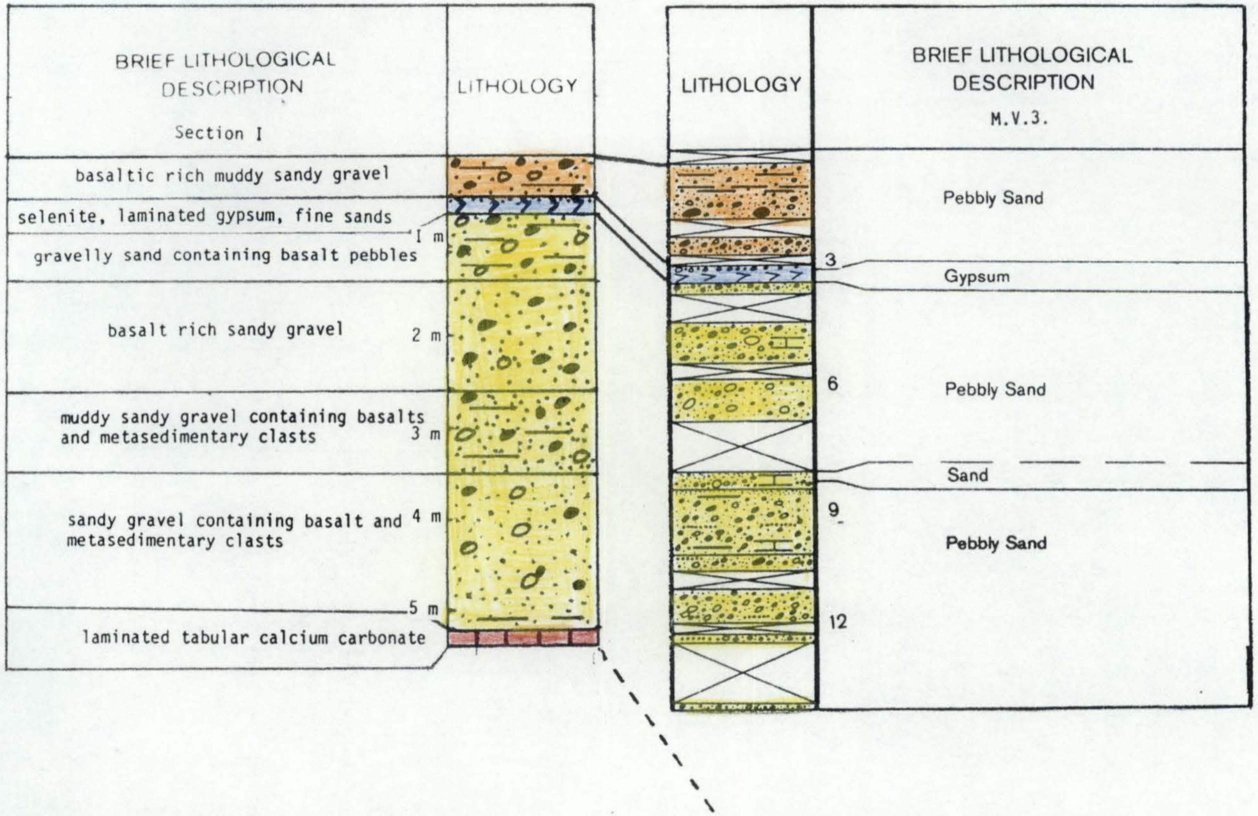


Fig. 6.3 Stratigraphic correlation of M.V.3. to Marshall Valley Stream Section I

CHAPTER 7

GLACIAL HISTORY AND ENVIRONMENT WITHIN MARSHALL VALLEY FROM THE ILLINOIAN TO THE PRESENT

Chapter 7

Introduction.

From the field observations and the laboratory analyses performed as a result of this thesis, it is possible to draw a number of conclusions about the glacial environment that has existed in the Marshall Valley during the last 180,000 years. The stratigraphy of the drift sheets in the Marshall Valley indicate the changing occupation of the valley by glacial ice and its associated proglacial lakes and sedimentary environment. The composition of the lacustrine sediments and evaporites enables the source of the sedimentary components and the mechanisms of deposition to be determined. The juxtaposition of the lacustrine sediments and the tills enables the relationship between the glacial ice and their proglacial lakes to be determined, along with the processes that led to the sediment deposition. The dating of the lake beds has enabled the chronology of the glacial lake occupation and that of the supporting ice sheets to be determined. This in turn has enabled the development of the last two Ross Sea Ice Sheets to be linked to episodes of world wide glacial eustatic sea level lowering.

(Dagel *et al*, in press) grouped the tills in the stream section into two groups based on their composition and grain size analysis.

The first group are classified as basal till units which are poorly sorted, compact and unstratified. They represent an environment in which the ice front was adjacent to and probably overran the site. In this environment both deposition and erosion are likely, with deformation and compaction due to superimposition of the glacial ice.

The second group of tills grade from basalt rich muddy sandy gravels to gravelly sands with massive to moderate stratification.

These units contain a range of bedding from sub horizontal to steeply dipping. Dagel *et al* (in press) classified these, as supraglacial to englacial debris, and postulated that they would have been deposited as outwash or debris flow fans close to the ice front, grading into fine sands and silts distal to the front, and including direct ablation.

Based on the sediment grouping above, we can construct a possible development sequence for the glacial incursion, retreat and formation of a proglacial lake.

The model is based upon the presence of glacial ice from the Ross Sea, with the glacier occupying the mouth of the valley and forming an ice front. Detrital sediments within the glacier ice (Fig.7.1) have been carried up valley and may be grouped into basal sediments (that are present at the base of the glacier and may be in contact with basement rocks) and englacial- supraglacial sediments (that are found in the glacier ice or on the glacier surface).

At the glacier front, melting occurs and the sediment is deposited in front of the glacier. Because of the low mean temperatures of Antarctica (≃-20°C) the glaciers are largely dry based at their margins and it is expected that there would have been little melting and subglacial flow. Therefore the greatest input of glacial sediments would be from the supraglacial-englacial debris as the surface of the glacier is subjected to ablation.

If the glacier front was in a stillstand position with a proglacial lake in front, then the smaller contribution of basal sediments would be deposited just in front of the glacier. The deposit would be coarse grained and have little sorting or stratification.

Depending on the nature of the boundary between the glacial front and the floating ice cover on the proglacial lake, the larger contribution of the supraglacial-englacial sediments could either be

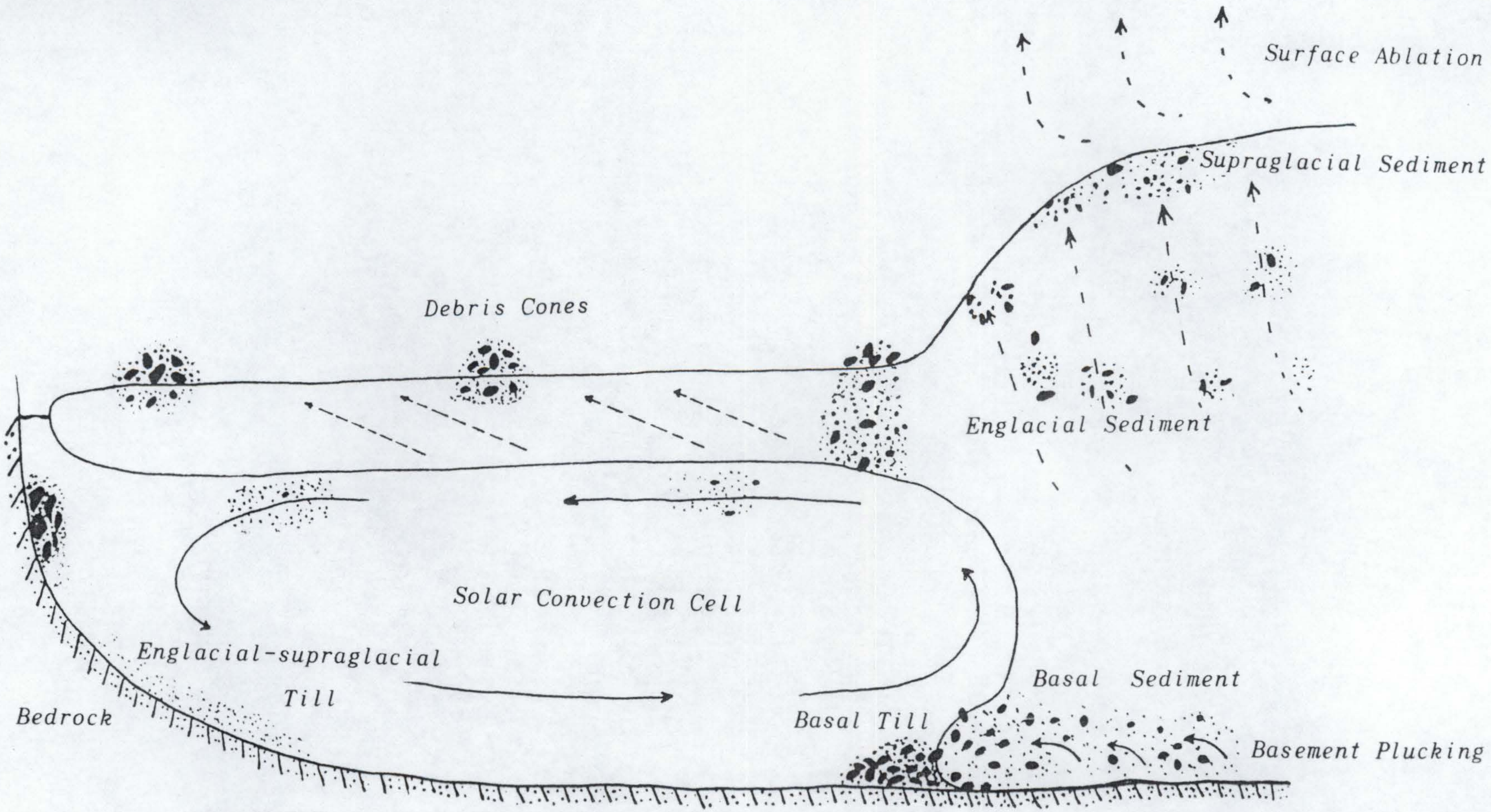


Fig. 7.1 Detrital sediments in a glacier front and proglacial lake

dropped through the proglacial lake ice in front of the glacier or be ice rafted across the lake. If the glacial sediments were dropped in front of the glacier, the coarser sediments would occur proximally to the glacier front, whilst the finer sediments would be carried by the lake convection system to the distal areas of the lake.

In the case where the glacial ice front (Fig.7.2a) was advancing the basal till and the coarse supraglacial-englacial till would be deposited proximally. This deposit would then be overrun by the glacier ice front and subjected to deformation and compaction. The fine distal sediments of the supraglacial-englacial till would eventually be covered by the coarser sediments of the basal tills. Therefore a site that had experienced glacial advance should show a sequence of finer sorted sediments overlain by coarse unsorted sediments.

Stillstands especially when the maximum advance of the glacier has been attained, may result in a dumping of sediment concentrated at the ice front. This results in the buildup of sediment into a terminal moraine.The downwasting of the ice surface will have allowed sediment to be concentrated behind the terminal moraine.

The retreat of the ice front (Fig.7.2b) would allow the deposition of the coarse basal tills to continue close to the ice front and the finer supraglacial-englacial tills over a wider area. Therefore the sequence of tills is reversed with coarse grained basal tills being overlain by fine grained supraglacial-englacial tills. In this way a glacial event, with advance and retreat phases, would be documented in the stratigraphy of the glacial tills, provided that sufficent sediment was deposited, and not subsequently eroded. In the model developed above, a proglacial lake would have been present in both the advance and retreat phases. The composition of the lake beds will be controlled by the chemical conditions and the clastic sedimentation occurring in the lake.

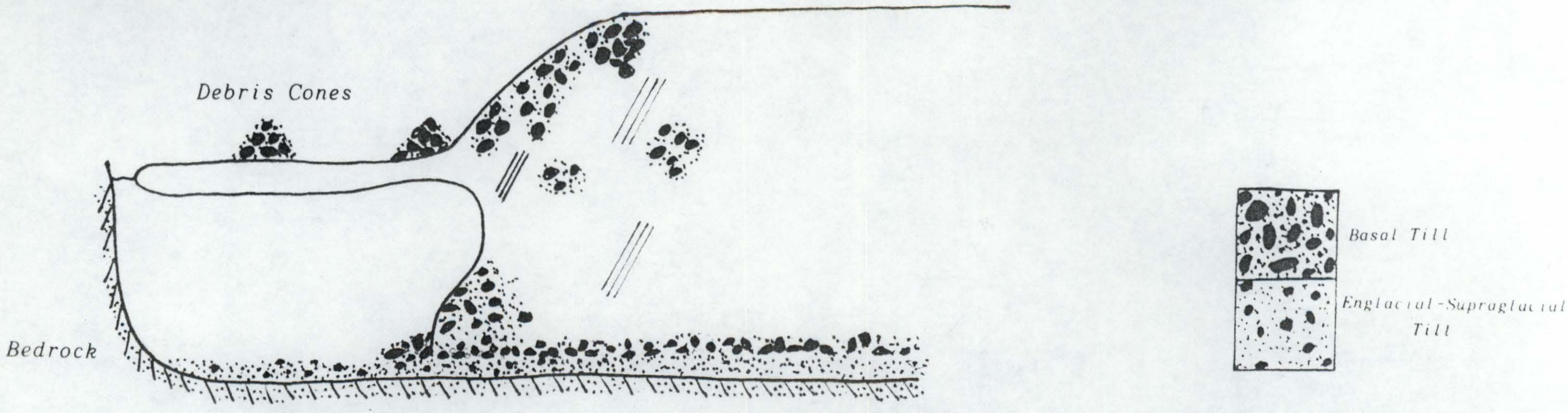


Fig. 7.2a Detrital sedimentation occurring at a advancing glacial ice front

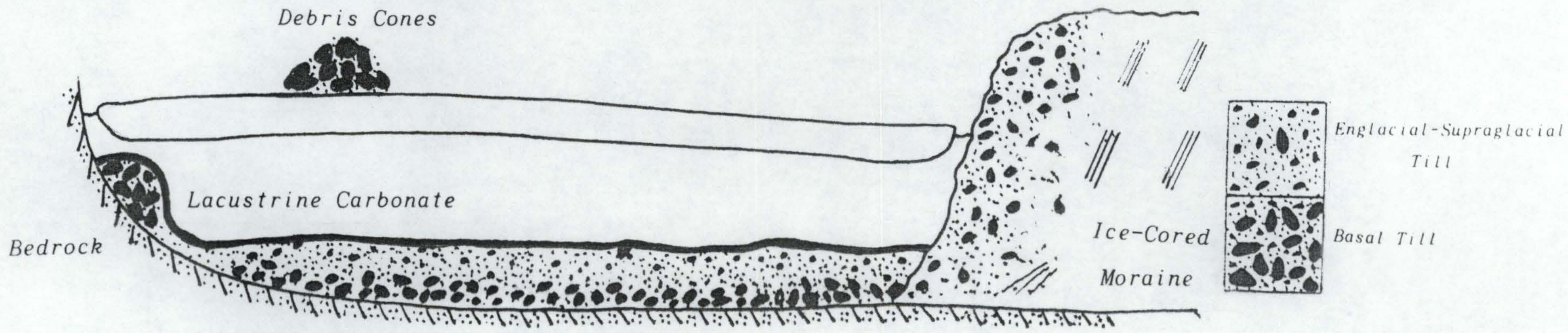


Fig. 7.2b Detrital sedimentation occurring at a stillstanding glacial ice front.

The carbonates are several centimetres thick and if formed at rates observed in Lake Fryxell, would take many decades to accumulate. They and their overlying gypsum are relatively free of clastic sediments, and suggest that little sediment was being transported into the valley at this stage. Advance phases of glacial incursions over a frozen base are likely to be characterised by an ice front bearing a very minor sediment load, which would allow the low sedimentation needed for lake bed deposition. By contrast, during retreat phases, especially those involving downwasting, sedimentation is characterised by the deposition of fine sediments into the lake so that only in the extreme distal areas of the lake, would clastic sedimentation be low enough for carbonate deposition to occur.

At the end of a retreat phase, the resulting stillstand and downwasting of the glacier ice front, may result in the build up of a thick lag over the ice as a letdown till is developed. This would significantly reduce the meltwater flow and and may severely limit the input rate of clastic sediments into the lake, again creating the clear water column necessary for carbonate deposition.

Chemical Conditions Necessary for Lake Bed Deposition.

Assuming the occupation of a proglacial lake with low clastic sedimentation, a model is constructed for the deposition of calcium carbonate and gypsum lake beds. Several conditions must be met:

1)Possible Sources of Lake Bed Salts.

Two lacustrine salts are found in Marshall Valley deposits. These are gypsum ($CaSO_4.H_2O$) and calcite ($CaCO_3$). A model of glacial lake formation has to be developed that has a source which can supply these salts. Several possible sources exist:

1) Seawater frozen into the glacier ice.
2) Salts acquired by the glacier from outside of the valley.
3) Weathering of the sediments and basement from within Marshall Valley.

Seawater Frozen into the Glacier Ice

Ca^{2+}, $SO_4{}^{2-}$ and $CO_3{}^{2-}$ are present in sea water. In our model the Ross Ice Sheet thickens and flows across M^c^Murdo Sound towards the Dry Valleys. During this process, it is possible that sea water becomes incorporated by freezing onto the base of the ice, or is trapped in the ice marginal embayments to be frozen onto the ice sheet margin. If incomplete freezing occurred, calcite, gypsum and mirabilite will be separated from the freezing seawater at an early stage, leaving chloride rich brines to drain away. The resulting ice of mixed origin is then transported into Marshall Valley. The later downwasting of the glacier ice releases the salts into the glacial lake.

If we consider the largest possible extent of a past glacial lake in Marshall Valley to be 10 km^2, we can approximate the volume of salts deposited in the valley. It is estimated that there is 4×10^8 kg $CaCO_3$ deposited in the valley.

If we first consider the supply of sufficent Ca^{2+} needed for deposition we must freeze 4×10^7 m^3 of sea water into the glacier ice.

The most limiting component to consider is the $CO_3{}^{2-}$ ion. This is because it has the lowest concentration in seawater of the three ions. For an adequate supply of this ion 1.3×10^9 m^3 of seawater must be incorporated into the sea ice. This is equivalent to freezing a layer 1m thick over an area of 1,300 km^2.

If we consider the area of seawater in the M^{c}Murdo Sound between Ross Island and the Dry Valleys, we have a sea ice surface of 3,000 km^2 in which to freeze in seawater. Therefore this source of lake bed salts is possible.

Salts Acquired by the Glacier from the Outside of the Valley

The second possible source of salts is from sediments incorporated in the glacier ice. As the thickening ice sheet increases in volume it will ground on the sea floor and erode the beds deposited there. The beds eroded and subsequently frozen into the sea ice may include salts frozen out of the chilled seawater. As the temperature of sea water drops, the solubility of the seawater components is decreased, causing first calcite and gypsum to precipitate out and then mirabilite to precipitate (Thompson and Nelson,1956). These salts would be subsequently incorporated in the ice sheet as sediment inclusions rather than the frozen liquid seawater as suggested in the first source.

These two sources should be considered together as one possibility as it is likely that both occur together in the formation of thickened sea ice, isolated from the open sea by an advancing ice sheet. As the freeze out salts may accumulate over a considerable period of time, the area required to provide the abundance of salts found in Marshall Valley is very much smaller than for direct freezing.

Weathering of the Sediments and Basement from within Marshall Valley

The third possible source is the weathering of basement rocks and sediments present in Marshall Valley and the addition of these salts into the lake system from streams. We shall consider the supply of Ca^{2+} into the lake.

A possible source of Ca^{2+} is marble identified in the suite of metasedimentary rocks found in Marshall Valley. However they form only a small proportion of the metasedimentary basement rocks, principally confined to the extreme landward end of the valley and valley walls at high altitudes.

In reconnaissance of the upper reaches of the Marshall Valley Stream, large blocks of marble were found in the stream adjacent to the outcrops of marble. However the expected quantity of cobbles and pebbles of marble (the result of stream action and weathering) are not found farther down the stream.

Because the exposed marble showed little visible evidence of weathering, (ie. no macroscopic pitting or rounding of the frost shattered edges) it is likely that at present the marble does not release a large supply of Ca^{2+} into the stream system.

In the 1983-1984 field season we collected a weathering deposit found on the surface of the valley, on the north side of the stream above Section III. This material was a bright white hard crystalline deposit. It showed evidence of physical and chemical weathering. In hand speciman this deposit suggested that it may have been marble found in a weathering regime. However laboratory analysis by XRD revealed that the Ca mineral was present in the form of aragonite and not calcite as would have been expected if the material had been marble. No trace of calcite was found and so the possibility of the deposit being weathered marble recrystallising to aragonite was not applicable, suggesting that it was another lake deposit.

A second problem with the weathering source for the lacustrine salts is the abundance of gypsum in the lacustrine salts, and the paucity of sulphur minerals in the basement rocks. Further more isotopic analysis of the lacustrine salts reveals a marine origin for the sulphur and the sulphate oxygen.

Gypsum ($CaSO_4.2H_2O$) amd mirabilite ($Na_2SO_4.10H_2O$) occur as common minerals throughout the Ross Sea I drift. Samples of these have been analysed (Lyon,1978, Black and Bowser,1970) and it has been found that the δO^{18} of the sulphate ions are identical to those of seawater, where as the δO^{18} and δD of the waters of crystallisation show that the sulphates were deposited from isotopically depleted glacial meltwaters. Therefore the sulphates, at least, have not been derived from weathering, and have not been emplaced as a direct deposit from a marine incursion. This means that salts are likely to have been derived indirectly from seawater.

Conclusions on the Origins of Lake Bed Salts.

The mechanism of freezing seawater and/or the freezing of seawater evaporites appears to be a mechanism that could supply all the volume of evaporite salts needed. However in considering the geochemistry of several aspects of the lake beds, we find that a seawater source cannot completely explain the results of such analysis.

Firstly the salts deposited, consist of carbonates and sulphates and lack the great bulk of marine salt components (sodium, magnesium, chloride etc). It is notable however that sulphates and carbonates are among the first salts to be deposited from freezing seawater, so that their abundance in Marshall Valley could be the result of fractionation through partial freezing.

Secondly the U^{234}/U^{238} ratios average 1.5, whereas the ratio in seawater is 1.15. Elsewhere in the M^{c}Murdo Dry Valley, ratios as high as 4.0 have been observed. The enrichment of U^{234} occurs as a result of the leaching of U bearing minerals with the preferential removal of U atoms which have undergone the process of α recoil. Thus at least 25% of the uranium present in the carbonates of the 130,000 to 180,000 year old lake beds and up to 50% of the uranium in the Ross Sea I lake bed has been derived through leaching.

Origins of the lacustrine waters.

Despite problems with the origins of the salts, the O^{18}/O^{16} ratios of the carbonates and the water of crystalisation of the gypsum show that they have been deposited from isotopically depleted meltwaters. Table 7a shows the δO^{18} values of the carbonates and the water of crystallisation, together with calculated values for the water from which they crystalised.

O^{18}/O^{16} ratios of oxygen containing minerals, reflect the O^{18}/O^{16} ratios of the waters from which they were derived. Likewise the D/H ratios can also be used to derive the composition of the original waters. Whenever a mineral is precipitated from solution, a fractionation of isotopes is possible, and this must be taken into account in the estimating of the composition of the original water. Calcite and aragonite preferentially accumulate O^{18}, and this fractionation is strongly dependant on temperature, with the fractionation decreasing by approximately 0.24‰/ 0°C. Thus the composition of the water from which calcium carbonate has precipitated can be given by

$$\delta^{18}O_{H_2O(SMOW)} = \delta^{18}O_{CaCO_3(PDB)} \ -3.25\ ‰$$

if the temperature was assumed to have been 4°C (Epstein *et al*,1953, Epstein, 1976).

The fractionation of oxygen isotopes between the waters of crystallisation of gypsum and the solution from which it had precipitated is less well known, with Gonfiantini and Fontes (1963) giving

$$\delta^{18}O_{H_2O(SMOW)} = \delta^{18}O_{gypsum\ hydrate\ (SMOW)} \ -3.49‰$$

Table 7a. Oxygen isotope ratios for carbonates and gypsum found in Marshall Valley.

		Carbonate	**	Sulphate	Sulphate	
Bed	Sample	$\delta_{18}O_{PDB}$	$\delta_{18}O_{SMOW}$	$\delta_{18}O_{SMOW}$ woc	$\delta_{18}O_{SMOW}$ woppt	δD
A-B	1	-41.2	-44.5			
A-B	2	-41.0	-44.3			
C-D	3	-37.8	-44.1			
C-D	5	-36.7	-40.0			
C-D	6	-39.2	-42.5			
C-D	7	-34.8	-38.1			
A-B	8	-38.8	-42.1			
A-B	9	-40.2	-43.5			
C-D	10	-41.2	-44.5			
C-D	11	-39.1	-42.4			
A-B	12	-39.4	-42.7			
C-D	14	-39.2	-43.2			
C-D	16	-33.1	-36.4			
E	18	-41.2	-44.5			
E	19	-38.9	-42.2			
F	20			-32.7	-36.2	-267
F	21			-34.8	-38.3	-280
F	22			-34.0	-37.5	-275
	23	-33.1	-36.4			
	24	-40.8	-44.1			
F	25			-34.6	-38.1	-276
F	26			-32.9	-36.4	-269
A-B	28			-39.1	-42.6	-310
A-B	30	-41.8	-45.1			
E	31	-39.2	-42.5			
	32			-40.1	-43.6	-316
A-B	33	-42.0	-45.3			
	34	-33.4	-36.7	-39.6	-36.1	-312
	35			-38.1	-41.6	-312
	36	-40.2	-43.5			
A-B	37	-39.7	-43.0			
A-B	38			-42.0	-45.5	-321
A-B	39	-26.7	-30.00	-37.5	-41.0	-310
E	40	-41.9	-45.2			
	41	-40.6	-43.9			
A-B	43			-39.5	-43.0	-284
A-B	44	-25.4	-28.7	-39.0	-42.5	-315
E	45	-40.0	-43.3			
A-B	46	-30.1	-33.4			
A-B	47	-40.1	-43.4			
A-B	48	-42.4	-45.7			
A-B	50	-38.4	-41.7			

** δO_{18} of the water in equilibrium with the carbonates.

woc water of crystallization.

woppt water of precipitation.

Table 7a.(cont)Oxygen isotope ratios for carbonates and gypsum found in Marshall Valley.

Bed	Sample	Carbonate $\delta_{18}O_{PDB}$	** $\delta_{18}O_{SMOW}$	Sulphate $\delta_{18}O$ woc SMOW	Sulphate $\delta_{18}O$ woppt SMOW	δD
E	53	-43.6	-46.9			
A-B	54	-32.4	-35.7			
A-B	55	-39.1	-42.4			
A-B	59	-40.3	-43.6			
A-B	60	-40.1	-43.4			
A-B	61	-41.2	-44.5			
	62			-33.1	-36.6	-273
A-B	64	-39.2	-42.5			
A-B	65	-41.6	-44.9			
A-B	66	-41.7	-45.0			
A-B	67	-42.2	-45.5			
A-B	68	-40.6	-43.9			
E	69	-39.3	-42.6			
E	70	-29.6	-32.9			
A-B	71	-41.9	-45.2			
A-B	72	-39.6	-42.9			
A-B	73			-39.5	-43.0	-312
A-B	74			-39.9	-43.4	-317
	75	-34.5	-37.8			
A-B	76	-28.5	-31.8			
	77			-39.5	-43.0	-316
A-B	79	-41.7	-45.0			
A-B	83			-41.0	-44.5	-323
A-B	84			-37.7	-41.2	-319
A-B	85	-42.4	-45.7			
F	86			-35.0	-38.5	-290
	87	-42.4	-45.5			
	88	-38.6	-41.9			
F	89			-27.0	-30.5	-248
	90	-38.5	-40.7			
A-B	91	-41.9	-45.2			
A-B	96	-41.8	-45.1			
A-B	97	-40.3	-43.6			

** δO_{18} of the water in equilbrium with the carbonates.
woc water of crystallization.
woppt water of precipitation.

between temperatures of 17-57 °C

and Fontes and Gonfiantini (1967) giving

$$\delta^{18}O_{H_2O(SMOW)} = \delta^{18}O_{\text{gypsum hydrate (SMOW)}} - 4.00‰$$

and

$$\delta D_{H_2O} = \delta D_{CaSO_4.2H_2O} + 15‰$$

(Fontes and Gonfiantini,1967)

The calculated δO^{18} values of the waters from which lacustrine carbonates of Ross Sea II age (A-B, C-D, E) is -δ43.8‰ and δD values of -300‰ wrt SMOW. These ratios are considerably more negative then any current environmental water (see Table 7.b) and are typical of snow precipitated on the East Antarctic Ice Sheet at altitudes of ≃2,500m.

In contrast the Ross Sea I glacial advance into the Marshall Valley did not result in deposition of any lacustrine carbonate, although it did cause gypsum to be deposited. The waters of hydration of gypsum from lake bed F indicate that this gypsum was deposited from waters with a mean δO^{18} value of -37.0‰ and a mean δD value of -257‰ wrt SMOW

The Miers Valley was also occupied by a Ross Sea I lake, from which both gypsum and calcite were precipitated (Clayton-Greene,1986). These appear to have been deposited in part through evaporation, but a large proportion of the samples cluster around the most negative end of the range and the calcite indicate a δO^{18} value for the lake waters of -36.7 ‰. It is notable that the oldest of these carbonates (C-14 age 26,000 yrs B.P.) had the lowest δO^{18} values, with the youngest (C-14 age 14,700 yrs B.P.) having the most positive δO^{18} values ($\delta O^{18}{}_{H_2O\ SMOW}$ -26.7‰). A few samples of gypsum have been analysed by Lyons (1978), who found that the waters from which these had precipitated ranged in δO^{18} from -36.6 to -37.2‰, and in δD from -285 to -288 ‰. These gypsums occurred immediately above and in contact with a calcite bed dated 23,000 yrs B.P. Thus the meltwaters which flowed into the Miers Valley showed a common origin with those which flowed into the Marshall Valley during the Ross Sea I Glaciation.

An enrichment in the O^{18} and D can result from evaporation off an open water surface, however plotting the δD values vs. the δO^{18} values for all of these gypsum samples, it can be shown that the samples lie in two clearly separated fields, one of which contains all of the Ross Sea II samples and the other, all the Ross Sea I gypsums. Both of these fields lie close to the Craig meteoric line (Craig,1961) and thus are distinct because the snow from which the glacial meltwaters were derived fell at different altitudes, and not as the result of evaporative concentration.

An examination of the δO^{18} values of surface snow collected from the Antarctic Ice Sheet, as a function of altitude, (Morgan,1982) shows a trend for increasing negative values with increasing altitude. Ignoring samples collected at low altitudes but far from the coast (ice shelf samples), and those collected at extreme northern locations, Morgan's data lies along a line from δO^{18} =-18‰ at sea level to δO^{18} =-53‰ at 3,500 m. Thus had the Ross Sea I and Ross Sea II ice formed at the present day they would have been derived from snow accumulated at 1,900 m and 2,500 m respectively.

Stuiver *et al*,(1981) have presented a recontruction of the late Wisconsin ice sheets covering Antarctica in which they postulate a fusing of the present East and West Antarctic Ice Sheets with the ice sheets extending into the Ross Sea embayment. Flowlines of that icesheet indicate that the meltwaters which occupied the Marshall and Miers Valleys at the time of "maximum glaciation" would have been derived from west of the Tranantarctic Mountains at altitudes of approximately 2,000 m (Fig.7.4).

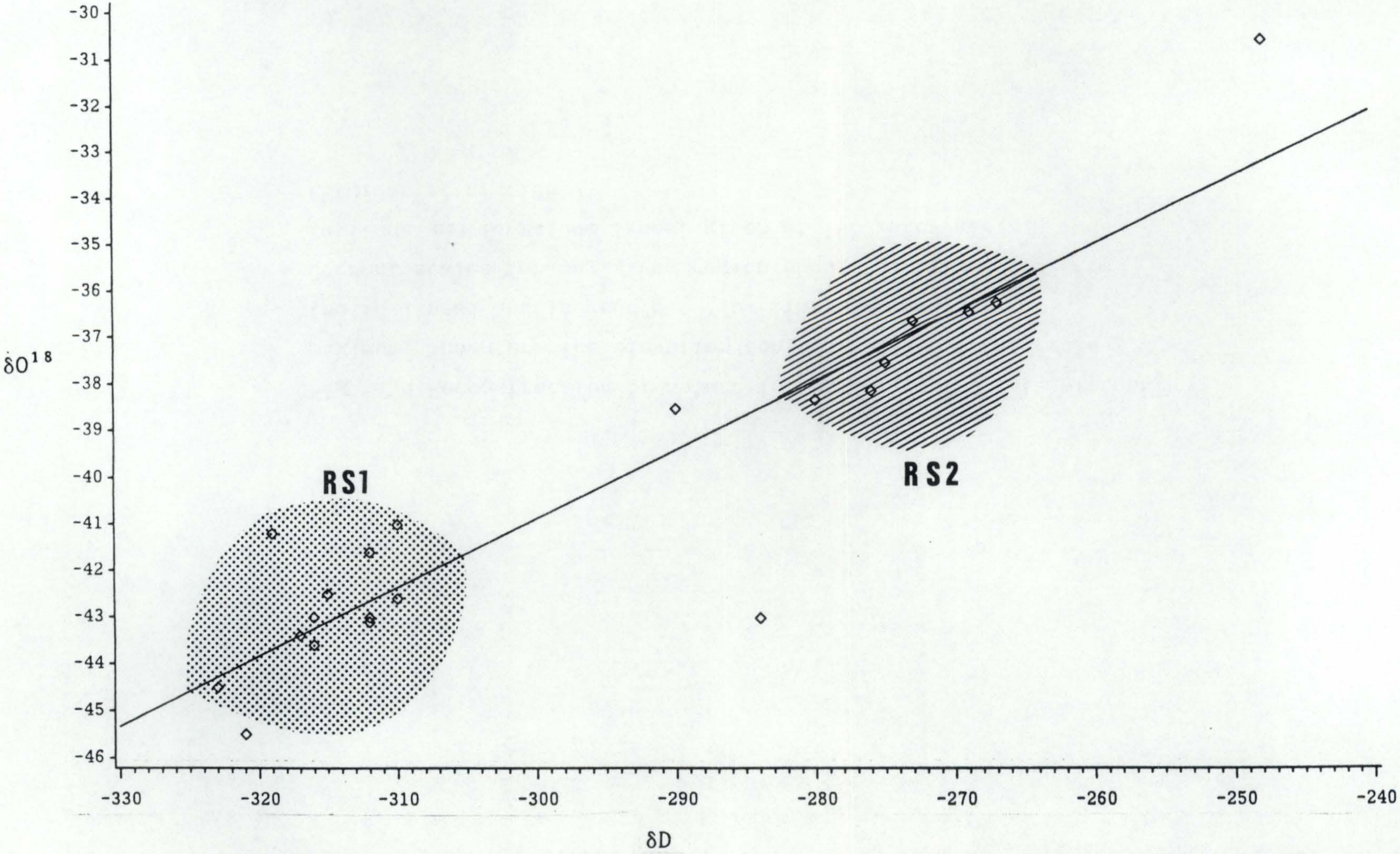
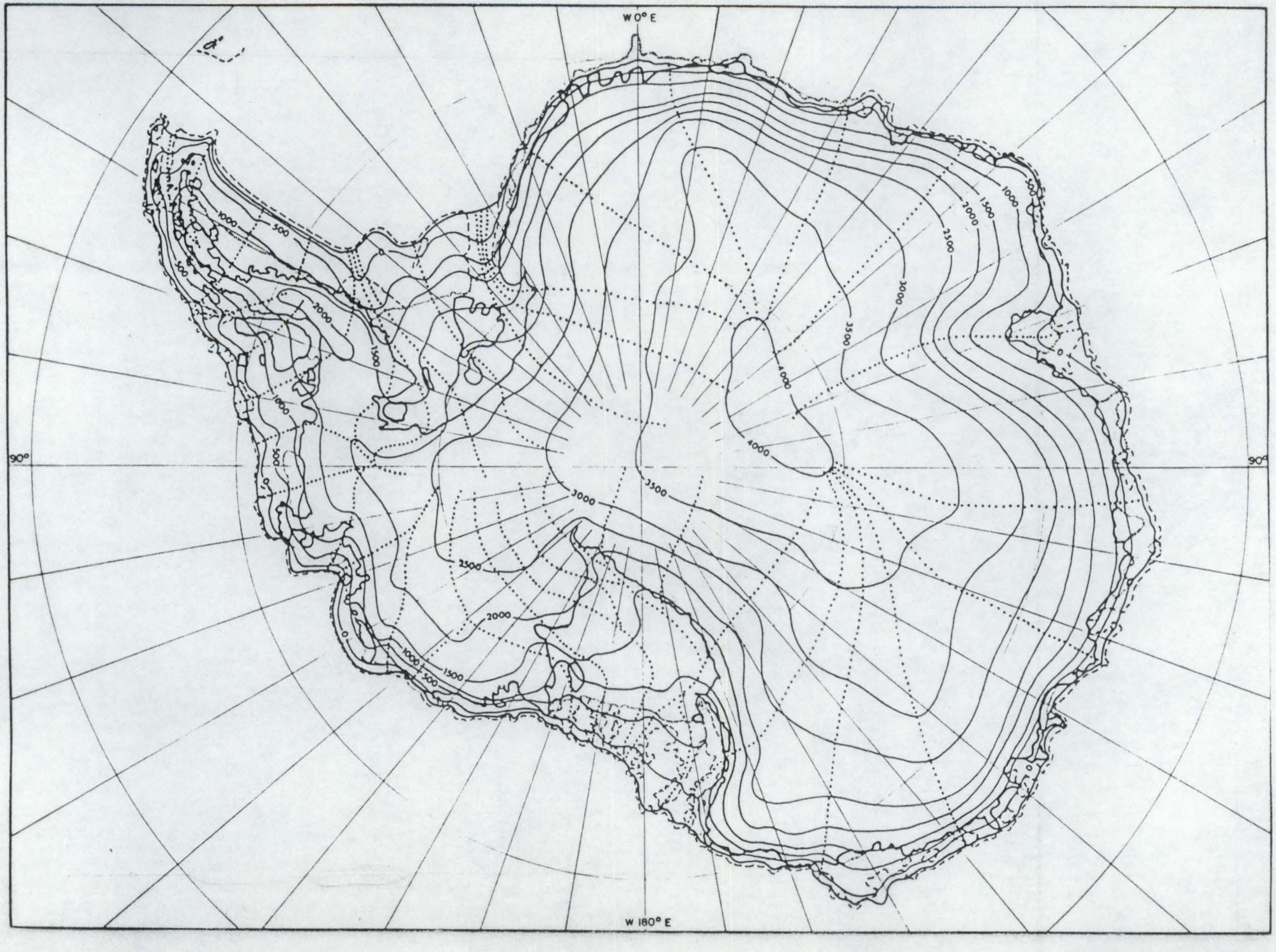


Fig. 7.3 Plot of δO_{18} vs δD for Marshall and Miers Valley gypsums

Fig. 7.4 Reconstruction of Antarctic ice during the late Wisconsin maximum. Shown are ice elevation contours at 500 m intervals (solid lines) and ice-surface flow lines (dotted lines). Correct scales for any given region should be calculated from latitude and longitude lines, given at 10° intervals. (Stuiver *et al* ,1981)


W 0° E
90°
90°
W 180° E
0
500
1000
1500
2000
2500
3000
3500
4000

Thus the δO^{18} and δD values of the Ross Sea I deposits are consistent with the Stuiver *et al*,(1981) model, but those of the Ross Sea II suggest that the resulting ice sheet may have been up to 600 m thicker in southern Victoria Land, or the flow lines for the ice draining into Koettlitz Glacier region may have extended further to the west to capture higher latitude precipitation. In neither case, could local precipitation have produced such isotopically depleted waters.

Mechanisms For Deposition.

If the mouth of the valley is dammed then the natural drainage of the valley is obstructed. The removal of water from the lake system is only possible by evaporation or ablation from the lake surface. The removal of water from the lake system is balanced by the melting of glacier ice and its addition to the lake.

Hence it is the evaporation and ablation rate and the degree of glacial melt that controls the concentration of salts within the lake.

If at sometime the evaporation/ablation rate of the glacial lake is not balanced by the input of melted glacier ice into the system, there will be an increase in the concentration of salts in solution and precipitation may occur. Therefore the continued evaporation of a lake will result in supersaturation of ions and deposition of evaporites.

In a glacial lake there is the added contribution of temperature to the system. A lowering in temperature decreases the saturation point of gypsum and calcite. This means that in polar climates, the change in conditions needed to trigger precipitation need not be as large as in a temperate climate.Therefore a glacial front containing pockets of frozen seawater and selectively precipitated marine salts, damming the valley mouth will provide ions that will become supersaturated, in a climate of high evaporation/ablation and therefore cause precipitation of evaporite lake beds.

The formation of lake evaporites is as a result of stable conditions for a sustained period of time. The stable conditions needed would be a supply of salts into the lake and a sustained high evaporation/ablation rate. This could result at the termination of a glacial lake cycle with the gradual decrease creating a shallow lake in which the saturation point could be exceeded and precipitation occur. The optimum example of these processes is the near complete evaporation of the glacial lake. Stable conditions could also be provided by a stillstand during a glacial advance, as the first ice to occupy a marine basin will remove most of the soluble salts.

In addition to inorganic precipitation it is possible that the conditions of precipitation were achieved by organic processes. The presence of living phytoplankton within the water column or benthic algal communities will have an effect on the saturation conditions in the lake. A glacial lake with a lake ice cover will create a closed system. The ice cover will isolate the water from the atmosphere and therefore there will be no CO_2 exchange. However light will be able to diffuse through the ice and the algae will be able to photosynthesise. The net effect of this photosynthesis is the removal of C in the form CO_2 from the system. This will increase the $ACO_3{}^{2-}$ and this increase may exceed the saturation point of the lake and therefore trigger precipitation.

A third possible mechanism is that precipitation was caused by the addition of one of the ions to the lake, already containing an abundance of the other ion(s), for example the addition of mirabilite to a lake already containing Ca^{2+} ions could cause the precipitation of gypsum by the common ion effect. That this glacially induced precipitation should have occurred during a period of very low clastic sedimentation would seem fortuitous.

Thus the environments for deposition of lake beds differ between the various precipitation mechanisms.

3) Triggering of Mechanisms.

Depending on the mechanisms controlling lakebed formation there are several possibilities as to the timing of such deposition.

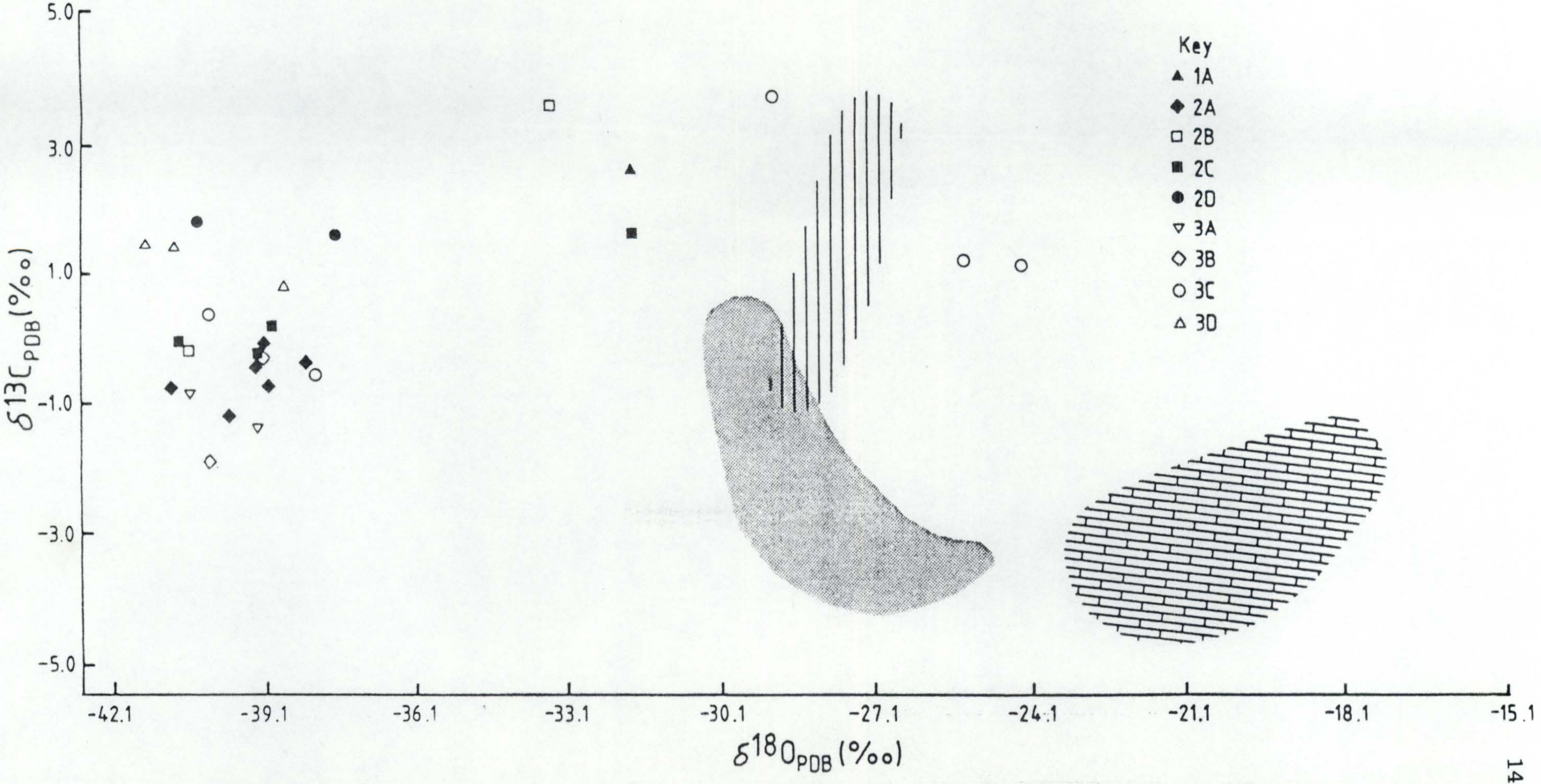
Within a glacial period there are seasonal fluctuations in conditions. Therefore the deposition of carbonates in Marshall Valley lakes could be an annual or regular event.

As shown for Lake Fryxell (Lawrence,1982) it would be expected that the summer period would be a time of peak photosynthesis for micro-organisms because of the 24 hours of daylight and the buildup of nutrients in the photic zone during the dark Antarctic winter. However the peak evaporation/ablation rate would be in winter as it is then that the removal of water is achieved by the freezing of water onto the bottom of the lake ice. This is balanced annually by the loss of water by evaporation/ablation at the lake surface. Thus biological precipitation is likely to be a summer event and accompanied by melting and clastic input, while evaporative precipitation would be a winter event and out of phase with a meltwater input of clastic sediments.

The biological triggering could occur within a deep water column, in conditions which fluctuated regularly to create a brief period of supersaturation and precipitation.

Lawrence (1982) showed that a plot of δO^{18} vs. δC^{13} could be used to distinguish between carbonates precipitated by phytoplankton depletion of CO_2 and those precipitated by evaporation. The range of

Fig.7.5 Stable isotope Plot for Marshall Valley carbonates in relation to Lake Fryxell (Taylor Valley) carbonates. Lake Fryxell unit B = rectangles field, Unit D = stippled field and Unit E = lined field. Most of the Marshall Valley samples plot about 0 ‰ PDB for δC^{13} but are isotopically lighter in δO^{18} than Lake Fryxell carbonates. The six samples which plot nearest to Lake Fryxell fields are all carbonate cements Lawrence, 1982).



Key
▲ 1A
◆ 2A
□ 2B
■ 2C
● 2D
▽ 3A
◇ 3B
○ 3C
△ 3D
5.0
3.0
1.0
-1.0
-3.0
-5.0
$\delta^{13}C_{PDB}$ (‰)
-42.1
-39.1
-36.1
-33.1
-30.1
-27.1
-24.1
-21.1
-18.1
-15.1
$\delta^{18}O_{PDB}$ (‰)

δC^{13} values obtained for the Marshall Valley carbonates lie within the field demonstrated by Lawrence (1982) to enclose the phytoplankton carbonates. Evaporites in Lake Fryxell showed progressive enrichment of O^{18} and depletion of C^{13} up through the beds. Such a trend is not apparent in any of the Marshall Valley carbonates (see Figure 7.5). So it appears that in the Marshall Valley at least carbonates were deposited as the result of biological CO_2 demand.

It is likely that the evaporation, freezing, sublimation process of deposition of lake evaporites occurred in the period well after the peak size of the glacial lake has been achieved, and is the result of the continued decrease in size of the lake by evaporation/ablation. It is also likely to occur in the surface waters close to the freezing front, and the resultant precipitate may not survive its descent to the bottom of the lake.

An investigation of the morphology of the lake bed deposits may suggest the conditions for their deposition.

Evidence for Mechanisms of Deposition from Calcium Carbonate Deposits.

The calcium carbonate lake beds found in the valleys exhibits several morphologies.

Several units of tabular calcium carbonate are found in the sections . The thickness of these beds varies from several millimetres to centimetres (Plate 7.I.). Calcium carbonate is also present as a micrite cement surrounding gypsum moulds and as cemented sands.

The tracing of these tabular calcium carbonate units in section, reveals that it is an extensive deposit. In reconnaissance of the surface of Marshall Valley, calcium carbonate beds were found cropping out in the surficial sediments. On the north side of the valley, calcium carbonate clasts were found on the valley slopes to a level

Plate 7.I Tabular calcium carbonate lake bed (photo C Hendy)

just below what appeared to be deltaic features (estimated elevation 400 m). On the south slopes of the valley calcium carbonate clasts were traced to an altitude that was below the level attained on the northern slope. These are not seen in section however and could possibly be glacial erratics.

As discussed in Chapter 4 the analysis of these calcium carbonate deposits reveals that the carbonate contains 2 polymorphs of calcium carbonate; calcite and aragonite. XRD analysis allows the relative abundances of each polymorph to be compared between lake beds. The trend emerging shows that aragonite becomes the more dominant mineral the younger the beds become.

With the use of SEM photography the morphology of the calcium carbonate lake beds were studied in greater detail. As shown in Chapter 4 the SEM photography reveals carbonates that have varying compositions of rhombic calcite crystals with acicular aragonite crystals or purely aragonite needles.

In general two distinct forms were recognized:

1) aragonite crystals that were randomly orientated and showed overgrowth or are entirely surrounded by calcite.

2) aragonite needles that showed radiating growth.

It is these distinct forms that may explain the deposition of the carbonates.

Aragonite is more soluble then calcite, however in highly supersaturated conditions the aragonite will precipitate before calcite. Therefore within a glacial lake with supersaturation conditions, aragonite will precipitate in the water column. The acicular needles of aragonite will fall through the water column to the lake bottom, where this rain of crystals will become a layer of randomly orientated crystals. Having lowered the saturation conditions of the lake, the deposition of calcite is favoured, which will infill around and enclose the aragonite needles. Repetitions in this order of deposition produced a laminated tabular calcium carbonate bed, in which can be found distinct bands of acicular aragonite crystals infilled with calcite rhombs. SEM examination does not show pitting of the aragonite needles, and hence it does not appear that the aragonite has undergone recrystallisation to the more stable calcite.

This mode of deposition suggests that the conditions of saturation within the lake must vary considerably with time. The bands of aragonite infilled with calcite, supports the theory of an environment with a relatively deep water column (measured in tens of metres), in which aragonite crystals settle in a randomly orientated position and the level of saturation conditions varies with a regular cyclicity. This appears to support the first mode of deposition where the precipitation of lake beds is the result of the interplay of the supply of the evaporite salts, and the degree of saturation of the lake together with the seasonal lifecycle of the micro-organisms found in the lake.

The second form of carbonate recognised, describes the cements found in the section. The cements are nearly pure aragonite and the crystals exhibit a radiating acicular habit into free space. This suggests that they have grown in place under conditions of intense saturation. This would be possible if the growth of the cement crystals was by evaporation. This would be the result of the near evaporation of the proglacial lake or evaporation from the sediments of a drained lake.

Evidence for Mechanisms of Deposition From Gypsum Deposits.

Gypsum units found in the stream sections have several morphologies. In Sections II, III, and IV, the gypsum crystals occur immediately above lake bed A-B, and are in the form of selenite. They form layers that are one crystal thick and measure up to 6 cm in length. Other units in the Sections I and II show very fine layers 2-3 cm thick which have selenite crystals interbedded with sands. The selenite crystals are measured in millimetres. Indications of the former presence of gypsum are found in the sections in the form of moulds. Cavities are found in grey micrite cemented beds and based upon the shapes of the crystal cavities these are identified as having been made of large (up to 2 cm) gypsum crystals (Plate 7.II.).

At the eastern end of Section III a transition occurs between selenite crystals encased in micrite, and moulds, with partially dissolved selenite crystals remaining in a honeycomb of aragonite (Plate 4.XVI, Chapter 4).

The large selenite crystals exhibit growth in 2 directions around a layer of medium sand. This deposition shows evidence of having grown in a downward direction into the sediments below as well as upwards into as overlying lake (Plate 7.III).

All the types of gypsum exhibit discrete crystal forms. This suggests that in order to grow to their respective sizes with such regular crystalline shapes, they require a sustained period of stable conditions and high saturation conditions. Conditions of a shallow depth of water would create high saturation and stable hydrological conditions.

Also the distribution of the selenite beds within the sections suggest that at times they occupied the entire length of the lake, however, the largest crystals are preserved at the highest altitudes to which lake bed A-B extends. With decreasing altitude (greater lake depth) the crystals diminish in size and preservation, suggesting that they were deposited by evaporative concentration, which would have been most effective in the shallow waters near the lake margin, and least effective in the deep waters closer to the glacier margin.

Plate 7.II Micrite cemented beds showing crystal cavities
(photo C Hendy)

Plate 7.III Units of selenite crystals exhibiting two directions of growth (photo C Hendy)

This could have occurred in two ways.

Firstly precipitation could have occurred when the lake was near its maximum extent with the thickness of the selenite and its degree of precipitation depending on the depth of the overlying water. This requires that the entire lake became supersaturated.

Secondly, precipitation may have been limited to the shallow waters, close to the lake margin, and the decreasing thickness with altitude reflecting the changing chemistry of the lake waters, as the lake drained or evaporated to progressively smaller size. If this were the case it would be expected that there would be a wedge of clastic sediments between the carbonate bed and its overlying selenite crystals, which would thicken down valley, and represent increasing time between the deposition of carbonates and deposition of sulphates. Unfortunately clastic lacustrine sediments associated with lake bed A-B are relatively thin throughout the entire length of the exposure, so that it is not possible to distinguish between the two options in this manner.

Model of Lake Bed Deposition.

It would seem that the deposits represent opposing methods of formation.

However both mechanisms for carbonate and gypsum deposition require some period of increased saturation conditions. The low abundance of clastic sediments, requires a stable ice front, since either an advance or retreat is likely to introduce sediment into the lake, or distort the carbonate or gypsum bed.

The first of the models presented above would probably have occurred at a stillstand during the advance of the ice into the valley,(since the bed A-B overlies an eroded surface and has itself been immediately overrun) when the lake was at near maximum extent.

In the second case, the carbonates would have been deposited in the lake near its maximum extent, but the gypsum, during the contraction of the lake.

Although the isotopic data points to a biological precipitation of carbonates, the paucity of diatoms and the association with gypsum would favour an evaporitive concentration. It is possible that both mechanisms work simultaneously.

The evaporitive mechanism which explains the deposition of gypsum, represents the final depositional event in the termination of the glacial lake. This occurred as the depth of water in the glacial lake became shallow . This deposition resulted in the growth of selenite crystals. In conditions of intense supersaturation and in stable hydrological environments, it is possible through evaporation to grow large selenite crystals. In such conditions the growth of the crystals may be achieved in decades.

However the relative positions of the evaporite units in the sections, allows the hypothesis of an order of deposition that can be satisfied by the fluctuations of the water levels of the lakes.

As suggested previously it is possible that the two separate mechanisms of deposition occurred at the same time in a glacial lake in the terminating phase as a result of the depth characteristics of the lake. Calcite would be deposited in the central portions of the lake at the appropriate depth whilst the first deposits of selenite would be deposited at the lakeshore in the very shallow water. As the volume of the glacial lake decreased the area of calcite deposition would become more centralised and would ultimately decrease. The deposition area of the gypsum would also become more centralised but would increase with the further creation of shallow water conditions.

Stratigraphic Evidence for Depositional Mechanisms.

Section III.

In Section III the Unit 3III exhibits the layering : from top to bottom, a matrix supported slurry of variable thickness, consisting of fine (< 1 mm) gypsum crystals, sand and occasional pebbles, in a porous calcium carbonate matrix; 1 - 4 cm selenite crystals having their long axes vertical; and 0.5 -1 cm thick tabular calcium carbonate.

The deposition of this layer within this unit can be explained by the depth of water present in the lake at the time and the subsequent effect on the saturation points of the evaporite salts.

In order of their deposition; the tabular calcium carbonate exhibits a randomly orientated rain of aragonite needles infilled with rhombic calcite crystals. This is consistent with a lake which through the introduction of salts and loss of solution by evaporation/ablation is undergoing fluctuations in the saturation of its solution. The seasonal growth of phytoplankton may be the triggering mechanism for deposition. This would produce alternating conditions for the deposition of aragonite and the calcite.

The next deposition of selenite with its large crystal size and regular crystalline growth suggests that the glacial lake has entered a phase of prolonged decrease in its volume . The thickness of the deposits of gypsum occurred in shallow water, when a small loss of water due to evaporation would have resulted in a considerable increase in the degree of supersaturation, and a prolonged stable hydrological regime necessary for the growth of larger selenite crystals.

The third and top most layer of calcium carbonate cement with small selenite crystals within, may represent a change of chemical conditions rather then a great fluctuation of the depth of the solution. The micrite is composed solely of aragonite which in many cases completely encloses the selenite having been deposited immediately after, or in some cases concurrently. The O^{18}/O^{16} ratios are slightly enriched in O^{18} compared to the underlying carbonate beds, which would be consistent with evaporitive precipitation. However the cements which occasionally extends several 10's of centimetres above this layer into the overlying drift, are considerably enriched in O^{18} and has a quite different origin. These have a O^{18}/O^{16} ratio much closer to that of local precipitation and have probably been formed from meltwaters or stream waters draining down through the drift.

Section IV.

The complicated interplay of depth of water and saturation conditions may also explain the transition of tabular calcium carbonate to micrite cemented gypsum to selenite crystals seen in Section IV. Section IV appears to lie at the furthest extent of the largest glacial lake . As the lake bed A-B rises to the valley surface within the section, exhibiting a transition from carbonates and gypsum to gypsum , it is likely that the lake beds found in the section are close to the lake shore. The transistion of evaporites can be seen as a response to the depth of the lake bottom.

The tabular calcium carbonate is representative of relatively deep water conditions whilst the selenite, the most near shore deposit, was formed at very shallow and saturated conditions. The micrite enclosed selenite moulds represents transitional conditions where selenite crystals were deposited but due to another fluctuation of conditions (possibly the termination of the lake) were redissolved.

Conclusion.

In summary therefore a complete description of either the mode, source of evaporite salts or sequence of deposition is not possible. However with a possible source of salts and an understanding of the saturation points of these salts, it is possible to construct a tentative sequence and mode of deposition.

Of importance to note is the effect that a change of glacial conditions has on the deposition of lake beds in Marshall Valley. The interplay of possible seasonal variations in climate and micro-organism populations alters the supply of lake bed salts, the volume of the lake and the chemical conditions of the lake waters. This may effect the mode of deposition of the calcium carbonate within the lake. A prolonged fluctuation within a glacial period or the termination of a glacial event may also alter deposition conditions to such an extent that gypsum becomes the favoured mineral for precipitation.

The test of this lakebed deposit theory is whether the spatial relationships of all glacial deposits are consistent with the glacial history of Marshall Valley discussed below.

Surficial Features of Marshall Valley.

(Dagel *et al*,in press) has described lateral moraine features which can be traced down to altitudes of 270 m. These features represent the lateral extent of glacial ice on the flanks of the valley mouth and can be traced around the headlands into adjacent valleys, representing the maximum extent of the Ross Sea I Drift. Because of this feature we can estimate that the damming of the valley mouth was at some point sufficient to create an ice wall of at least 20 m in height. The geomorphology of the features seen on the valley walls further inland suggests possible deltas and strandlines relating to past ice margins.

On the valley floor there are several large moraine deposits. The moraines present at the valley mouth have a height of 12 m (Dagel *et al*,in press). Based on their morphology and position relative to the lateral moraines, these floor moraines are thought to occupy a position close to the hingepoint of the glacier. This is assumed to be the point where the glacier ice floated onto the lake surface. The development of these moraines will be discussed later.

7 km further up the valley there are a series of moraines . These moraines have been dissected by the present stream and by a former stream (south of its present position). Because of its dissection the internal structure of the moraine can be seen and several lake beds have been discovered. The moraines series are thought to be evidence of the furthest extent of ice during a previous glacial event (assumed to be Ross Sea II).

A Glacial History For Marshall Valley.

A glacial history can be interpreted from the stratigraphy of the Marshall Valley sections, based on the geochronology obtained from the U/Th dating of the calcium carbonate lake beds found in the sections. From the analysis of the U/Th dates (see Table 3a, Chapter 3) we are able to differentiate 4 separate lake formations within the timespan represented in the sediments. The geochronology extends back to 180,000 years B.P. which is assumed to include from the Ross Sea II Glaciation to the present.

180,000 Year Bed.

The oldest dated bed encountered in the Marshall Valley Sections is a tabular calcium carbonate unit which gives a mean age of 182,000 years B.P. σ 28,000 years. This dated bed is identified at Sections IV, III, II and possibly at Section I.

Similarity exists within the beds morphology and geochemistry as well as the age correlation. Sections IV, III and II all show a similar 3 layer sequence within their lake bed units. Section II contains a calcium carbonate layer sandwiched between two layers of micrite cemented moulds. Section III contains a calcium carbonate layer; a layer of selenite crystals and micrite cemented selenite moulds. The Section IV unit shows a lateral transistion from calcium carbonate to micrite cemented moulds to selenite crystals. This suggest that the beds show a correlation based upon their depositional environment- allowing for lateral fluctuations .

The 180,000 yr beds from Section IV, III and II also exhibit similarities in their mineral compositions. All three units show similar trends in their relative % weight of aragonite versus calcite.

All three units are the oldest lake beds encountered in their respective sections and can be traced with only short gaps in outcrop between sections. They rest unconformably on beds exhibiting a variety of lithologies and structure. Therefore their stratigraphy suggests that they represent a single continuous lakebed present in a lake occupying Marshall Valley at 180,000 years B.P. (Plate.7.IV).

The lake bed dated in Section I; Unit 7I (180,000 years B.P. σ 26,000 years) is only tentatively correlated to the lake beds found in the other sections, as there exists certain discrepancies in its morphology and geochemistry. Unit 7I does not contain a similar 3 layered morphology as found in the other sections. Stratigraphically

Plate 7.IV Marshall Valley Stream Section III (photo C Hendy)

Plate 7.V Marshall Valley Stream Section I (photo C Hendy)

it is not the lowest calcium carbonate bed found in the section. Unit 9I is a calcium carbonate bed found lower in the section. The results of XRD analysis shows that Unit 7I has a different composition to the other proposed correlated units (Plate 7.V).

This evidence places uncertainty on the date obtained from this unit. We suggest that this bed is possibly representative of a later lake. The further dating analysis of both Units 7I and 9I could verify the situation.

Within Section III, the stratigraphy is complex due to the deformation of the beds , this has previously been interpreted as an isoclinal fold (Dagel *et al*,in press). As detailed in Chapter 5, it appears more likely that the folding was due to the gravity sliding of beds down slope on a non moving toe. This has reverse folded the bed but with only minor deformation. This slump could have been triggered by an instability resulting from the retreat of the ice front.The plastic nature of the deformed beds suggests that slumping occurred subaqueously.

From the stratigraphy of the Marshall Valley Sections the presence of the 180,000 year lake beds marks the extent of a large lake that reached the entire length of the sections. This created a lake of dimensions of 3.3 km by 1.6 km by 140 m. The transition of evaporites seen in the lake bed in Section IV has already been discussed in the chapter and may represent the edge of the lake. Here the gradation in depth of water is represented in the type of evaporite deposited.

From the dating of beds in the section we have evidence that the lake extended beyond the large terminal moraines now found at Section III. However due to the transistion in composition of the lake beds in Section IV it is likely that the lake did not extend far beyond this point.

Reconnaissance of the surface to the west of Section IV gives further information as to the extent of the lake. At a point less the 0.5 km west of Section IV the desert pavement changes from a high basalt composition to a high metasediment composition. This change marks the extent of the Ross Sea Glaciations that introduced Ross Island basalt volcanics to the valleys. The metasediment represents the original sediment of the valley. We are unable to date the basaltic drift and so the basaltic pavement may be the result of earlier glaciations or may be a combination of several glacial events. However it does further provide evidence as to the greatest possible extent of the 180,000 year lake.

Analysis of the sediments, associated with the 180,000 year lake bed in the section, are sequences of gravels and sands in mud and sand matrices. All units are poorly sorted. The high percentages of basalt derived material found associated in the deposits is further evidence that this deposit is the result of a Ross Sea glacial incursion. Thus on the basis of the dating of the lakebed and its association with the surrounding tills, the deposits are assumed to be from the previously undated Ross Sea II Glaciation .

Immediately above the 180,000 year old bed, in Sections II,III and IV but below the next lake bed, there is a basalt rich sandy gravelly drift which culminates in a series of pronounced terminal moraines bisected by Section IV. These are unstratifed at the base but exhibit muddy/silty laminations at the top. Dagel *et al* (in press) interpreted this as supraglacial or englacial debris which indicated an ice advance to Section IV, but not beyond, as the 180,000 yr bed forms the desert pavement at the valley beyond Section IV. This deposit may include a thin basal drift overlain by a thicker englacial/supraglacial drift, and was implaced soon after 180,000 yr B.P., as the lake beds have not been subjected to subaerial weathering.

160,000 Year Bed.

The next dated lake bed found in the Marshall Valley Sections is a tabular calcium carbonate with a mean age of 155,000 years B.P. σ 38,000 years. This identified bed is found in Sections II and III.

Similarity exists between the beds in each section in morphology and geochemistry. The weight % compositions show a trend towards an equal calcite:aragonite composition. SEM photography reveals that this bed contains calcite and randomly orientated aragonite crystals, with aragonite crystals infilling cavities and calcite overgrowths on the aragonite crystals.

This morphology suggests that the lakebed was laid down as a rain of aragonite crystals, triggered by a change in saturation point of the lakewaters. Then by the subsequent alteration of precipitation conditions the calcite was laid down around the aragonite.

Section I contains no dated calcium carbonate bed that corresponds to the 160,000 year lake bed.

This lake bed is not represented in Section IV and so allows for the greatest extent of the lake to be mapped. The lakebed is found near the surface of Section III but it is not obvious whether it is below the large terminal moraines overlying Section III. It can be traced on the surface almost to Section IV. The estimated size of the 160,000 year B.P. lake is 1.5 km by 1 km by 100 m. Therefore this lake was of smaller size then the previous lake.

Within Section II there are two lake beds which give mean ages of 142,000 years B.P. σ 19,000 years and 154,000 years B.P. σ 8,000 years respectively. Based on the stratigraphy of the section the two lakebeds could have been discrete depositional events even though they cannot be distinguished by radiometric dating or their geochemistry. Thus they could have been reworked by a glacial advance, as they are encased in basal drift (Dagel*et al*, in press). Identified as beds C and D they could also represent fluctuations within the glacial period.

Ice may have advanced up the valley producing glacial debris deposits and moraines. The subsequent recession and downwasting of ice created a lake and there was the possibility of deposition of lake beds. In this case however no gypsum was deposited.

The stratigraphy of Section II between Units 5bII and 7II represents the fluctuation of an ice advance and retreat. During this period the ice front advanced past the section after the deposition of a lake bed, retreated and was followed by another short advance of ice beyond Section II before the subsequent retreat.

At Section III there is only a single layer of lake bed dated at 160,000 years and therefore the two Section II units represent the furthest extent of this small fluctuation in glacial conditions, and it is probable that Lake Bed D represents glacially transported clasts of Lake Bed C from some location down valley.

The sediments within this glacial sequence contain a high basement rock content but are separated by a lens of basalt sand. Because of the 160,000 year age of the lake bed it is assumed that this lake formation event was also the result of a Ross Sea II Glaciation with M^{c}Murdo Volcanics providing the supraglacial/englacial sediments and Transantarctic Mountain basement rocks the basal sediments.

130,000 Year Bed.

The third dated lake bed is a laminated tabular calcium carbonate which gives a mean age of 129,000 years B.P. σ 12,000 years. This dated unit is identified in Sections III and II. As well as a correlation in dates there exists a similarity between the beds morphology and geochemistry. The lake bed is almost entirely aragonite. SEM photography reveals that the aragonite crystals have grown in place.The uranium analyses (Table 3.a, Chapter 3) show this bed to have considerably higher U content (20 ppm) than beds C-D and A-B (9 ppm).

There is no evidence of a lakebed of this age in the stratigraphy of Section IV and so Section III marks the extent of this lake bed. This dated lake bed can be traced along Section III and the bed crops out on the surficial sediments of the large moraines deposited at Section III.

The 130,000 year old bed overlies the 160,000 year bed at Section III and extends west to within 100 m of Section IV, with only a few centimetres of silts and sands separating the carbonate beds. Thus it can be concluded that no glacial advance overrode Section III between 130,000 years and 160,000 years B.P. As these beds form the lag surface between Sections III and IV, no subsequent glacial advance has penetrated as far up the valley as Section III

Reconnaissance of the area surrounding Section III shows that a similar two layered calcium carbonate crops out on the surface of all the dissected moraines in the region. This field work was carried out in the 1982/1983 field season when Mrs Laurie Doolin (nee M^{c}Leod) produced several sections through moraines as the result of trenches dug in the area.

East of Section III, the 160,000 and 130,000 year beds become buried under a thin layer of basalt rich clasts. One explanation for this may be the transport of sediment by ice rafting. Clayton-Greene (1986) has presented a model of a dammed glacial lake onto which the glacier will introduce frozen sediment. These packets of sediment radiate outward to the edge of the lake by the action of the repeated addition of ice at the ice front, and melting of lake ice at the moat, supported by the continued freezing of ice from the meltwaters below and ablation from the surface (Fig. 7.2b). Therefore as the result of this movement or due to the final recession of the glacial ice, the debris is carried to the moat zone or is left stranded as the lake drains.(Fig. 7.6a and 7.6b) This is the presumed explanation for the patchy shallow deposits found on the surface of Marshall Valley between Section I and II (Plate 7.VI.).

Therefore by the dating and stratigraphy of the sections we can identify a 130,000 year old lake bed which occupied the valley with dimensions 2.5 km by 1.5 km by 100 m. Based on the chronology of the lake bed it is also assumed to be the result of a Ross Sea II Glaciation. This lake represents the last of the lakes found within this timespan.

The history of the terminal moraines may be linked to, the advance and retreat of the Ross Sea glacial ice front. The ice front first advanced into Marshall Valley about 182,000 Years B.P., formed a proglacial lake, then continued to advance to form terminal moraines at Section IV, and deposit basal and supraglacial/englacial drifts containing > 90% basalt. The ice front retreated beyond Section III prior to 160,000 years B.P. and perhaps out of the valley. A second advance occurred at 160,000 years B.P., and a second proglacial lake was formed. The advance continued up the valley overrunning Section II but terminating between Sections II and III. This ice front retreated prior to 130,000 years B.P., when a third lake occupied the valley. This lake evaporated and was not overrun until the advance of the Ross Sea I Glaciation.

Plate 7.VI Ice rafted debris cone

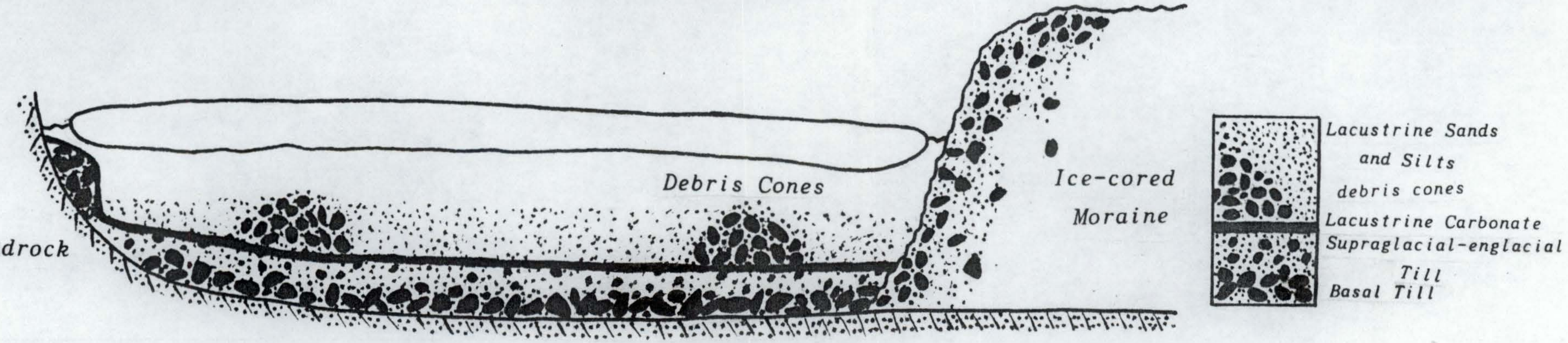


Fig. 7.6a Ice rafting of debris cones

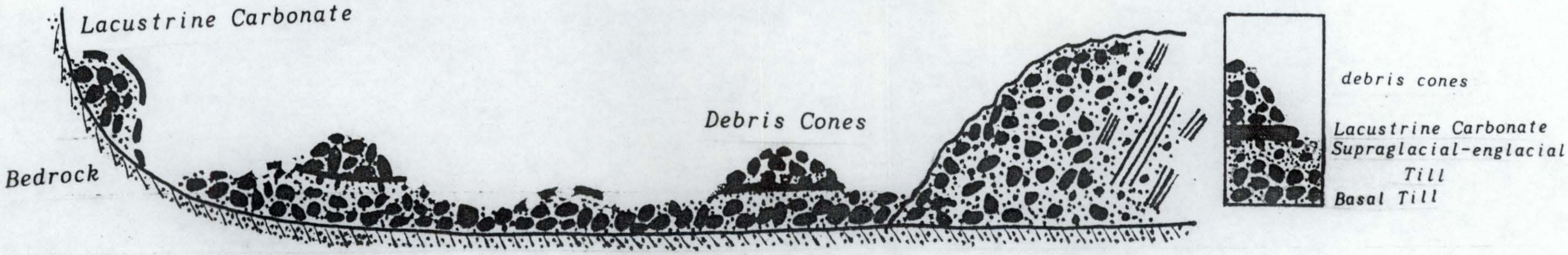


Fig. 7.6b Erosion of lake bed stratigraphy

Ross Sea I Glaciation.

The youngest dated lake bed found in the Marshall Valley stream section is a laminated gypsum unit of mean age 33,000 σ 19,000 years B.P. The dated bed is identified at Sections I and II. This unit is not found in Sections III and IV and this defines the extent of the lake. Further evidence of the limited size of this lake can be found in a survey of the geomorphology of the surface features of the valley.

As discussed previously lateral moraines can be identified on the valley walls and traced down the valley slopes. On the valley floor between Sections II and III there exists a subdued hummocky topography. The desert paving of this topography exhibits a patchy covering of basalt rich sediments which contrast with the less basalt rich surface. The lateral moraine can be traced down the valley slope from the valley mouth and extrapolates to encompass the patchy drift. This marks the extent of the most modern bed.

The subdued topography is hypothesised to have been created by the ice rafting of frozen sediment across a lake surface and the downwasting of ice cored drift. This produces the roughly circular deposits of the sediments previously collected by the glacier ice.

The reliability of U/Th dates derived from gypsum is not as good as those derived from calcium carbonate. However the U/Th dates for Units 2I and 2II agree with the radiocarbon dates from algae taken from the drift deposited on the coastline of Marshall Valley. These dates suggest that the downwasting of ice in the mouth of Marshall Valley was occurring 8,000 to 13,000 years ago at the termination of the Ross Sea I glacial period (Dagel *et al*, in press).

The final lake bed is estimated to have had an extent of at least 2 km by 1 km by 50 m. The basalt rich drift of which the lake deposit is a part, can be traced continuously around adjacent headlands to the type section at the Bulwork (Denton *et al*,1971), and has been identified by Denton as Ross Sea I Drift.The dating of this lake bed confirms that it is representative of a Ross Sea I Glaciation.

Glacial History of Marshall Valley.

In summary: (Fig. 7.7) At a period prior to 180,000 years B.P. there was a period of erosion of the older glacial drifts in Marshall Valley. At about 180,000 years B.P., ice advanced into the valley to form a lake which deposited a sequence of carbonate beds and gypsum (dated 180,000 years B.P.). The lake was extensive and it appears that the tabular calcium carbonate lakebed had a wide distribution throughout the lake. This was followed by an advance up valley to Section IV in which basalt rich basal and supraglacial/englacial drift was deposited throughout the valley.

The next glacial advance dated 160,000 years B.P. extended to Section III. A slump then occurred at Section III. The thick calcium carbonate/selenite unit slid down the slope contorting at the non moving toe into a series of folds.

A lake was then created which was smaller in extent then the last but still contained extensive lake beds. Within this glacial event there may have been a fluctuation in ice advance and retreat and as a result at Section II there are two closely dated beds . These might have been part of the same bed superimposed by glacial erosion.

The last ice retreated from the valley prior to 130,00 years when an evaporating lake occupied the valley . Downwasting of lake ice and ice rafting produced the patchy basalt rich drift that overlies this lacustrine sediment.

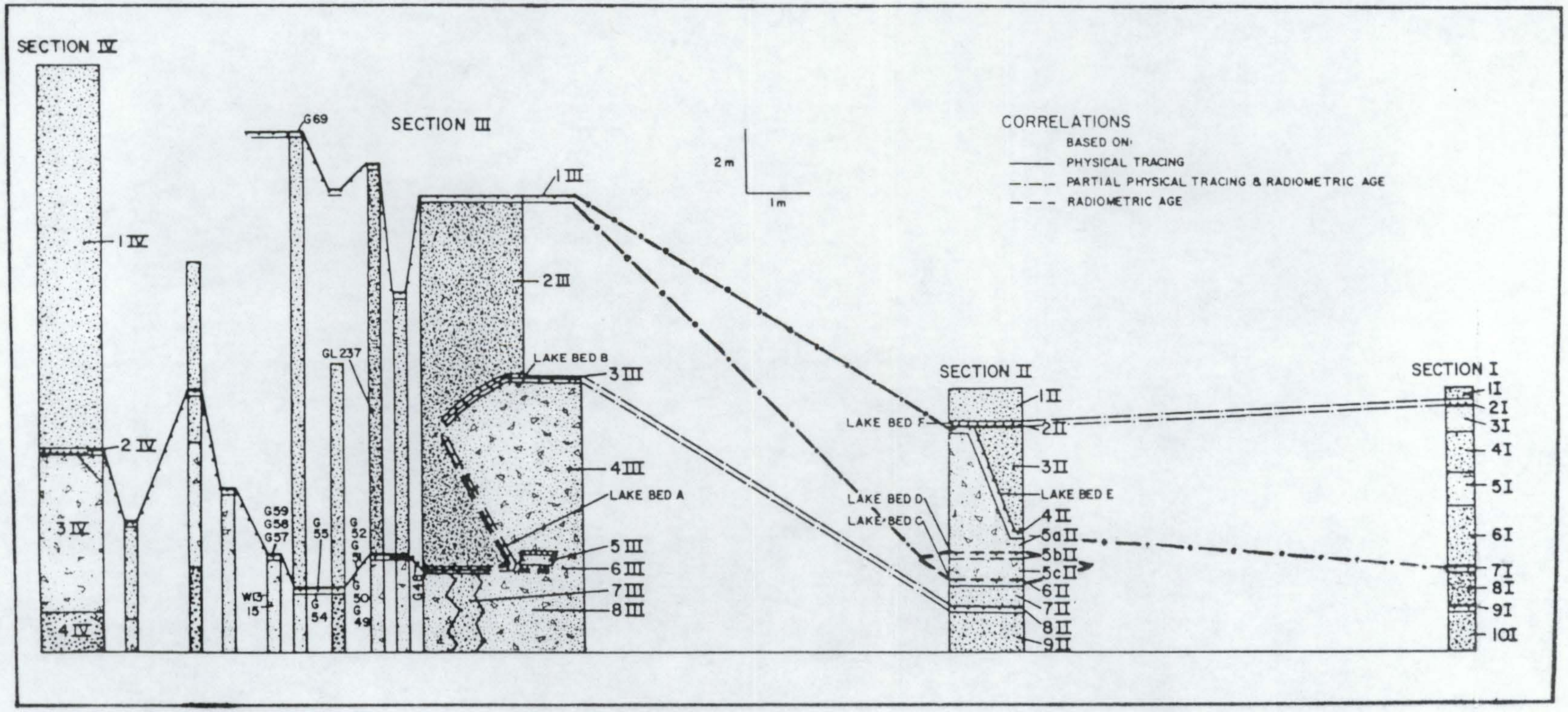


Fig. 7.7 Stratigraphic correlation of Marshall Valley Stream Sections

These three glacial events are all grouped together as parts of the Ross Sea II Glaciation, representing fluctuations within this glaciation.

These two glacial advances exactly fit the dated low sea levels determined from O^{18}/O^{16} analysis of Deep Sea cores (Cuthbertson,1985 , Nelson *et al*, ,1985a,1985b) of 185,000 years B.P. and 154,000 years B.P. The two periods of low sea level apparently initiated the grounding of the Ross Ice Shelf and development of Ice Sheets in the Ross Sea. The sea level started to rise rapidly at 135,000 years B.P., and is recorded in the valley by the deposits of evaporites at 130,000 years B.P.

The last lake present in the valley is represented by the gypsum beds . The lake bed is dated at 15-30,000 years B.P. and was the Ross Sea I Glaciation. In the smallest of the studied glacial incursions, the lake formed, occupied the area east of Section II. This glaciation deposited ice rafted debris between Section I and II resulting in a subdued topography and patchy drift pavement. The Ross Sea I Glaciation also deposited the large area of ice-cored moraines still found at the east end of Marshall Valley. The dating of algae from this drift gives ages which agree with the U/Th dating of this glacial period.

CHAPTER 8

CONCLUSIONS

Chapter 8

Conclusions

1) The Marshall Valley Stream Sections reveal gypsum and calcium carbonate lake beds interbedded between glacial drift. These lake beds can be differentiated from one another on the basis of their stratigraphy and physical morphology, mineralogy and isotopic composition (Table 8.a).

Four discrete lake beds can be differentiated by the use of U/Th dating. The 3 older calcium carbonate lake beds date between 180,000 and 130,000 years B.P. and have been assumed representative of the penultimate glaciation- Ross Sea II. The youngest gypsum bed has a U/Th age of 33,000 years but C^{14} dating of fossil algae gives an age of 14,000 years and represents the last glaciation- Ross Sea I. The O^{18}/O^{16} and D/H ratios of gypsum from Marshall and Miers Valley, indicate that the waters filling the proglacial lakes come from two distinct origins. These coupled with the O^{18}/O^{16} ratios of the carbonates indicate that both originated as snow precipitated at high altitudes, such as would be expected in a fused East and West Antarctic Ice Sheet, 300 km to the south. The Ross Sea I waters were deposited at 1,900 m but the Ross Sea II were deposited at 2,500 m, probably west of the Tranantarctic Mountains.

The Ross Sea II lake beds can be further differentiated from one another through XRD analysis since the beds show increasing aragonite:calcite ratio as they young. S.E.M. analysis also shows distinct trends in the crystal morphology exhibited between the lake beds. The older lake beds contain calcite enclosed, aragonite needles. The trend changes to predominantly aragonite needles grown *in situ*, as the beds young.

Table 8.a Characteristics of Marshall Valley Lake Beds

Lake Bed	U/Th Age	σ	U_{234}/U_{238}	U ppm	% Calcite	% Aragonite	Crystal Habit	selenite/ Selenite moulds	Lakewater $\delta O_{18\,H_2O}$	σ
F	33	19	2.22	3.5	----	------		gypsum	-37.0	2.8
E	129	12	1.44	21.6	5	94	radiating acicular	absent	-43.9	1.6
C-D	155	8	1.51	8.5	53	44	randomly orientated acicular	absent	-42.7	2.2
A-B	182	28	1.49	8.9	73	35	randomly orientated acicular	present	-44.1 -43.4 *gypsum	1.2 1.2*

2) By the use of chemical analysis it is possible to construct several models of the environment needed to deposit the glacial sequence seen in the valley. The lake beds are the result of deposition within a proglacial lake, created by the incursion of the Ross Sea Ice Shelf into the valley.

3) The origins of the salts needed for lake bed deposition have several possibilities. On a quantitaive basis it is possible that salts derived from a seawater origin can supply all the lake bed salts needed. However, analysis of the U isotope ratios of the lake beds show enrichment from the leaching of crustal rocks. This suggests that there is a terrestial contribution of components from the weathering of basement rocks.

4) 3 models of deposition are possible for the lake beds. The interplay of the energy and chemical fluxes within a proglacial lake system, controls the saturation conditions of the lake. An increase in saturation conditions within the lake may trigger the chemical precipitation of the lake beds. The presence of phytoplankton within the water column may control the chemical saturation conditions of the lake biologically and trigger precipitation. The addition of an excess of one of the lake bed components may trigger precipitation by the common ion effect. This third mechanism appears unlikely.

The timing of deposition of the lake beds is dependant on the mechanisms causing deposition. If the deposition of the lake beds is by an inorganic mechanism, then the triggering of deposition is dependant on the balance between the input of meltwaters and salts and the output of meltwater. This mechanism of deposition is therefore dependant on the evaporitive conditions prevailing at the time. Thus deposition is favoured in the winter periods, when there is peak evaporation/ablation. If there is an inorganic triggering mechanism to deposition, then the optimum conditions for precipitation, occur in summer, when there is peak CO_2 demand for photosynthesis by the phytoplankton. Plots of $\delta O^{18}/O^{16}$ versus δC^{13} for Marshall Valley carbonates suggests that deposition was the result of CO_2 demand. However it is probable that increased saturation of the lake by evaporation is still necessary for the accumulation of the tabular calcium carbonate bed. The gypsum beds were not biologically deposited and require an evaporitive environment for growth.

5) The deposits of gypsum and carbonates exhibit several different morphologies which may be related to the depth of water in which the particular deposit formed. The carbonates appear to form in deeper water environments, with a rain of aragonite needles alternating with encasing calcite crystals and the gypsum appears to form in a much shallower environment. Therefore the transition within a lake bed of calcium carbonate, through to selenite crystals (with possible transistion deposits) is related to the depth profile of the lake. Through time, with evaporation, this depth profile may decrease with a subsequent decrease in the aerial distribution of the precipitating deposits.

6) Coring to basement in the Marshall Valley has shown that the carbonate deposition is unique to the Ross Sea II Glaciation, although there is evidence of up to 8 previous glacial incursions into the Marshall Valley. At present there does not appear to be any practical means of dating these sequences.

7) As the result of the analysis of this thesis, it is possible to construct, a glacial history for the Marshall Valley for the last

180,000 years. Within this time period, there has been, 3 glacial incursions into the Marshall Valley, which has deposited the 4 lake beds seen in the Sections, and the lateral, terminal and ice-cored moraines found on the surface of the valley.

a) Prior to 180,000 years ,there was erosion of the glacial drift. As a result of this erosion, the 180,000 years glacial deposits form an unconformity with the older drift. At 180,000 years B.P., ice advanced up the valley, stillstanded and formed a proglacial lake which deposited the lake bed A-B. The ice then advanced farther up the valley and deposited the basal drift and terminal moraines beyond Section IV. Previous investigation of the folded lake beds A-B found at Section III, had suggested thrust faulting as the result of glacial advance over the site. Further study shows that the lake bed has folded by downslope slumping, as the result of reoccupation by a glacial lake (160,000 years). A hummocky topography created as the result of slumping extends downslope above Section III.

b) In the next glacial advance (dated 160,000) years the ice advanced to Section II, again stillstanded and deposited lake bed C-D in the resulting proglacial lake. It is possible that the ice front retreated, readvanced, then retreated again within Section II, to produce the two layered lake bed C-D, but it is also possible that a readvance/retreat phase of the ice front, redistributed the lake bed into the overlying drift deposit.

c) After the retreat of ice from Section II, the ice margin stillstanded once more within the valley, east of Section I, and the 130,000 years lake bed was deposited. At Section III the 130,000 years lake bed overlies the 160,000 years lake bed with no coarse grained drift. Therefore the ice did not readvance up the valley to this point. After the deposition of the lake bed, the ice front retreated from the valley. These three lake bed depositional episodes are representative of the same glacial period and involve ice precipitated at the same altitude. This is assumed to be the Ross Sea II Glaciation.

d) The youngest lake bed in Marshall Valley is the gypsum lake bed , deposited as the result of ice advance into the valley at about 30,000 years B.P. The ice advance produced the lateral moraines seen in the valley. These features can be interpolated to contain the lake bed outcrops and the patchy subdued topography found on the surface of the valley. The glacial drift which forms the lateral and ice-cored moraines found at the mouth of the valley, can be traced along the Victoria Land coast. This material is termed Ross Sea Drift. The basalt rich patchy topography seen in the Marshall Valley may be the result of ice rafting of debris cones over the lake surface. On the downwasting of the lake ice, these deposits form a capping deposit on the finer lake silts, preserving the stratigraphy from the surrounding erosion. This results in the basalt rich, subdued, roughly circular topography seen in the valley.

8) The U/Th dating of Marshall Valley lake beds shows a correlation to the eustatic sea level curve. The three lake beds grouped as Ross Sea II correlate with Illinoian sea level lowering (Isotope Stage 6). Within this period, the 2 glacial advances (that deposited the 3 lake beds) exactly match the lowest sea levels. The final retreat of ice after the deposition of the last lake bed, coincides with the recorded rapid rise in sea level. The dating of the Ross Sea I lake bed also correlates with a low sea level in the Wisconsin (Isotope Stage 2). Therefore within the Marshall Valley for the last 180,000 years a drop in sea level has initiated the grounding of the Ross Sea Ice Shelf and the development of an ice sheet in the Ross Sea. This has resulted in Ross Sea Glaciations.

APPENDIX II

Appendix I.

U/Th Techniques used in dating Marshall Valley Carbonates.

1). Separation of U and Th Isotopes

(1) Weigh out about 35grams of powdered sample and take a record of the amount used. Dissolve it in a sufficient amount of conc. HNO_3.

(2) When all the sample has dissolved , pipette 10mls of U^{232}/Th^{228} spike and heat the resulting solution close to boiling for one half hour. This will bring about the equilibrium of the sample to spike and drive off any CO_2 remaining. (Note 1).

3) Allow the solution to cool, then filter the solution through a buchner funnel and dilute to 1 litre . Collect residue in a watch glass, dry it and weigh it. Record the weight of the residue.

4) Add to the supernate 8-10 drops of $FeCl_3$ which is to act as a carrier to subsequent precipitation. (Note 2).

5) Adjust the pH of the solution to approximately 5 by the titration of conc. NH_4OH. Do not add excess NH_4OH. (Note 3). The resultant iron(III) hydroxide precipitate quantitatively removes the U and Th isotopes from solution.

6) Centrifuge (using 250ml plastic centrifuge bottles) , collect supernatent . Wash the precipitate with distilled water , centrifuge and collect supernatent.

7) Take the collected supernatent , add $FeCl_3$ and reprecipitate at pH 7.5. Repeat centrifuge steps and collect and add precipitate 2 to precipitate 1.

8) Dissolve the precipitation by the dropwise addition of conc. HCl. Add twice as much conc. HCl as is needed to dissolve the precipitate to ensure that normality of the solution > 8N.

9) Add this > 8N solution to an Anion Exchange column of AG1-X8, 50-100 mesh, chloride form analytical grade anion exchange resin (Note 4)

10) In the 8N solution, TH isotopes are eluted while U isotopes and Fe(III) are absorbed onto the resin. After initial 8N HCl containing precipitate has eluted through, add a further elution 8N solution until approximately 1 litre has passed through column. Collect all fractions in a large beaker. The Th fraction will now be ready to be boiled to dryness and further purification. (Note 5)

11) The U isotopes and Fe(III) are eluted from the column by the addition of 1 litre volume of a solution of 0.1M HCl. The resulting combined fractions (which should have an orange red colouration indicating the presence of Fe(III)) should now be boiled down to dryness and further purification.

2. Thorium Purification

(1) Evaporate the Th eluate (from the column) to dryness. Remove from heat immediately as it is difficult to dissolve the residue if it is caked hard to the beaker. Allow to cool.

(2) Dissolve the white residue in a minimum amount of $Al(NO_3)_3$ in 0.1M HNO_3 and transfer the resulting solution to 10ml test tubes. The dissolution of the residue in $Al(NO_3)_3$ will require time and perseverance.

(3) Add 3-4 mls of Hexone to the saturated $Al(NO_3)_3$ solution and emulsify with a pasteur pipette. Centrifuge and remove the hexone phase to a separate test tube. Repeat this phase at least 3 more times and combine Hexone fractions.

4) Add 3-4 mls of distilled water to the Hexone fraction and emulsify the two phases with a pasteur pipette. Centrifuge and remove

the aqueous layer (bottom layer) to a clean 100ml beaker. This back extraction removes the Th isotopes to the aqueous phase. Repeat the back extraction at least twice more and combine the aqueous fractions.

(5) Add 2 drops of conc. HNO_3 to the aqueous fraction in order to destroy any organic matter and evaporate to dryness. There should be a small amount of white residue left.

6) Take up the white residue in a minimum amount of 0.1N HNO_3 (pH 1.5). Rinse the beaker twice again and combine the washings in a 10ml test tube. The pH of the solution should be 1.5. (Note 7)

7) Add 1ml of TTA solution (0.4M thenoyltrifluoroacetone in benzene) to the aqueous solution and emulsify with a pasteur pipette. Centrifuge and remove the TTA fraction to another test tube. Repeat twice again and combine the TTA fractions. The TTA phase will have an orange red colour. This TTA solution contains the purified Th isotope which is now ready to be plated.

3. Uranium Purification

1) Evaporate the uranium fraction to dryness.

2) Take up the residue using a minimum amount of 8N HCl and place in a 10ml test tube using a pasteur pipette. Rinse the beaker several times with a few drops of 8N HCl to ensure that all Fe(III) and uranium is taken up (until the HCl rinse is clear indicating absence of iron). Total volumes should be kept low. (Note 8)

3) Remove the iron from solution by adding an equal volume of di-isopropyl ether to the HCl solution. Emulsify the two phases using a pasteur pipette, centrifuge, draw off the ether layer and discard. Repeat until there is no yellow colouration in the aqueous phase.

4) Remove the aqueous phase to a clean beaker and evaporate to dryness.(Note 9).

5) Rinse the beaker three times with a minimum amount of water and 2-3 drops of 6N HCl. Combine the washings in a test tube. Adjust the pH to 3-4 with the dropwise addition of 3N NaOH. (Note 10)

6) Add 1ml of TTA solution to the aqueous solution, emulsify and centrifuge . Remove the TTA phase to a separate test tube using a clean pasteur pipette. Repeat twice and combine TTA fractions. The TTA solution containing the purified uranium isotopes is now ready to be plated. (The TTA phase is usually orange red in colour.)

7) If after plating, the plate appears "thick" (contains a lot of iron), the uranium must be replated. (Note 11)

4)Sample Mounting.

The TTA-Benzene solution containing either the purified uranium or thorium isotopes is mounted on 7/8" aluminium discs. Tweezers should be used to handle these discs throughout the process.

1) Wash the disc in acetone and allow to dry in air.

2) Flame the disc over a bunsen flame to a cherry red colour. Allow to cool.

3) Place disc on hotplate and warm gently.

4) Use a pasteur pipette and add the TTA solution dropwise. Add sufficiently often so that there is some solution on the disc at all times . Do not heat the disc strongly as this will result in loss of sample due to splattering.

5) After total evaporation, flame the disc again to a cherry red colour with a bunsen flame.

6) Allow the disc to cool down and label it using a marker pen on the underside of the disc. The sample should now be ready to be counted.

(5) Notes

(1) If too much spike is added compared to the naturally occurring isotopes, there may be considerable tail effect. For carbonates, 10mls of spike to approx. 20mls of sample, appeared to be the best spike to sample ratio, for the Antarctic samples.

2) Most of the Antarctic samples have sufficient amounts of iron in them and do not require the addition of further Fe^{3+}.

3) If excess NH_4OH is added, problems arise from the formation of NH_4Cl crystals when eluting the sample through the anion exchange column. This will block the column. If this has taken place, a hot air blower may be used to help redissolve the NH_4Cl crystals by heating the sides of the column.

4) There are two columns sizes in use. For most 10g samples, the 75 ml column is adequate. However for the Antarctic samples, especially those containing larger amounts of iron, the use of the 200ml columns may be required.

5) The Th fraction should be colourless. Slight yellow colouration would indicate the presence of Fe(III) and hence possibly also U isotopes. This will occur if the column is overloaded. The larger 200ml column would therefore be required to effectively separate the U and Th isotopes. All the > 8N solution containing the U and Th isotopes must be recovered by passing through the column usual amounts of 8N HCl and then 0.1 HCl. Boil the combined fractions to dryness, and take up with 8N HCl until the HCl rinse is clear, indicating no iron is left in the beaker. Pass through the 200ml column following the procedure as for the 75ml column.

6) If there is a large amount of residue, a lot more $Al(NO)_3$ solution will be required to dissolve the residue and hence require the use of more centrifuge tubes.

7) Remove the beaker from the hotplate as soon as the solution has evaporated to dryness otherwise it will be difficult to remove the residue from the beaker. Dissolution of the residue is easier if the beaker is rinsed with 0.1M HNO_3 while the beaker is still hot after the evaporation to dryness.

8) Some samples have a considerable amount of iron. If more than 6 mls of HCL is required to take up the residue, use more then one test tube - this makes the iron extraction simpler.

9) Presence of iron(which will result in a thick plate) will be indicated if the residue has a yellow brown appearance. Before proceeding further, take up the residue with a minimum amount of 8N HCl and treat again with di-isopropyl ether to remove the iron still present.

10) The pH should be adjusted to at least 3 in order to be able to extract the U isotopes from aqueous solution.However at pHs >5 it is noted that some precipitation may occur (possibly due to Ca and Al salts). Low yields of U isotopes result if this precipitate is not redissolved by the lowering of the pH and extracting of the U isotopes again with TTA. If precipitate forms at pH 3-4 then the sample should be evaporated to dryness, taken up in 8N HCl and passed through the anion exchange column once again to remove the iron causing precipitation. The U fraction will once again be obtained as the 0.1M HCl fraction.

11) Add sufficient aqua regia (1 part HNO_3 to 3 part HCl) to a small beaker to cover the "thick " plate. Allow sufficient time for at least all of the plated material to dissolve off as well as some of the plate. Evaporate the resulting solution to dryness and take up the residue with 8N HCl. Pass the solution through a 200ml anion exchange column (there will be a lot of residue)since the residue will now contain substantial amounts of nickel and chromium salts as well a

iron (III) and uranium isotopes. After eluting with 8N HCl (at least 3 times) and discarding the resulting fractions, the uranium isotopes and iron (III) is eluted through with several column volumes of 0.1N HCl (at least 3 column volumes). Evaporate the 0.1N HCl fraction to dryness, take up with a minimum amount of 8N HCl and remove iron(III) by extracting with di-isopropyl ether. (A distinct green colour indicating the presence of some nickel salts will remain in the aqueous solution.) Without further purification, adjust the pH of the aqueous solution to 3-4, extract the U isotopes with TTA and plate the sample (following steps 5 and 6 of U purification).

Note: There is usually a substantial amount of Ni present in the aqueous solution of the TTA stage (distinct green solution). However very "thin" samples with good yields have been obtained without further purification.

12) The resin has to be regenerated by passing 2 column volumes of 8N HCl when the column is saturated with SO_4^{2-}. The condition of the column can be checked by taking test portions off the column and adding to 0.2M $BaCl_2$ solution. A white precipitate of Ba SO_4 will result if the column is saturated and is allowing the SO_4^{2-} to pass through.

APPENDIX II

Appendix II (Dagel *et al*, in press)

Valley-Floor Stratigraphy.

Deposits of selenite, calcium carbonate, and unconsolidated sediments crop out discontinuously along steep banks of the valley-floor stream. Four stratigraphic sections excavated along the north bank of the stream were mapped in detail (Figs. 4 through 7). They are designated Section I through IV (Fig.3). Additional excavations were made opposite Section IV, as well as between Section III and IV. Table 2 and Figures 8,9 and 10 summarize physical characteristics of the stratigraphic units exposed in these excavations.

Descriptions of units in each section are given below. The unit numbers correspond with those in Figures 4 through 7. The grain size classification is from Folk (1954), the sorting classification is from Folk and Ward (1957), and the roundness classification is from Pettijohn(1957). Colors are of the <2mm fraction and follow the Munsell system. Several units contain sandsized basalt clasts and light-coloured mineral fragments that give the units a salt and pepper appearance. From a distance these units appear very dark grey to dark grey (N3/0-N4/0) and are termed "variegated".

Unit Discriptions.

Section I

Unit 1I. Unit 1I extends to the surface of Section I. It is a light brownish gray (2.5Y6/2), muddy sandy gravel containing a few percent cobbles. The gravel contains approximately 80% basalt, 10% granite, and 10% metasedimentary rock fragments. In section, this unit has a sharp contact with Unit 2I: however, excavation behind Section I revealed that Unit 1I directly overlies Unit 3I in some places. In other places Unit 1I is absent and Unit 2I is the surface unit. A desert pavement of coarse sand to cobblesize clasts is developed on Unit 1I.The clasts are wind polished; a few have a light desert varnish.

Unit 2I. Unit 2I underlies Unit 1I with a sharp contact and consists of three layers. The upper layer is a 0.5-cm-thick bed of fine sand mixed with 1 mm selenite crystals. It overlies with a sharp contact a selenite bed , one crystal thick (1-2 cm). This bed overlies a 2 to 8 cm thick layer of laminated gypsum with a sharp, disconcordant contact. The laminae are domed into parallel hemisphereoids 1 to 3 cm in diameter.

Unit 3I. Unit 3I underlies Unit 2I with a sharp contact. It is a gravelly mud. The <2 mm fraction is pale olive (5Y6/4), very poorly sorted (s.d. = 3.34 ϕ), and has a mean grain size of 6.21 ϕ. The sand fraction consists of angular to subrounded grains of feldspar, basalt, quartz, metasedimentary rock fragments, and olivine. The gravel fraction consists almost entirely of subrounded basalt pebbles. Most of the sand and gravel is scattered throughout the mud matrix; some is concentrated in 0.25 cm beds within the upper few cm of the unit. The unit is unconsolidated, uncemented, and exhibits a blocky fracture.

Unit 4I. Unit 4I underlies Unit 3I with a sharp contact. Unit 4I is a sandy gravel. The <2mm fraction is variegated, poorly sorted (s.d. = 1.45 ϕ), and has a average mean grain size of 1.20 ϕ. The sand is subrounded to well rounded and contains basalt, olivine, and quartz grains. The gravel is subangular to rounded and contains approximately 91% basalt (2% of which is kenyite), 3% granite, and 6% metasedimentary rock fragments.

Unit 4I contains several sub-horizontal, undulating layers approximately 10-40 cm thick that contain different amounts of gravel. The gravel clasts exhibit bedding-parallel fabric. The unit is slightly finer grained and more massive towards the bottom. A prominant layer near the top of the section consists almost entirely of 2 to 4 cm gravel. Fossil algae occurs in this gravel layer.

Unit 5I. Unit 5I underlies Unit 4I with a sharp contact. It is a compact, muddy sandy gravel and contains approximately 4% cobbles and 1% boulders. The <2 mm fraction is pale brown (10YR6/3), very poorly sorted (s.d.= 2.46 ϕ), and has an average mean grain size of 2.21 ϕ. The sand fraction is composed of subangular to rounded grains of basalt, quartz, and feldspar. The gravel is subangular to subrounded and contains 61% basalt, 23% granite, and 12% metasedimentary rock fragments. A few clasts are striated. Unit 5I is largely unstratified although it contains a few discontinuous, horizontal bands of gravel that are one clast thick.

Unit 6I. Unit 6I underlies Unit 5I with a sharp, discordant contact. Unit 6I varies in color, grain size and stratification from west to east. It is a sandy gravel containing cobbles and boulders. At the western end of the section, Unit 6I contains 10% cobbles and 5% boulders; the <2 mm fraction is grayish brown (2.5Y5/2), very poorly sorted (s.d.=2.39 ϕ) and has a mean grain size of 1.94 ϕ. At the eastern end of the section, Unit 6I contains 3% cobbles and no boulders; the <2 mm fraction is variegated, poorly sorted (s.d. = 1.83 ϕ) and has a mean grain size of 1.70 ϕ. The sand fraction of Unit 6I is angular to subrounded and is composed of equal proportions of quartz, feldspar, basalt, and metasedimentary rock fragments. The gravel is subangular to rounded and contains approximately 58% basalt, 21% granite, and 21% metasedimentary rock fragments. Several striated clasts occur in Unit 6I. The western two-thirds of the units exhibits an undulating , subhorizontal banding due to several thin , discontinuous, subhorizontal layers and lenses of muddy sand and sandy mud and poorly defined horizontal beds containing varying amounts of gravels and cobbles. Clast fabric is parallel to this banding. The eastern end of Unit 6I is stratified. At the top of the unit the strata have an apparent dip of approximately 30° to the west and consists of 50-cm-beds of sandy gravels alternating with 2-to-4-cm-thick beds of sandy mud. These steeply dipping beds are truncated at the top by Unit 5I and become tangential to the base of Unit 6I. The base of Unit 6I consists of a few cm of laminated sandy silt containing pebbles. At the east end of the section the silt laminae are continuations of the steeply dipping beds of sandy silt and silty sand. The thickness of the laminated silt layer decreases from about 15 cm at the eastern end of Section I to 4 cm at the western end of the section.

Unit 7I. Unit 7I underlies Unit 6I with a sharp contact. Unit 7I consists of 1-to-2 cm of laminated , tabular calcium carbonate that is underlain in the east half of the section by 5-15cm of chalky calcium carbonate. The contact between the chalky and tabular calcium carbonate is gradational in some places and exhibits a shallow angular unconformity in others. The tabular calcium carbonate is thinner at the western end than at the eastern end of the section. Unit 7I contains pebble size to boulder size dropstones.

Unit 8I. Unit 8I underlies Unit 7I with a 1 to 5 cm transition zone of muddy gravelly sand. Unit 8I consists of muddy sandy gravel containing approximately 15% cobbles and 2-5% boulders. The <2 mm fraction is grayish browm (2.5Y5/2), very poorly sorted (s.d. = 2.19 ϕ), and has a mean grain size of 1.94 ϕ. The sand fraction consists of subrounded to rounded and consists of approximately 76% basalt, 13% granite and 12% metasedimentary rock fragments. A few percent of the clasts are striated. Unit 8I is crudely stratified . The western three-quarters of the unit exhibits faint, undulating, subhorizontal stratification that is concordant with that of overlying Unit 7I.

The layers are 1 to 20 cm thick and are recognizable because they contain slightly different amounts of mud,sand , gravel or cobbles.The layers are commonly separated by laminae of mud and fine sand. The gravel and cobbles exhibit a strong preferred orientation parallel to the stratification. The eastern one-quarter of the unit exhibits stratification and fabric similar to that of the western three-quarters of the unit, except that the layers are thinner and more distinct.

Unit 9I. Unit 9I is a thin unit of silty calcium carbonate and calcium carbonate-cemented pebbly sand that separated Units 8I and 10I. It's upper contact is gradational; its lower contact forms an angular unconformity with Unit 10I.

Unit 10I. Unit 10I is a sandy gravel containing about 5% cobbles. It is a grayish brown (2.5Y5/2) matrix. Unit 10I is similar to the east end of Unit 8I. Unit 10I exhibits 5 to 10 cm parallel plane beds that have apparent dips of 25° to 30° to the west and are truncated by Unit 9I.

Unit Discriptions; Section II

Unit 1II. Unit 1II forms the surface of Section II. It is a sandy gravel containing 3% cobbles. The <2mm fraction is variegated, poorly sorted (s.d. = 1.87 ϕ), and has a mean grain size of 1.78 ϕ. The sand fraction consists of medium to fine, subrounded to well-rounded grains of basalt, olivine, and quartz. The gravel is subangular to rounded and contains approximately 99% basalt, 1% metasedimentary rock fragments, and no granite. The gravel forms a desert pavement at the top of the section where it exhibits light desert varnish, slight pitting and ventification. A thin layer of calcium carbonate coats the buried surfaces of the clasts. Unit 1II is unstratified.

Unit 2II. Unit 2II is a layer of selenite crystals that is one crystal thick. Unit 2II exhibits a pronounced dip component towards the present stream cut; in places it lies parallel to the present stream bank. Unit 2II is underlain by Unit 4II in the west part of the western part of the section, and by Unit 3II in the eastern part of the section.

Unit 3II. Unit 3II is a gravelly sand containing approximately 5% cobbles and one large boulder. It is interbedded with thin layers of selenite crystals and gypsiferous sandy silt. The gravelly sand beds are up to 1 m thick. The <2 mm fraction is variegated, poorly sorted (s.d. = 1.85 ϕ), and has an average mean grain size of 1.73 ϕ. The sand fraction is composed of subrounded to rounded grains of basalt, quartz and olivine. The gravel is subangular to rounded and contains approximately 73% basalt,17% granite, and 10% metasedimentary rock fragments. Three-to-five centimetres masses of 0.5 cm of selenite crystals occur within the gravelly sand beds near the top of Unit 3II. At the east end of Section II the beds have an apparent dip of 15° to 20° to the east. The dip becomes shallower to the west, where the beds become broadly undulating. The bedding beneath the large granite boulder in the middle of the section curves downwards in a dropstone-like structure. The gravelly sand beds are themselves laminated. They commonly exhibit climbing ripple lamination having 2 to 4 cm laminae with in-phase crests. Locally, the beds exhibit 2 to 4 cm plane lamination. The laminations are composed of alternating layers of coarse and fine sand; The gravel is evenly dispersed throughout each bed.

The interbedded layer are 1 to 6 cm thick and consists of 2 mm to 1 cm long selenite crystals and laminated (1 to 5 cm) calcareous fine sand and silt that exhibits close-linked lateral hemisphereoid structure. Unit 3II thins to less then 1 m at the eastern end of Section II. It pinches out completely near the middle of the section.

Unit 4II. Unit 4II consists of 1 to 4 cm of laminated gypsiferous fine sand and silt that overlie up to 4 cm of laminated, calcium carbonate that contains enclosed pebbles. These layers are locally separated by fine selenite crystals. The laminated calcium carbonate is locally underlain by a thin bed of micrite honeycombed with selenite crystals molds. Clasts that project up from the top of underlying Unit 5II are draped by the basal laminated carbonate bed of Unit 4II. In the eastern 10m of Section II, where Unit 4II directly overlies Unit 6II, the basal layer of laminated carbonate is oxidized reddish-orange.

Unit 5II. Unit 5II consists of an upper subunit (5aII) and a lower subunit (5cII) separated by subunit 5bII. Subunit 5aII underlies Unit 4II with a sharp contact; it is a compact, unstratified, muddy sandy gravel containing 5 to 15% cobbles and 1% boulders. The <2mm fraction is pale brown (10YR6/3), very poorly sorted (s.d.= 3.00 ϕ), and has an average mean grain size of 2.60 ϕ. The sand fraction is very angular to rounded and is composed of basalt, feldspar, calcium carbonate-cemented silt aggregates, quartz, and metasedimentary rock fragments.The gravel is subangular to rounded and contains 11% basalt,730% granite and 16% metasedimentary rock fragments.

Subunit 5cII is a compact, unstratified muddy, sandy gravel to gravelly, muddy sand containing 10 to 20% cobbles and 1% boulders. The <2 mm fraction is pale brown (10YR6/3), very poorly sorted (s.d = 2.83 ϕ) and has an average mean grain size of 2.36ϕ.

The sand fraction is very angular to rounded and is composed of basalt, feldspar, calcium carbonate-cemented silt aggregates, quartz, and metasedimentary rock fragments. The gravel is subangular to rounded and contains 28% basalt, 50% granite, and 22% metasedimentary rock fragments.

Subunits 5aII and 5cII are separated in places by a thin layer of basalt-rich sand and in other places by a thin layer of tabular calcium carbonate (subunit 5bII).

The tabular clasts are 0.5 to 2.0 cm thick and up to 15 cm in diameter. The clasts have sharp, broken edges; they are not weathered or rounded. Unit 5II contains many conglomerate clasts. These clasts consist of a gravelly sand that is moderately cemented with light gray (N7/0) carbonate. Within Unit 5II near the west end of Section II is a 1 m by 6 m lens of heavily cemented coarse breccia. The lens directly overlies Unit 6II and is capsulated with a nearly continuous layer of tabular calcium carbonate. Unit 5II thins towards the east end of the section and pinches out entirely about 15 m from the west end of the section.

Unit 6II. Unit 6II is a discontinuous unit consisting of a 0.5 cm layer of micrite honeycombed with models of small selenite crystals, underlain by a 0.5 to 1.5 cm layer of laminated tabular calcium carbonate. In places the calcium carbonate of Unit 6II has cemmented the top few cm of Unit 7II. Unit 6II sharply defines the contact between Units 5II and 7II throughout much of Section II.

Unit 7II. Unit 7II consists of two layers. The upper layer is 5 to 20 cm thick, variegated and is composed of interlaminated coarse and fine basalt-rich sand. The lower layer is 25 to 80 cm of unstratified sand containing 1 to 5% small cobbles. The <2mm fraction is variegated, very poorly sorted (s.d.= 2.12 ϕ) and has a mean grain size of 1.65 ϕ. The sand fraction is subrounded to well-rounded and consists of basalt, quartz, and olivine. The gravel is subangular to rounded and contains approximately 86% basalt (3% of which is kenyite), 9% granite, and 5% metasedimentary rock fragments.

Unit 8II. Unit 8II consists of a thin, continuous layer of light gray tabular calcium carbonate on the top surface of which are preserved molds of 1 to 5 mm long selenite crystals. It is underlain by a 0.5 cm layer of grayish brown micrite honeycombed with molds of selenite crystals. Unit 8II has a sharp upper contact; its lower surface drapes over clasts projecting from underlying Unit 9II. Unit 8II can be traced in float and in isolated outcrops from Section II to Section III where it is equivalent to Unit 3III.(Fig 12)

Unit 9II. Unit 9II is a poorly exposed deposit of gravelly sand containing a few cobbles. It is moderately cemented with light gray calcium carbonate. The clasts are composed of basalt, metasediments, and granite; many are ventifacts. The gravel is subangular to subrounded and exhibits a horizontal fabric. The unit is unstratified, although it locally displays 1cm horizontal parting. Unit 9II forms the base of Section II. THe base of Unit 9II is not visible.

Unit Discriptions; Section III.

Unit 1III. Unit 1III consists of several mineral layers. It crops out just beneath the desert pavement in Section III as well as through the desert pavement in places behind Section III. From top to bottom the layers consist of silty, finely laminated tabular calcium carbonate; calcium carbonate-cemented sand; white, hard, 0.5 to 1.0 cm anhedral blades of calcium carbonate that in places grade laterally to chalky white calcium carbonate; and brownish gray micrite honeycombed with molds of selenite crystals. The most laterally continuous layer is the laminated tabular calcium carbonate. The laminae consist of 0.25 to 0.50 cm thick layers of gray micrite alternating with <1 mm thick layers of fine, black detritus. The laminated calcium carbonate layer drapes over clasts projecting up from the underlying unit. Unit 1III is discontinuous but can be traced along the top of the stream bank to the west of Section III.(Fig 12).

Unit 2III. Unit 2III is a muddy sandy gravel containing <1% cobbles. The <2 mm fraction is variegated, very poorly sorted (s.d.=2.77 ϕ) and has an average mean grain size of 2.37 ϕ. The sand fraction is composed of angular to subrounded grains of basalt with minor olivine and quartz. The gravel is subangular to rounded and contains approximately 98% basalt (5% of which is kenyite), 1% granite and 1% metasedimentary rock fragments. Fragments of the barnacle Hexelasma antarcticum were found in Unit 2III. Unit 2III is unstratified except for a few 10 to 20 cm thick, wavy, discontinuous, subhorizontal beds of muddy sand.

Unit 3III. Unit 3III consists of three layers. The general sequence is , from top to botttom, a matrix supported slurry of variable thickness consisting of fine (<1mm) gypsum crystals , sand, and occasional pebbles, in a porous calcium carbonate matrix; one to four cm selenite crystals having their long axes vertical; and one half to one centimetre thick tabular calcium carbonate. The tabular calcium carbonate layer is locally underlain by another thin layer of selenite crystals. Where it is not underlain by such a layer, the tabular calcium calcium carbonate drapes over clasts projecting from the top of the underlying unit. The contacts between these layers, as well as between the entire unit and the adjacent units, are sharp and distinct.

A few centimetres of calcium carbonate-cemented sediment locally underlies Unit 3III. This appears to be simply a thin cemented zone of the underlying unit.

Unit 3III can be traced as a continuous bed for several metres east of Section III. It can be traced in float and in outcrops between Section III and Section II, to Unit 8II. (Fig.12)

Unit 4III. Unit 4III is a compact,muddy sandy gravel containing 10 to 15% cobbles. The <2mm fraction is pale brown (10YR6/3), very poorly sorted (s.d.=3.17ϕ) and has an average mean grain size of 2.82ϕ. The sand fraction consists of very angular to subangular grains of quartz, feldspar,mafic minerals, and metasedimentary clast. Unit 4III is unstratified. Clasts locally exhibit a subhorizontal fabric.

Unit 4III overlies and contains deformed fragments of Unit 6III. At the west end of the section a long piece of underlying Unit 6III is rotated out of place and incorporated into Unit 4III. An 8 m long zone at the base of Unit 4III contains broken pieces of Unit 6III that are overturned and folded. Near the middle of Section III Unit 6III has been deformed into a tight 2 to 3 m overturned isoclinal fold with its axis perpendicular to the valley trend. Selenite crystals within this folded layer are deformed. Unit 4III sediment surrounds the folded layer. At the east end of the section, another exposure shows deformed pieces of Unit 6 III. Here the mineral layer has been folded and imbricately faulted. Unit 4III sediment lies adjacent to and above these deformed beds. Unit 4III pinches out abruptly at the west end of Section III.

Unit 5III. Unit 5III is a muddy sandy gravel containing approximately 1% cobbles. The <2mm fraction is variegated and is very poorly sorted. The sand fraction is composed of subangular to subrounded grains of basalt with minor olivine and quartz. The sand is subangular to rounded and is composed almost entirely of basalt.

Fragments of the barnacle Hexelasma antarcticum Borradaile occur in Unit 5III. The unit is massive except for a few thin, wavy stringers of muddy sand. Unit 5III is discontinuous and lenses out both to the east and the west.

Unit 6III. Unit 6III consists of three layers and is identical to Unit 3III. In several places within the section, Unit 6III is folded and ripped up into Unit 4III (see above). Unit 6III crops out almost continuously along the north bank of the stream cut between Sections III and IV and is the equivalent of Unit 2IV in Section IV.(Fig.12)

Unit 7III. Unit 7III is exposed in a few places within Section III and between Sections III and IV where it underlies Unit 6III with a sharp contact. It consists of gravelly sand that contains approximately 10% cobbles and that is moderately cemented with light gray (N7/0) calcium carbonate. The sand is subangular to subrounded and is composed of metasediment, granite,basalt fragments, quartz, feldspar, and minor olivine. The gravel and cobbles consist of angular fragments of local igneous and metasedimentary bedrock and a few basalt clasts. The unit exhibits no stratification or fabric.

Unit 8III. Unit 8III is exposed within Section III and between Sections III and IV, where it directly underlies Units 4III,6III and 7III with sharp contacts. Unit 8III is a compact, muddy sandy gravel containing approximately 15% cobbles. The <2mm fraction is pale brown (10YR6/3), very poorly sorted (s.d.=2.47 ϕ) and has a mean grain size of 1.94 ϕ. The sand fraction consists of angular to subrounded grains of quartz, feldspar, mafic minerals, metasedimentary rock fragments, basalt fragments and mica. The gravel is subangular to rounded and consists of approximately 76% granite and 24% metasedimentary rock fragments.

Unit Discriptions; Section IV.

Unit 1IV. Section IV exposes the interior of a large moraine Fig 3. Unit 1IV makes up the upper part of the moraine. Unit 1IV is a sandy gravel that contains no boulders or cobbles. The <2 mm fraction is variegated, poorly sorted (s.d.=1.80 ϕ), and has a mean grainsize of 1.39 ϕ. The sand is very angular to subrounded and is composed primarily (>90%) of basalt fragments. Minor constituents are feldspar, olivine, and quartz grains. The gravel is subangular to rounded and contains approximately 89% basalt, 5% kenyite and 5% metasedimentary rock fragments. Unit 1IV contains faint, 10 to 40 cm thick, subhorizontal beds of coarse and fine sand.

Unit 2IV. Unit 2IV is a thin mineral layer. It forms a sharp contact between basalt-rich Unit 1IV and basalt-poor Units 3IV and 4IV. At the west end of Section IV, Unit 2IV consists of a single layer of 2 to 6 cm selenite crystals. To the east the crystals are smaller (1 to 2 cm) and are partially replaced by selenite crystal molds in grayish brown micrite. Farther to the east the gypsum is almost entirely replaced by micrite that is honeycombed with selenite molds. At the east end of the section, Unit 2IV consists of tabular calcium carbonate and scattered fragments of micrite honeycombed with selenite crystal molds. Unit 2IV crops out almost continuously along the north stream bank between Sections IV and III and is the equivalent of Unit 6III in Section III.(Fig 12)

Unit 3IV. Unit 3IV is a compact, muddy sandy gravel that contains approximately 15% cobbles and 10% boulders. The <2 mm fraction is pale brown (10YR6/3), very poorly sorted (s.d.=2.51 ϕ) and has a mean grain size of 2.28 ϕ. The sand fraction is very angular to rounded and consists of metasedimentary rock fragments, feldspar, quartz and a few (<10%) basalt fragments. The gravel is subangular to rounded and contains approximately 2% basalt, 79% granite and 19% metasedimentary rock fragments. Unit 3IV is unstratified. In places it exhibits horizontal clast fabric.

In the easternmost exposure in Section IV, Unit 3IV is separated from Unit 2IV by two subunits. The upper subunit is a 60 to 70cm thick layer of pale brown (10YR6/3) sandy gravel containing approximately 15% cobbles. The sand consists of subangular to subrounded grains of basalt, quartz and olivine. The gravels and cobbles are angular and contain metasedimentary , granite and a few basalt clasts. The lower subunit is a 40 cm thick layer of pale brown (10YR6/3) parallel laminated (0.5 to 1.0 cm) medium sand. The grains are angular to subrounded and are composed of quartz and dark lithic fragments.

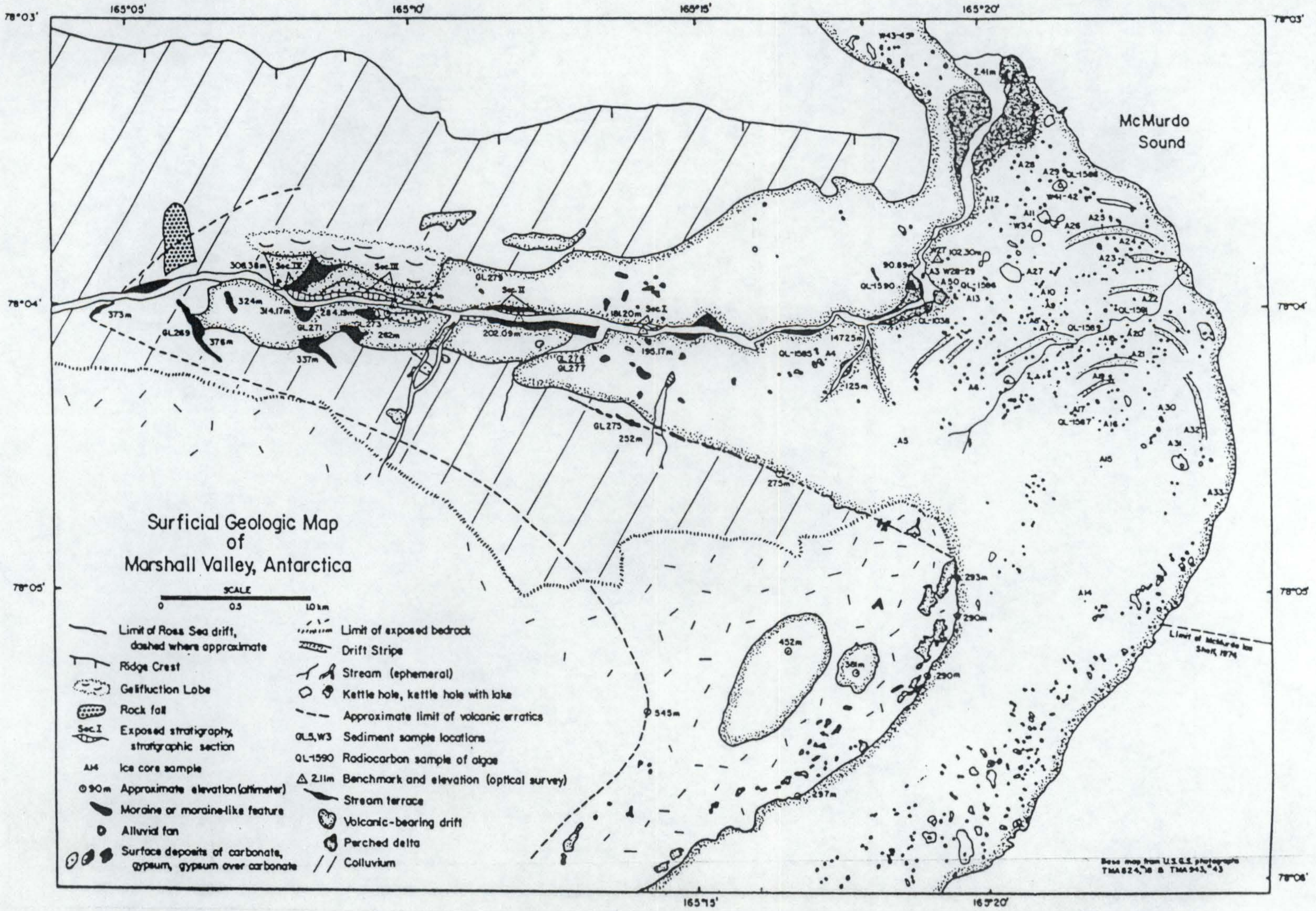
Unit 4IV. Unit 4IV is a sandy gravel containing 15% boulders and 10 to 20% cobbles. The <2 mm fraction is variegated to pale brown (10YR6/3), poorly sorted (s.d.=1.56 ϕ) and has an average mean grain size of 1.27 ϕ. The sand consists of medium to coarse, subangular to rounded grains composed of about equal proportions of quartz and basalt with minor amounts of feldspar and metamorphic rock fragments. The gravel is subrounded to rounded and contains approximately 2% basalt, 75% granite, and 23% metasedimentary rock fragments. The unit is unstratified. Unit 4IV is visible in the western three exposures in Section IV. The base of Unit 4IV is not visible in the section.

Table 2

Unit	Sample	Gravel-Sand-Mud (wt. %)	Sand-Silt-Clay (wt. %)	Verbal Classification	Mean (ϕ)	Standard Deviation (ϕ)	Skewness	Kurtois	Sample	Clast Lithology: Basalt (Kenyite)-Granite-Metasediment (ϕ)	Roundness
3I	GL100	10-21-69	23-52-25	gM	6.21	3.24	-0.21	-9.55	GL100	86(0)-05-09	0.39
4I	GL202	36-63-01		sG	1.39	1.46	2.10	9.83			
	GL203	23-75-01		sG	1.19	1.43	2.74	13.30	GL246	90(2)-03-07	0.35
	GL204	40-53-06	89-07-03	msG	1.95	2.28	2.17	6.05			
5I	GL205	39-53-08	87-08-04	msG	2.27	2.47	2.09	5.12	GL245	61(0)-23-16	0.35
	GL206	39-52-10	85-10-05	msG	2.40	2.63	1.95	4.10			
	GL207	36-58-06	91-06-04	sG	1.94	2.39	2.37	6.83			
6I	GL208	37-60-04	94-02-03	sG	1.59	2.18	2.85	10.45	GL244	61(0)-19-19	0.35
	GL209	40-56-03	95-04-02	sG	1.70	1.83	2.56	11.26	GL247	56(0)-24-24	
8I	GL212	43-51-07	88-09-03	msG	1.94	2.19	2.12	6.18	GL243	76(0)-13-12	0.39
1II	GL213	31-66-03		sG	1.78	1.87	2.26	6.87	GL249	99(0)-00-01	0.35
	GL214	28-70-02		gS	1.63	1.55	2.47	10.36			
3II	GL219	23-72-05	93-03-04	gS	1.98	2.22	2.69	8.85	GL250	73(0)-17-10	0.36
	GL223	27-71-02		gS	1.58	1.79	2.16	7.17			
	GL216	38-52-11	83-10-07	msG	2.39	2.90	1.69	2.43			
5aII	GL218	58-36-06	85-09-06	msG	2.25	2.51	1.74	3.12			
	GL222	35-50-15	77-13-10	msG	2.86	3.27	1.47	1.34	GL253	11(0)-73-16	0.34
	GL226	38-48-14	77-13-10	msG	2.89	3.31	1.42	1.12			
5cII	GL215	29-61-10	85-09-06	gmS	1.98	2.73	1.10	3.80			
	GL221	59-30-11	73-21-06	msG	2.73	2.91	1.26	1.20	GL252	28(0)-50-22	0.32
	GL217	35-58-07	90-06-04	msG	2.03	2.54	2.34	5.97			
7II	GL220	33-65-02		sG	1.16	1.72	2.45	8.97	GL251	86(1)-09-05	0.39
	GL225	18-77-06	93-04-03	gS	1.75	2.11	2.93	10.26			
2III	GL232	40-50-10	83-12-05	msG	2.39	2.68	1.63	2.63			
	GL234	41-48-11	81-13-06	msG	2.34	2.86	1.64	2.68	GL257	98(5)-01-01	0.31

Table 2 (Cont'd)

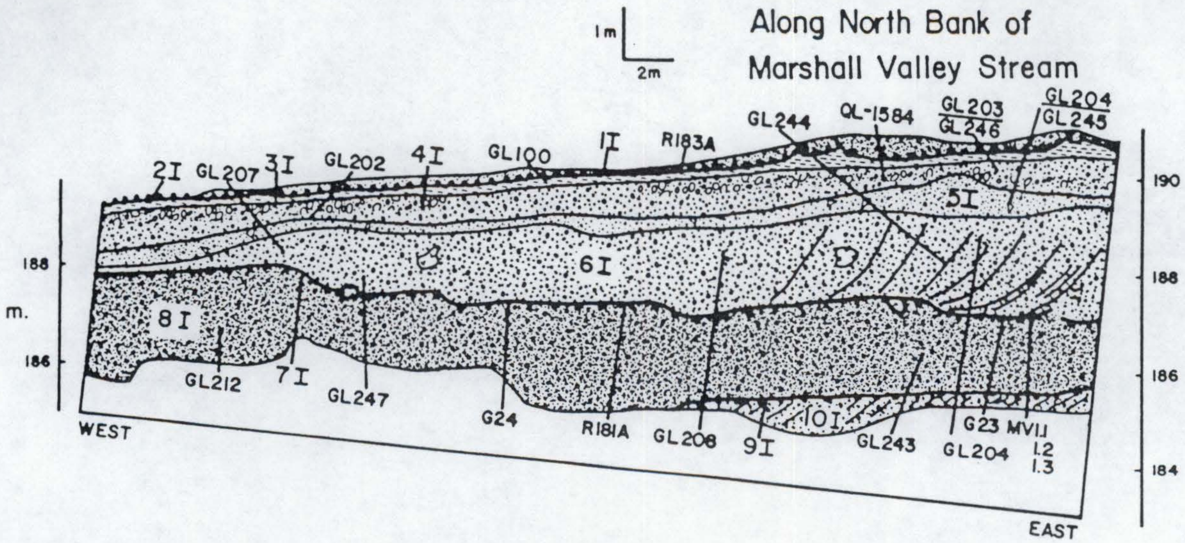
Unit	Sample	Gravel-Sand-Mud (wt. %)	Sand-Silt-Clay (wt. %)	Verbal Classification	Mean (ϕ)	Standard Deviation (ϕ)	Skewness	Kurtosis	Sample	Clast Lithology: Basalt (Kenyite)-Granite-Metasediment (%)	Roundness
2III[a]	GL237	31-67-03		sG	1.52	1.85	2.45	7.95			
4III	GL230	38-50-12	81-09-10	msG	2.75	3.22	1.52	1.58			
	GL231	44-44-12	78-12-10	msG	2.79	3.11	1.50	1.62	GL256	00(0)-79-21	0.34
	GL235	39-48-13	79-11-10	msG	2.92	3.17	1.49	1.47			
8III	W19-21	56-39-05	89-07-05	msG	1.94	2.47	2.21	5.48	W19-21	00(0)-76-24	0.33
8III[a]	W13-15	38-54-09	86-10-04	msG	2.28	2.47	1.95	4.46	W13-15	00(0)-73-27	0.35
1IV	GL242	42-56-02		sG	1.39	1.80	2.44	8.29	G1263	95-00-05	0.34
3IV	GL239	44-47-09	85-12-04	msG	2.17	2.34	1.71	3.36	GL260	00(0)-75-25	0.36
	GL240	38-54-08	88-07-05	msG	2.10	2.49	2.27	5.90	GL261	02(0)-79-19	0.35
	GL241	48-42-10	81-14-05	msG	2.60	2.66	1.58	2.80	GL262	02(0)-84-14	0.34
	GL282	36-55-09	85-10-04	msG	2.24	2.55	1.86	3.77			
4IV	GL238	46-52-02		sG	1.32	1.73	2.86	10.50	GL259	02-75-23	0.36
	GL281	53-47-01		sG	1.21	1.38	2.88	14.39			
RSD	GL275	47-51-01		sG	1.59	1.61	2.38	9.46			
	W27	03-94-02		(g)S	1.47	1.47	3.19	14.66	W27*	100(6)-00-00	0.40
	W28-29	15-77-08	91-06-03	gS	1.85	2.23	2.54	8.04	W28-29	95(6)-03-02	0.36
	W34	07-33-60	35-50-15	gM	5.02	3.15	0.51	-0.08	W34	100(20)-00-00	0.40
	W41-42	27-69-04	94-03-02	gS	2.06	1.91	2.52	10.66	W41-42	96(1)-03-01	0.43
	W43-45	33-65-01		sG	1.18	1.53	2.84	12.60	W43-45	95(0)-04-01	0.39
OD	GL242	42-56-02		sG	1.39	1.80	2.44	8.29	GL263	95(3)-00-05	0.34
	GL269	54-44-02		sG	1.56	1.83	2.41	7.91			
	GL271	43-53-03	95-05-00	sG	1.53	1.52	1.17	2.64			
	GL273	38-61-01		sG	1.74	1.44	2.21	10.33			
	GL279	16-82-02		gS	1.57	1.46	2.35	10.73			



Surficial Geologic Map of Marshall Valley, Antarctica

Fig. 4

Stratigraphic Section I
Along North Bank of
Marshall Valley Stream

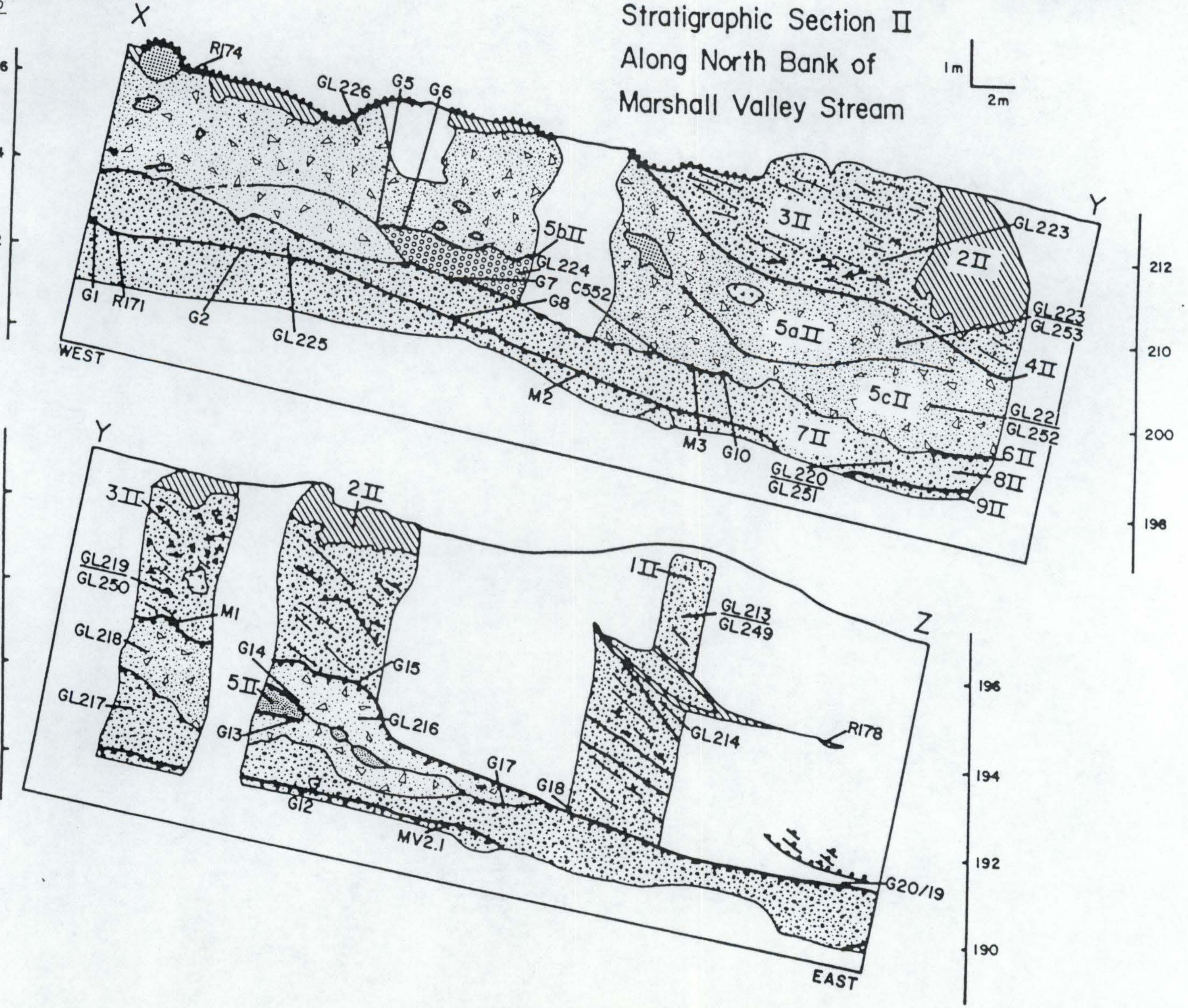


EXPLANATION

- TILL
- SAND
- VOLCANIC-RICH GRAVEL
- VOLCANIC-RICH GRAVELLY MUD
- VOLCANIC-RICH MUDDY SANDY GRAVEL; STRATIFIED WHERE SHOWN
- VOLCANIC-RICH SANDY GRAVEL & GRAVELLY SAND; STRATIFIED WHERE SHOWN
- MODERATELY CEMENTED CONGLOMERATE, OCCURRING AS CLASTS
- HEAVILY CEMENTED COARSE BRECCIA, OCCURRING AS CLASTS
- VOLCANIC-POOR SANDY GRAVEL & GRAVELLY SAND; STRATIFIED WHERE SHOWN
- COBBLE OR BOULDER
- SLUMPED SEDIMENTS
- CARBONATE
- SELENITE GYPSUM
- SELENITE GYPSUM DRAPING STRATIGRAPHIC SECTION
- BROKEN FRAGMENTS OF SELENITE GYPSUM OVER CARBONATE

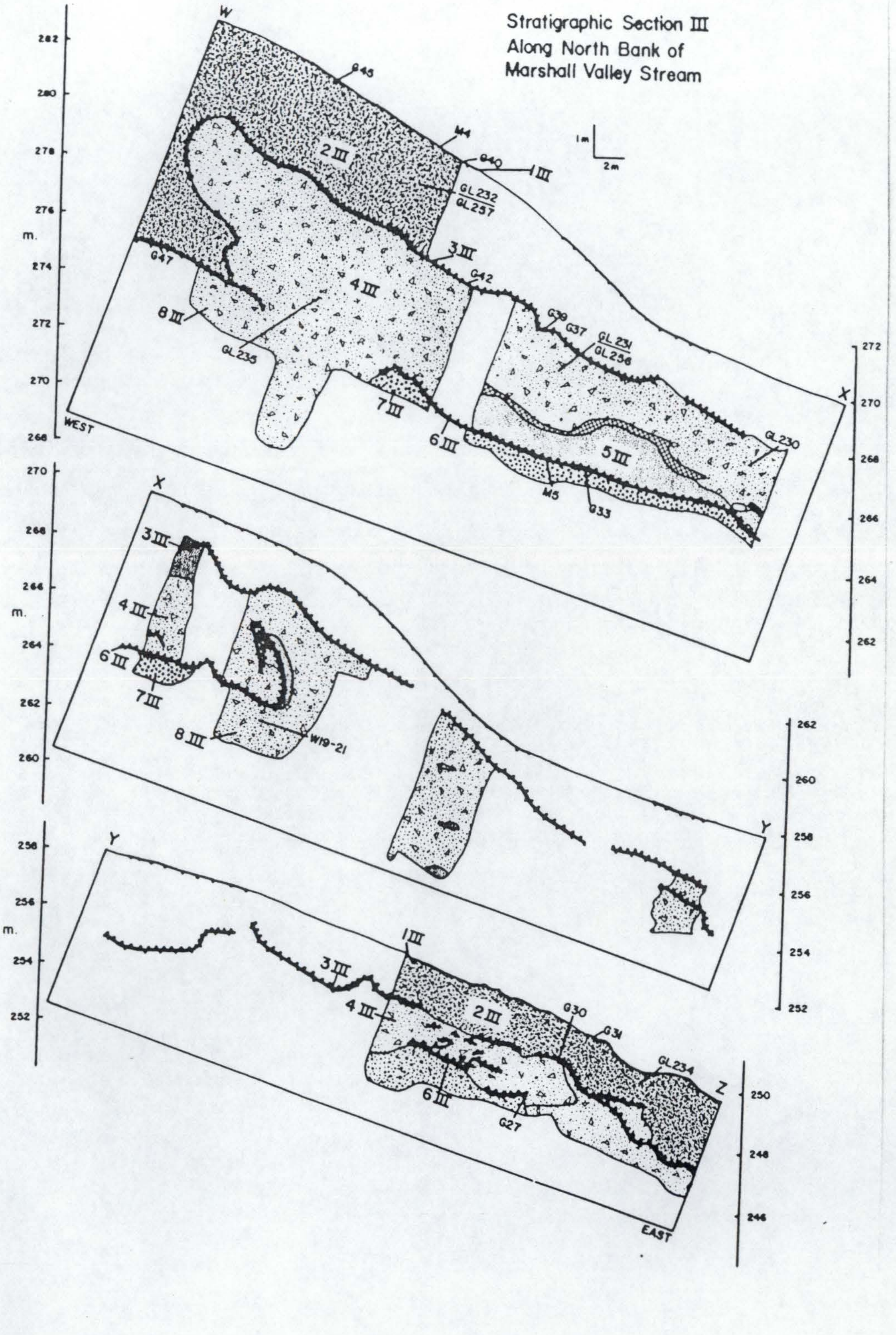
G42, R123, C552	CARBONATE OR GYPSUM SAMPLES	W10-12, GL209 GL245	SEDIMENT SAMPLE	M4	DIATOM SAMPLE
QL-1584	RADIOCARBON SAMPLE	MV1.2	PALEOMAGNETIC SAMPLE	3II	UNIT DESIGNATION

Fig. 5



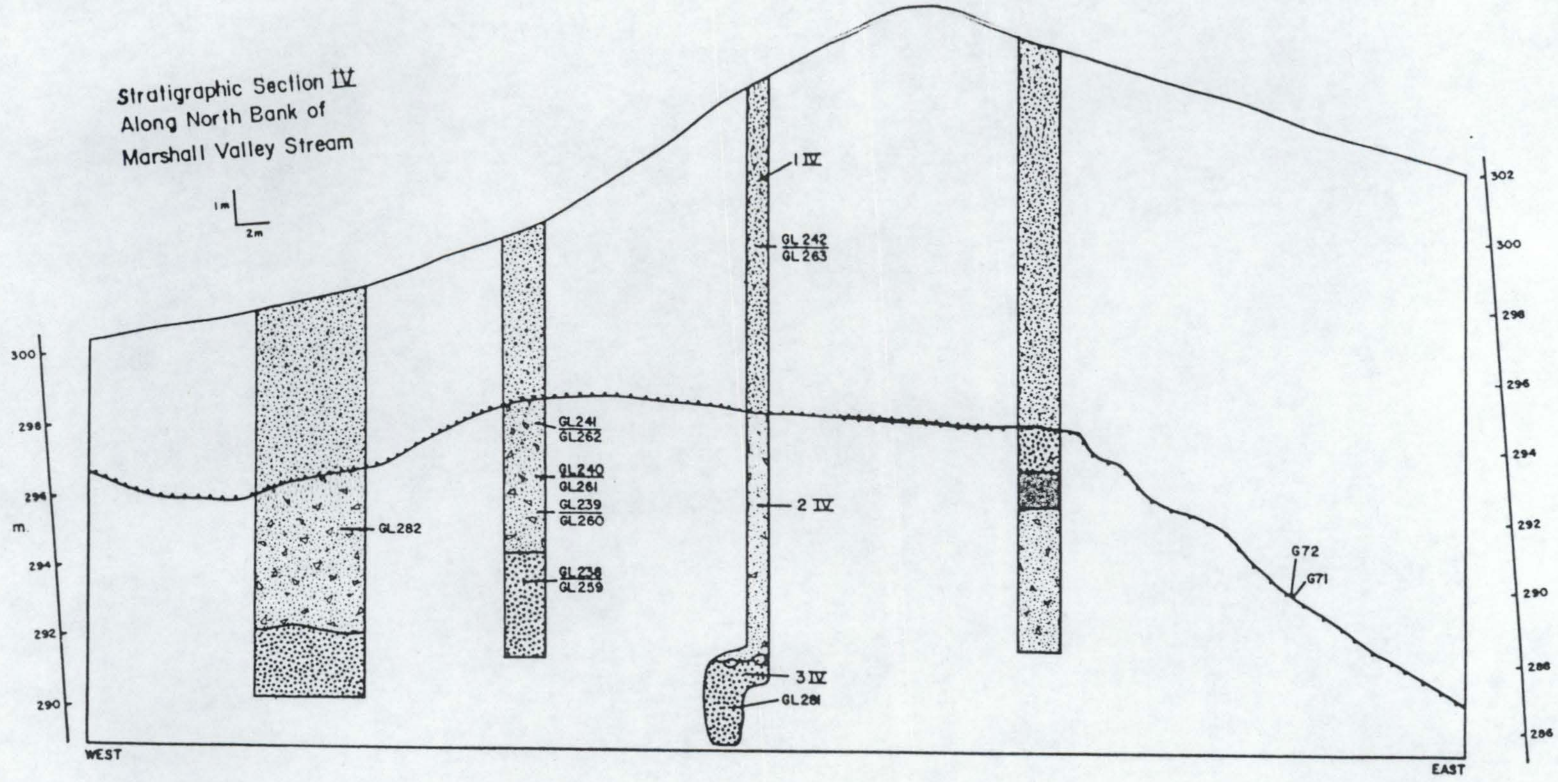
Stratigraphic Section II
Along North Bank of
Marshall Valley Stream

Fig. 6



Stratigraphic Section III
Along North Bank of
Marshall Valley Stream

Fig. 7



Stratigraphic Section IV
Along North Bank of
Marshall Valley Stream

Fig. 8

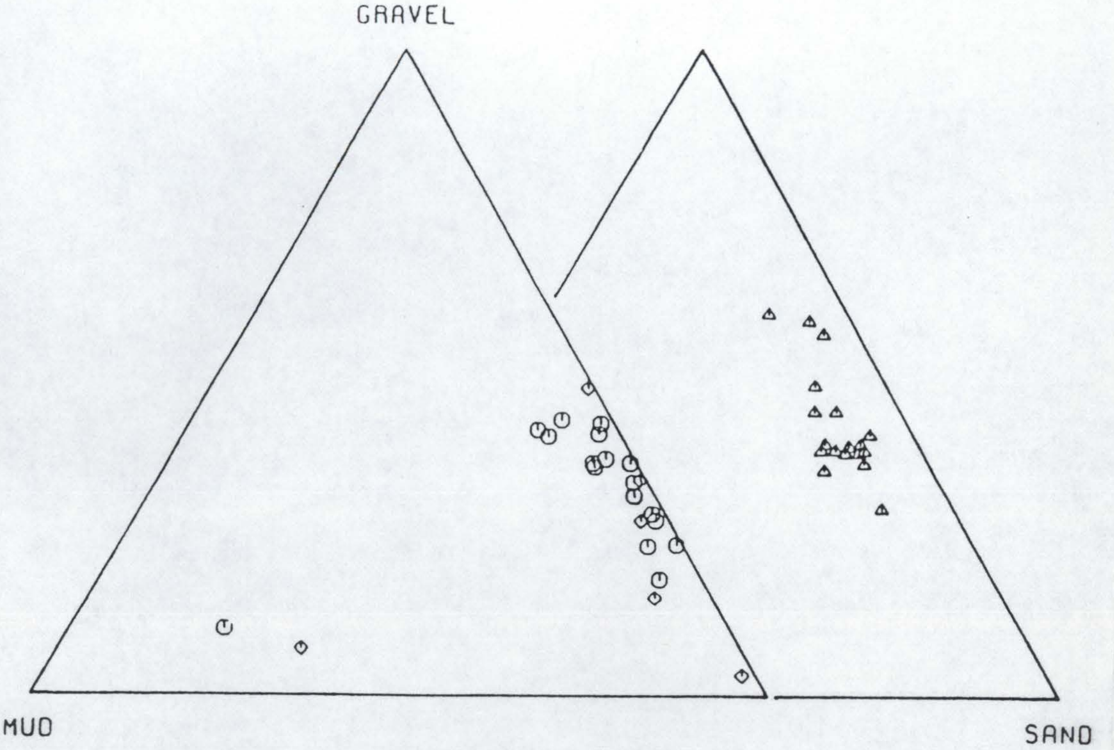


Fig. 9

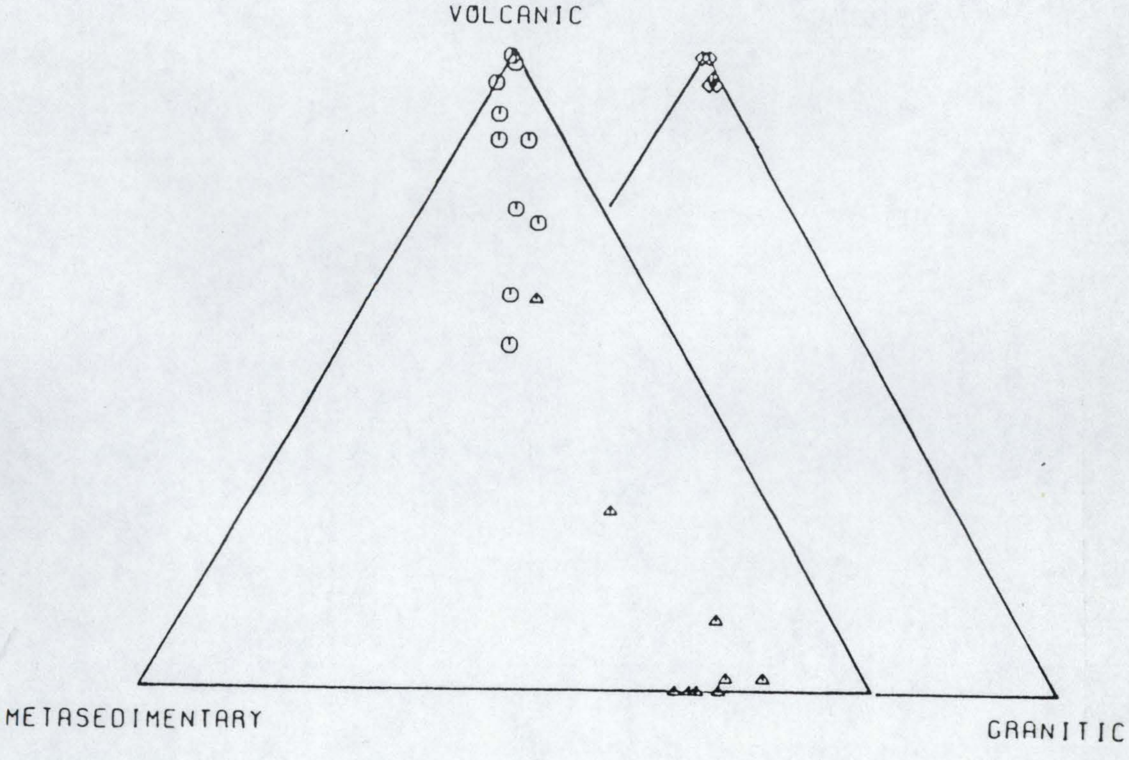
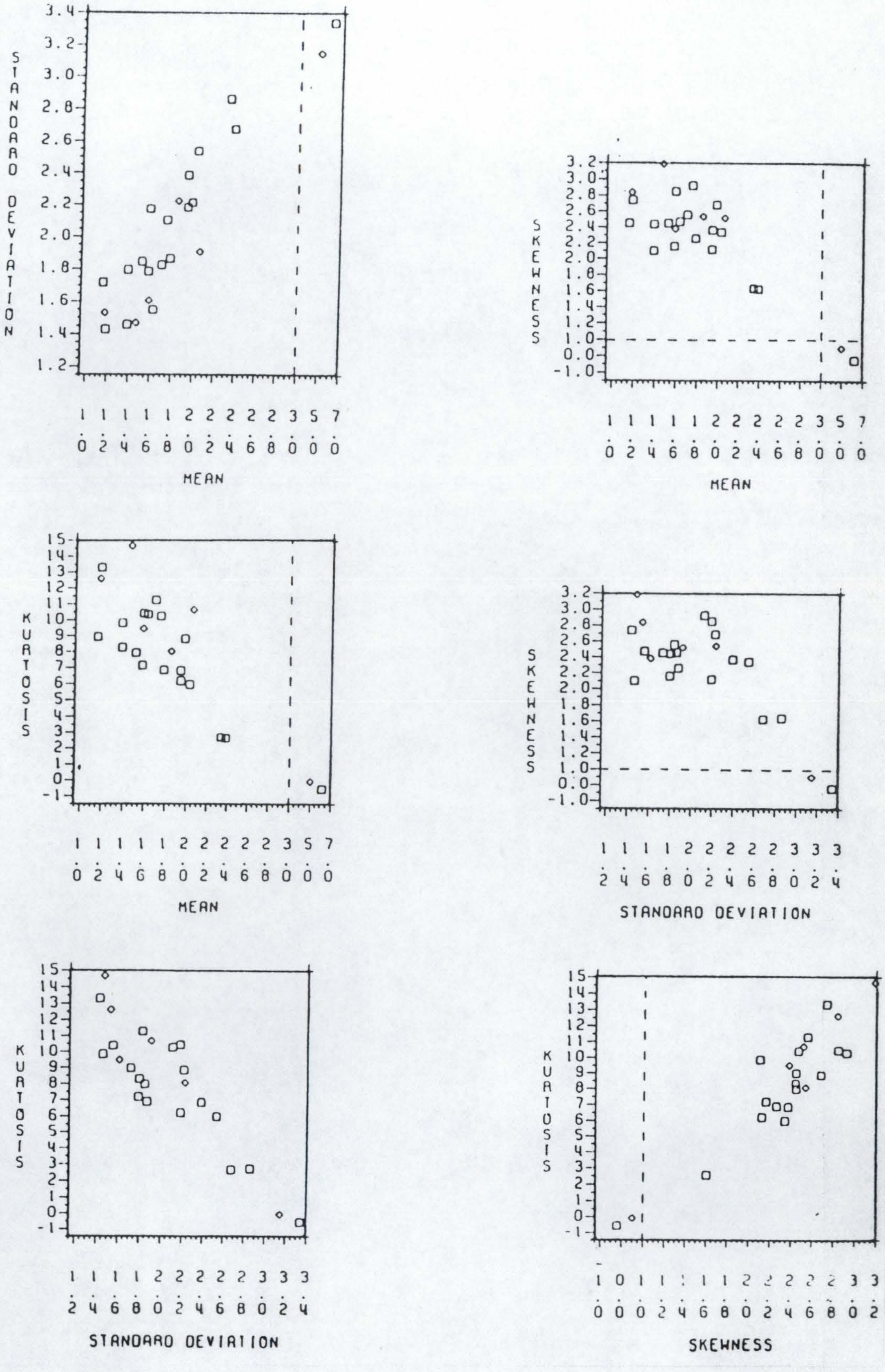


Fig. 10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3.2
3.0
2.8
2.6
2.4
2.2
2.0
1.8
1.6
1.4
1.2
MEAN
10 12 14 16 18 20 22 24 26 28 30 50 70
SKEWNESS
3.2
3.0
2.8
2.6
2.4
2.2
2.0
1.8
1.6
1.4
1.2
1.0
0.0
-1.0
MEAN
10 12 14 16 18 20 22 24 26 28 30 50 70
KURTOSIS
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
-1
MEAN
10 12 14 16 18 20 22 24 26 28 30 50 70
SKEWNESS
3.2
3.0
2.8
2.6
2.4
2.2
2.0
1.8
1.6
1.4
1.2
1.0
0.0
-1.0
STANDARD DEVIATION
12 14 16 18 20 22 24 26 28 30 32 34
KURTOSIS
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
-1
STANDARD DEVIATION
12 14 16 18 20 22 24 26 28 30 32 34
KURTOSIS
15
14
13
12
11
10
9
8
7
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3
2
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0
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SKEWNESS
-10 00 10 12 14 16 18 20 22 24 26 28 30 32

Fig. 10 (continued)

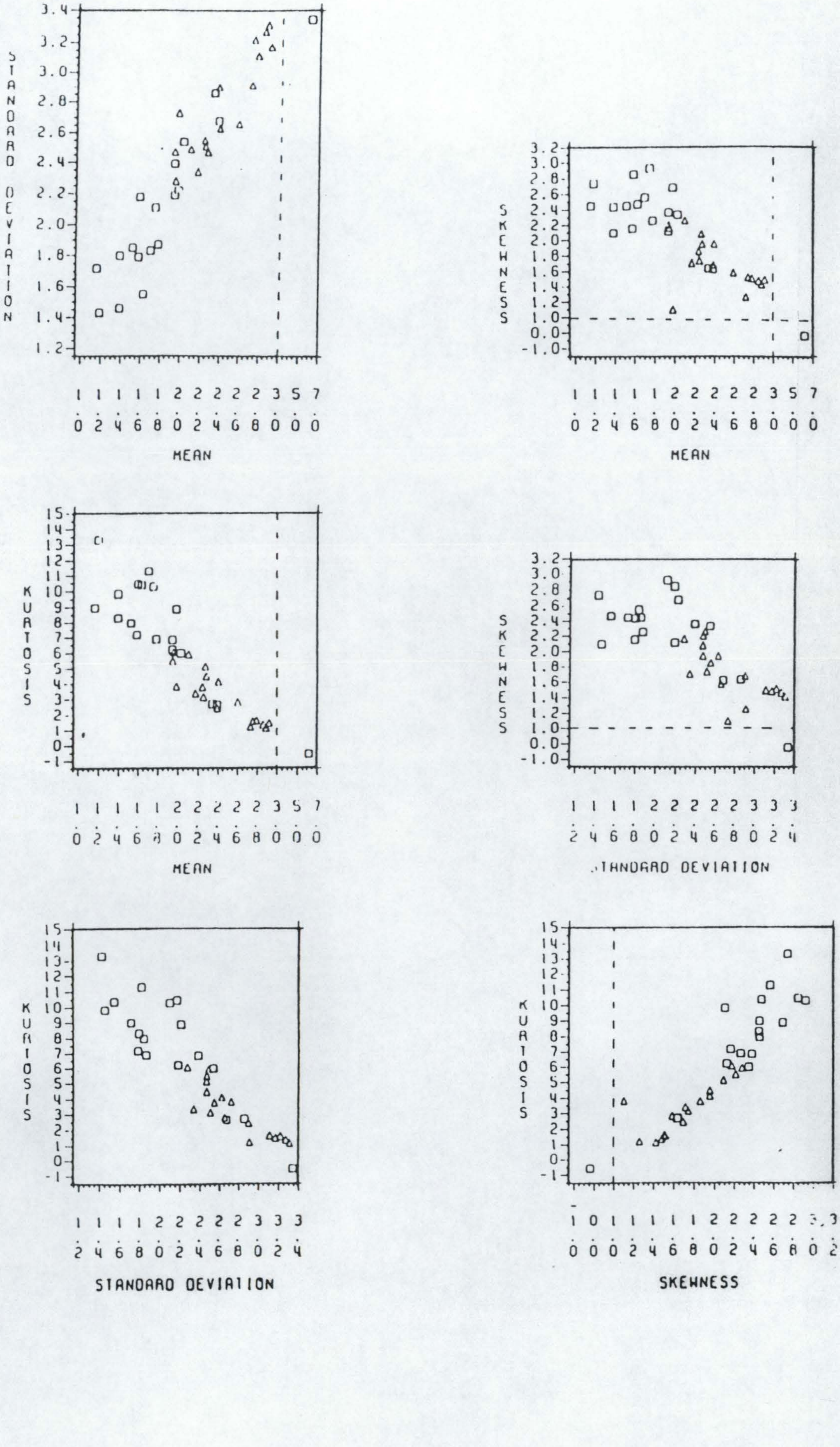
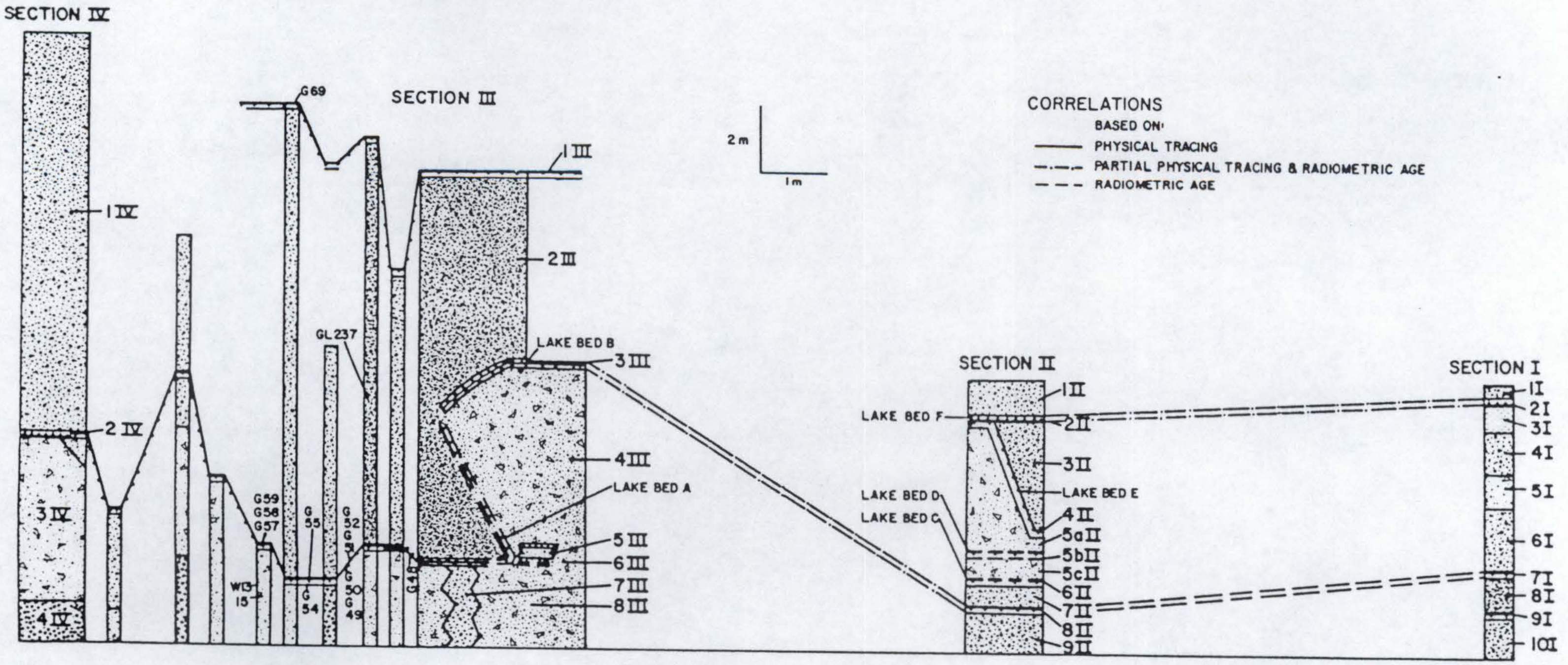


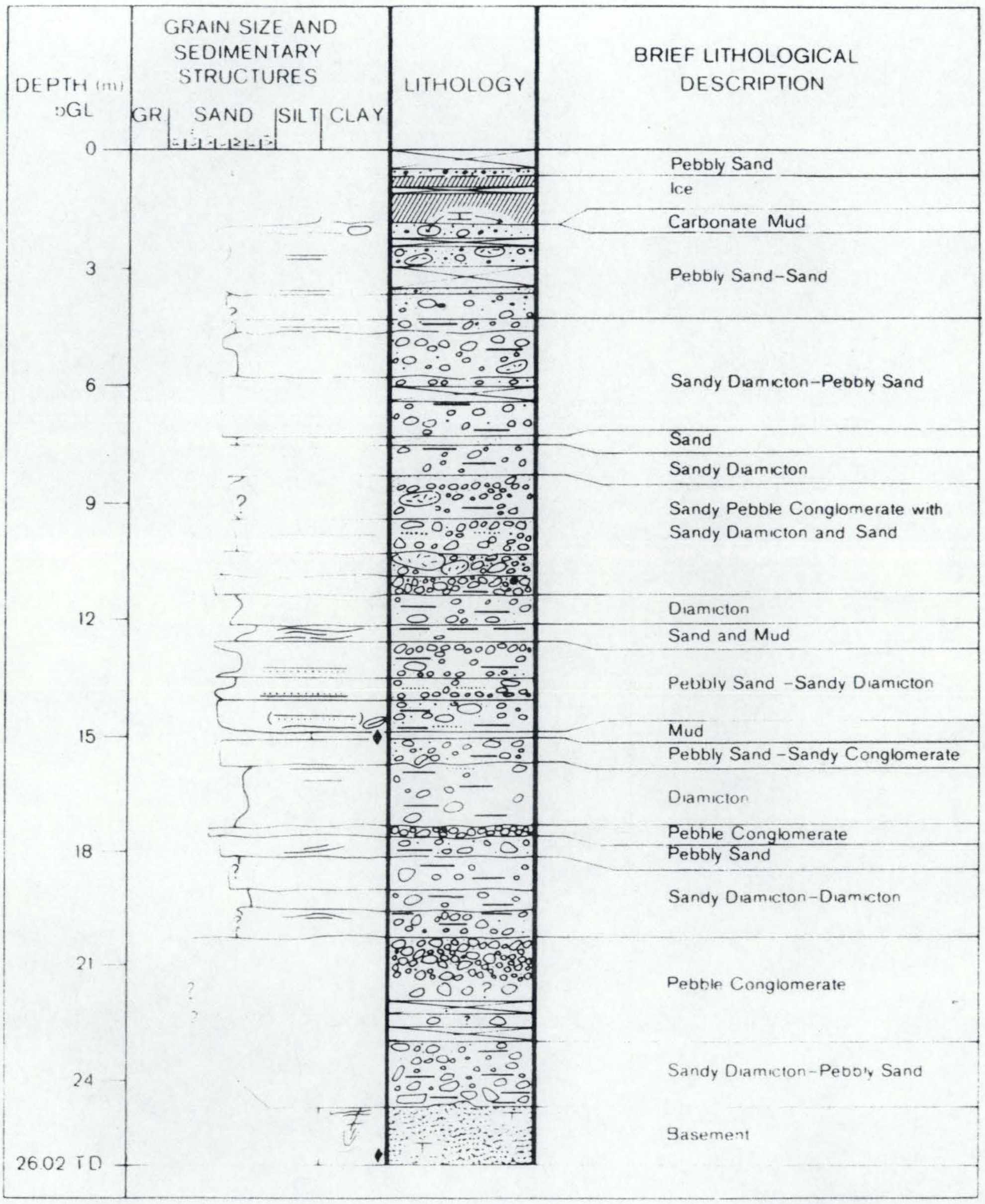
Fig. 12


SECTION IV
1 IV
2 IV
3 IV
4 IV
G69
SECTION III
1 III
2 III
GL 237
LAKE BED B
3 III
4 III
LAKE BED A
5 III
6 III
7 III
8 III
G59
G58
G57
W13-15
G 55
G 54
G 52
G 51
G 50
G 49
G 48
2 m
1 m
CORRELATIONS
BASED ON:
PHYSICAL TRACING
PARTIAL PHYSICAL TRACING & RADIOMETRIC AGE
RADIOMETRIC AGE
SECTION II
1 II
LAKE BED F
2 II
3 II
LAKE BED E
LAKE BED D
LAKE BED C
4 II
5a II
5b II
5c II
6 II
7 II
8 II
9 II
SECTION I
1 I
2 I
3 I
4 I
5 I
6 I
7 I
8 I
9 I
10 I

APPENDIX III

(from Robinson,1986)

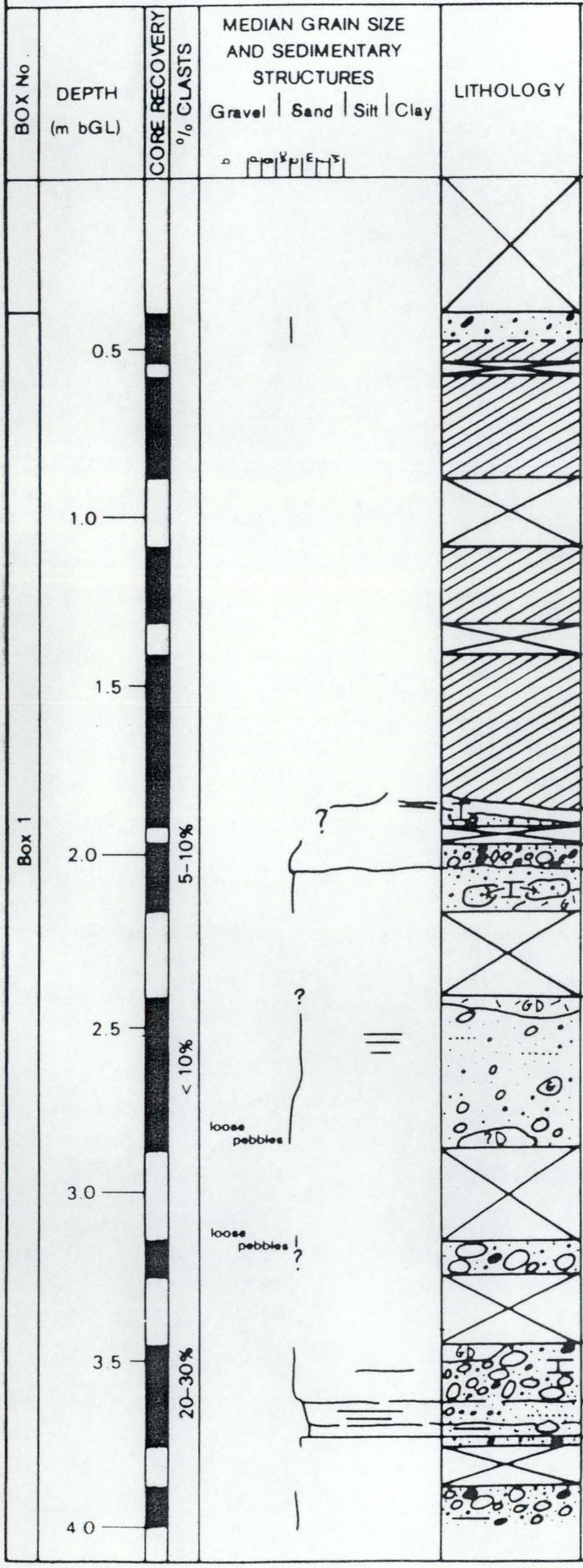
MARSHALL VALLEY 1 SUMMARY LOG



PROJECT Marshall Valley MV-1

SHEET No. 1

SCALE 1:20



DESCRIPTION

1 PEBBLY SAND non-stratified, poorly sorted, fine to coarse, volcaniclastic sand with subangular to subrounded, predominantly volcanic, small pebbles.

2 ICE clear to weakly bubbly ice, containing diffuse, medium to coarse sand grains in the upper portion; becoming almost debris-free and bubble-free in the lower portion. Bubble foliae are inclined up to 40° and bubbles appear elongate (or stretched). Total ice thickness c. 1.39 m; lower contact sharp and inclined 30°.

3 CARBONATE MUD pale greyish yellow to buff, weakly stratified, weakly indurated, strongly calcareous mud overlying calcareous sandy (to gritty) mud.

4 PEBBLY SAND see unit 1.

5 PEBBLY SAND non-stratified, moderately well sorted, coarse sand containing granules and pebbles, including two carbonate cemented, mud clasts containing sand and granules, also one granitic clast (c. 50 mm).

6 PEBBLY SAND very weakly stratified, poorly sorted, medium to coarse, volcaniclastic sand with abundant granules and scattered angular to subrounded, small pebbles to small cobbles.

7 PEBBLY SAND non-stratified, very poorly sorted, medium to very coarse sand containing angular to subangular, small to large pebbles of predominantly volcanic composition.

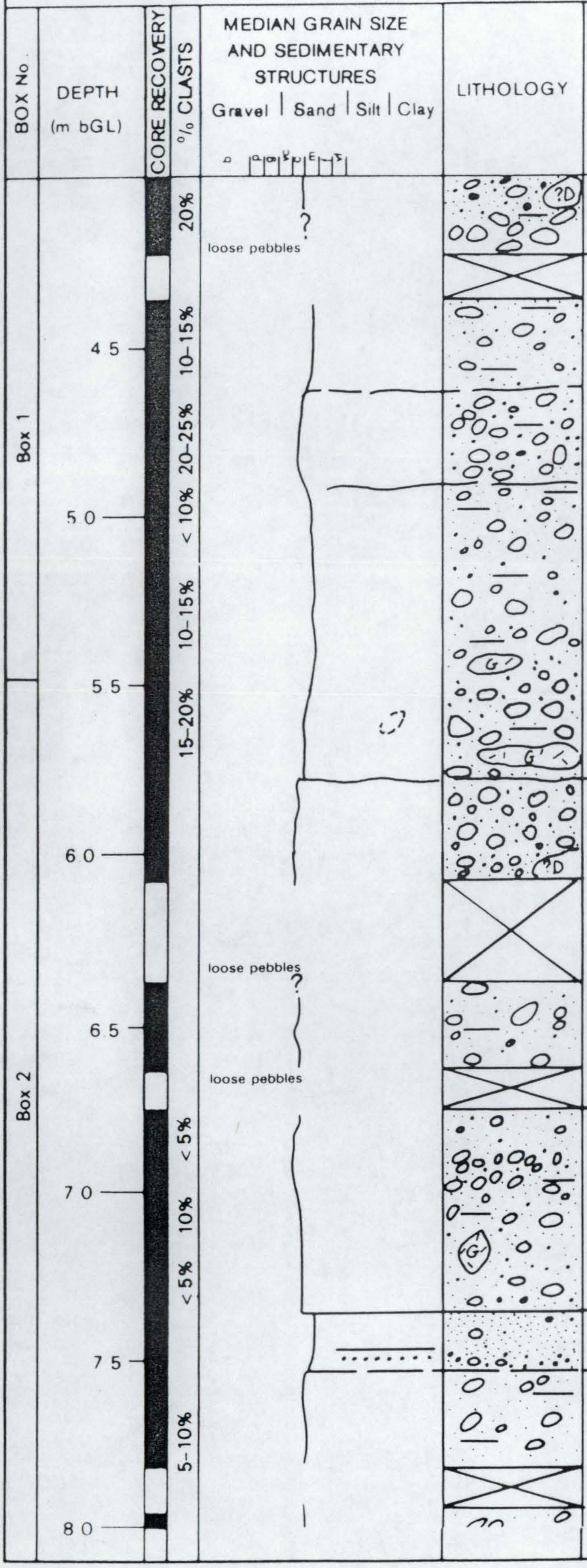
8 SAND very weakly stratified, moderately well sorted, fine to medium sand with few small pebbles.

9 PEBBLY SAND see unit 7, also contains minor mud giving a sandy diamicton character in parts (3.70–3.74 m).

PROJECT Marshall Valley MV-1

SHEET No. 2

SCALE 1:20



DESCRIPTION

10 PEBBLY SAND (?SANDY DIAMICTON): non-stratified, non-calcareous, poorly sorted, medium to coarse sand, with minor mud and granules and scattered angular to subrounded, very small to large pebbles of predominantly plutonic and volcanic composition. A few pebbles are coated, or partly coated, by a hard, non-calcareous, thin (c. 1 mm) veneer

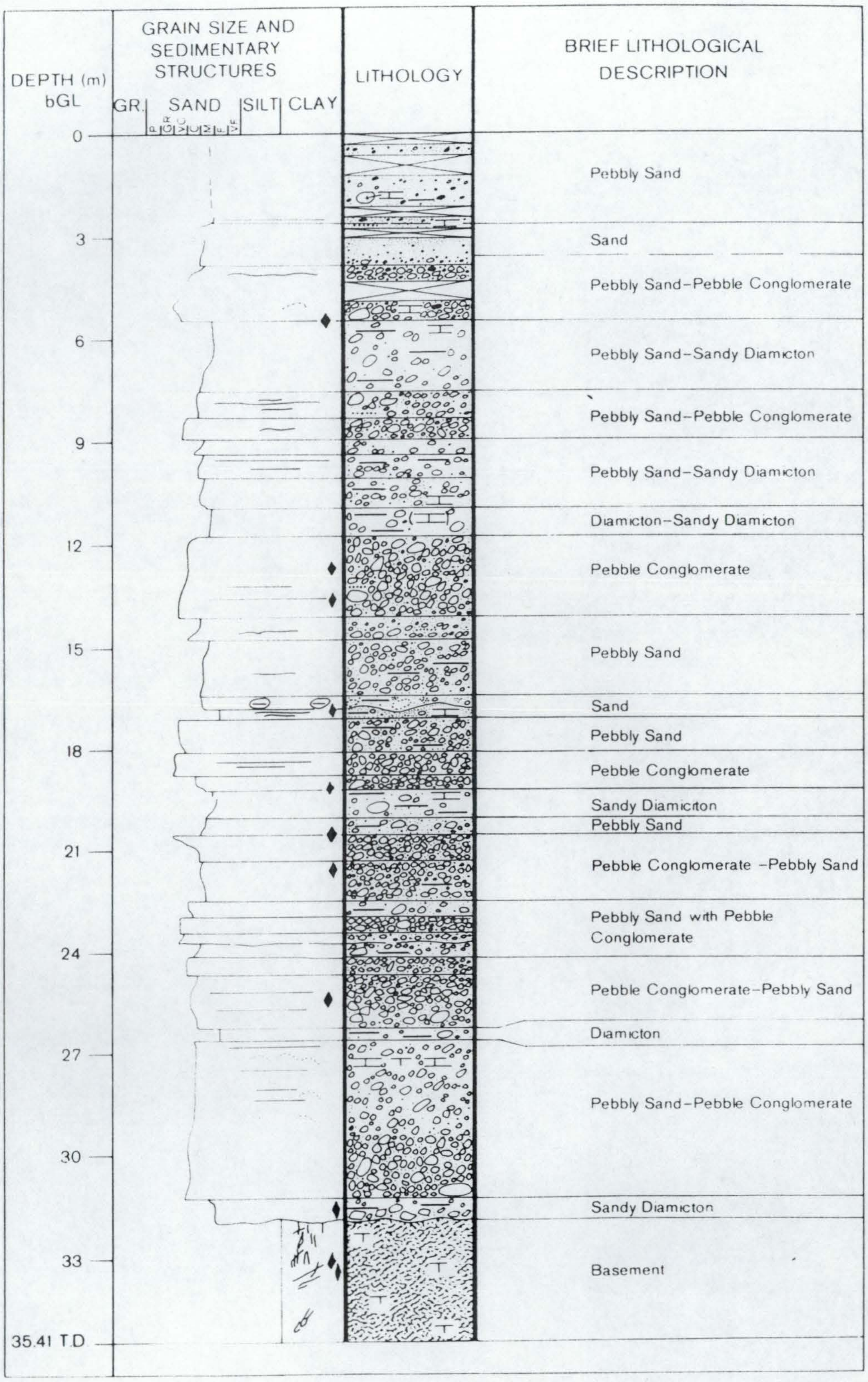
11 PEBBLY SAND non-stratified, non-calcareous, poorly sorted, coarse sand with scattered granules and pebbles which are occasionally coated (as described above - unit 10)

12 PEBBLY SAND (?SANDY DIAMICTON) non-stratified, non-calcareous, very poorly sorted slightly muddy, medium to very coarse sand with granules, and angular to subrounded, very small pebbles to small cobbles (predominantly of granitic and volcanic composition)

13 SAND non-stratified, poorly sorted to moderately well sorted, medium sand with minor scattered granules and pebbles; small pebbles and granular sand at the unit base grade upwards to medium sand.

14 PEBBLY SAND (SANDY DIAMICTON) non-stratified, slightly indurated, non-calcareous, poorly sorted, muddy, medium to coarse sand, with angular to subrounded very small to large pebbles; (predominantly of granitic and volcanic composition)

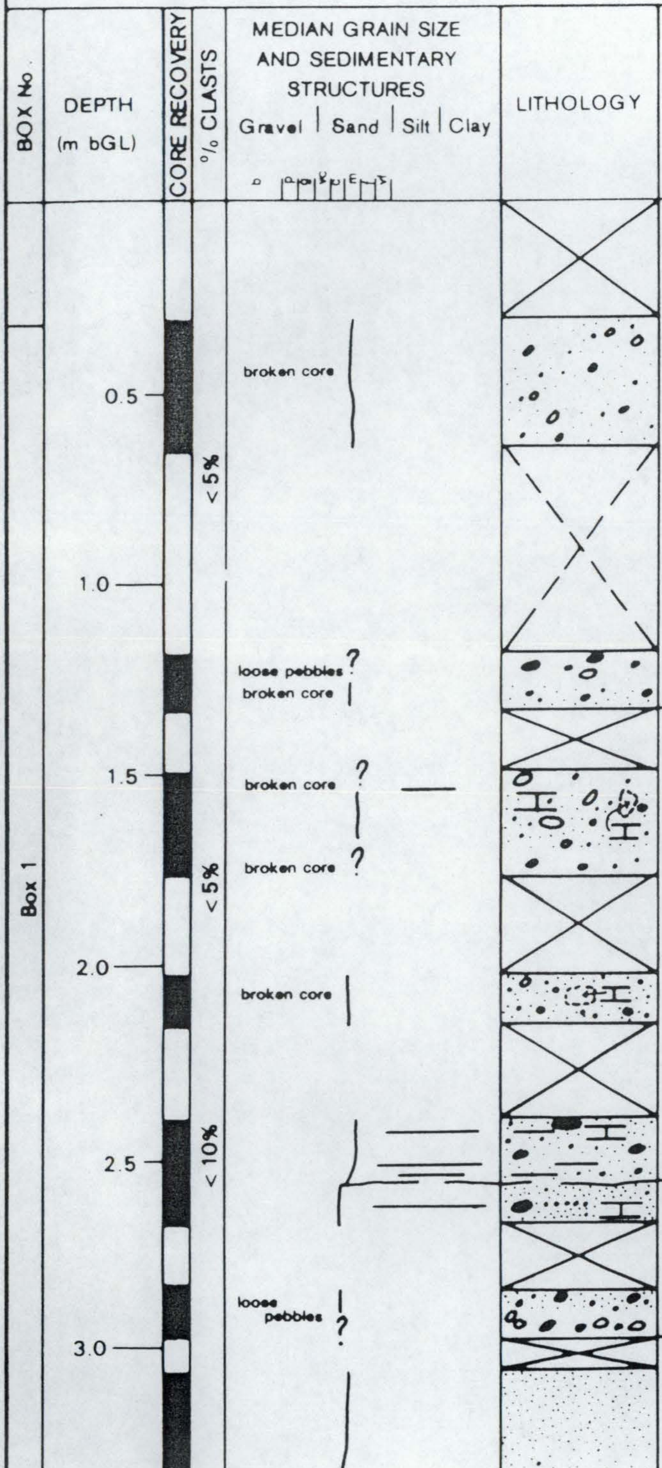
MARSHALL VALLEY 2 SUMMARY LOG



PROJECT Marshall Valley MV-2

SHEET No. 1

SCALE 1:20



DESCRIPTION

1 GRANULAR SAND: non-stratified, non-calcareous, poorly sorted, very fine to medium, volcaniclastic sand with angular to subrounded, small to large pebbles (up to 26 mm) of predominantly volcanic and granitic composition.

2 PEBBLY GRANULAR SAND: non-stratified to weakly stratified, calcareous to strongly calcareous, poorly sorted to very poorly sorted, very fine to coarse, volcaniclastic sand with angular to subrounded, small to large pebbles (up to 20 mm) of volcanic, granitic and ?marble composition. Carbonate cement and clasts occurs throughout. Muddy laminae near unit base give pale yellow mottling.

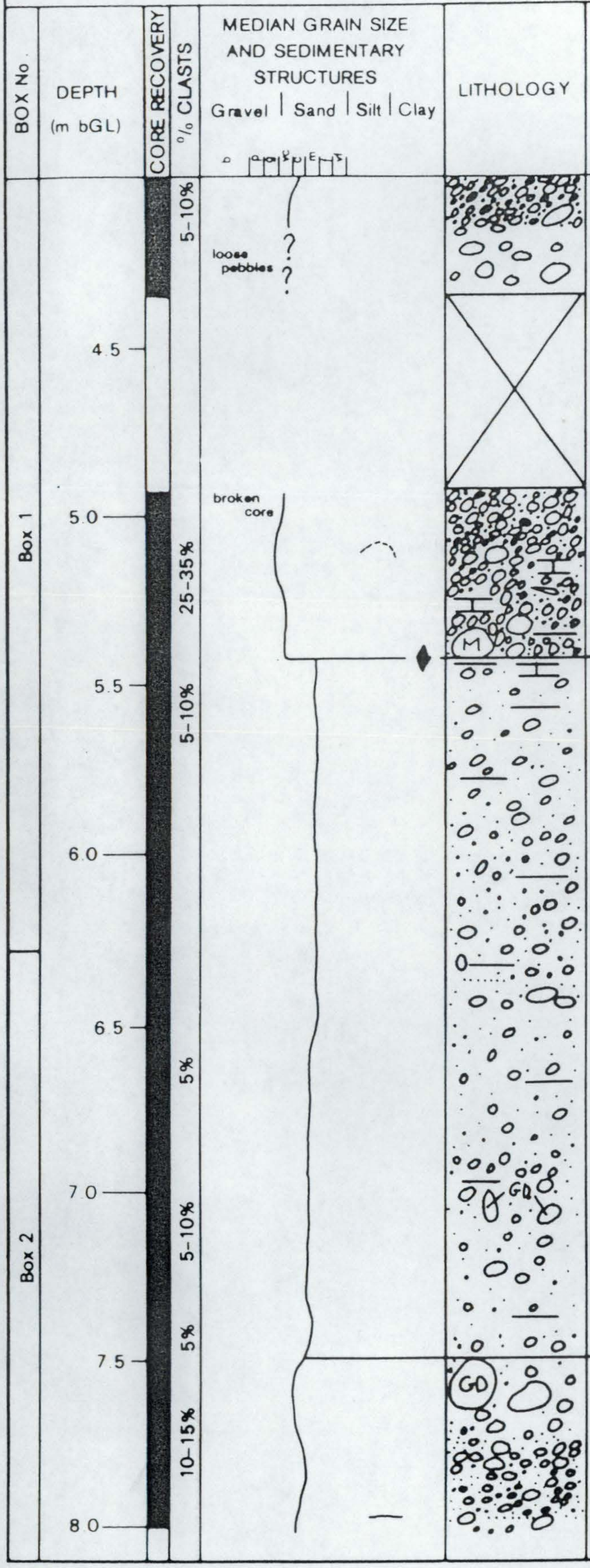
3 PEBBLY GRANULAR SAND: similar to unit 2, with slightly coarser matrix and non-stratified.

4 SAND: non-stratified, non-calcareous, moderately well sorted to poorly sorted, medium volcaniclastic sand; grading upwards from a coarse sand base to medium sand; no pebbles.

PROJECT Marshall Valley MV-2

SHEET No. 2

SCALE 1:20



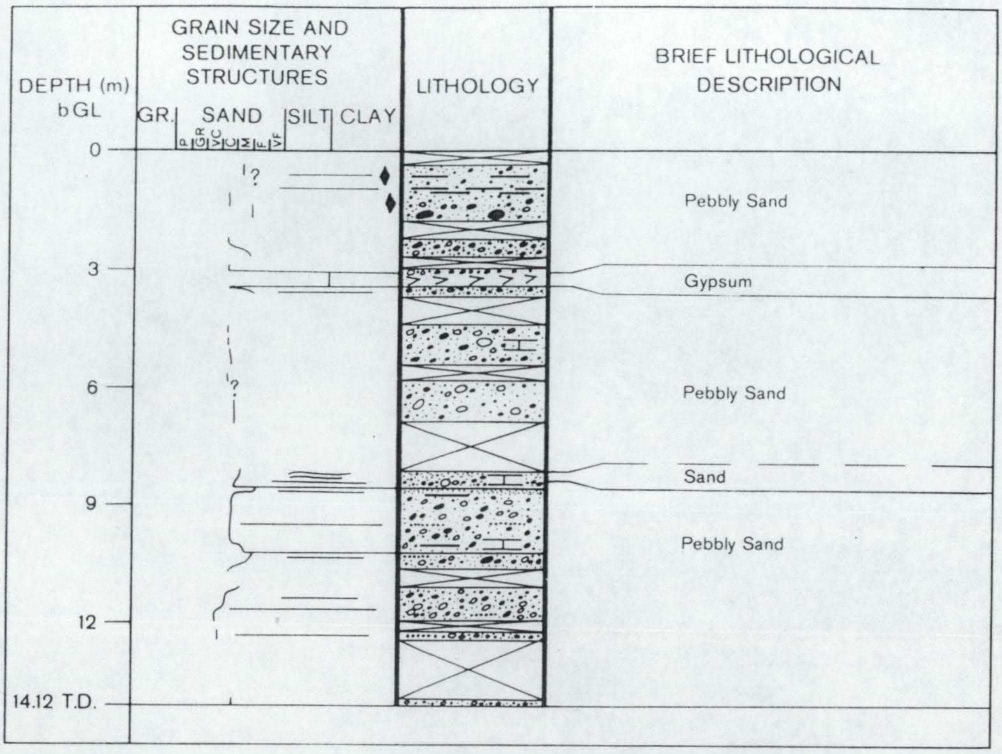
DESCRIPTION

5 PEBBLY GRANULAR SAND TO GRANULAR CONGLOMERATE non-stratified, non-calcareous to very calcareous, poorly sorted to very poorly sorted, slightly muddy, medium to very coarse sand matrix. The clasts are angular to subrounded, small pebbles to large cobbles of predominantly granitic composition and are both matrix- and clast-supported. A hard, thin (< 1 mm) mud coating is formed around a few large pebbles (at c. 5.0 m). High intergranular ice occurs near the unit base and the sandy matrix is absent

6 PEBBLY MUDDY SAND TO SANDY DIAMICTON: non-stratified, non-calcareous to very calcareous, poorly sorted to very poorly sorted, slightly muddy fine to medium sand forming a matrix around angular to subrounded, small to very large, granitic and ? metamorphic pebbles. The unit grades upwards from a pebbly muddy sand at the base to a sandy diamicton

7 PEBBLY GRANULAR SAND non-stratified to weakly stratified, non-calcareous, poorly sorted, medium to coarse sand with abundant granules and angular to subrounded, small pebbles to large cobbles of granitic and ?dolerite composition

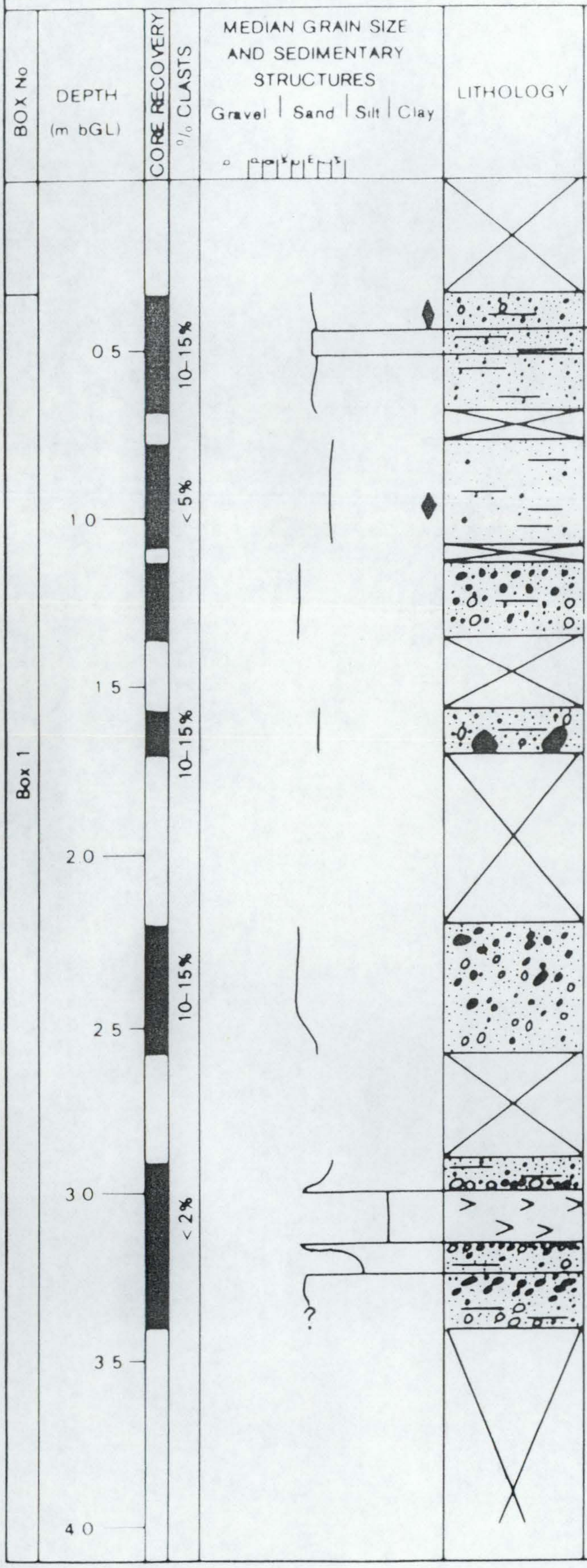
MARSHALL VALLEY 3 SUMMARY LOG



PROJECT Marshall Valley MV-3

SHEET No. 1

SCALE 1:20



DESCRIPTION

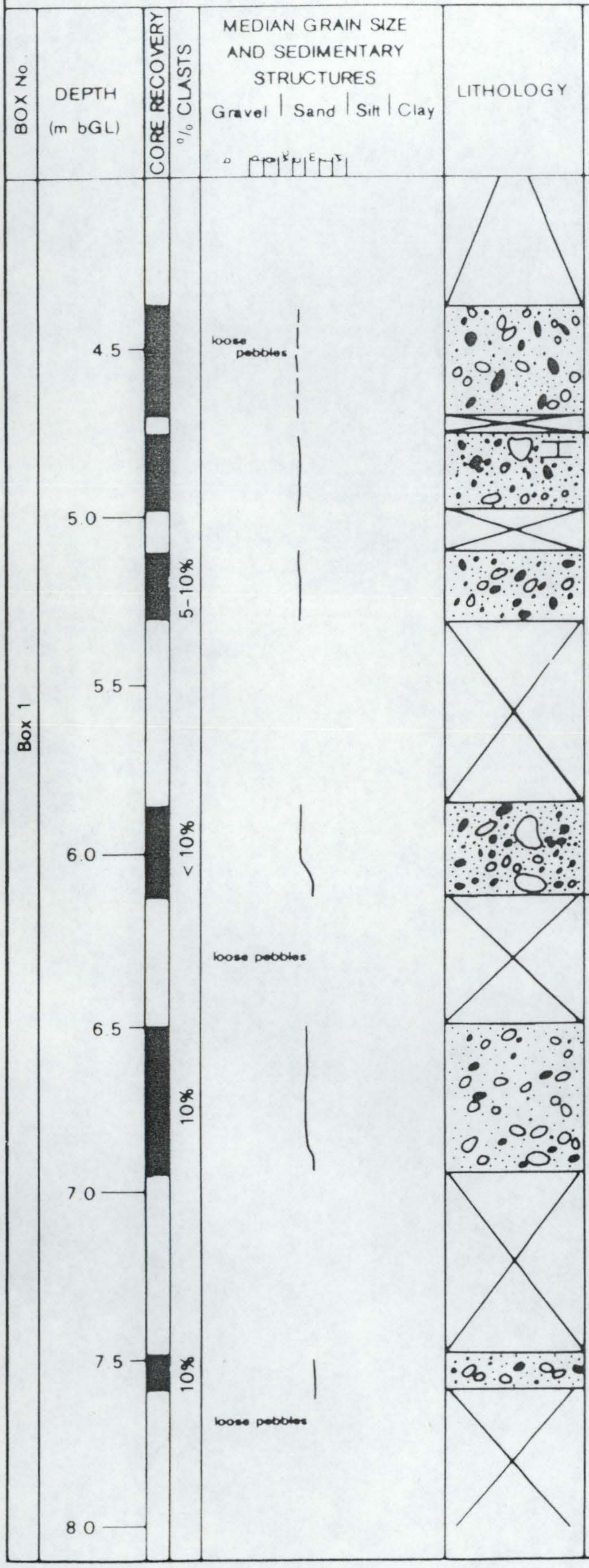
1 **PEBBLY SAND** non-stratified, non-calcareous, poorly sorted, ice cemented, fine to coarse sand with a high volcaniclastic content, granules and small to medium pebbles occur throughout. Between 0.76 and 1.08 m there is a non-stratified calcareous, moderately well sorted, fine sand bed with minor small basalt pebbles. Gypsum (150 mm thick) and sandy mud beds occur between 2.9 and 3.25 m. Poor core recovery.

PROJECT Marshall Valley MV-3

SHEET No. 2

SCALE 1:20

DESCRIPTION



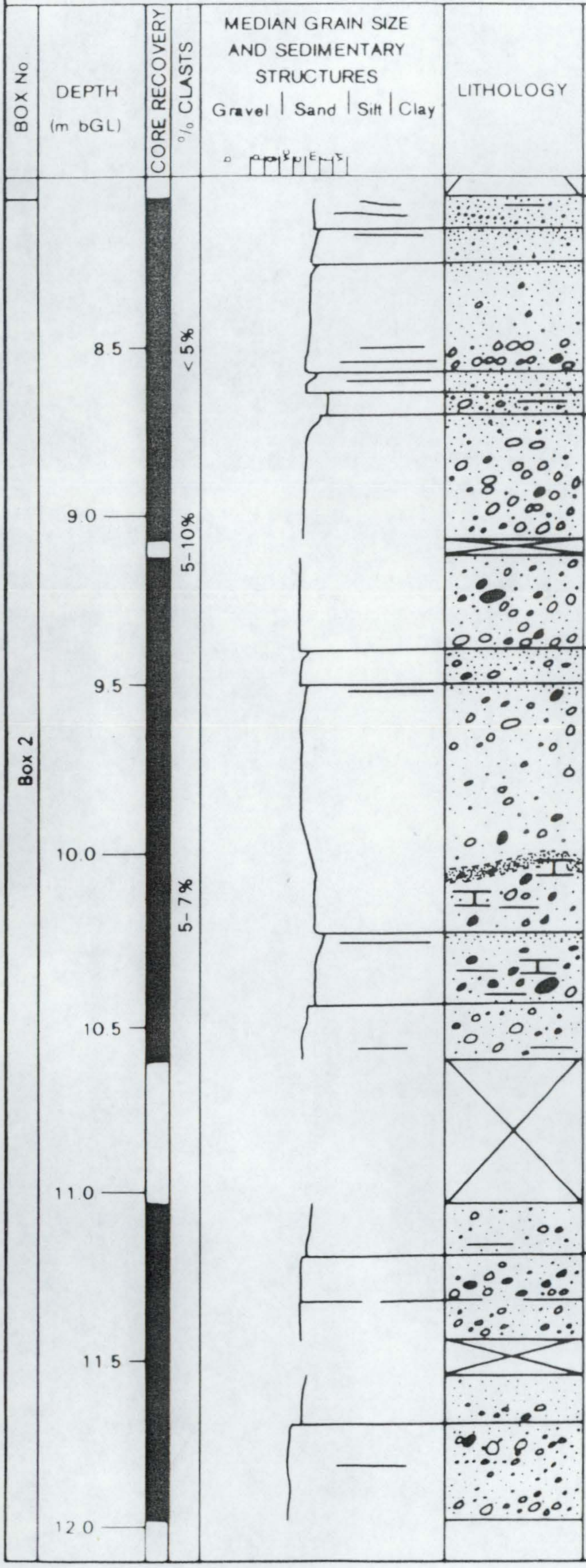
2 PEBBLY SAND non-stratified, slightly calcareous, poorly sorted, medium to coarse sand with abundant subangular to subrounded, small to large basaltic pebbles. Poor core recovery

3 PEBBLY SAND similar to unit 2, but non-calcareous

PROJECT Marshall Valley MV-3

SHEET No. 3

SCALE 1:20



DESCRIPTION

4 GRANULAR SAND weakly stratified to well stratified, thin-bedded, calcareous moderately well sorted to poorly sorted, fine to medium volcaniclastic, sand with numerous granules and a few small pebbles Mud laminae near the top of the unit are irregular wavy and inclined up to 45°

5 PEBBLY SAND non-stratified, slightly calcareous poorly sorted to very poorly sorted, medium to coarse volcaniclastic sand with abundant volcanic and granitic subangular to rounded, small to very large pebbles

6 PEBBLY SAND weakly stratified, calcareous to strongly calcareous, poorly sorted to very poorly sorted, muddy, medium to coarse, volcaniclastic sand with scattered volcanic and granitic, angular to subrounded, small to very large pebbles A calcite cemented-layer occurs c 10.1 m

7 SAND TO PEBBLY SAND weakly stratified, non-calcareous to slightly calcareous poorly sorted medium to coarse volcaniclastic sand with scattered volcanic and granitic subangular to subrounded, small to large pebbles Poor core recovery

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