LATE CAINOZOIC GLACIATION

AND

MOUNTAIN GEOMORPHOLOGY

IN THE

CENTRAL HIGHLANDS OF TASMANIA.

by Kevin Kiernan

BA(Hons.)

Submitted in fulfilment of

the requirements for the degree of

Doctor of Philosophy

UNIVERSITY OF TASMANIA

HOBART

1985

(source unknown)

" Among all of nature's phenomena, not a single one seems to me to be more worthy of the interest and curiosity of the naturalist than glaciers."

-L. Agassiz, 1840

Etudes sur Les Glaciers : 2.

TABLE OF CONTENTS

DECLARATION

This thesis contains no material which has been accepted for the award of any degree or diploma in any university and, to the best of my knowledge and belief, contains no copy or paraphrase of material previously published or written by another person, except where due reference is made in the text.

Kevin Kiernan

December 1985.

ACKNOWLEDGEMENTS

I acknowledge with considerable gratitude the the assistance of Dr. Eric Colhoun throughout the project. I am also deeply appreciative of the efforts of Dr. Les Wood. It is with gratitude that I remember too the encouragement and assistance I received from the late Dr. Joe Jennings, and also from Dr. Mike Macphail.

My thanks are extended to those who have helped me in the field at various times, including Peter Robertson, Barry Davey, Richard Cosgrove, Karen and Ellen Kiernan, Steve Harris, Norm Sanders and my benumbed companion in the "Rotten to the Core" expedition to drill the Gould Plateau tarn, Mick Iloski.

The Department of Geography, University of Tasmania, provided a base and some logistical support and also funded the radiocarbon dating. The study was undertaken while I was a recipient of an award from the Australian Department of Education under the Commonwealth Postgraduate Research Awards Scheme.

The National Parks & Wildlife Service (Tasmania) provided some accomodation at Lake St. Clair and I would like to thank Dick Dwyer, Steve Hepworth and Bob Tyson of the Service in particular. Kevin Ellis provided boat transport on Lake St Clair when needed.

Kate Charlesworth patiently and skilfully drafted figures 4.2, 5.3, 6.2, 10.2 and 12.2 and many of the other figures are the better for her advice. I would also like to thank Nel Gill for typing the tables. Dr. A.R. Martin and Dr. J.I. Raine kindly allowed me to use unpublished data from the Mt. Kosciusko area. For help in other ways I should also like to thank Dennis Charlesworth, Therese Hughes, Jamie Kirkpatrick, Sib Corbett, Tom Errey, Professor Martin Williams and Lily Hughes. The assistance given by my wife Karen in proof reading and other tasks during the final preparation of this thesis was invaluable.

Perhaps the greatest debt of all is owed to the friends who encouraged me, and to Karen and Ellen who tolerated it all.

ABSTRACT

The broad topographic framework and erosion surface morphology of west central Tasmania predates the early Pleistocene. The valley systems, however, have been emphasised by glacial erosion which has played a major role in shaping the detailed geomorphology of the mountains.

Part of an extensive ice cap that developed in the Tasmanian Central Highlands during the late Cainozoic discharged southwards via a major outlet glacier that occupied the valley of the Derwent River.

The heart of the Central Tasmanian ice cap probably lay west of the Du Cane Range. When the ice cover was most extensive the Derwent Glacier was up to 500 metres thick. It may have extended to as low as 230 metres above sea level, 70 kilometres downstream from its source in the cirques of the Du Cane Range. Two diffluent lobes of this glacier spread eastwards to merge with other glaciers in the Nive Valley. Other diffluent lobes extended southwards into the upper Gordon Valley, and westwards into the upper Franklin and Alma valleys. At the maximum phase the Franklin and Alma glaciers were confluent around Mt. Alma, near the present junction of the Collingwood and Franklin rivers.

The more westerly glaciers displayed the highest rates of mass throughput hence glacial landforms are more abundant and better developed in the west.

Analysis of the post-depositional modification of the glacial landforms and sediments suggests that at least three glaciations took place. The first glaciation was probably early Pleistocene or late Pliocene in age while the most recent and smallest occurred during the late Last Glacial Stage.

Glaciation would have demanded colder temperatures and an increased solid precipitation budget, but no major shift in the direction of snow bearing winds is neccessitated. At no stage was the mean annual air temperature likely to have been more than 9° C less than present.

The glaciations were probably broadly contemporaneous with those at similar southern latitudes in Andean Patagonia and South Island New Zealand. Like the glaciers of those areas the ice masses of west central Tasmania were mainly of temperate maritime character.

The glaciations were accompanied by periglacial activity beyond the limits of the ice. The development of rock glaciers suggests that localised areas of permafrost existed during the Last Glaciation.

The glacial oversteepening has greatly facilitated slope retreat in areas of high structural anisotropy, particularly under periglacial conditions. Interglacial weathering and erosion was comparatively innocuous, although the presence of a substantial vegetation cover seems to have been critical to the maintenance of slope stabilty, particularly in steeper and more elevated terrain. The geomorphic evidence does not demand any climate deterioration during the Holocene. The most active geomorphological agent of the Holocene interglacial is humankind.

LIST OF FIGURES

Section C

LIST OF TABLES

INTRODUCTION

The most extensive valley glacier systems known to have existed in Tasmania during the late Cainozoic were associated with an ice sheet at least several hundred square kilometres in extent that inundated the western part of Tasmania's Central Highlands. The principal southern oulet for this ice cap was the Derwent Glacier. This glacier excavated a major glacial trough that is now partly occupied by Lake St. Clair, Australia's deepest inland water body. Smaller alpine glaciers moulded the adjacent mountains and valleys.

This thesis examines the legacy of glaciation on the southern margin of the ice cap. It seeks to establish the areal and altitudinal limits of the glacier systems, the patterns of ice movement, and to differentiate the sequence, duration and ages of the glaciations. It also investigates the impact of glaciation upon the mountain geomorphology.

This study has benefitted from the contributions of earlier workers who established firstly that the area had been glaciated and secondly the broad outlines of the glacial geomorphology and glacial climates. Most importantly their work has highlighted issues which require resolution.

However, there are impediments to working in this area. These include terrain which is not only difficult to traverse but is also often cloaked in dense vegetation which limits geomorphological visibility. Few useful sections are available because streams have seldom incised deeply into the sediments which floor many of the valleys, while steep slopes preclude the survival of much surficial sediment. There are few tracks and even fewer roads. Indeed, a large part of the area has recently been placed on the United Nations World Heritage List in recognition of its wilderness values. Elsewhere difficulties have been imposed by the filling of a hydro-electric reservoir in the very critical southeastern part of the study area.

Nonetheless, a substantial area has been examined in detail for the first time. I count it my great privilege to have had the opportunity to undertake such a task in an area as beautiful as this one, generally with the intensity of wonder and appreciation that only solitude amid wild magnificance will allow.

PART A

ON THE EDGE OF AN ICE - CAP

chapter one
REGIONAL ENVIRONMENT

"During a long life spent in wandering and observation it has never been my good fortune to see so magnificant a picture"

-Sir John Franklin, Mt. Arrowsmith, 1842.

LOCATION AND CHARACTER OF THE STUDY AREA.

The island of Tasmania is the most mountainous region in Australia. This study focusses on some of the loftiest among its mountains. Elevations in the study area range from 1470 m. in the Du Cane Range down to 230 m further south in the Derwent River Valley. Lying between 41° 55' - 42° 25' south latitude, and 145° 55' - 146° 25' east longitude, these temperate mountains are wet and moderately windy, subject to the westerly regime of the Roaring Forties and within 80 km. of the Southern Ocean (Figure 1.1).

Although it presently lacks active glaciers this is a mountain environment in the classical sense, with considerable elevation, steep gradients, rocky terrain, seasonal snow and ice, a legacy of Pleistocene glaciation and considerable evironmental diversity. The relative relief between valley floors and summits is generally less than the 1000 m. regarded by Barsch and Caine (1984) as characteristic of high mountains. However, it constitutes a "High Mountain System" according to the criteria of Troll (1972, 1973) as it encompasses multiple vegetation belts

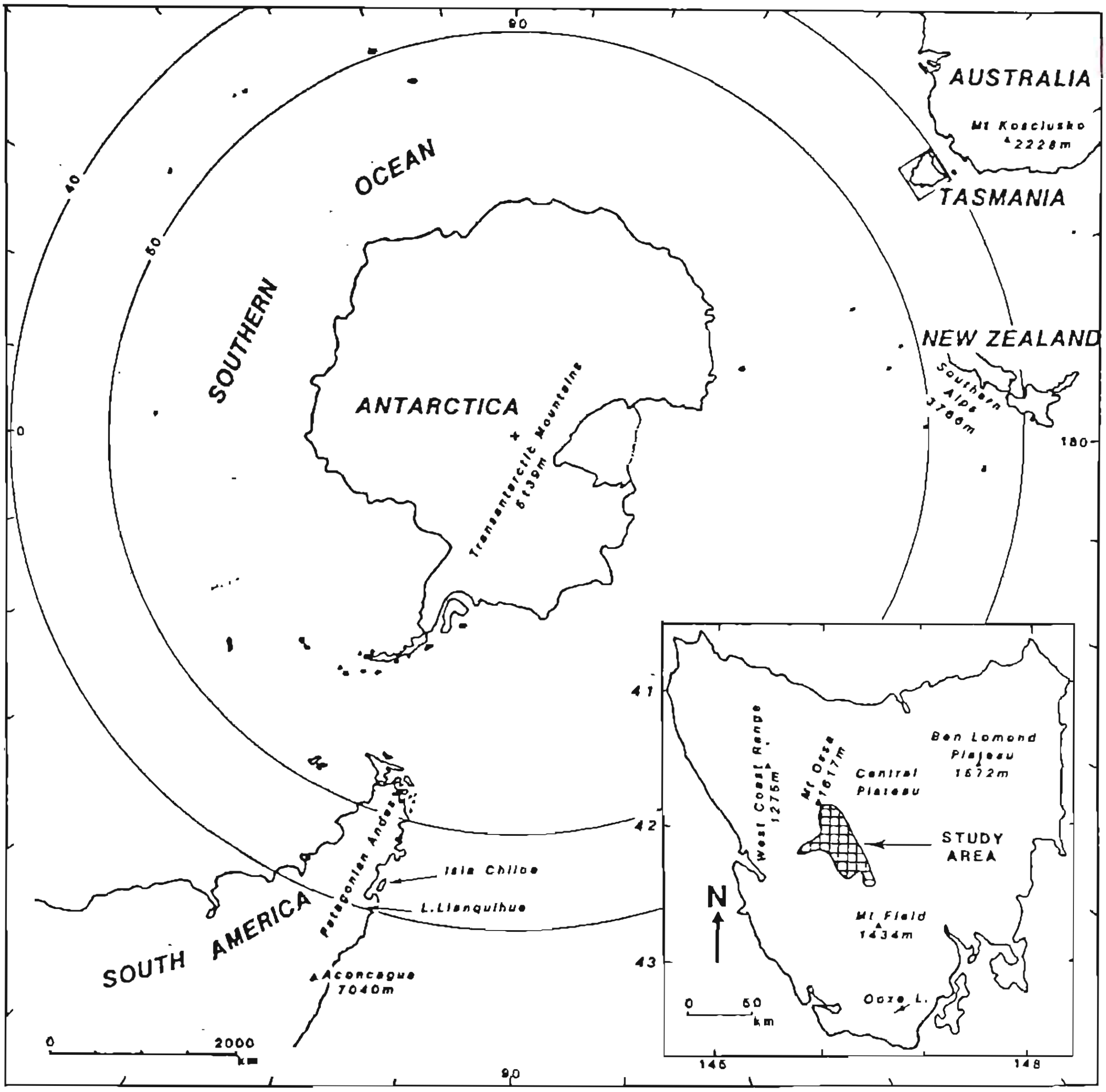


Figure 1.1 Location of the study area.

including a considerable area which lies above treeline.

For the most part the relief is of Rocky Mountain type (Barsch and Caine, 1984). Interfluves are rounded, summits are flat, the local relief is low and numerous accordant levels appear to represent ancient erosion surfaces. The glacial legacy of the area is most striking in a few localised areas of Alp type relief where steep cirque headwalls and glacial arêtes are prominent. The glacial legacy is also apparent on the lake bestrewn western part of Tasmania's Central Plateau which bears comparison with the Barren Grounds of Canada, and parts of Finland, the Tibetan Plateau and West Greenland (Jennings and Ahmad, 1957; Banks, 1973a; Sugden and John, 1976).

The study area sits astride Tasmania's principal drainage divide. Three major river systems - the Mersey to the north; the Pieman to the west; and the Derwent to the southeast - all have their sources in the Du Cane Range (figure B1). Tasmania's largest river, the Gordon, together with its principal tributary, the Franklin, also rises in the study area. This major divide corresponds with an abrupt change in the whole character of the environment. This transition has profound implications for human activity; for the geomorphic environment; and for the processes of weathering and erosion in particular. There is, according to Tyler (1981):

" a line passing through the Derwent Bridge pub, a natural geologic, climatic, edaphic and vegetational watershed...the frontier separating the grazing lands to the east from the dark forests of the west...Here possums become darker and butterflies change colour to become different subspecies...The water chemistry (is) a distinctive weak seawater..brewed to the colour of tea in the peats of the buttongrass..its rivers (run) brown and acid, its pools (lie) black on the plains...The waters retain the ionic character of their westerly ocean, acidified by plant remains till bicarbonate goes, so starved of calcium that local crayfish reserve it for the most essential parts of their skeletons."

GEOLOGY AND STRUCTURE

In large measure this striking change in the landscape reflects a morphostructural boundary between Tasmania's fold structure province to the west and fault structure province to the east (Figure 1.2) (Davies, 1967). Precambrian quartz schists and quartzites occur west of the Derwent-Franklin divide where they are overlain by dolomite (Figure 1.3). These rocks were deformed during uplift of the Tyennan Geanticline from which subsequent rocks were eroded. Although they have not been recognised in the study area, the occurrence further to the northwest and to the south of diamictites that may be glacial in origin hints that this part of Tasmania may have been glaciated as long ago as the Precambrian. These diamictites are associated with dolomite and lavas on King Island (Carey, 1946; Scott, 1951; Jago, 1981) and along the Arthur and Julius rivers in northwestern Tasmania (Griffin and Preiss, 1976). They also occur in the Wedge River Valley (Corbett and Banks, 1974) and Weld River Valley (C.Calver, pers. comm.) in southwestern Tasmania.

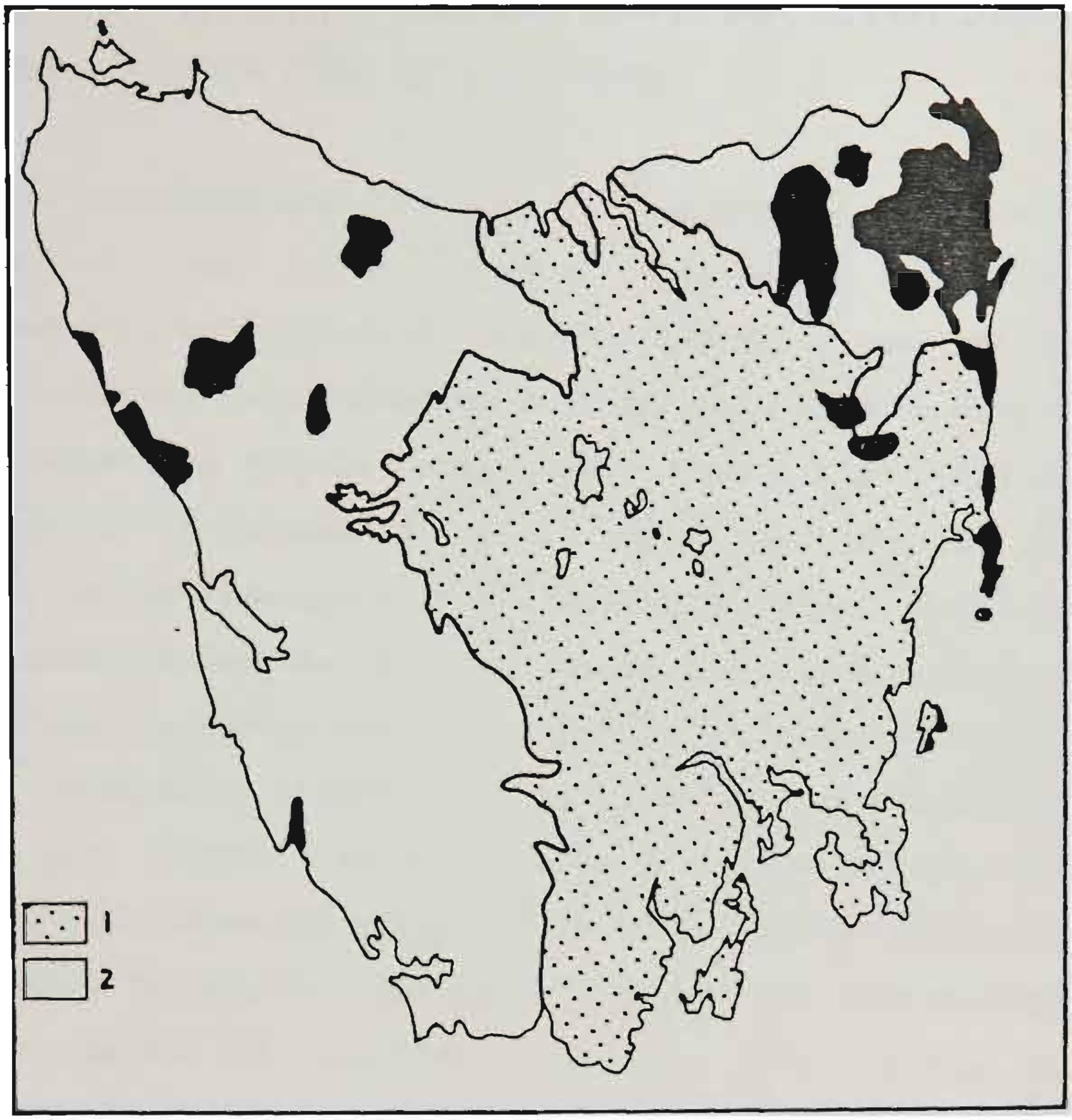


Figure 1.2 Tasmanian morphostructural provinces (after Davies, 1965):

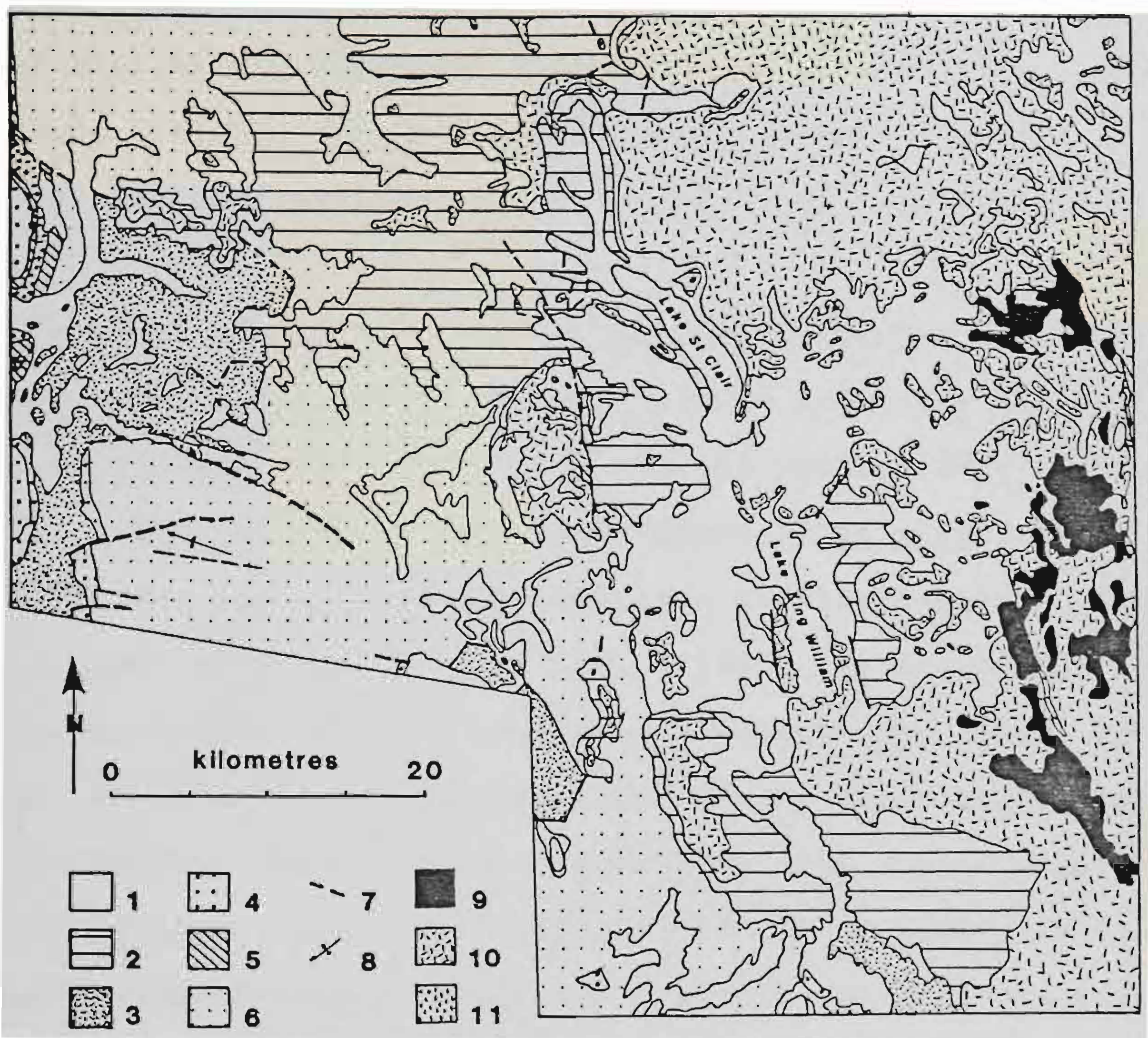
(1) fault structure province; (2) fold structure province with areas of granite shaded.

These rocks may be equivalent to the late Precambrian (Adelaidean) glacial deposits that are widespread on mainland Australia (Dunn et al., 1971).

Rocks of the Ordovician Junee Group crop out on the western margin of the study area. Siliceous terrestrial conglomerate, marine quartz sandstone and siltstone of the Owen Formation were deposited into actively eroding troughs that flanked the Tyennan Geanticline. Tasmania probably lay in tropical latitudes at this time. Sandstone and warm shallow marine limestones of the Gordon Limestone Formation were later deposited as the sea transgressed across the landsurface, possibly due to the decline of glaciers and ice sheets elsewhere (Fairbridge, 1979) or to a slowly sinking shelf area (Banks, 1962a). Siluro-Devonian quartzites. slates, siltstones and shales of the Eldon Group conformably overlie the Ordovician limestone. These rocks were deformed into a series of parallel flattened folds during the Tabberabberan Orogeny which occurred in middle Devonian times.

Essentially flat-lying rocks of Permian and younger age predominate within the fault province. The most widespread rocks are Permo-Carboniferous glacio-marine and marine sediments. These occur in the headwaters of virtually all the valleys tributary to the upper Derwent. During the late Carboniferous Tasmania probably lay within 10^{o} of the South Pole. Continental ice sheets spread out northeastwards into Tasmania from highlands bordering the region to the southwest, now oceanic terrain but probably then occupied by

Figure 1.3 Geology of the study area (simplified from Corbett and Brown, 1975):
(1) Quaternary sediments; (2) Permian glacio-marine sequences and Triassic fluvio-lacustrine sequences; (3) Siluro-Devonian quartz sandstones, siltstones and shales; (4) Ordovician limestone, siliceous terrestrial conglomerate, marine quartz sandstone and siltstone; (5) Cambrian shallow water deposits and orthoquartzite sequences; (6) Precambrian dolomite, metaquartzite sequences, pellitic sequences and garnetiferous rocks; (7) fault; (8) hinge line with plunge of major fold; (9) Tertiary basalt; (10) Jurassic dolerite; (11) Cambrian acid igneous rocks with intermediate volcanic and associated rocks.


Lake St Clair
Lake King William
N
0 kilometres 20
1
2
3
4
5
6
7
8
9
10
11

Antarctica. The Carboniferous-Permian boundary occurs in the upper part of the Wynyard Tillite, which belongs to the lower part of the Parmeener Supergroup (Banks, 1962b). The glaciers reached their maximum extent in the early Permian but ice rafting continued to occur until the late Permian.

The Permo-Carboniferous strata are both marine and non-marine, only gently deformed, and overlie an eroded basement with a relief of 600-1000 m. (Clarke and Banks, 1975; Banks, 1978). The tillites are the product of in situ decay of debris laden ice and contain some thin mudflow lenses. The grain-size distribution of the tillite is comparable to sediments presently being deposited beneath the Ross Ice Shelf in Antarctica. The remainder of the deposits consist of mudstones, diamictites, sandstones and conglomerates (Banks et al., 1955; Frakes and Crowell, 1975; Banks, 1978). The glaciomarine rocks are overlain by fluvial and paludal sediments, and by freshwater sandstones, silstones and mudstones of Triassic age that were deposited in a low energy fluviatile environment.

The Central Plateau and a number of mountains close to its western margin is capped by a sheet of Jurassic dolerite that penetrated the Parmeener Supergroup. This dolerite represents a response to the initial fracturing of the Gondwana supercontinent, and the geology in the study area is very similar to that in the dry valleys of Antarctica and parts of southern Africa. The intrusion of the dolerite was probably associated with some raising of the landsurface. Subsequent erosion has removed most of the cover rocks.

Uplift and southeastward tilting of the plateau commenced about 65 Ma BP as Antarctica separated from the Australian continent (Griffiths, 1971; Banks, 1973b; Sutherland, 1973). Basaltic eruptions occurred into valleys during late Oligocene and early Miocene times (Sutherland, 1973). The basalts are confined to the easternmost part of the study area where they locally overlie minor alluvial and lacustrine sediments of Tertiary age (Prider, 1948).

TOPOGRAPHIC FRAMEWORK

The alpine landscape of the fault structure province is one of erosional uplift mountains (Fairbridge, 1968a). Tensional block faulting and widespread sheets of resistant dolerite have given rise to a landscape that is characterised by tabular dolerite interfluves that separate valleys of gentle gradient which have been cut in Permian and Triassic rocks. The drainage pattern is predominantly rectangular. The fold structure province is dominated by fold mountain belts (Fairbridge, 1968a). Here a series of shallowly arcuate ridges trend broadly north to south. The mountains consist of resistant Precambrain quartzites and Ordovician conglomerates that rise above valleys incised in limestones, phyllites and schists. The interfluves are sharp and the drainage pattern is a dense trellis (Davies, 1967).

From the summit of Mt. Arrowsmith Calder (1849) noted that "..the greater portion of the surface lying between (the mountain crests) is so thrown up into hills as to impart to the country generally an appearance of wonderful irregularity of surface, which however presents on examination extensive levels; over some of these I traced the track to Macquarie Harbour". Clemes (1925) drew attention to three such surfaces in the Lake St. Clair area, which he named the Lower, Intermediate and Higher plateaux surfaces. Early views attributed these to the faulting of at least one erosion surface (Lewis, 1945a; Carey, 1947; Hills and Carey, 1949). The accordance of some summits east of Lake St Clair was attributed to the presence of flat-lying sills of resistant dolerite (Giblin, 1928; Fairbridge, 1949).

Davies (1959) argued that the surfaces were multicyclic in origin. He observed that they were continuous across dolerite injected at different levels and at different altitudes, and across other rock types in adjacent terrain. From this he argued that the surfaces had been preserved on, rather than caused by, the dolerite. Davies recognised at least six (and possibly seven) levels (Figure 1.4). He interpreted five of these as erosion surfaces which had developed at or near base level, with the uppermost level being monadnocks. He named the lowermost the Lower Coastal Surface and attributed it to marine processes. To the accordances above this he gave the names Upper Coastal Surface (365 - 460 m); the St. Clair Surface (730 - 825 m); the Lower Plateau Surface (915 - 1065 m); the Higher Plateau Surface (1200 - 1350 m); and the Higher Monadnocks. Davies remained hesitant about a possible further surface at 550 - 610 m. The origin, age and significance of these accordances remains problematic.

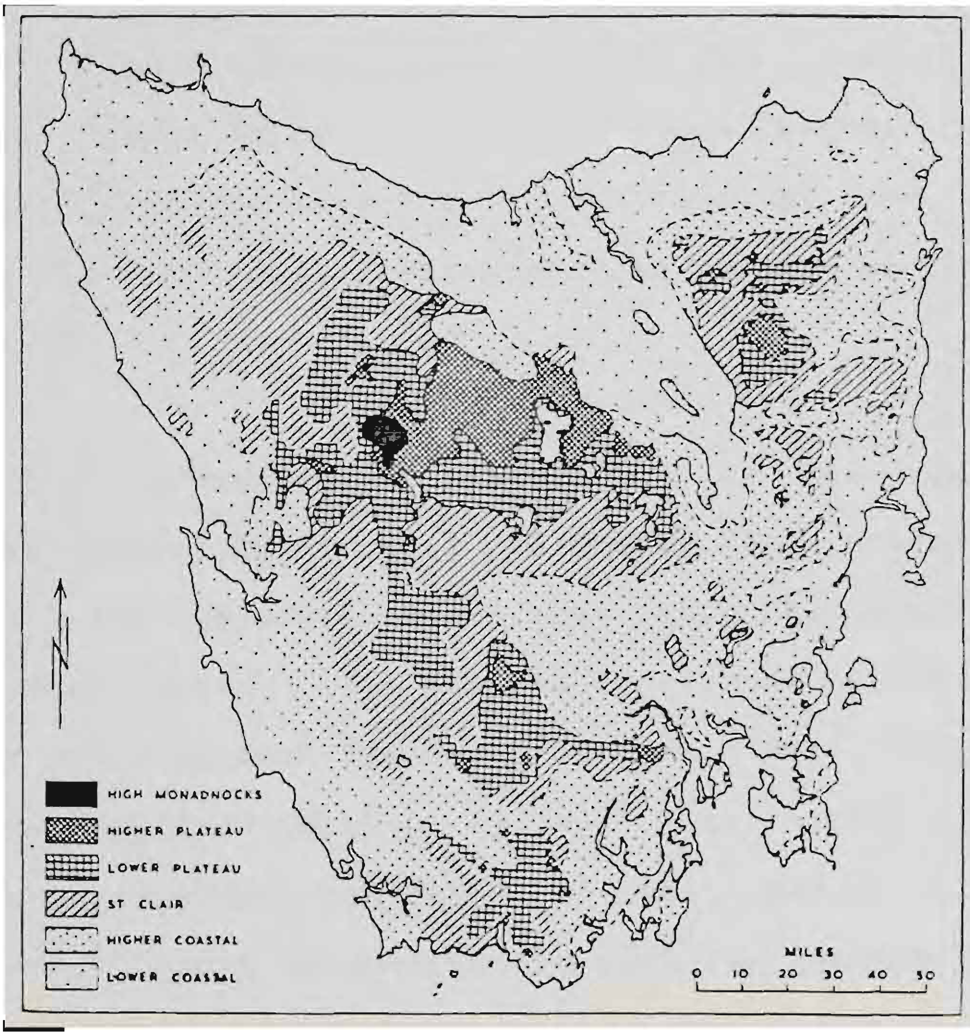


Figure 1.4 Distribution of erosion surfaces (after Davies, 1959).

The Lower Coastal Surface is not present in the study area but all the others are well represented. The Collingwood, Alma and lower Loddon Plains all lie at about 400 m and probably represent part of the Upper Coastal Surface. The St. Clair Surface takes its name from an extensive plain between Lake St. Clair and Butlers Gorge, but further west is represented by accordant summits. The Nive Plains, the floor of the Cuvier Valley, the floor of the Franklin Valley between Lake Dixon and Lake Undine and some of the hills north-west of the Alma River appear to be part of the St. Clair Surface.

The Lower Plateau Surface is discernible on the southern part of the Central Plateau, in the Cheyne Range around Lake Hermione, at the Labyrinth, Mt. Arrowsmith, Mt. Mary, and several other sites. The stepped appearance between this level and the Higher Plateau Surface may represent intermittent uplift (Banks, 1973a). The Higher Plateau Surface is represented by parts of the Central Plateau itself. The Higher Monadnocks are well represented in the Du Cane Range and elsewhere.

A pre-Carbon iferous erosion surface has been exhumed in some localities. It may only have been extensively stripped in areas which are accordant with more recent erosion surfaces (Davies, 1959). This pre-Carboniferous surface descends from 912 m on Last Hill to 851 m on Pyramid Hill and 730 m in the Franklin Valley upstream of Lake Dixon (Banks, 1962b). Lithological benches that are formed on flat-lying Permian and Triassic rocks also induce broadly accordant levels.

CLIMATE AND VEGETATION.

The fold province is the most consistently wet area in Australia (Figure 1.5). More intense rainfall events are experienced only in parts of tropical Queensland. It also experiences fewer sunshine hours per day (average four hours) than anywhere else in Australia. It is one of only three areas in Australia where mean annual runoff exceeds 1250 mm., the national average being only 45 mm while less than 5% of Australia experiences in excess of 250 mm (Watson, 1976, 1978).

Orography has a marked impact upon the distribution of rainfall in western Tasmania. The heaviest rainfall, around 3400 mm pa., is experienced some 20 km inland on the most westerly of the ranges. There is a strong precipitation gradient east of this point. Within the study area the rainfall probably ranges from about 2600 mm near the Collingwood River to about 1100 mm on the eastern side of the Wentworth Hills. Strong precipitation gradients also exist down the lengths of the valleys, and are particularly evident in the Derwent and Nive valleys. Detailed climatic data are available for only two sites within the study area, namely Cynthia Bay at the southern end of Lake St. Clair some 75 km from the west coast, and Butlers Gorge 18 km SSE of Cynthia Bay (Table 1.1). Both sites lie adjacent to the Derwent River. Snow and frost may occur at any time of the year above 1000 m , snow generally being associated with

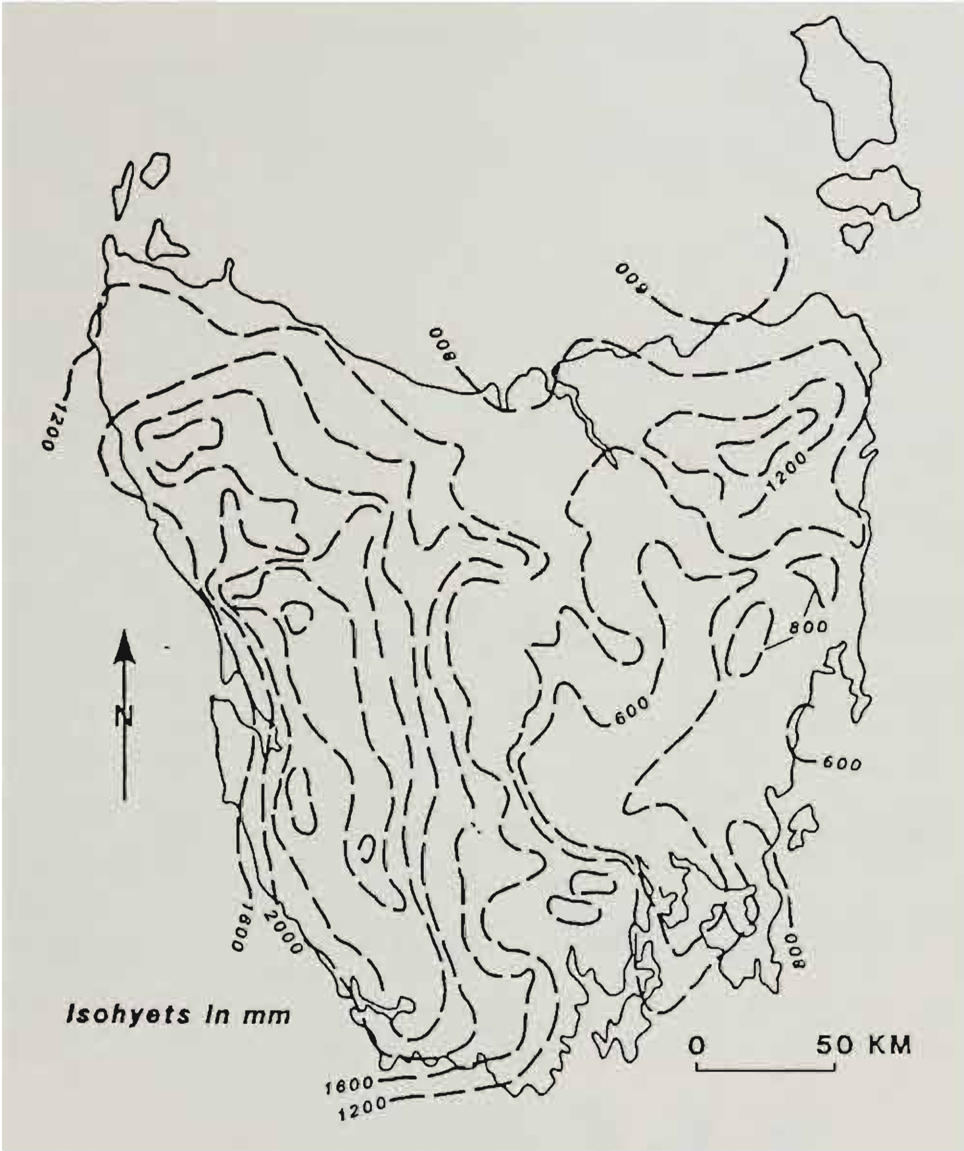


Figure 1.5 Annual precipitation in Tasmania. Based on data supplied by the Tasmanian Bureau of Meteorology and compiled by J.B. Kirkpatrick.

Station	Period	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sept	Oct	Nov	Dec	Annual Total or Mean
a) Average Monthly Rainfall (mm.)														
St.Clair	1937-75	77	81	85	126	141	148	170	166	158	140	127	116	1533
Butlers Gorge	1941-75	92	85	101	149	170	160	182	176	164	154	145	122	1708
Tarraleah	1935-75	70	70	73	103	115	112	124	129	118	107	103	95	1219
Cradle Valley	1917-75	147	135	154	228	280	275	320	307	275	251	219	183	2774
Frenchmans Track	1968-1973	-	-	-	-	-	-	-	-	-	-	-	-	2093 (est.)
b) Number of Rain Days														
St.Clair	1937-75	13	13	15	18	21	19	23	22	20	21	19	18	221
Butlers Gorge	1941-75	14	16	17	21	22	20	23	23	22	21	19	19	233
Cradle Valley	1917-75	16	14	18	20	21	21	24	23	22	21	19	18	237
c) Mean Frequency of Snow (days)														
St.Clair	1957-73	0.2	0.1	0.4	0.8	2.5	3.3	4.4	4.5	4.2	3.6	1.7	0.6	26.3
Butlers Gorge	1957-73	0.1	0.1	0.5	0.9	2.6	3.2	5.1	4.9	4.8	2.9	1.5	0.6	27.2
Cradle Valley	1957-73	-	-	-	-	-	-	-	-	-	-	-	-	52.0
d) Mean Relative Humidity 0900 hrs (%)														
St.Clair	1937-75	64	69	73	76	79	81	83	82	75	67	64	64	73
Butlers Gorge	1941-75	72	78	83	83	91	95	95	93	80	77	73	74	83
Cradle Valley	1917-1975	73	78	89	87	86	95	99	100	86	76	72	72	84
e) Mean Relative Humidity 1500 hrs (%)														
St.Clair	1937-75	45	47	57	59	65	66	65	64	56	57	53	53	57
Butlers Gorge	1941-75	52	56	61	67	74	79	77	72	64	60	60	57	65
Cradle Valley	1917-1975	61	64	76	78	80	95	92	86	75	69	65	64	75
f) Mean Cloud Coverage (eighths)														
St.Clair	1937-73(0900hrs)	4.6	5.1	5.7	6.3	6.7	6.8	6.8	6.6	6.2	5.7	5.7	5.5	5.9
Butlers Gorge	1957-73(0900hrs)	4.6	5.4	6.0	6.3	6.6	6.6	6.7	6.4	6.4	5.9	6.0	5.6	6.0
Butlers Gorge	1957-73(1500hrs)	4.5	4.6	5.1	5.8	6.0	6.0	6.2	5.9	5.9	5.6	5.7	5.3	5.5
g) Frequency of Fog (days)														
St. Clair	1957-73	0.2	0.8	1.1	1.6	2.2	3.0	2.4	1.8	1.5	0.2	0.1	0.1	15.0
Butlers Gorge	1957-73	0.2	1.1	3.1	2.8	3.0	4.0	3.5	2.7	1.9	0.4	0.5	ND	23.0
h) Mean Sunshine (hrs/day)														
St.Clair	1964-74	7.9	7.6	5.7	3.8	2.8	2.4	2.6	3.2	3.6	5.8	6.3	6.3	4.8
i) Daily Maximum Temperature (°C)														
St.Clair	1937-75	18.3	18.5	15.9	12.5	9.1	7.6	6.5	7.6	9.3	11.7	13.8	15.8	12.2
Butlers Gorge	1941-75	19.2	19.1	16.6	13.2	9.6	7.9	6.9	8.0	10.0	12.6	14.5	16.5	12.8
Cradle Valley	1917-75	17.2	17.6	14.7	11.2	7.5	5.1	4.6	5.2	6.6	9.8	12.3	13.7	10.3
j) Daily Minimum Temperature (°C)														
St.Clair	1937-75	7.1	7.6	6.0	4.3	2.5	1.1	0.4	0.5	1.5	3.0	4.5	6.1	3.7
Butlers Gorge	1941-75	6.3	6.6	5.2	3.5	1.7	0.3	-0.2	-0.1	0.8	2.4	3.9	5.4	3.0
Cradle Valley	1917-75	6.3	7.8	5.9	4.3	1.5	0.1	0.1	0	0.3	1.6	3.0	4.0	2.9
k) Daily Median Temperature (°C)														
St.Clair	1937-75	12.7	13.05	10.95	8.4	5.8	4.35	3.45	4.05	5.4	7.35	9.15	10.95	7.95
Butlers Gorge	1941-75	12.75	12.85	10.9	8.35	5.65	4.1	3.35	3.95	5.4	7.5	9.1	10.95	7.9

Station Locations:	St.Clair	666.0m	42°6'S	146°11'E
	Butlers Gorge	735.2m	42°7'S	146°13'E
	Cradle Valley	914.0m	41°38'S	145°57'E
	Tarraleah	550.0m app.	42°18'S	146°26'E
	Frenchmans Track	340.0m	42°13'S	145°59'E

Table 1.1 Climatic data for the study area (after Watson, 1976;1978; ; Bureau of Meteorology, 1980).

cold polar air masses from the south. An increasingly pronounced summer dry period is experienced east of Lake St. Clair.

Vegetation patterns in the area (Figure 1.6) reflect an interplay between climate, geology, soils and fire frequency (Sutton 1929; Jackson 1965, 1973; Davies, 1978; Kirkpatrick 1980, 1984). The montane flora of Tasmania has similarities with that of New Zealand and Patagonia, and is considered to be more Antarctic than Australian in character (Minchin, 1981). Endemism in the flora reaches up to 60% in the west but diminishes to less than 15% in the east, and reflects the climatic and geological significance of the western mountains to the vegetation (Brown, 1981). The predominant vegetation formations within the present study area are temperate closed rainforests, eucalypt forests, sedgelands, heathlands and herbfields.

With low fire frequencies closed rainforests of Nothofagus cunninghamii (Myrtle), Atherosperma moschatum (Sassafras), the conifer Phyllocladus aspleniifolius (Celery Top Pine) and Andopetalum biglandulosum (Horizontal Scrub) develop at lower elevations. Nothofagus cunninghamii, the deciduous N. gunnii (Fagus) and the native conifers Athrotaxis selaginoides (King Billy Pine) and A. cupressoides (Pencil Pine) are present at higher elevations.

The eucalypt forest and woodland is dominated by Eucalyptus delegatensis (White-top Stringybark) on good sites, with E. dalrympleana (Mountain Gum), E. pauciflora (Cabbage Gum) and

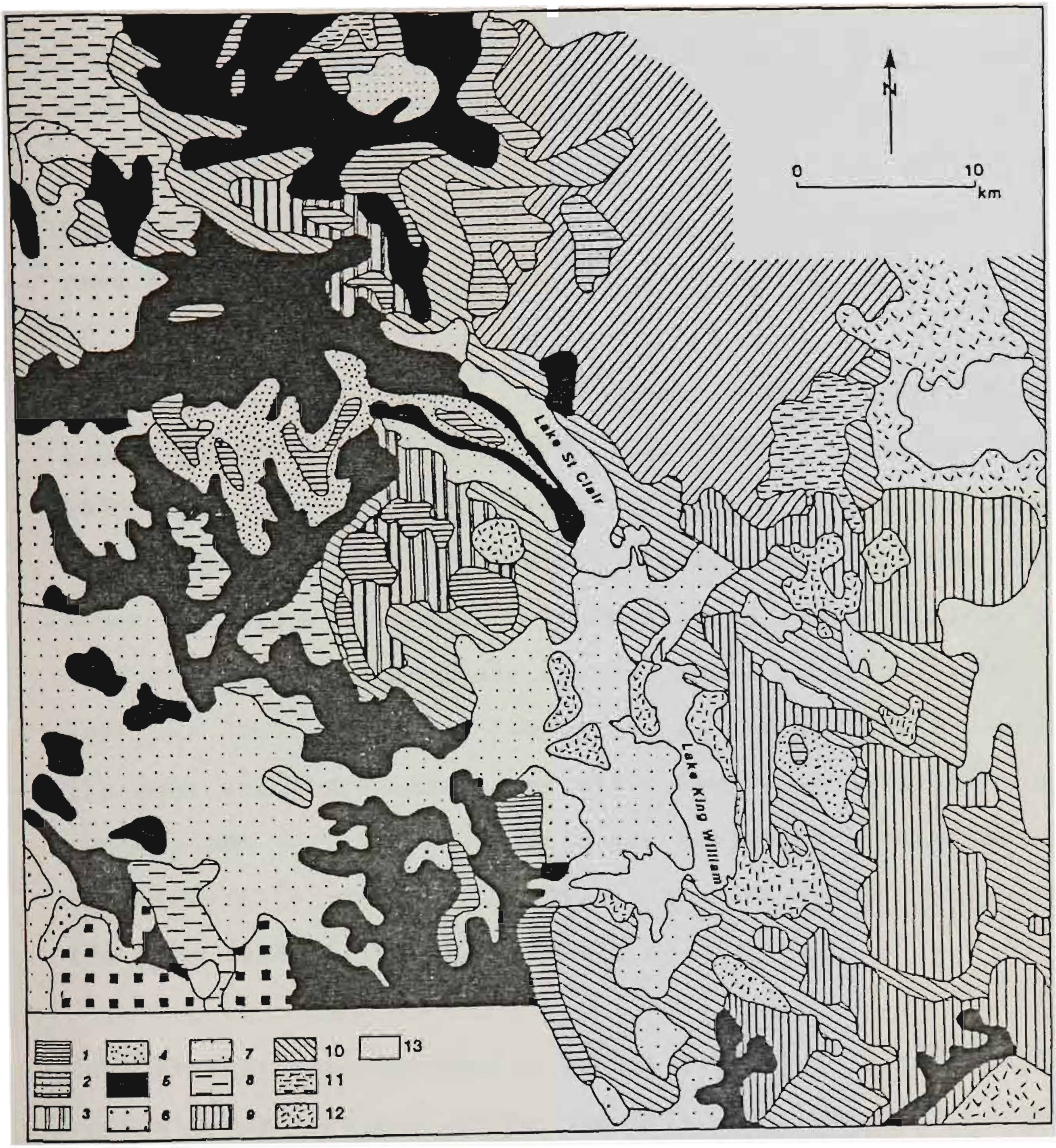


Figure 1.6 Vegetation in the study area (redrawn from Kirkpatrick and Dickinson, 1984): (1) central alpine complex; (2) eastern alpine complex; (3) Labyrinth subalpine complex; (4) Eucalyptus coccifera forest; (5) rainforest; (6) wet scrub; (7) Buttongrass moor; (8) Eucalyptus obliqua forest; (9) Eucalyptus delegatensis forest taller than 40 m.; (10) Eucalyptus delegatensis forest less than 40 m. tall; (11) Buttongrass moor-grasslands mosaic; (12) montane grassy forest; (13) grassland.

E. amygdalina (Black Peppermint) on poorer sites. Eucalyptus nitida (Smithton Peppermint) is common in the west where it is sometimes emergent over rainforest. Above 1000 m E. coccifera (Tasmanian Snow Gum) is widespread. Poa grasslands commonly intrude where trees are excluded by cold air drainage or repeated fires (Jackson, 1965, 1973).

The sedgelands are dominated by Gymnoschoenus sphaerocephalus (Button-grass) on poorly drained and infertile sites. Leptosperum and Bauera scrubs are common near sedgeland margins and along watercourses, and may grade into Eucalypt woodland. The heathland and herbfields are dominated by australmontane species above treeline (1200 m approx.) and sometimes down to as low as 1000 m where drainage and micro-climate permit. Complex mosaics of local communities are common. Fjaeldmark communities occur mainly on mountains capped by fissile Permo-Triassic sediments. (Jackson 1973; Kirkpatrick 1984).

The present distribution of forest represents a response to climatic amelioration during the Holocene. MacPhail (1981) has identified two broad phases in postglacial forest development at Mt. Field, 40km to the south: an Early Temperate Phase of rainforest expansion that occurred from 10 - 6 ka BP when conditions were interpreted as being warmer and wetter than present; and a Late Temperate Phase that saw the forests become more open in structure and dominated by sclerophyll species, in response to climates which were interpreted as being more variable and prone to drought and fire.

The area presently supports a varied marsupial fauna including four species of macropods, possums, dasyurid carnivores and Tasmanian devils.

HUMAN HISTORY

Although the area remains largely in a natural condition, fires lit by human beings have probably long had an impact upon vegetation patterns and slope stability. The Tasmanian aborigines knew Lake St.Clair as Leeawuleena, the Sleeping Water, and the people of the Big River Tribe who frequented the area may have travelled westwards into the upper Franklin (Ling Roth, 1899; Kiernan, 1982a). A preliminary archaeological survey further east on the Central Plateau has revealed a site density of 4.9 sites/km^2 . These sites are probably of Holocene age (Cosgrove, 1984). No detailed archaeological survey has been conducted in the study area but a number of sites have been recorded during fieldwork for this thesis.

The time at which the first campfires were kindled beside the Sleeping Water is unknown, but aborigines were present by 20 ka BP further down the Derwent Valley (Murray et al., 1980) and even in the rugged Franklin Valley to the south-west (Kiernan et al., 1983). The antiquity of this occupation is emphasised by the abandonment of the Franklin Caves about 15 ka BP, still 5 ka prior to any reliable evidence for humankind at equivalent southern latitudes in the Americas and a similar time prior to the painting of Altimira and other celebrated European cave sites

(Ortiz-Troncoso, 1981; Jones, 1982).

Despite the fact that their tool kit was the simplest ever recorded in the world's ethnographic literature (Jones, 1968) it is to be anticipated that the aborigines have left a legacy in the landscape of the study area because many mountain environments exist in a metastable state and hence are particularly vulnerable to disturbance (Barsch and Caine 1984). Some circumstantial evidence exists to link aboriginal firing with slope instability and sedimentation in various parts of Tasmania (Davies, 1967, 1974; Kiernan *et al.*, 1983).

Europeans first settled near the mouth of the Derwent River in 1804. From the summit of Mt. Charles surveyor William Sharland became the first European known to see Leeawuleena in 1823. Mistaking the identity of the river which drained from it he renamed Leeawuleena as Gordon Lake. Eight years later a party led by Surveyor Frankland recognised the error and renamed it Lake St. Clair after the St.Clair family of the Scottish Loch Lomond. From the summit of Mt. Olympus Frankland sprinkled the area with classical Greek nomenclature before descending into the Cuvier Valley to camp beneath the "Pines of Olympus" (later renamed Pencil Pines) beside the sandy beach of Lake Petrarch (Frankland 1835).

In 1837 Surveyor Calder found the true source of the Gordon to be Lake Richmond, 22 km. south of Lake St. Clair. Three years later he ventured westward of Lake St. Clair, and from the summit of Fatigue Hill (now Mt. Arrowsmith) found that "...a wilder prospect could scarcely be imagined, for in almost all directions the landscape ends in groups of broken mountains." Governor of the colony and noted arctic explorer Sir John Franklin later accompanied Calder across the ranges to Macquarie Harbour at the mouth of the Gordon River on the west coast.

These may not have been the first Europeans to have seen the western mountains. There were numerous escapes from the harsh conditions at a convict station that was established on an island in Macquarie Harbour in 1822, where 7000 lashes were inflicted upon 169 prisoners in one year alone. While at least sixty two of the escapees perished in the bush, at least nine of them eaten by their comrades, a few made it back to Hobart Town to be given the option of the gallows or service with the surveyors. In 1835 Surveyor McKay found convict clothing at some aboriginal huts near the confluence of the Surprise and Franklin rivers. This suggests that at least one escapee may have passed through the Lake St. Clair area (Binks, 1980).

PRESENT LAND-USE

For many years the area west of Lake St. Clair was regarded with awe and even fear by the European settlers. The name Transylvania was given to it on early maps. Much of the western landscape remains largely unchanged from that encountered by the early surveyors. As early as 1885 the Executive Council withdrew from sale all Crown Land within half a mile of Lake St. Clair and Lake Petrarch, in recognition of the potential of the area for public recreation. Work commenced on an hydro-electric dam across the Derwent River at Butlers Gorge shortly after the second world war. The resulting impoundment is known as Lake King William. The level of Lake St. Clair was later raised by 3 m for the same purpose. That lake is now part of the Cradle Mountain - Lake St. Clair National Park (137,274 ha). The scenery produced by past geomorphic processes has proven a greater economic asset than the putative gold that initially drew people to this area. Today upwards of 100,000 people visit the southern shoreline of Lake St. Clair each year.

A road from Lake St. Clair to the mining settlement of Queenstown near the west coast was completed in 1932. It remains the only formed route across the western ranges. There are no permanent settlements between Lake St.Clair and the Queenstown area, only overgrown tracks and traces left by prospectors and timber cutters. The Franklin River catchment now forms part of the Franklin - Lower Gordon Wild Rivers National Park that adjoins the Cradle Mountain - Lake St.Clair National Park. Proposals by the state government to construct hydro-electric dams on the Franklin River within this park were blocked in the High Court of Australia by the national government in 1983. Both parks now form part of the western Tasmanian Wilderness National Parks complex of 7,693 km^2 that is protected under the United Nations World Heritage Convention. The remainder of the study area is mainly State Forest or uncommitted Crown Land, with some small freeholdings and leaseholdings in the east

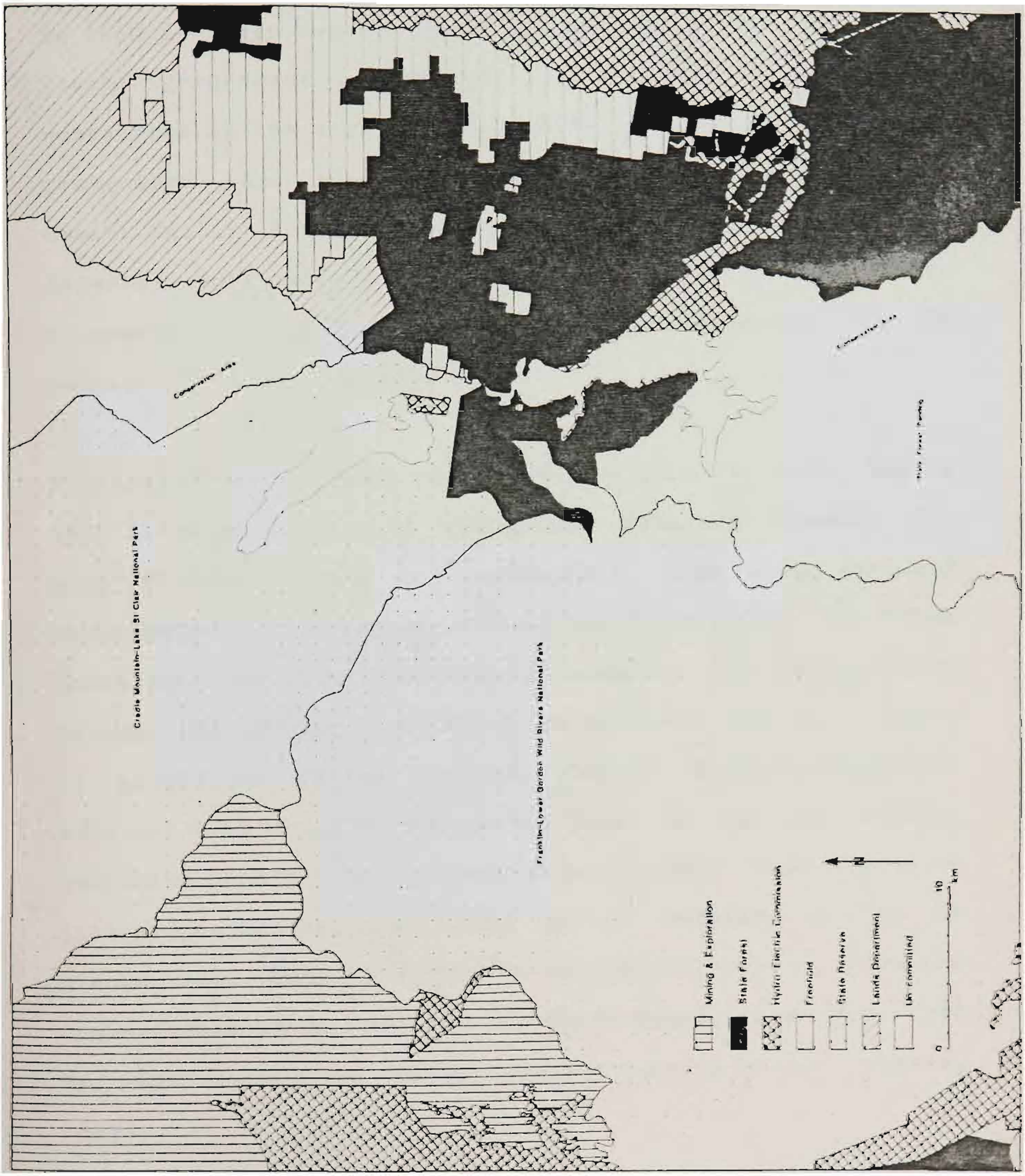


Figure 1.7 Present land tenure in the study area.

(figure 1.7).

RECOGNITION OF THE GLACIAL LEGACY

In 1860 Surveyor Charles Gould was appointed as Tasmania's first government geologist. Two years later he was despatched to the study area in search of gold and first recognised that glacial features were present: "..In the Cuvier Valley I was struck...by the similarity to the terminal moraine of a glacier presented by an enormous accumulation of boulders which chokes the lower end of the valley .." (Gould, 1860).

Exploration of the wild country to the west was slow, but in 1883 Surveyor T.B. Moore recognised erratics between Mt. King William I and Mt. Arrowsmith. Continuing westward Moore noted: "..Painters Plains are described by Count Strzelecki as an accumulation of pebbles. The accumulation is composed of every variety of rocks, with large boulders of greenstone strewn over the plains. These boulders are also met with cropping out on the tops of the surrounding quartzite hills. It is quite possible that these masses of greenstone, occurring as they do in solitary blocks or groups have been brought, in the Glacial Period, from the higher lands of Mt. Gell or the Eldon Range, and deposited by that agency in their present resting places" (Moore, 1883).

Another surveyor, C.P. Sprent, had also recorded the presence of glacial features since commencing work in the area in 1875, and was soon able to report that "..traces of glacial action are common all over the west coast in localities close to the high mountains (Sprent, 1887). According to the government geologist of the day "..no-one had a more intimate knowledge of this wild country, for he (Sprent) was the pioneer who first opened out the greater part of this country to miners by a process of track-cutting, almost like tunnelling in the horizontal, bauera and other scrubs - the terrible barriers to progress in this region " (Johnston, 1894a). Those same barriers have continued to impede geological mapping and the investigation of the glacial record in the Gordon - Franklin basin.

In his comprehensive volume The Geology of Tasmania Johnston (1888) amplified the case for previous glaciation in the area. In a later paper (Johnston, 1894a) he argued that the lakes of the upper Franklin, together with Lake Petrarch and other lakes east of the King William Range, were of glacial origin. However, he followed Gould (1860) in attributing Lake St. Clair to the Derwent having been dammed by a lava flow. Although Tasmania had by now been cited in the pages of Nature as a locality in which rock basin lakes of glacial origin were present (Wallace, 1893) Johnston also contended that the "..innumerable lakelets and lagoons on the upper levels (of the Central Plateau) are due to original surface irregularities produced by the flow of greenstones on erupting" (Johnston, 1894b).

Officer (1894) almost totally rejected the idea of any glaciation in the Lake St. Clair area. He discarded both the lava dam and earlier crater lake explanations for Lake St. Clair, attributed both Lake St. Clair and Lake Petrarch to valley floor subsidence, and contended that the dolerite rubbles on the floor of the Cuvier Valley were lag deposits. In a subsequent paper Officer, Balfour and Hogg (1895) rejected the erosional evidence for glaciation east of Mt. King William I but conceded that "..the possibility of (Lake St. Clair) being a glacial lake is worth considering." Three decades were to pass before Clemes (1925) reinterpreted the ridges of dolerite rubble south of the Lake St.Clair as moraines and firmly established the glacial origin of that lake and many other features in the study area.

Reconaissance observations by a number of workers (eg. Lewis, 1945b; Derbyshire, 1963, 1967; Peterson, 1968, 1969; Davies, 1969) enable a number of issues to be identified. These issues are reviewed in chapter two. Some are explored in detail later in this thesis. Others require a detailed study in themselves and this thesis cannot realistically seek to do more than contribute to the data base available for future workers.

chapter two

SOME ISSUES IN THE MOUNTAIN GEOMORPHOLOGY OF THE TASMANIAN CENTRAL HIGHLANDS

Issues raised by previous studies of this mountain landscape include the areal extent of past glaciation; the patterns of ice movement; the geomorphic consequences of glaciation; the age of the glacial events; the palaeoglaciological and palaeoclimatic environments; and the effect of extraglacial and nonglacial processes.

1 The areal extent of glaciation

Difficult terrain and a consequent reliance upon broad morphological criteria which were open to alternative explanations has impeded attempts to define the areal extent of the maximum ice cover in these mountains.

On the basis of the erosional morphology represented by cirques, glaciated troughs and roches moutonnées, and also on the basis of moraine distribution, Lewis (1934) suggested that between 33% and 50% of Tasmania had once been covered by ice (Figure 2.1). Lewis envisaged that three glaciations had occurred, and that during the earliest of these glaciers from the Lake St. Clair - upper Franklin area had extended down the full length of the Franklin Valley (Lewis, 1939, 1945b). Prider (1947) proposed that the glacier in the Derwent Valley had reached Butlers Gorge. Fairbridge (1949) argued that ice extended down to 600 m. in the Waddamana area. This was 300 m. lower than Lewis had suggested.

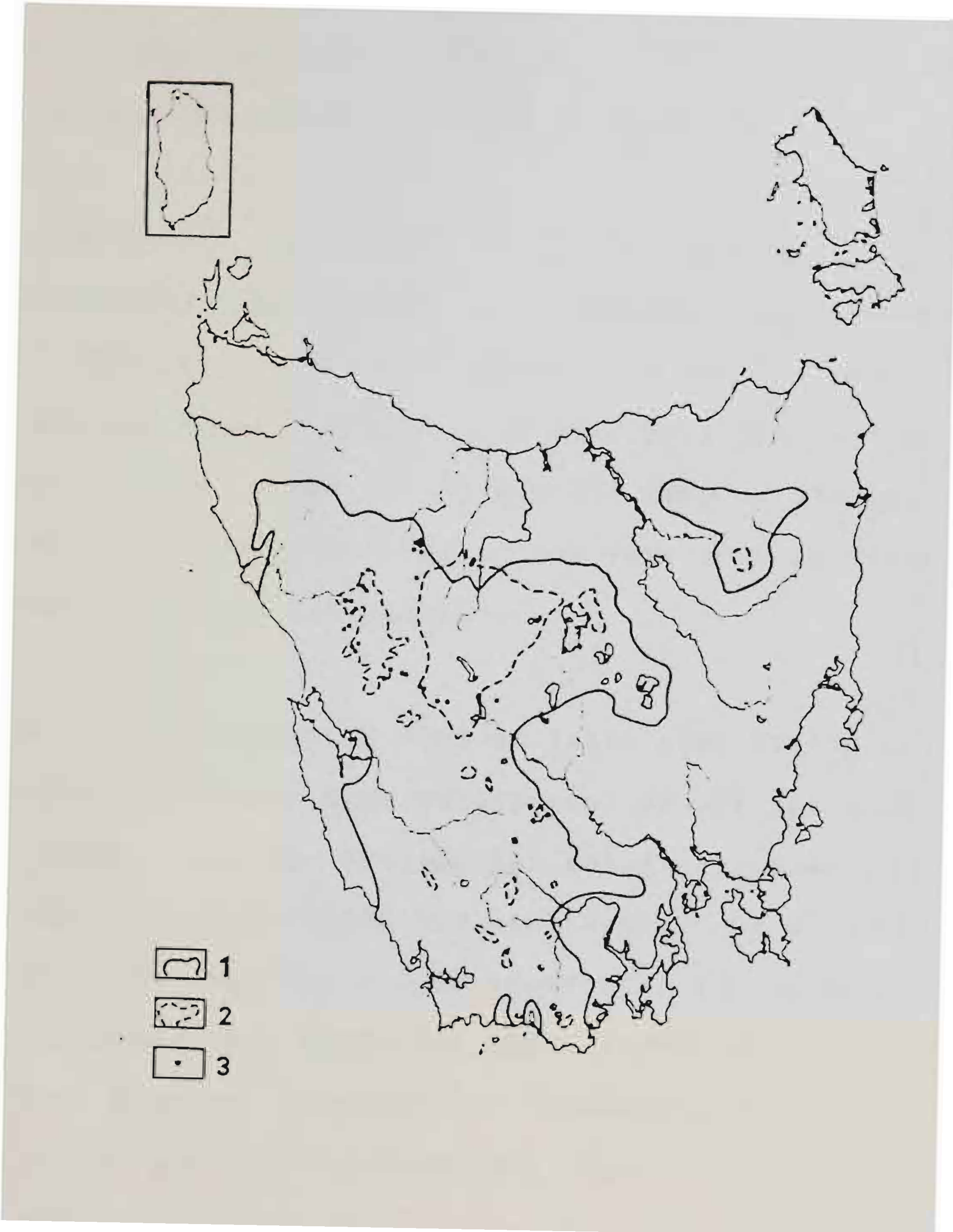


Figure 2.1 Limits of Pleistocene ice in Tasmania: (1) according to Lewis (1945b); (2) according to Banks (1962); (3) erratics beyond the limit of continuous drift. (after Derbyshire, 1966).

Fairbridge based this suggestion on erosional landforms and attributed the absence of drift deposits that would have supported his interpretation to their deposition in overdeepened bedrock depressions 20 m deep.

The origin of the lakes on the plateau east of Lake St. Clair remained contentious for decades (eg. Montgomery, 1894; Officer, 1894; Johnston, 1894b; Clemes, 1925). Jennings and Ahmad (1957) argued that only the smaller lakes on the western part of the plateau were of glacial origin and that other lakes further to the east such as Great Lake and Lake Echo were not glacial.

Davies (1958) suggested that at least some of the sediments which had previously been interpreted as glacial drift were more likely to be periglacial solifluction deposits, and that the ice limits which had previously been proposed were suspect. The case for a more restricted ice cover was given added credence when Banks and Ahmad (1959) argued that the putative glacial deposits at Malanna in western Tasmania were not of glacial origin, but were fault scarp scree breccias and fluvial, lacustrine and palludal sediments that had been deposited in the lowland produced by faulting. Spry and Zimmerman (1959) proposed that the Franklin Glacier had not extended beyond the confluence of the Franklin and Collingwood rivers. These developments led to a dramatic revision of the proposed maximum ice limits to depict a small ice sheet on the Central Plateau, a smaller ice cap in the West Coast Range, and several isolated cirque and short valley glaciers (Jennings and Banks, 1958; Davies, 1962)(figure 2.1).

However, Davies (1962) cautioned against total rejection of the possibility that glaciation had been much more extensive. Subsequent research compounded this ambivalence (eg. Blissett, 1962; Threader, 1962). Unequivocal glacial deposits and erosional landforms were subsequently documented in the Derwent Valley to slightly beyond the ice limit which had previously been accepted (Macleod _et al._, 1961; Derbyshire, 1963, 1965, 1967, 1968a; Gulline, 1965; Peterson, 1969). A growing body of evidence and assertion was synthesised by Fish and Yaxley (1966, 1972). The emerging picture was presented on _The Glacial Map of Tasmania_ (Derbyshire _et al._, 1965). This indicated that some ice margins were uncertain as isolated erratics had been recorded beyond the limits of continuous drift in areas such as Redan Hill in the Collingwood Valley.

Subsequent studies confirmed the past glaciation of the upper Franklin Valley and King William Range (Derbyshire, 1968; Peterson, 1969); and of the Collingwood Valley and adjacent areas to the west (MacIntyre, 1964; Read, 1964).

Later investigation of cirque morphology and the discovery of anomalous conglomerate clasts above cirque headwalls in the Raglan Range led Peterson (1969) to conclude that the Central Tasmanian ice sheet had extended as far west as the Raglan Range, while Derbshire (1972) theorised that the Central Tasmanian and West Coast Range ice caps may have been continuous.

It is now approaching two decades since the question of ice limits in the southwestern part of the Central Highlands has been reviewed. The recent suggestions that massive boulder deposits 70 km downstream in the Derwent Valley are probably glacial (Kiernan, 1983a) and that extensive outwash gravels existed in the lower Franklin Valley (Kiernan et al., 1983) highlight the fact that many questions regarding the extent of the maximum ice cover remain unresolved simply because much of the terrain still has not been traversed in a detailed search. Moreover, little is known of the ice limits which prevailed during different phases of glaciation.

2 The patterns of ice movement

Jennings and Ahmad (1957) demonstrated that a major glacial divide between north and south flowing ice on the Central Plateau approximately corresponded with the position of the present fluvial divide. They reported that much of the the south-flowing ice spilled into Lake St. Clair while the remainder flowed into the Nive catchment (Figure 2.2). Lake St. Clair clearly acted as a major channel for ice from the Central Plateau and Du Cane Range. Some of the Du Cane ice has been claimed to have flowed westwards into the Murchison Valley (Derbyshire, 1963)(Figure 1.2). Diffluent ice from the Derwent Valley also passed into the head of the Cuvier Valley from which further diffluence took place into the Franklin and Alma valleys (Derbyshire, 1963; Kiernan et al., 1983). The probability that the Derwent and Franklin ice was confluent south of Mt. Rufus was recognised by Lewis

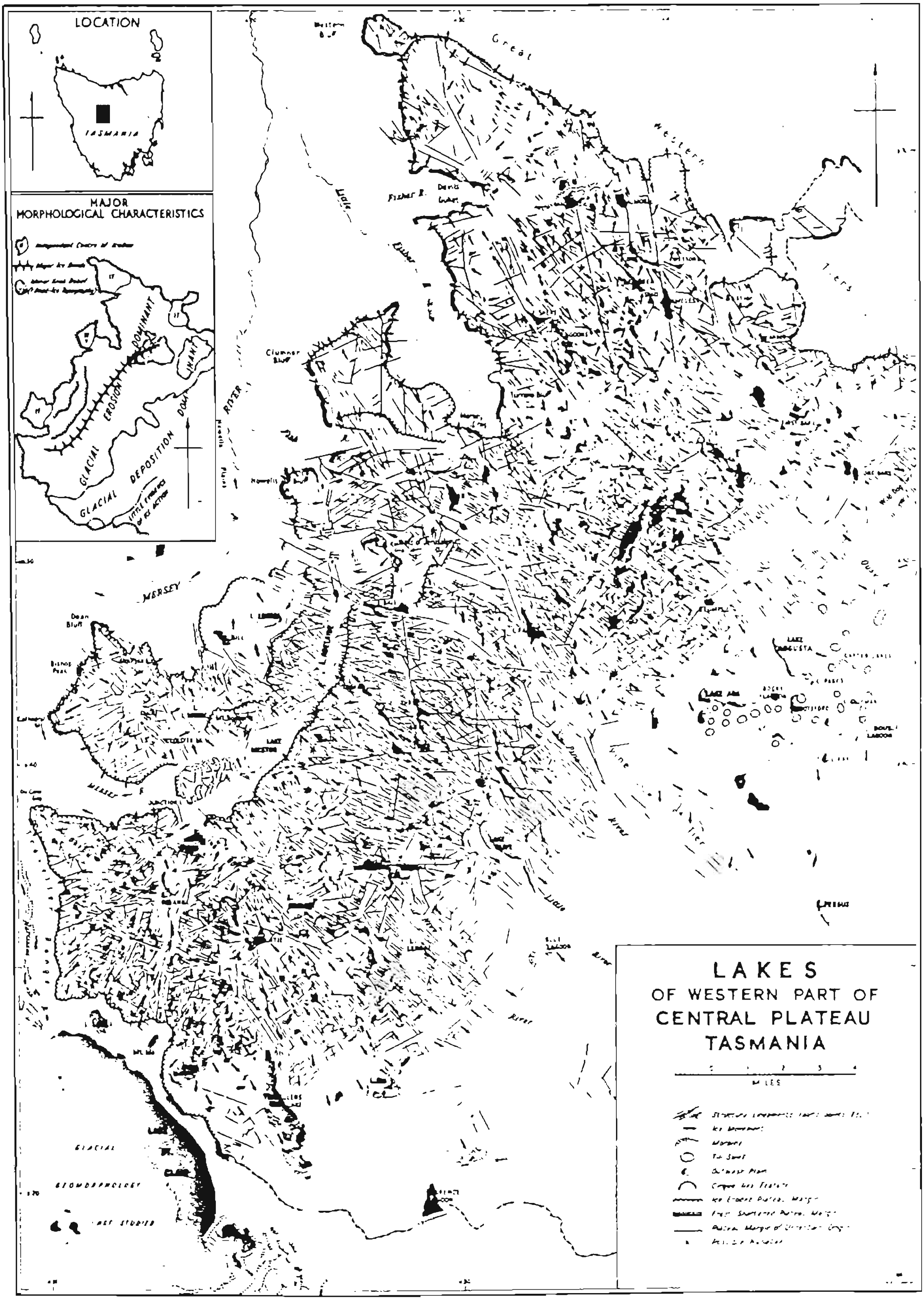


Figure 2.2 Erosional morphology of the Central Plateau (after Jennings and Ahmad, 1957).

(1939). Derbyshire, (1963) contended that this ice discharged westwards via King William Saddle into the Surprise Valley.

Spry and Zimmerman (1959) recognised glacial deposits in the Franklin as far downstream as the Collingwood confluence, and contended that a distributary lobe of the Franklin Glacier extended up the Collingwood Valley. This seems inconsistent with the alleged failure of the glacier to extend further downstream into the Franklin Gorge. Alternative interpretations of the till in the Collingwood basin were provided by Read (1964) who considered that a glacier moved southwards down the tributary Balaclava Valley, and by MacIntyre (1964) who argued that the glacial deposits had been laid down by an ice sheet as it retreated towards the highlands.

Derbyshire (1968a) suggested that the tills in the upper Gordon Valley were depositied by glaciers that developed on the eastern slopes of the King William Range south of Lake Montgomery.

Prider (1948) attributed erratics near Butlers Gorge to a glacier that had reached this point from the northwest. These erratics were not observed by Derbyshire (1968a) following the filling of Lake King William which drowned the deposits in which the erratics occurred. Derbyshire attributed tills as far down the Derwent Valley as the headwaters of Mossy Marsh Ck. to the glacier from Lake St. Clair. Therefore, some uncertainty has existed with respect to ice movements in this area.

Clemes (1925) argued that glaciers also descended from the Central Plateau via the Travellers Rest and Clarence valleys but did not define their maximum extent. It has since been suggested that till in the Clarence River basin south of the plateau was deposited by a distributary lobe of the Derwent Glacier (Derbyshire et al., 1965; Derbyshire, 1968a). If a glacier reached the Waddamana area (Fairbridge, 1949) it would seem reasonable to assume that it was a southeasterly extension of the Central Plateau ice cap.

3 The geomorphic consequences of glaciation

A further set of issues relate to the impact of glaciation upon the landscape. Debate in this regard has questioned the extent to which landforms that have been regarded as glacial in origin might be attributed to preglacial or extraglacial processes; the controls which conditioned glacial erosion and deposition; and the extent to which subsequent modification of the landscape has been the result of glacial preparation.

(i) Glacial versus nonglacial

Several workers have argued that the broad form of the landscape considerably predates the late Cainozoic glaciations and is largely the result of preglacial fluvial erosion (Johnston, 1894a; Clemes, 1925; Davies, 1959; Derbyshire, 1967). However, Lewis (1934) contended that the earliest of three glaciations to affect the area had so modified the preglacial landscape as to make its reconstruction difficult. Lewis (1939) attributed the carving of the Surprise, Franklin and Collingwood valleys to this early glaciation. Davies (1959) suggested that Lewis may have mistakenly attributed some landforms to glacial erosion when in reality they represented older erosion surfaces and valleys. While this might be true in some cases, it should be recognised that Lewis was himself aware of the presence of erosion surfaces in this landscape (Lewis, 1945a).

The relative contributions of fluvial and glacial erosion to the deepening of some of the valleys was one of the first issues to arise in Tasmanian glacial geomorphology. Montgomery (1894) argued that glacial erosion had been a major factor in the evolution of the landscape but Johnston (1894a) contended that " the more mobile gravitation and the infinitely greater dissolving power of water in motion" had been far more important.

Similarly, Clemes (1925) considered the Lake St. Clair Valley to have resulted from the "combined river action and glacial action along a great fault plane...the wide valleys are due to glaciers working along the previously eroded river beds, and lakes have been formed behind the morainal dams. The lakes on the Traveller Range are probably due to glacial scooping." Lewis (1934) also recognised the great importance of fluvial erosion and attributed gorges up to 600 m. deep to erosion by rivers following the earliest glaciation.

In each of these cases the fluvial erosion was regarded as an interglacial phenomenon. A major change came when Lewis (1939) recognised that some of this erosion may have been accomplished by proglacial meltwater. Subsequently Derbyshire (1971a) suggested that the present gorge of the Derwent River east of Bedlam Walls had been initiated as a subglacial meltwater channel.

(ii) The genesis and morphology of the glacial landforms

Derbyshire (1967) argued that only the finer details of the landscape were due to glaciation but acknowledged that some distinctive erosional forms had been produced where the ice cover was thick. He identified 265 cirques in north-west central Tasmania and differentiated between what he termed "high discrete cirques" which had been formed solely by snowfall and snowdrift to form local glaciers above the general level of the ice in the valleys, and other cirques that had been over-ridden by ice or that lay at valley heads (trough ends).

Jennings and Ahmad (1957) proposed that a gradient of increasing glacial erosion southwestwards could be recognised on the Central Plateau. Davies (1967) and Peterson (1968) showed that this trend continued across the island, cirque floors being appreciably lower in the southwest. This climatic gradient meant that towards the north-east cirque glaciers became increasingly dependant upon the location of snowfences and shading ridges, and hence upon the preglacial topography and bedrock geometry that had given rise to it.

The apparent absence of cirques from some more easterly ranges such as the Wentworth Hills was attributed to the interception of snowfall by ranges further to windward (Derbyshire, 1968a). Derbyshire found that the mean orientation of the cirques was in the southeasterly quadrant (figure 2.3). Ranges that are oriented north-south show evidence of greater cirque development than those that trend northwest-southeast which suggests that the orientations of the snowfences were an important factor.

A contrast has been recognised between the predominantly shallow and open cirques around Lake St. Clair and the deeper cirques and rock basins further west. This has been attributed to differences in glacial climate, and to less vigorous ice movement in the east where snowfall would have been less than in the mountains nearer the coast. It has been argued that the cirque glaciers around Lake St. Clair generally failed to extend far beyond the cirque threshholds (Derbyshire, 1967,1968a; Peterson, 1968,1969).

Peterson (1968,1969) attributed the generally poor development of cirques in the fault structure province to less advanced preglacial dissection which had resulted in less favourable sites for glacier growth; to a shortage of shady sites at the foot of essentially straight fault-line scarps and snowfences; and to lesser ice discharge. He theorised that the susceptibility of dolerite to interglacial chemical weathering might have resulted in the

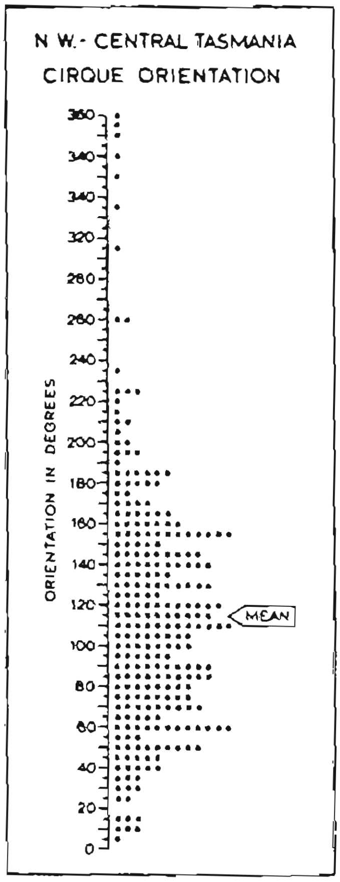


Figure 2.3 Orientation of 265 cirques in north-west central Tasmania (after Derbyshire, 1967).

eastern cirques having been subject to greater interglacial degradation than those in the west with the result that differences in form were progressively emphasised by the ice of successive glaciations.

Peterson (1969) argued that cirques had been over-ridden by ice and as a result were poorly developed in areas that had been subject to ice sheet glaciation. He also considered that in other cases cirque glacier activity had been stifled by large trunk glaciers which impeded discharge from the cirques.

The Central Plateau is an ice abraded plain on which roches moutonnées and rock basin lakes are abundant (Jennings and Ahmad, 1957; Davies, 1969). Some of the lakes on the plateau are believed to be of depositional or composite origin. Lakes become fewer but larger in areas of steeper terrain. It has been suggested that some marshes further to the southeast may be rock basins that have been filled by sediment (Fairbridge, 1949; Jennings and Ahmad, 1957).

Spry and Zimmerman (1959) pointed out that U-shaped troughs were well developed in the upper reaches of the Surprise, Stonehaven, Alma and Franklin valleys but were not present further downstream where the gradient was more gentle. They took the view that this was due to "marginal conditions" near the glacier terminus. However, Derbyshire argued that high longitudinal valley gradients had generally inhibited the development of U-shaped cross profiles. He considered that the overdeepened glacial trough of Lake St. Clair had resulted from a large imput of ice from the Central Plateau that accentuated the difference in form between the deep trunk valley and its shallow tributary valleys (Derbyshire, 1968a, 1971). Read (1964) attributed the lack of erosional landforms in some valleys tributary to the Collingwood simply to the glaciation having been short-lived. Fairbridge (1949) argued that the paucity of glacial erosion near Victoria Valley was due to thin and stagnant ice.

Landforms that have resulted from glacial deposition are widespread but less visually striking than their erosional counterparts. Derbyshire (1963, 1967) considered most of the moraines to be recessional in origin. Lake St. Clair is partly dammed by a moraine barrage which accounts for about 24% of its volume of 1.69 km3. Estimates of the thickness of drift south of the lake between Bedlam Walls and Mt. Rufus have ranged from 220 m (Lewis, 1939) to 120-150 m (Clemes, 1925) and 45 m (Derbyshire, 1971). The last figure is based on a bathymetric survey of the lake. Glacial deposits are less abundant east of Lake St. Clair than to the west and this has been attributed to a more sluggish glacier regimen (Davies, 1967; Derbyshire, 1972).

(iii) Extraglacial and nonglacial deposits

Clemes (1925) and Derbyshire (1967) have argued that glacial erosion has conditioned the subsequent development of the landscape, while Derbyshire (1973) has based part of his climatic interpretation upon the character of extraglacial and nonglacial deposits. An attempt has been made in this thesis to record surficial slope sediments which reflect upon the character of geomorphic processes although the sediments are seldom sufficiently widespread to depict on the accompanying maps (Section B). The classification of White (1981) has been adopted. White recognises three basic forms of non catastrophic alpine mass movement deposits : (1) talus, including rockfall, alluvial and avalanche talus, avalanche boulder tongues and protalus ramparts; (2) rock glaciers; and (3) block slopes, fields and streams.

Talus originates solely or by some combination of rockfall, flowage or avalanche, impact scattering, creep and settlement. This leads to the formation of a debris slope or felsenmeer, which is defined by Church _et al._, (1979) as "any slope upon which clasts have accumulated by mass wasting processes". The nature of the debris reflects lithology, dislodgement processes and inheritance. Clasts may be dislodged by direct climatic factors such as frost or snow wedging, heating or cooling, rain or waterflow, or by geological factors such as unloading, debris wedging, hydration or seismic activity (Rapp, 1960a; White, 1976; Church _et al._, 1979). Movements of the debris result from impacts, mass failure, downslope creep and avalanche (Luckman, 1971; 1978).

Clemes (1925) has claimed that "massive deposits which were undermined [by glacial erosion] gradually lean over at greater and greater angles until they topple over and pile up in a tangled mass of boulders below." This process of slab toppling involves rotation about an axis at the base of the block and has received most attention in northeastern

Tasmania where Caine (1979, 1982, 1983) also concluded that glacial steepening was a critical factor. Some topple masses can be very large. One partly detached mass 1.5 km. long, 100-150 m wide and at least 60 m broad has been recorded on the Great Western Tiers in central northern Tasmania (Kiernan, 1984) while slab topples have moved as far as 2 km. downslope at Ben Lomond (Caine, 1982).

The initial stage in this process involves the formation of 'rock crevasses' by dilation in response to unloading (Harland, 1957). Large dilation trenches which may be tens of metres wide and deep and hundreds of metres long progressively develop. Derbyshire (1973) recognised that several stages in this process were preserved on Mt. Olympus, among them "landslipped rock masses which have preserved their initial form" and "large landslips which have spread themselves down the mountain sides." Derbyshire recognised that this process was at least assisted by frost action. Caine has suggested that cliffs which have developed by slab failure with tension cracking are characterised by a two facet concave profile. Other cliffs which have developed by cambering due to the failure of underlying sediments have a straight, near vertical profile (Caine, 1982). After their collapse the dolerite columns generally disintegrate into masses of large boulders to leave block aprons and block slopes (White, 1981) which consist of prismatic joint-bounded blocks that have an imbricate relationship to one another.

Debris texture commonly reflects its origin or mechanism of transport (Church *et al.*, 1979). Hence, intact imbricate blocks within the present study area are probably the result of slab toppling while slopes formed of poorly sorted angular clasts are probably the result of rockfall. Rockfall mantles commonly rest at an angle of 20-35 degrees and have an abrupt outer rim. Caine (1983) has argued persuasively that the dolerite rockfall deposits at Ben Lomond are probably the result of frost wedging. This is consistent with the observation that large rockfall events in alpine environments elsewhere have generally occurred under conditions of glacial or neoglacial cold (Grove, 1972; Porter and Orombelli, 1980). Alluvial talus is characterised by the inwashing of fine materials of silt and sand size and the creation of crudely bedded cones.

Avalanche talus occurs beneath areas where snow can accumulate by drifting, often at a major re-entrant in a cliffline or scarp, and may be associated with localised disruptions to the distribution of vascular plants due to protracted snowlie. Avalanche deposits may be sorted by momentum such that the largest blocks travel furthest (Gray, 1973; Martinelli, 1974; Perla and Martinelli, 1976; Luckman, 1977). Protalus has been recognised in a number of cirques in the Lake St. Clair area (Derbyshire, 1964, 1968b, 1973). Bryan (1934) proposed the term protalus rampart to describe an accumulation of single rocks which mark the downslope margin of a snowbank across which impacting rocks have slid. Such features commonly lie forward of later talus and may be arcuate or linear in plan with a frontal angle of 40-50 degrees (White, 1981).

Derbyshire (1973) has recorded the presence of fossil rock glaciers on the slopes of Mt. Olympus and Mt. Gell. From the presence of transverse ridges and furrows he argued that the rock glaciers had been ice cemented. Ice-cored rock glaciers commonly exhibit collapse pits and meandering furrows (Vernon and Hughes, 1966; White, 1976). Block streams extend downslope and consist of interlocked blocks. Their fabric commonly suggests periglacial movement with the assistance of interstitial fine materials that have later been removed by piping (Caine, 1968a; White, 1981).

Unsorted diamictons mantle many slopes in the study area. These consist of fragments of local bedrock up to 2 m or more in size set in a silty clay matrix. They have moved over slopes of as little as 4 degrees. The clasts generally have a strong downslope fabric and are commonly angular. However, they may be spherical where they are derived from deeply weathered dolerite. These characteristics suggest that frost action played a major role in fragment formation and movement of the mass. Similar mantles in many areas of Tasmania have previously been interpreted as solifluction deposits (Davies, 1958, 1967; Nicolls and Dimmock, 1965; Derbyshire, 1973). Notwithstanding the difficulty in proving solifluction to have occurred by inference from the characteristics of fossil deposits (French, 1976; Washburn, 1979; Wasson, 1979; Soons, 1980) their geographical continuity with periglacial block slopes supports the proposition that they moved under periglacial conditions, while radiocarbon dating indicates that they were last active during the late Last Glacial Stage (Davies, 1967; Colhoun, 1979a; Wasson, 1977). They are particularly well developed on the dolerite mountains where the regolith is rich in clay.

Alluvial, palludal and peat sediments have been recorded in the text but are seldom sufficiently widespread to be depicted on the maps.

(iv) Glacial climates

Attempts to reconstruct the glacial climate of the study area have been based upon the pattern of glacial erosion and the character of the glacial and periglacial deposits. Derbyshire considered that the high discrete cirques would have been the first to have been glacierised and deglacierised. He argued that the progressive change in cirque form from west to east and the severity of glacial erosion in the west pointed to a stronger west-east precipitation gradient than that which prevails today.

The cirques around Lake St. Clair have been characterised as "subtropical continental" in form, and similar to those in the mountains of northeastern Spain, the low-mid latitude Andes, and parts of the Sierra Nevadas (U.S.A.) and the mountains of Japan. They have been attributed to strongly localised wind-drift accumulations of snow, high sun angles and low air temperatures (Derbyshire, 1967; Peterson, 1968, 1969). The better formed cirques and rock basins further west in the Frenchmans Cap massif have been attributed to higher glacier energy and a more maritime glacial environment, similar to that of maritime Norway (Peterson 1969).

The character of glacial deposits has also been suggested to point towards increasing continentality to the east. Peterson (1968, 1969) and Derbyshire (1969, 1971b) have suggested that protalus and terminal moraine ridges which bound many of the eastern cirques are more typical of continental than maritime conditions, and that the abundant protalus would have demanded many freeze-thaw cycles. Derbyshire (1966, 1971) argued that a progressive eastward reduction in the size of end moraines and outwash plains also suggested increasing continentality. This appears to have been confirmed by the descent of glaciers to lower elevations and more abundant washed drifts in the west. Derbyshire argued that the more abundant cryergic phenomena in the east further supported this, but their frequency may stem from the abundance of dolerite clays.

From the close association between rock glaciers on the western slopes of Mt. Olympus and the shallow cirques with protalus which occur on its eastern flank Derbyshire (1973) concluded that the summit area was cold and dry. However, he believed that such conditions may have been limited to the higher peaks as vigorous valley glaciers had produced a few washed drifts in the adjacent valleys. Some of the Lake St. Clair moraines comprise up to 75% bedded materials which suggests that meltwater was abundant in the terminal zones. Others exhibit undisturbed false bedding and only localised slumping of debris-charged ice with little meltwater

(Derbyshire, 1963, 1965). On the other hand the moraines in the central West Coast Range imply high rates of accumulation and ablation with rapid mass throughput (Kiernan, 1980).

Derbyshire (1967) noted a broad similarity between those areas of Tasmania which were glaciated and those which presently receive in excess of 250 mm/pa (water equivalent) of snow (Figure 2.4). As the Central Plateau did not fit this pattern he proposed that a greater proportion of the snow must have been associated with winds from north of west than is presently the case.

The calculation of palaeotemperatures from the equlibrium line altitudes of the valley glaciers is complicated by the presence of extensive erosion surfaces which impose altitudinal constraints upon glacier limits. A temperature depression of 7.3° C at Lake Dixon during the most recent glaciation was proposed by Kiernan _et al._, (1983). Derbyshire (1973) used the rock glacier limits on Mt. Olympus to calculate that the mean annual temperature when they were formed was at least 6.5° C less than at present. This latter figure is likely to provide a better estimate as there is likely to be less interference by the topography over which the rock glaciers moved.

(iv) Glacial chronology

The original proponent of multiple late Cainozoic glaciation in Tasmania was government Surveyor T. B. Moore who, following more than a decade of observations in the Franklin Valley and West Coast Range, arrived at the conclusion that

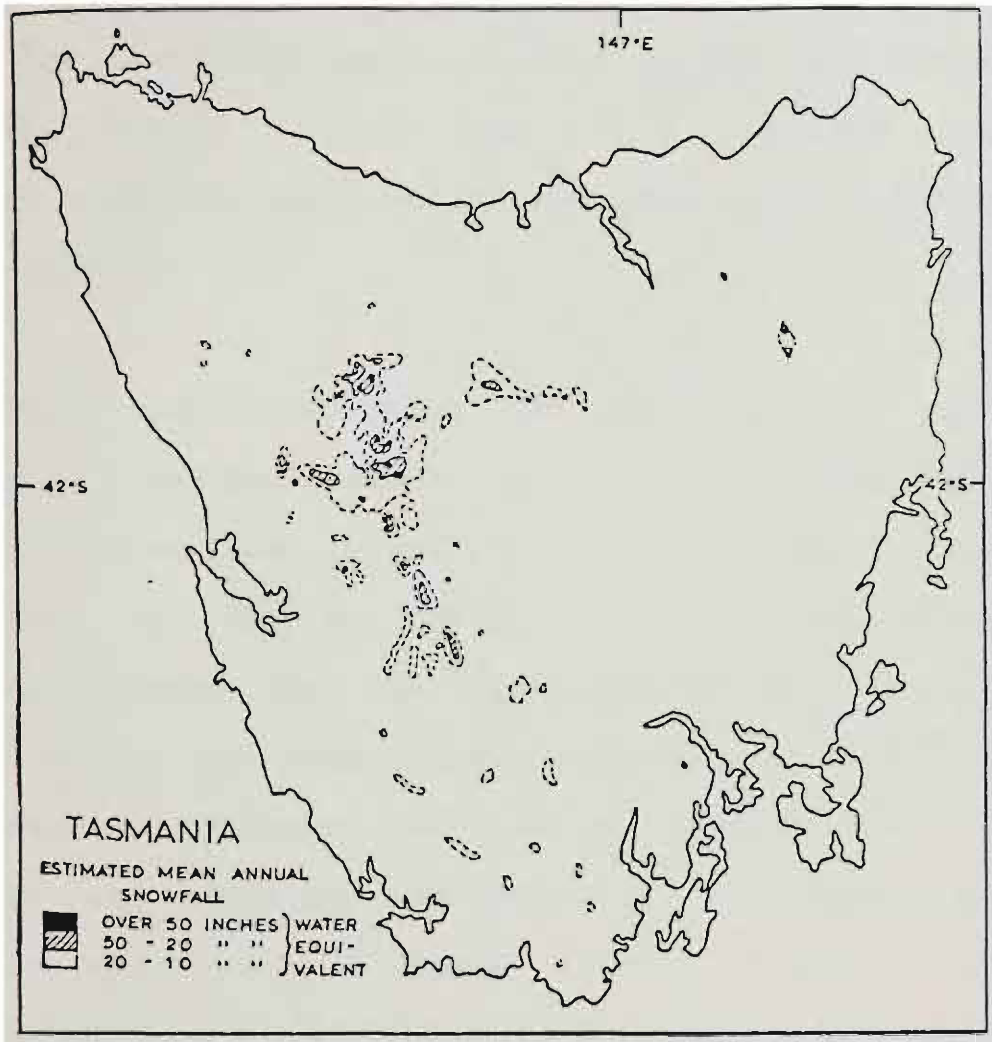


Figure 2.4 Estimated present snowfall in Tasmania, based on a Bureau of Meteorology formula relating total precipitation to orography (after Derbyshire, 1969).

two glaciations had taken place (Moore, 1883, 1894, 1895; Lewis, 1934). Moore's suggestion was subsequently taken up by Johnston (1894a), Taylor (1922) and David (1923) to account for distinct nivation levels in the West Coast Range and at Mt. Field. Hills came to a similar conclusion sometime prior to 1922 but did not publish his observations (Lewis, 1934).

Lewis later concluded that the younger of the two glaciations between which Moore, Hills and David had distinguished was the older of the two which had been recognised by Taylor (Lewis, 1923, 1924, 1934). He therefore proposed that three glaciations had taken place (Lewis, 1923) and subsequently both he and David confirmed this proposition (Lewis, 1926; David, 1923, 1926). Lewis envisaged an early ice cap glaciation followed by a valley glaciation and finally a cirque cutting stage. He named these the Malanna Glaciation, the Yolande Glaciation and the Margaret Glaciation respectively (Lewis, 1939) (figure 2.5).

Clemes (1925) was the first to apply a multiglacial perspective to the Lake St. Clair area. He argued that four glaciations had taken place and boldly correlated these with the Gunz, Mindel, Riss and Wurm of alpine Europe. Clemes suggested that "the preliminary work of dissecting the plateau and widening the great valleys" could be attributed to the Gunz. He attributed the lack of deposits from this period to erosion by the Mindel glaciers which had produced the most striking glacial landforms. Clemes believed that the Riss stage saw a cirque glaciation that had produced glacial horn peaks at Mt. Byron and Mt. Ida, and that the Wurm Glaciation had been very minor and affected only the highest sites such as the tarns on Mt. Olympus.

It was another decade before Lewis followed his former geology master with a tentative and temporary extension of his own model to embrace a fourth glaciation, based upon evidence from southern Tasmania (Lewis and Murray, 1935). Lewis considered that major valleys in the present study area such as the Surprise Valley were the result of erosion during the Malanna Glaciation. He suggested that Clarence Lagoon was attributable to the Yolande Glaciation. Although he relied mainly on erosional evidence he also recognised that there was considerable soil development between the Yolande and Malanna ice limits.

This multiglacial model was subsequently rejected on the basis that in focussing on erosion rather than upon stratigraphy it was methodologically unsound, and also following more detailed investigation of a number of critical sites. The first challenge came when a radiocarbon assay of 26,480 +- 800 BP (W-323), obtained on derived wood collected from within the Malannan deposits in the Linda Valley (Gill, 1956) suggested that the Linda Moraines dated from the last glaciation. A detailed study of the morphology of the Central Plateau found nothing to support multiple glaciation there, and from the freshness of the glacial erosion it was concluded that the plateau ice cap had occurred no earlier than 30-40 ka BP (Jennings and

Ahmad, 1957). Subsequent work at the Malanna type site suggested that the deposits were not glacial in origin but were associated with the development of fault-line scarps (Banks and Ahmad, 1959). Finally, a study of the Lake St. Clair area "failed to unearth any evidence of multiple glaciation " (Derbyshire, 1963). All the evidence seemed to be explicable in terms of a single glaciation during the Last Glacial Stage (Jennings and Banks, 1958; Derbyshire et al., 1965).

Other studies during this period continued to show uncertainty about the number of glaciations. Spry (1958a) proposed that an ice cap and subsequent, possibly recessional, valley glacier phase had occurred in the Mersey-Forth area of northern Tasmania. Spry encountered difficulties in extending this model to the Franklin Valley where the glacial deposits were much more deeply weathered (Spry and Zimmerman, 1959). This discrepancy was attributed to differences in postglacial climate, a contention which seemed supported by the similar deep weathering of the deposits in the Linda Valley. However, Spry and Zimmerman acknowledged the alternative possibilities that the Mersey-Forth deposits were either more recent or that the dating of the Linda deposits might be in error. Grant-Taylor and Rafter (1963) provided a radiocarbon assay of > 40,000 BP (NZ 348) from the Linda Valley but like another unpublished infinite date on wood obtained by Banks (R 488) its significance went unrecognised (E.A.Colhoun, pers. comm.)

Read (1964) postulated an ice cap and subsequent valley glaciation further west around the King-Franklin divide south of Eldon Bluff, while MacIntyre (1964) adopted a similar position following work in the Collingwood Valley. MacIntyre's observations that the Collingwood River had incised 10m. into bedrock since the ice had retreated, that many metres of slope deposits overlay till at Redan Hill and that terraces had been constructed up to 60 m above present river level all hint at a greater antiquity than the Holocene age she ascribed to them. The Glacial Map of Tasmania (Derbyshire et al., 1965) noted the presence of erratics beyond the limits of continuous drift in areas such as Redan Hill and reopened the possibility of multiple glaciation.

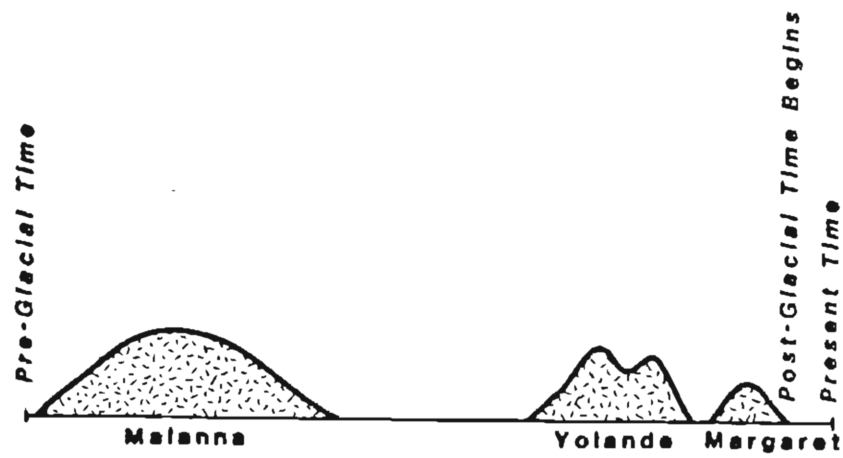
In the absence of dateable organic deposits, till weathering became increasingly critical to the search for evidence of multiple glaciation. In his study of the Frenchmans Cap area in the middle Franklin Valley Peterson (1966) showed the reserve towards weathering evidence that prevailed at that time. Peterson expressed caution that some clasts in the till might have been preweathered (Hale, 1958; Derbyshire, 1965). He later observed that the weathered clasts in the glacial deposits had "almost certainly decayed after emplacement in the till as glacial transport in mountain valleys would be unlikely to allow the retention of cohesion in weathered boulders." He acknowledged that "higher tills are in some cases obviously less weathered than tills in lower areas" but argued this to indicate "nothing more than the usual single phase sequence of cirque glaciers extending to a maximum as valley glaciers followed by retreat into the original cirques." He concluded that the deposits at Frenchmans Cap were comparable to the "equally fresh evidence of glaciation in the nearby Linda Valley" and concluded that all the deposits were referrable to the last glaciation.

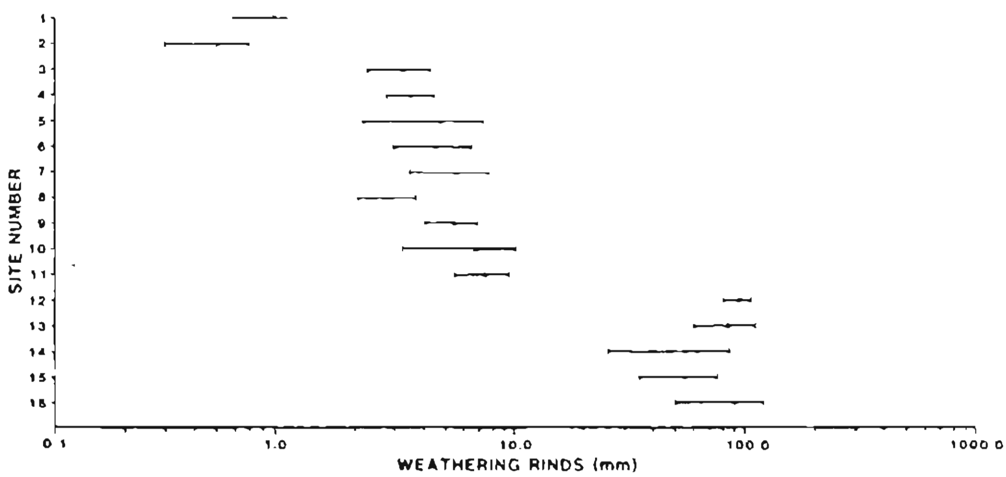
Subsequently Paterson (1965, 1966, 1969) used differences in the degree of lithification and chemical weathering of glacial deposits to argue that two separate glaciations had taken place in the Mersey-Forth area, the younger indicated by unconsolidated and unweathered drift and the older by weathered drift and highly lithified basal till. Derbyshire (1967, 1968a) also distinguished between two advances in the Derwent Valley on the basis of till consolidation and the development of secondary clay minerals, notably those of the kaolinite family, in the older till. Secondary clay minerals were generally absent from the younger till. He did not define the limit of the younger advance. While he found no fossil evidence to indicate whether interglacial or merely interstadial warming had intervened between the two glacial phases he argued that they were "well separated" in time. Derbyshire and Peterson (1971) attempted to categorise all the Tasmanian tills in terms of this two-fold model.

Colhoun (1975) obtained a radiocarbon assay of >36,400 BP (Gak-5595) on wood overlying till in the Henty Valley, reported one of the infinite assays from the Linda Valley (R 488) and reinterpreted the earliest of the Mersey-Forth

Figure 2.5 Relative dating and duration of Tasmanian glacial stages according to A.N. Lewis (1945b). This model was based primarily upon erosional landforms.

Figure 2.6 Thickness of dolerite weathering rinds from the glacial deposits of the central West Coast Range. Mean and standard deviation are plotted on a three cycle semi-log base (after Kiernan, 1983b). Compare with figure 2.5.



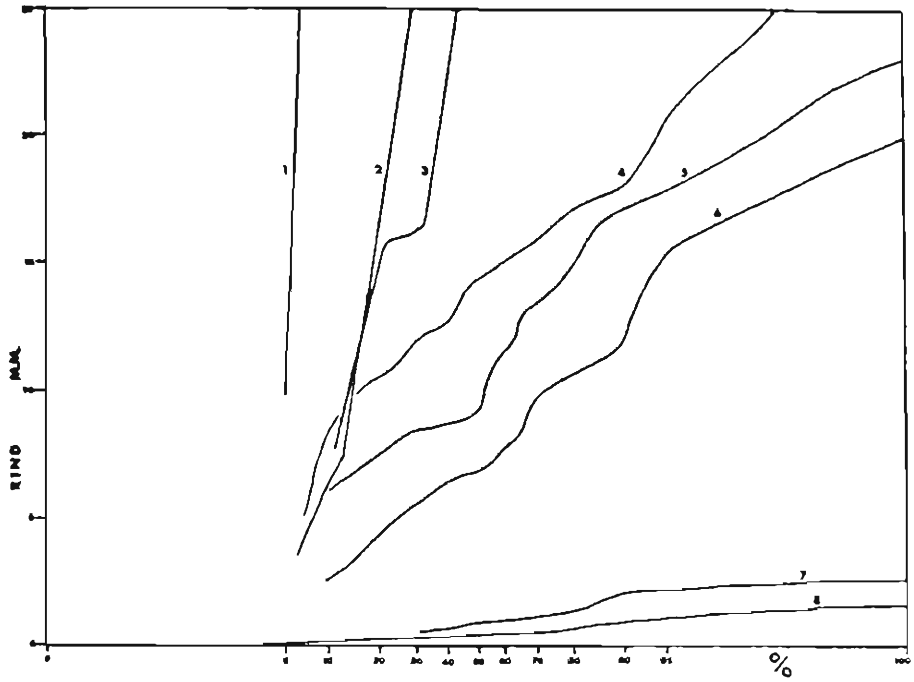


Sites
1. lower Dante
2. Sedgwick II moraine
}Dante Glaciation
3. lower gravels at Dante fan
4. Nelson River caves
5. Western Comstock
6. Sedgwick I moraine
7. King Valley
8. King railway bridge
9. Nelson Valley
10. lower Linda
11. Lynchford
}Comstock Glaciations
12. Karlsons Gap
13. Mine office
14. Football ground
15. Linda Hotel
16. Comstock Spur
}Linda Glaciations

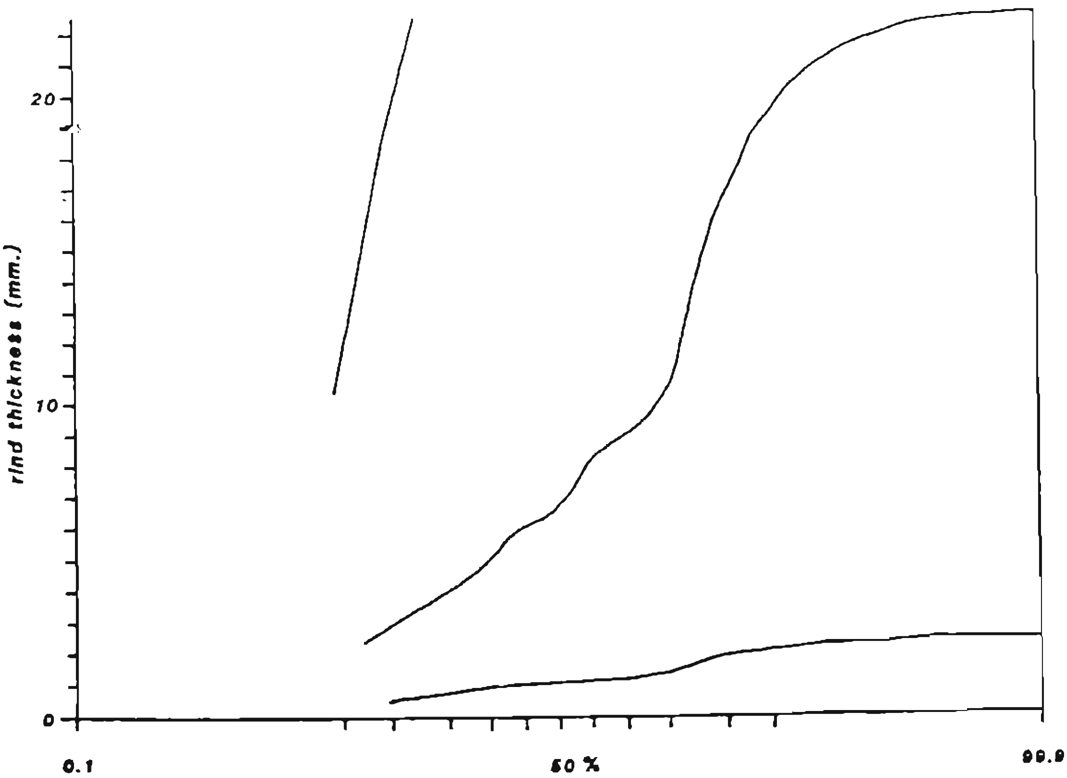
glaciations as possibly being of late Tertiary age. In subsequent papers Colhoun (1976) argued on the basis of till weathering and stratigraphic relationships that till in the Forth Valley predated the Last Interglacial, globally defined (Shackleton and Opdyke, 1973); argued from the weathering of the Linda deposits that they too must predate the Last Interglacial Stage (Banks et al., 1977); and showed that two glaciations had occurred in the middle Huon Valley in southern Tasmania (Colhoun and Goede, 1979). Subsequent fieldwork in the Pieman Valley provided radiocarbon evidence for a glaciation prior to the late Last Glacial Stage (Sansom, 1978; Colhoun, 1979a).

A radiocarbon assay of 18,800 +- 500 BP (ANU 2533) was subsequently obtained on driftwood in proglacial silts beneath lightly weathered outwash in the King Valley (Kiernan, 1980, 1983a, 1983b). This date indicated that a small glaciation, the Dante Glaciation, had taken place in the central West Coast Range during the late Last Glacial Stage. Two more extensive glaciations, the Comstock and Linda glaciations, were recognised on the basis of morphostratigraphic and weathering evidence. The difference in the thickness of dolerite weathering rinds in the glacial deposits amounted to one order of magnitude between each of these glaciations (Figure 2.6). Weathering rinds up to 300mm. thick on dolerite clasts and the presence of Tertiary type pollen in the sediments overlain by the moraines was interpretted as evidence that the Linda Glaciation was ancient. Two zones in the weathering of the Comstock deposits and presumed interstadial type pollen in deposits

Figure 2.7 Cumulative percentage frequency of dolerite weathering rinds from some glacial deposits: (1) in the Mersey River Valley, central northern Tasmania (after Kiernan, 1982b); (2) in the Weld River Valley, southwestern Tasmania (based on Kiernan, 1983a and unpublished data. Maximum recorded rind thickness in the Mersey River Valley is 250 mm and the maximum recorded in the Weld Valley is 185 mm.



1. Railton-till. Maximum recorded thickness is 250mm. 2. Mersey Hill, Mole Creek - till. 3. Tarleton - outwash terrace. 4. Mersey River bank 200 metres upstream from Croesus Cave - till. 5. Union Bridge, Mole Creek - till. 6. Parangana damsite - till. 7. Latero-terminal moraine 2.3 kilometers east from Rowallan Dam - till. 8. Last Glacial Maximum limit 1 kilometer north from Rowallan Dam - till.



between these two zones suggested that a degree of climatic amelioration occurred during the Comstock glaciation (Kiernan 1980).

Weathering evidence for a similar three-fold division of the glacial sediments in the Mersey Valley (Kiernan 1982b) and in several other Tasmanian valleys was subsequently recognised (Figure 2.7) and a three-fold model strikingly similar to that proposed by Lewis (1945b) was therefore re-erected (Kiernan, 1983a). The Linda Stage was tentatively dated at between 600 ka BP and late Pliocene in age on the basis of weathering evidence (Kiernan 1983a). The deposits at Linda, and others in the Pieman Valley that are weathered to a similar degree (Augustinus, 1982) have recently been shown to be palaeomagnetically reversed (Colhoun and Augustinus, 1984; Colhoun, 1985a). This confirms that they can be no younger than Early Pleistocene.

Radiocarbon assays of Tasmanian glacial deposits and some other diamictons are presented in tables 2.1, 2.2 & 2.3. Assays from a number of areas indicate that final deglaciation of Tasmania was complete by 10 ka BP (Macphail and Peterson, 1975). An early deglaciation date of 17,700 +- 400 BP (SUA 1359) from the Ooze Lake cirque in southern Tasmania is problematic because of the posibility of older organic material having been washed into the lake deposits following retreat of the ice (Macphail and Colhoun, 1985).

Few organic samples suitable for radiocarbon assay have been

SITE	MATERIALS AND POSITION	DATE (BP)	LAB.NO.	SOURCE	COMMENTS
West Coast Range	(Tasmap 7913,8013 and 8014)				
Map ref. 902459	Dante site - twigs in silt beneath palaeosol on outwash	20,100 ± 470	SUA 2155	this thesis	minimum age of outwash
" 815406	Conglomerate Ck - wood in palaeosol between tills, outside last glaciation ice limits.	30,050 ± 2000	ANU 2535	Kiernan, 1980	minimum age
" 785525	Langdon R. - organic deposits with interglacial type pollen between	41,900 + 1000 - 900	SUA 2277	Colhoun, 1985a	
" 785525	- " -	43,000 + 1200 - 1100	SUA 2278	Colhoun, 1985a	
" 734495	Henty Bridge - wood in clay overlying till	> 34,600	Gak 5595	Colhoun, 1979	
"	- " -	> 34,190	Gak 6294	Colhoun, 1979	
" 879843	Tullabardine Ck - wood overlying till	> 43,800	SUA 1047	Colhoun, 1985b	
" 872825	Mackintosh Dam - charcoal overlying till	> 35,000	SUA 1289	Colhoun, 1985a	
" 869822	Lower Mackintosh Valley - wood overlying till	36,200 + 3400 - 2400	SUA 1287	(Colhoun & Sanson)	
" 869822	Lower Mackintosh Valley - charcoal overlying till	> 31,700	SUA 1288	(Colhoun & Sanson)	
" 835422	Linda Ck - wood from rhythmites	26,480 ± 800	W 323	Gill, 1956	palaeomagnetically reversed (Colhoun, 1985b)
" "	" - wood from proglacial sands	> 40,000	NZ 348	Grant Taylor & Rafter, 1962	
" "	" - *Phyllocladdus* stump beneath glacial sands	> 40,000	R 488	Banks *et al.*, 1977	
" "	" - " -	> 48,500	ANU 3413	Colhoun, 1985b	
" "	" - wood in glacial sands	27,800 ± 700	ANU 2480A	Kiernan, 1980	
" "	" - α cellulose of sample ANU 2480	23,100 ± 600	ANU 2480B	Kiernan, 1980	
Mt. Mueller Map ref. 510657	Pontoon Hill - base of ice-pushed organic clays	33,240 + 3370 - 2370	Gak 5625	Colhoun, 1985b	probably minimum age

Table 2.1 Radiocarbon dating of Tasmanian glacial deposits that predate the late Last Glacial Stage (isotope stage 2 of Shackleton and Opdyke, 1973)

SITE	MATERIAL AND POSITION	DATE (BP)	LAB. NO.	SOURCE	COMMENTS
West Coast Range	(Tasmaps 8013, 8014, and 7913)				
902459	Dante site - driftwood in proglacial silt beneath outwash	18,800 ± 500	ANU 2533	Kiernan,1980,1983	predates glacial maximum
902459	Dante site - *Donatia novae zealandie* beneath silt	21,180 ± 370	SUA 2154	this thesis	"
784795	Chester Ck - charcoal in soil at 5.7m depth in scree inside earlier ice limit	18,190 ± 340	SUA 1042	Colhoun, 1985b	dates slope instability
734495	Henty Bridge-organic clay beneath slope deposits	23,640 ± 1030	Gak 5597	Colhoun, 1979	predates slope instability
Central Highlands	(Tasmap 8114)				
373746	Fish River - charcoal in lateral moraine	> 28,000	SUA 1938	(Hannan)	charcoal probably predates glaciation
Franklin Valley	(Tasmap 8012)				
989907	Kutikina Cave - charcoal beneath roof-fall deposit	19,700 ± 850BP	ANU 2785	Kiernan *et al.*, 1983	onset of frost shattering?

Table 2.2 Radiocarbon dating of the onset of conditions of maximum cold during the late Last Glacial stage in Tasmania.

SITE	MATERIAL AND POSITION	DATE (BP)	LAB.NO.	SOURCE	COMMENTS
Adamsons Peak 867004	(Tasmap 8211) carbon fragments at 2.5m in cirque core	9,080 ± 200	I 7571	Macphail & Peterson, 1975	minimum age of deglaciation
Denison Range 405900	(Tasmap 8112) L.Wurawina upper cirque (1000m) - peat at 1.65-1.75m.	10,360 ± 170	SUA 1354	(Colhoun & Macphail)	
Mt. Field 640758	(Tasmap 8212) unnamed tarn (1150m) - organic mud at 3m. in cirque	9,725 ± 180	I 8008	Macphail & Peterson, 1975	"
641757	James tarn (1158m) - organic mud at 1.5m in cirque	8,280 ± 460	Gak 1158	"	"
705752	Beatties Tarn (975m) - lake mud at 3.2-3.4m	11,420 ± 205	SUA 325	Macphail, 1979	"
664785	Eagle Tarn (1033m) - organic mud at 4.5-4.6m in cirque	12,960 ± 950	ANU 3118	(Green)	"
West Coast Range 824545	(Tasmap 8014) Tydnall Plateau (1000m) - organic clay at 0.82-0.94m	9,050 ± 120	SUA 1358	(Colhoun & Macphail)	"
807508	Poets Hill (Intermoraine) Lake (600m) - organic sediment	11,420 ± 770	Gak 6297	Colhoun, 1979	"
Frenchmans Cap 032207	(Tasmap 8013) Lake Nancy (1036m) - wood and leaf fragments on cirque lip	8,720 ± 220	Gak 671	Peterson, 1968.	"
078193	Lake Vera (560m) - organic clay at 2.7-2.8m	11,530 ± 240	I 7683	Macphail, 1979	"
Franklin Valley 989907	(Tasmap 8012) Kutikina Cave - hearth on top of roof fall deposit	14,840 ± 930	ANU 2781	Kiernan *et al.*, 1983	end of frost shattering
Central High-lands 264665	(Tasmap 8114) Warragarra Rockshelter - archaeological site inside last glaciation ice limit	9,760 ± 720	Beta 4757	Lourand os, 1982	minimum age of major ice retreat
Southern Ranges 758837	(Tasmap 8210) Ooze Lake (880m) - lake clay at 5.7m	12,590 ± 230	SUA 1943	Macphail & Colhoun, 1985	inwashed charcoal may predate deglaciation
758837	Ooze Lake (880m) - charcoal in lake clay at 6.4-6.65m	17,700 ± 400	SUA 1359	"	

Table 2.3 Radiocarbon dating of climate amelioration and deglaciation in Tasmania.

SITE & MAP REF.	MATERIAL AND POSITION	DATE(BP)	LAB.NO.	SOURCE	COMMENTS
Monpeelyata Canal (8213) 730385	charocal in palaeosol below solifluctate	30,400 ± 230	Gak 1163	Derbyshire, 1968c	maximum age of slope instability
	Astelia alpina fragments below solifluctate	2,900 ± 80	Gak 1020	Derbyshire, 1968c	same level as Gak 1163 - suspect
Browns Marsh (8213) 643305	peat at 3.1m in core	8,575 ± 125	I 9558	Macphail, 1979	inception of peat formation
Dyes Marsh (8113) 485385	peat at 5.2m depth	4,930 ± 180	Gak 784	Derbyshire, 1967	suspect; see Chapter 15
Tarraleah (8113) 556162	*Sphagnum* peat at 0.4-0.5m in core	7,000 ± 160	ANU 2496	Macphail, 1984	
	Sedge peat at 1.1-1.2m in core	7,970 ± 180	ANU 2745	Macphail, 1984	
	organic clay at 2.5-2.54m	9,080 ± 195	I 9559	Macphail, 1984	
Laughing Jack Lagoon (8113) 415320	wood in organic silt at 0.45-0.5m	810 ± 60	SUA 1957	this thesis	see Chapters 12 and 15
	wood in minerogenic silt at ~1m	1,540 ± 60	SUA 1958	this thesis	see Chapters 12 and 15
Narcissus River (8113) 256484	charcoal at base of silt over outwash	7,650 ± 250	SUA 2079	this thesis	see Chapters 4 and 15
Mt.Gould Plateau tarn (8113) 220512	*Athrotaxis selaginoides* fragments in organic silt, 1.6m. in core	7,920 ± 250	SUA 2080	this thesis	see Chapters 4 and 15
Mt. Arrowsmith (8113) 255261	charcoal in palaeosol on till beneath scree	13,000 ± 640	SUA 1959	this thesis	see Chapters 12 and 15

Table 2.4 Radiocarbon dates from the Tasmanian Central Highlands.

obtained from the Tasmanian Central Highlands and none of these have provided useful evidence concerning the age of the glacial events (Table 2.4). Some major objectives of the present study are to re-assess the evidence for multiple glaciation in the Tasmanian central highlands and to define the limit of the younger drift recognised by Derbyshire; to investigate the status and complexity of the "older advance" (Derbyshire, 1967); and to assess the relationships that existed between the glaciations of the Central Highlands and those that occurred in the West Coast Range.

chapter three

MAPPING AND DATING THE GLACIATIONS

" Of all these criteria only a few can usually be applied in any one place. One of them affords only a tentative opinion, but when several of them point to the same conclusion confidence is much strengthened"

- Blackwelder, 1931.

A major aim of this study is to expand on the work of Derbyshire (1967) and to differentiate the sequence, duration and age of the glacial events. Data assembled during a lengthy field mapping programme were plotted on maps of 1:50,000 and on aerial photographs at a variety of scales. Aerial photography aided the identification and plotting of the landforms which formed the basis for mapping several of the glacial units, but was of limited assistance in heavily forested country.

Landforms were identified on the basis of conventional criteria (Embleton and King, 1975) and the identification of glacial deposits follows Dreimanis (1976). The topographic position of moraines and their associated outwash terraces provides a basic means of establishing the relative sequence of glacial advances and retreats (Peterson, 1966). The moraine and outwash together provide a single mappable morpho-stratigraphic unit (Frye and Willman, 1962). More detailed field investigations sought to locate exposed sections through the deposits that would enable the development of a stratigraphic model. Organic deposits suitable for radiocarbon assay or which might provide palynological evidence from which past climates might be interpreted were also sought.

Few useful sections were located. The little organic material that was obtained all proved to be Holocene or latest Pleistocene in age (c. 13 ka BP) (Table 2.2) and so failed to provide any useful radiocarbon dating of the glacial events or to provide any insight into Pleistocene climates. Time-dependant differences in the form and character of the landforms and deposits have, therefore, assumed considerable importance in the development of a stratigraphic model. Such characteristics are commonly referred to as relative dating (RD) criteria (Birkeland _et al._, 1979). They might more accurately be referred to as post-depositional modifications (PDM) as they exist independant of whether they are used as an aid to relative dating. They comprise changes to the morphology of landforms, and physical and chemical changes to the deposits that occur progressively over time. Most of these techniques focus upon glacial deposits and depositional landforms but a few examine changes that have occurred to erosional landforms within ice limits defined by glacial deposits.

1. Post-depositional Modification as a Dating Tool: Some General Considerations.

Evidence of a post-depositional kind has long been used to date landforms and deposits (eg. Leverett, 1909), and has gained increasing acceptance in the study of glacial landscapes (Blackwelder, 1931; Richmond, 1962; Sharp, 1969; Burke and Birkeland, 1976).Despite some misgivings due to the possibility of some characteristics being a legacy of original deposition rather than being post-depositional in origin (Hale, 1958; Derbyshire, 1965; Peterson, 1966) such criteria have been applied to Tasmanian glacial problems by a number of workers. Some have adopted essentially qualitative approaches with varying degrees of success (Gill, 1956; Derbyshire, 1967). In recent years others have achieved more consistent results through the employment of more quantitative methods (Caine, 1968, 1983; Kiernan, 1980, 1982, 1983a,b; Augustinus, 1982).

The use of such criteria as dating tools is complicated by the fact that the characteristics of any one landform or deposit are influenced by factors other than age. Almost all post-depositional modifications are a response to the same range of determinants that was identified by Jenny (1941) as fundamental to soil development, that is, not time alone but also the nature of the parent material, climate, vegetation and topography. Ideally the dating of landforms and deposits should be based upon inter-site differences which are solely a function of age, but in reality this is difficult to achieve. The validity of conclusions based on PDM criteria rests heavily on the extent to which the influence of factors other than time has been minimised in the selection of sites, and on the parameters chosen for scrutiny.

A wide range of parameters has proved useful for dating glacial sequences, but generally only a few of these will be available in any particular area, and even fewer will be applicable at all sites within it. Excessive reliance upon any criterion is unwise because other factors may have been more influential than time and give results which depart from a time dependant sequence. Very often some criteria provide results that appear to be meaningless or are even contradictory to other lines of evidence. Greatest confidence can be placed in those conclusions that are supported by several different parameters (Blackwelder, 1931; Burke and Birkeland, 1979).

Further difficulties that arise in the use of PDM as a dating tool include the identification of the most useful criteria and the quantification of the results in a manner that facilitates useful comparison. Comparison is often hindered by differences such as lithological variation which militate against the collection of data on the same parameters at all sites. Computer manipulation of PDM data is often not justified because the original imput is too incomplete, too coarse, or even too clouded by operator variance. A further impediment is inadequate knowledge of the rates at which post-depositional modification takes place and the relative impact of different environmental factors upon those rates of change (Birkeland, 1984).

Due to the probable intervention of the local factors of parent material, topography and organisms, especially vegetation, any correlation between different areas would seem to be most sound if based upon similar relative sequences rather than on absolute numerical similarities in the PDM values. If erosion or weathering rate equations can be derived then it may prove possible to estimate the ages of glaciations in numerical terms (Colman and Pierce, 1981; Caine, 1983; Kiernan, 1983a).

Despite these limitations the employment of PDM criteria may enable better differentiation between glacial deposits that are probably of different ages where stratigraphic or radiometric evidence is not available. If no differences in PDM can be demonstrated then deposits are more likely to be of the same age than of different ages. The terrestrial record of glaciations is fragmented. This is partly due to the complete or partial obliteration of earlier deposits during subsequent glaciations. A recent probability analysis by Gibbons _et al._, (1984) suggests that the most probable terrestrial legacy of the glaciations that are inferred from the deep sea record to have taken place during the last 900ka, is a sequence of only two or three moraine belts. This is consistent with the glacial record already evident in Tasmania (Kiernan, 1983a; Colhoun, 1985a). The real importance of using PDM techniques may ultimately prove to lie more in the matching of glacial deposits in different areas than in enabling a significant number of further glaciations to be recognised.

2. Relative Dating by PDM: Applications and Limitations.

In developing a stratigraphic model morphologic and stratigraphic evidence has been supplemented by the evidence derived from a variety of PDM criteria. The assumption that all the landforms and deposits were originally in an identical or very similar condition underlies all these techniques, and undoubtedly provides their greatest weakness. In the case of the techniques which focus on the nature of deposits the condition of the original material can often be cross-checked to some extent by reference to the unaltered material at greater depth. Not even this low level of certainty is available for morphological techniques that assume very similar original slope angles and crest widths, or demand an absence of syngenetic modification.

Because the principal valleys in the study area extend from dolerite uplands, dolerite occurs in all the glacial deposits. This facilitates the aquisition of PDM data using comparable material. However, because the proportion of dolerite in some of the deposits is low, and it is desirable to compare as many parameters as possible, other rock types have also been investigated. Almost all of the glacial troughs are incised through Permo-Triassic rocks for at least part of their course. Hence, these rocks are also ubiquitous in the glacial deposits. Some of the Permo-Triassic clasts are highly siliceous and very resistant to weathering. Others comprise incoherent sandstones or mudstones which break down so readily that they have totally decomposed in all but the most recent glacial deposits.

A brief resume follows of the specific PDM parameters that have been used in this study to aid dating. Their

	Criterion	Rationale/Assumptions	Principal Limitations	References	Methodology of present study
1.	Weathering rinds on igneous clasts	rinds thicken inwards over time	all soil forming actors important especially climate and clast mineralogy	Benedict, 1981; Coleman & Pierce, 1981 Porter, 1981	25 clasts broken; caliper spike in pit of deep, soft rinds
2.	Weathering rinds on secondary clasts	- ditto -	- ditto -	Anderson & Anderson, 1981	minimum of 10 clasts broken
3.	Weathering rinds on metamorphic clasts	- ditto -	- ditto -	Icole, 1973	minimum of 10 clasts broken
4.	Hardness of weathering rinds	rinds become softer over time	softeness may be a function of retained moisture rather than time		scale 0-hammer needed to penetrate 1-knife, 2-sharp end of pencil 3-blunt end of pencil, 4 finger, 5-fist
5.	Clay mineralogy of weathering rinds	secondary clay minerals will be best developed in oldest rinds	clay mineral development may occur too slowly to be a useful tool	Kiernan, 1980; Colman, 1982a,b	XRD of powdered rind
6.	Pebble coherence	chemical weathering will destroy pebble structure over time	all soil forming factors impinge, especially climate and mineralogy	Drake, 1971	qualitative; scale of 0(coherent) to 5(barely recognisable)
7.	Clast surface colour	weathering of rocks will deepen oxidation colours over time	till geochemistry and other soil forming factors important	Whitehouse *et al.*, 1980	Standard Soil Color Charts

Table 3.1 Post-depositional modification of subsurface clasts and its use as a relative dating tool.

applications and principal weaknesses are tabulated. The factors that underlie these limitations are addressed more fully in the following section. It has only been possible to record the majority of the listed criteria at a very few sites. Rather than incorporate a full review of the assumptions, applications and methods relevant to each technique the reader is referred to the works cited.

Techniques which are based on PDM of subsurface clasts have provided the most consistent results of all relative dating methods (Burke and Birkeland, 1979; Colman and Pierce, 1981) (Table 3.1). One reason for this lies in the fact that the subsurface environment tends to be less prone to non-temporal influences than the surface. The most frequently used technique involves comparison of the thickness of weathering rinds.

A second set of methods focuses upon the modification of the subsurface till matrix (Table 3.2). These examine the degree of pedogenic alteration which is achieved over time. Virtually all require numerical expression to be used effectively. Difficulties in differentiating between pre-consolidated till and those which have undergone post-depositional lithification can impede this line of investigation.

A third set of methods examines the progressive modification of clasts that occur on the surface of moraines (Table 3.3). These methods assume that older moraines will be more impoverished in intact clasts than will younger moraines.

Criterion	Rationale/Assumptions	Principal Limitations	References	Methodology in present study
1. Colour of B horizon	pedogenic colour change most advanced on oldest tills	C horizon not always accessible for comparison; varying significance of individual soil forming factors	Nelson, 1980; Birkeland *et al.*, 1980	field condition, Standard Soil Color Charts
2. Depth to C horizon	pedogenic alteration will have most deeply penetrated oldest tills	varying significance of individual soil forming factors	Flint, 1971; Birkeland, 1978; Kiernan, 1983a	measured in metres to deepest clast with rind
3. Depth to lowest Cox horizon	oxidation will have extended to greatest depths in oldest tills	geochemistry - matrix density, vegetation and climate important	Burke & Birkeland, 1979	measured in cm. to lowermost iron-pan.
4. pH of matrix at fixed depth in profile	oldest tills will be most acid at depth	varying significance of individual soil forming factors	Porter, 1976	field determination at fixed depths
5. Percentage loss on ignition	organic enrichment will be greatest in oldest till profiles	- ditto -	Nelson, 1980	sampled at fixed depths, combusted for 7.00hours at 350-400°C
6. Clay mineralogy	secondary clay mineral development most advanced in oldest tills	secondary minerals may have been deposited with tills	Derbyshire, 1967; Quigly & Ogunbadejo, 1974	X-ray diffraction of samples from fixed depths
7. Development of clay films	clay void-linings develop progressively over time	microclimate and site drainage important	Birkeland, 1964	qualitative, scale of 0(nil) to 5 (thick)
8. Grain size below sand - laboratory determination	chemical weathering will produce smaller grains over time	orginal size range important	Birkeland, 1964; Rutherford, 1971	hydrometer and seive methods
9. Grain size below cobble grade - field determination	- ditto -	- ditto -	Sharp, 1969	pit excavation; seived 0.5 and 1.0cm; weight in field cond.
10. Degree of lithification	oldest deposits are the most indurated and lithified	preconsolidation, geochemistry and over burden pressures important	Patterson, 1966	qualitative; scale of 0(pliable) to 5(hammer to modify)
11. Clast socket staining	sockets in oldest tills will be most discoloured by illuviated products	till geochemistry and other soil forming factors important	Porter, 1981	qualitative; scale of 0(same colour as till) to 5(v.discoloured

Table 3.2 Post-depositional modification of subsurface till matrix and its use as a relative dating tool.

Criterion	Rationale/Assumptions	Principal Limitations	Reference	Methodology in present study
1. Ratios of surface to subsurface clast lithology	readily weatherable surface clasts assumed to be progressively depleted over time	till compositions in valley moraines seldom likely to be uniform	Sharp & Birman, 1963; Birkeland, 1964	50 surface, 50 subsurface clasts; super-imposed locations
2. Number of remnant surface boulders on moraines	surface boulders are progressively broken down over time	running water may concentrate surface boulders as lag deposits	Blackwelder, 1931; Sharp, 1969; Burke & Birkeland, 1979	quadrat 10m x 3M along moraine crest
3. Percentage of clasts with remnant abrasion surfaces	abrasion surfaces become more pitted over time	abrasion surface frequency may never have been uniform	Sharp, 1969	quadrat 10m x 3m along moraine crest
4. Clast weathering rind thickness	weathering rinds extend inwards and thicken over time	may reach an equilibrium thickness due to rind breakdown on exterior	Jackson & Keller, 1970; Thorn, 1975; Porter, 1970; Chinn, 1981	25 clasts broken for the purpose
5. Ratio of split to non-split clasts	clasts on oldest moraines assumed to have undergone more physical weathering	microclimate and fire history important	Shroba, 1977; Burke & Birkeland, 1979	quadrat 10m x 3m
6. Clast surface oxidation colours	weathering of rocks will deepen oxidation colours over time	mineralogical factors important	Whitehouse *et al.*, 1981	field condition, Standard Soil Color Charts
7. Depth of surface pitting on clasts	progressive weathering out of mineral crystals will cause deeper pits over time	mineralogical factors important	Carroll, 1974; Miller & Birkeland, 1974	measured in mm. with vernier caliper spike

Table 3.3 Post-depositional modification of surface clasts and its use as a relative dating tool.

Criterion	Rationale/Assumprions	Principal Limitations	References	Methodology in present study
1. Local relief of moraine	assumes that moraine height will diminish over time due to erosion	differences may be depositional in origin		field estimate or aneroid determination
2. Width of moraine crest	assumes that crests become broader as they are eroded down	moraines may be deposited with broad crests	Mahaney, 1973; Burke & Birkeland, 1979	measured in metres by tape
3. Moraine slope angles	moraine slopes assumed to become more gentle as erosion proceeds	original moraine slopes may be gentle	Benedict, 1981; Burke & Birkland, 1979	measured in degrees by clinometer
4. Periglacial disturbance of moraine slopes	disturbance will be greatest on oldest moraines	possibility of syngenetic disturbance	West, 1969; Clapperton *et al.*, 1978	terrace trend and riser measured by tape; range or average if group
5. Dimensions of patterned ground within glacial limits	large scale patterning indicates subsequent phase of periglacial climate	dependent on substrate;	Grant, 1977	measured by tape; range or average if group
6. Extent of solifluctate within glacial limits	solifluctate will be thickest within oldest limits	dependent upon bedrock or other substrate	Potter and Moss, 1968	measured thickness at natural exposure
7. Extent of talus accumulation within glacial limits	accumulation will be greatest within oldest limits	may be conditioned by different rock types and microclimate	Brookes, 1977	qualitative assessment
8. Survival of primary ice melt topography	kettle holes will be more degraded on older moraines	kettle holes may never have existed	Clapperton *et al.*, 1978	qualitative assessment; scaled 0 (no change) - 5(advance)
9. Degree of fluvial dissection	general measure of progressive moraine erosion	dependent upon topographic condition	Nelson, 1954; Birkeland, 1968	qualitative assessment, scaled 0-5

Table 3.4 Post-depositional modification of depositional landforms and its use as a relative dating tool.

Criterion	Rationale/Assumptions	Principal Limitations	References	Methodology in present study
1. Width and depth of joint opening	weathering will have penetrated joints most deeply on oldest surfaces	original condition, rock type and microclimate important	Caine, 1983	measured in millimetres
2. Depth of fluvial or glaciofluvial incision	older moraines will be more dissected	incision proceeds at different rates on different rock types	Morrison, 1968	depth measured in metres
3. Width and depth of solution pans	pan size will increase with time	rock type factors may cause variation	Pheasant, 1971; Caine, 1983	average of 5 pans for site if possible
4. Presence, size and extent of tors	large tors assumed to be removed by strong glacial erosion	not all may be removed during glaciation	Pheasant, 1971; Sugden, 1968	qualitative, or measured in metres
5. Heights of solution pedestals beneath erratics	height of pedestal measures extent of landsurface lowering since deposition	rock type, microclimate and soil characteristics important	Peterson, 1982	measured in millimdtres

Table 3.5 Post-depositional modification of erosional landforms and its use as a relative dating tool.

Factors other than time significantly impinge upon some of these parameters, for instance fire may open the ground surface to erosion and split clasts.

Some of the first relative dating techniques to have been used are those that rely upon apparently time-dependant degradation of original depositional morphology (Table 3.4). A final set of methods focuses not upon deposits but upon erosional landforms that occur between successive glacial limits, and compares the extent to which the erosional morphology has been degraded (Table 3.5).

Relative dating in this thesis is based primarily upon sub-surface criteria.

3. The Determinants of Post-depositional Modification.

Apparent post-depositional differences in the character of landforms and deposits are the result of a variety of factors that can be placed in four broad groups. These are: (1) primary factors, which stem from the condition, character and origin of the site from the outset; (2) environmental factors, whereby different sites are subject to different modifying forces or similar forces operating at different rates; (3) temporal factors, particularly the period of time over which any particular site is subject to modification; (4) operator factors, such as operator perception and operator variance; and (5) any combination of primary, environmental, temporal and operator factors.

Jackson and Sherman (1953) have argued that the factors responsible for chemical weathering are essentially those enunciated by Jenny (1941) as responsible for soil development. This perspective has been extended by Colman and Pierce (1981) to explain the development of weathering rinds. It is applicable to all forms of PDM. In terms of the model of Jenny (1941) parent material is a primary factor; climate, vegetation and topography are environmental factors; and time is a temporal factor.

Serious difficulties confront any attempt to isolate the effects of the individual factors upon weathering and erosion. All are complexly inter-related. For example, parent material, climate (both past and present) and topography may all condition the nature of the vegetation cover. The vegetation cover may in turn condition the local climate and the nature of weathering processes which subsequently modify the parent material. This modification leads to further changes in the vegetation that will influence soil acidity. Providing that its limitations are recognised, the basic four-fold model outlined above provides a convenient framework within which to consider the origin of site differences and the safeguards required in any application of PDM dating techniques.

A. Primary factors.

Primary factors that influence the apparent response of any landform or materials to PDM include the original morphology of the landform and the composition of the rocks. Variability in the response of those materials may be the result of their mineralogy, prior weathering and packing during or after deposition. Any application of PDM parameters to dating problems must ensure that comparisons are made only between materials which are likely to have responded to like processes in similar ways and at similar rates.

1. General considerations.

There is no guarantee that all parts of a bedrock surface were uniformly abraded by ice as local variations in glacier flow influence the intensity of ice erosion from place to place. The failure of glaciers to remove deeply weathered krasnozem and bauxite soil profiles in the Tasmanian highlands (Hale, 1958) together with the presence of autometasomatic alteration products to very great depths in the dolerite joint networks (Hale, 1958) might easily give an impression that a greater degree of postglacial modification has taken place within former ice limits than has actually been the case. In the same way, any technique that is underscored by an assumption of original morphological uniformity among landforms as potentially diverse as moraines produced by ice cap and outlet valley glaciers must be regarded as highly suspect.

Marked differences also exist in the rate at which erosion will proceed on different types of rock. In the central part of Tasmania's West Coast Range postglacial streams have incised far more deeply into Cambrian volcanics than they have into the more resistant siliceous rocks of the Owen Formation (Kiernan, 1980). Similarly, weathering and erosion processes may proceed at a greater pace and with greater effect upon ablation and meltout tills than upon highly compacted basal tills.

Some writers have also expressed concern that the entrainment of preweathered materials may impart a false appearance of age to some glacial deposits (Derbyshire, 1965; Peterson, 1966). However, it seems unlikely that weathered clasts would remain intact for long during glacial transport (Peterson, 1966). Preliminary investigations by the writer suggest that weathering rinds are fairly rapidly lost by abrasion from dolerite clasts entrained in solifluction deposits on Mt. Wellington in southern Tasmania.

Differences in the grain size distribution of the clasts in glacial deposits are also likely to affect the rates of weathering. In the central part of the West Coast Range dolerite weathering rinds in outwash deposits are generally at least 20% thinner than those in the associated moraines, while in the western United States the difference is about 11% (Kiernan, 1980; Colman and Pierce, 1981). The precise cause of this relationship is unknown but it seems likely that the greater retention of moisture in the more clay-rich tills may be the basic control. The matrix texture also influences the surface vegetation and, hence, can influence variations in the acidity of the soil moisture.

Differences in the size of mineral grains also exert a major influence upon weathering rates. Although it has been argued that finer grained clasts break down more rapidly than coarse grained clasts (Derbyshire, 1967) the opposite seems more likely to be the case. Coarse grained igneous rocks such as granites generally succumb to grussification within the space of a few thousand years while finer grained hypabyssal rocks may persist for tens of thousands or even hundreds of thousands of years (Icole, 1963; Birkeland et al., 1979). In the western United States Colman and Pierce (1981) found that rinds on fine grained andesites are 84% as thick as those on coarser grained clasts.

Mineralogical factors may reverse the trend for preferential weathering of coarse grained rocks. Mafic rocks weather more rapidly than felsic rocks (Goldich, 1938; Loughnan, 1969; Bikeland, 1984). On the other hand it has been shown in at least one case that rind development involved no preferred order of element loss (Donner and Anderson, 1962). A feedback process may also take place in some tills that are rich in clasts susceptible to comparatively rapid breakdown. This is likely to operate through the cumulative effect upon the proportion of clays and hence moisture retention, and possibly also through leachates that are aggressive to other minerals in the till. Acid clays may enhance weathering (Graham, 1941; Grant, 1969).

In summary, the impact of primary factors upon PDM parameters is invariably great. Colman and Pierce (1981) found their effect on PDM of basalt and andesite clasts exceeded that of climate between different areas, although climate was dominant within each area. The present thesis restricts comparisons to sites of similar lithology and till facies. The parameters recorded for dolerite are only compared between sites where this rock type makes up at least 50% of the clasts.

2. Dolerite and its weathering products.

In view of the importance to this study of dolerite weathering it is appropriate to examine the nature of this rock and its decomposition. The dolerite is a mafic igneous rock that occurs in the form of sills up to 500 m thick and as steep-sided dykes. It is generally highly coherent, the principal points of weakness being cooling joints that are vertical and hexagonal (or less frequently, rectangular and platy as at the eastern end of Laughing Jack Lagoon) together with horizontal faulting or unloading joints which intersect the cooling joints to produce hexagonal prisms (Hale, 1958).

The dolerite originated from a tholeiitic quartz dolerite association and its basic constituents are plagioclase feldspar (average 42%), pyroxenes (augite and pigeonite) (34%), and a mesostasis (24%) that is generally quartz and alkaline feldspar (McDougal, 1961, 1962). The texture is generally ophitic to sub-ophitic, although in more acid situations there may be a duplex texture with finer grained crystalline patches occurring in association with radial intergrowths of alkalai feldspar and minute pyroxene crystals. Analysis of the dolerite at chilled margins indicates a remarkable lateral uniformity of composition. However, <u>in situ</u> vertical differentiation of the magma on cooling has produced variation that can extend to andesitic composition in extreme cases.

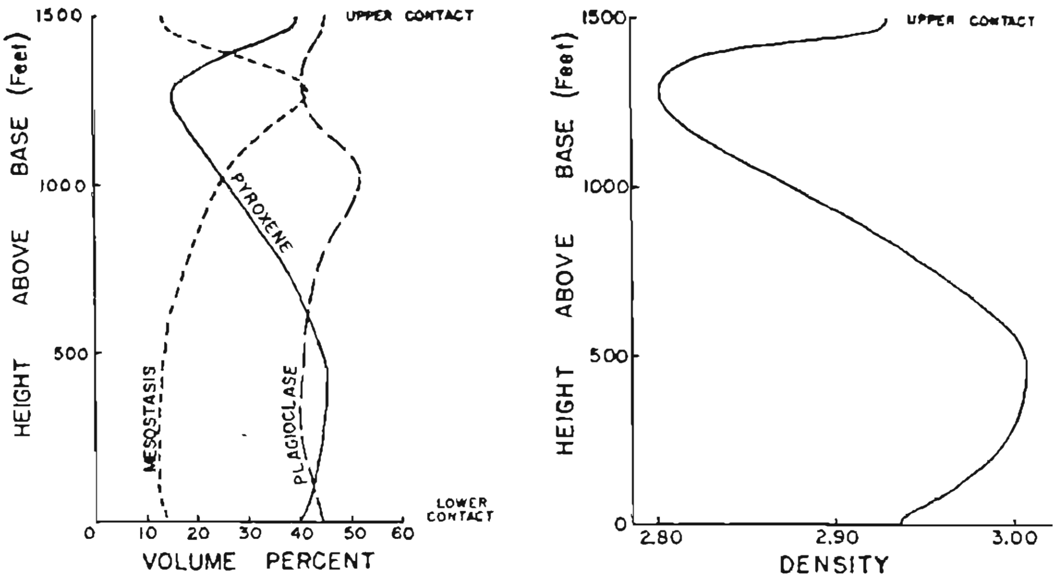


Figure 3.1 . Vertical variation in composition and density of Jurassic dolerite through a thick sheet (after McDougall, 1962).

This vertical differentiation means that there is considerable variability in chemical composition, mineralogy, grain size, fabric, density and magnetic properties (Edwards, 1942; Joplin, 1957; McDougal, 1962).This differentiation is sufficiently pronounced to allow the dolerite to be zoned into four mappable units (lower, central, pegmatitic and upper) on the basis of grain size and pyroxene : plagioclase : mesostasis ratios (Spry, 1958b) (Figure 3.1 and Table 3.6). In the lower zone of the dolerite near Great Lake pyroxenes (orthopyroxene, auigite and pigeonite) predominate over plagioclase (bytownite) and there is little mesostasis. The central and upper zones show increasing predominance of plagioclase and more mesostasis. In some cases there may be a granophyre zone where late crystallising residues lost volatiles through a fractured roof. Here the mesostasis may exceed the combined proportion of pyroxene and plagioclase (Sutherland, 1973).

The mineralogical and grain-size differences are of great significance to how the rock responds to weathering and erosion. Some dolerites which are more mafic and MgO rich than the original magma can result from the early settling and crystallisation of magnesium rich pyroxenes from higher parts of the intrusion (McDougal, 1962). The mesostasis breaks down particularly rapidly (Spry, 1958b) hence the relative proportion of the mesostasis exerts a major control upon the rate at which the dolerite weathers. Grain-size is also important. Nicolls (1958a) argues that the influence of varying differentiation products on gross soil development

MINERAL	Lower Zone	Central Zone	Granophyre
Clinopyroxene	43.0	23.9	13.1
Orthopyroxene	5.0	-	-
Plagioclase	42.5	44.3	16.6
Mesostasis	10.1	29.5	65.8
Iron Oxide	9.5	2.3	4.2
Olivine	-	-	0.3

Table 3.6 Mineralogy of dolerite differentiates, Great Lake area (after Sutherland, 1973).

is slight, but Spry (1958b) considers the influence on clast weathering to be considerable, coarse grained dolerites being the most rapid to succumb to attack.

The chemical weathering of dolerite clasts initially leads to the development of a weathering rind. This may be defined as a concentric zone of weathered rock that parallels the outer surface of a clast and extends inward over time (Colman and Pierce, 1981). Such rinds are easily distinguished on most dolerite clasts. The thickness of the rinds is easily measured because the transition from the brownish or reddish colour of the rind to the blue-grey of the unaltered dolerite is generally sharp. Occasionally it may be diffuse particularly in coarse grained clasts, with a whitish zone of discolouration underlying the rind proper. The initial rind results from the development of a greenish chlorite-like mineral in fractures and cleavages within the pyroxene. This oxidises to a brown material. The same material may form patches within the interstices between feldspar crystals, while limonite produces a reddish rind in some specimens (Edwards, 1955).

Comparison of magnesium rich bedrock and its weathering products from the Butlers Gorge area (Table 3.7) reveals only slight oxidation of the iron in the pyroxenes, and hydration in the initial rind. As weathering proceeds lime, magnesia and silica are leached out while alumina concentrates in the chlorite seams. Limonite in the clay that coats the bedrock at Butlers Gorge results from the conversion of pyroxene. Lime, soda and silica are reduced

	A	B	C	D	E	F	G	H
SiO_2	53.3	52.74	52.90	52.60	47.13	53.40	43.40	22.6
Al_2O_3	15.5	12.28	12.10	14.24	14.44	14.68	22.47	34.42
Fe_2O_3	0.8	0.24	1.04	0.17	6.85	1.04	13.63	15.20
FeO	8.3	9.35	8.24	10.71	8.90	8.36	0.64	0.51
MgO	6.8	10.35	10.35	7.51	8.95	7.50	1.29	0.93
CaO	11.1	11.33	11.25	11.70	6.61	10.40	2.56	tr.
Na_2O	1.7	1.36	1.71	1.10	0.60	1.77	2.08	0.14
K_2O	1.0	0.72	0.60	0.62	0.86	0.77	1.11	0.05
H_2O	-	-	-	-	-	0.10	3.88	4.32
H_2O+	0.7	0.27	0.46	0.30	3.70	1.21	7.84	20.19
MnO	0.1	0.31	0.28	0.25	0.35	0.15	0.04	0.08
TiO_2	0.6	0.56	0.56	0.60	1.03	0.63	1.50	0.90
P_2O_5	0.1	0.16	0.10	0.22	0.12	0.07	0.05	0.07
CO_2	-	-	-	-	-	tr.	0.15	0.21
Cr_2O_3	-	-	-	-	-	-	-	-
SO_3	-	0.04	0.03	0.07	tr.	0	0	0
S	-	-	-	0.04	tr.	-	-	-
Cl	-	0.20	0.53	0.21	0.11	tr.	tr.	tr.
	100.0	99.96	100.17	100.34	99.65	100.08	100.64	99.63

A - average Tasmanian chilled dolerite

B - fresh dolerite from Sullivans Quarry, Butlers Gorge

C - weathering rind from dolerite, Sullivans Quarry, Butlers Gorge

D - fresh dolerite, Butlers Gorge damsite

E - clayey coating on fresh dolerite, Butlers Gorge

F - fresh core of dolerite boulder, St. Leonards, central northern Tasmania - similar composition to Butlers Gorge

G - 2cm weathering rind from same boulder as F.

H - low grade doleritic bauxite surrounding boulder F.

(*Sources:* Edwards, 1942; McDougall, 1962)

Table 3.7 Composition of weathered and unweathered dolerite from two Tasmanian sites.

through the conversion of the feldspar to kaolin, while oxidation of the iron in the pyroxenes (probably due to feldspar leachates) increases the proportion of FeO (Edwards, 1955). Feldspar weathers rapidly to illite and thence by ion exchange along layer boundaries to montmorillonite (Birkeland, 1964). Dolerite may then weather slowly to kaolinite, providing there is complete removal of all alkalis, alkaline earths and iron, together with at least 50% of the silica (Loughnan and Golding, 1958). The divalent ions (Ca, Mg, Fe) inhibit desilicification by flocculating silicic acid, while Ca and Mg interfere with the kaolin lattice which does not contain them (Ollier, 1969).

As weathering proceeds still further silica is greatly reduced; brownish or brownish yellow fibrous limonite replaces almost all of the pyroxene and reprecipitation of Fe occurs on the original grain boundaries (Edwards, 1955). Gibbsite has been recorded in weathering rinds at Linda (Kiernan, 1980). Gibbsite is produced following the initial hydrolysis of feldspar to kaolinite, an interchange reaction and a second hydrolysis which creates silicic acid and aluminium hydroxide. (Douglas, 1977).

The expansion of weathering grains of feldspar and oxidised or hydrolised iron-rich minerals (Bloom, 1978) may mean that the progressive inward extension of weathering rinds is supplemented by some expansion outwards where confining pressures permit. This may underlie slight differences in rind thickness at some sites. It may also be responsible for the flaking of rinds and for the development of concentric shells in the rinds that surround deeply weathered clasts exposed by excavation at sites such as Linda and in the Wentworth Hills. The observation that surface rinds are generally thinner than those that develop in the subsurface environment (Colman and Pierce, 1981) may partly be the result of rind destruction in this manner.

Finally, it has been argued by some workers that secondary clay minerals in soils formed on tills are due at least in part to the weathering of primary minerals (Birkeland, 1964; Morrison, 1967). Derbyshire (1967) has argued that the presence of kaolinite and halloysite in some Tasmanian tills suggests that they are older than other tills that do not contain them. However, a number of studies have revealed an absence of secondary clay minerals in rinds on basalt and andesite clasts that are imbedded in argillic B horizons which revealed good clay mineral development (Crandell, 1963; Colman, 1982b). The dominant constituent in these rinds was allophane, an early stage in the transition from primary to secondary minerals (Fieldes, 1966; Swinedale, 1966; Colman, 1982a). Colman suggests from this that either the secondary minerals in the tills did not develop from in situ decomposition, or that the clay minerals develop more rapidly in the soil environment than they do in the rinds.

Although this thesis focusses upon dolerite weathering as an aid to dating the glaciations it would appear that the variation within dolerite itself is sufficient to bring about considerable variation in its response to weathering and erosion. Comparisons between clast weathering at different sites are confined to medium grained dolerites, in the hope that some co-variance between grain size and mineralogy will help to minimise the errors induced by both factors.

B. Environmental factors.

Environmental influences upon the response of a site to PDM include topography, organisms and climate. Once again the complexity of inter-action between each of these sub-factors means that the precise impact of each cannot be isolated. This discussion attempts to suggest at least the order of magnitude of the influence each may have upon PDM and safeguards which are required as a consequence.

1. Topography.

The topography of a potential sample site is significant in terms of drainage and the possible interruption of weathering by erosion or burial. The topography is important because the specific weathering process is a function of the balance between the loss of ions associated with hydrolysis and gains of ions from upslope (Icole, 1963) Clearly, moraines that are situated in a topographic position that favours runoff over their surface will be more rapidly dissected than those that lie in less erosion-prone situations. Topography may also influence sub-surface parameters through its effect on drainage of the regolith.

The presence of fresh striations on ice abraded bedrock need not indicate a comparatively recent event if the site has been subject to burial by subsequent deposits and later been exposed (Dyke, 1977, 1979). Colman and Pierce (1981) found that surface erosion had only a limited effect on sub-surface weathering rinds, but clearly the effect of surface erosion on surface rinds is likely to be considerable.Burial by subsequent deposits may be more of a problem. It would seem hazardous to assume a total cessation of weathering in the subjacent deposits in all cases.

In this thesis PDM data other than moraine slope angles have only been recorded from sites on or beneath moraine crests to minimise local differences brought about by variable topography. Sites obviously subject to erosion or burial have been avoided unless noted otherwise.

2. Organisms.

While some erosion or change of drainage patterns may be induced by fauna, vegetation generally exerts the most significant biological control over PDM. This role derives from the importance of vegetation in maintaining slope stability but at the same time conditioning the acidity of soil moisture and the nature of the soil micro-climate.

Observations in several Tasmanian mountain areas suggests that rinds developed under very acid conditions at shallow depth beneath peaty button-grass bogs appear to be slightly thicker, and certainly to develop more rapidly, than those beneath more open sclerophyll forests. Because the button-grass tends to develop in swampy low lying areas and the forest on better drained moraines this difference is perhaps more properly ascribed to topographic factors. An apparent tendency towards enhanced rind development also seem to exist beneath Nothofagus forests. Colman and Pierce (1981) found that rinds under developed grassland (or sage) in the western USA tended to be around 13% thinner than those beneath a forest cover. Rinds on surface clasts also tend to be thicker where a lichen cover is present (Jackson and Keller, 1970).

An attempt is made in this thesis to minimise local differences brought about through biological activity by avoiding sites which have obviously been disturbed by burrowing animals or fire; by excluding data on surface clasts in areas where spalling is apparent; and by restricting comparisons to areas bearing similar vegetation today. These steps will not solve the problem. Many of the deposits have been present in the landscape sufficiently long to have supported a wide variety of different vegetation types under different climates, while man and his fires are likely to have been present for at least twenty millenia (Kiernan, 1983a; Colhoun and Augustinus, 1984; Kiernan et al., 1983).

3. Climate.

The principal climatic parameters of concern here are temperature and precipitation. The Van't Hoff temperature rule states that for every 10^{o} C rise in temperature the velocity of a chemical reaction doubles (Jenny 1941). Temperature conditions whether physical or chemical processes will be dominant in the environment at any particular time. It will also condition the vegetation cover and the stability of slopes. Local variations in temperature brought about by aspect or other factors may promote marked differences in the degree of mechanical breakdown of surface clasts. A number of workers have inferred past temperature conditions from the degree of pedogenesis and the soil colours imparted by the resultant weathering products (Matsui, 1967; Colhoun and Goede, 1979). The great difficulty with such approaches is that similar characteristics to those imparted by warm climates may come about simply in response to more lengthy exposure under cooler conditions. The only routes out of this appear to lie with supporting stratigraphic or palynological evidence.

From his study of the weathering of alluvial terraces in the Pyrenees Icole (1963) concluded that the impact of possible Quaternary palaeoclimates upon palaeosols was not at all apparent and that such features were the result of prolonged rather than enhanced conditions. Icole found that the increase in weathering with age reflected progressive hydrolysis related to the total rain imput since deposition, with no other factor evident. He argued that the sole cause of weathering was the permanent leaching action of percolating rainwater. Colman and Pierce (1981) found that there was a two-fold increase in the thickness of subsurface weathering rinds along a 45 km long transect subject to a decrease in precipitation by a factor of three, and an increase in temperature of less than 1^{o} C.

Wind may exert some influence if it has a pronounced evaporative effect, but in the present study area is more likely to be significant through its impact upon snow deposition. Certain limestone karren forms suggest that snow may act like a temporary soil cover, albeit a biologically sterile one (Jennings, 1971). Evidence exists to indicate that weathering rind development is significantly enhanced beneath snowbanks (Colman and Pierce, 1981).

Climate is also important in determining which secondary clay minerals develop as weathering proceeds, as different clays tend to develop under specific conditions (Keller, 1964). Icole (1963) claims that because high leaching environments are inimicable to the preservation of gibbsite the clay mineral associated with the oldest deposits in the Pyrenees is kaolinite.

Within the present study area temperature and precipitation gradients are frequently more marked down the lengths of the valleys than they are between valleys. As a consequence correlation on PDM grounds between valleys probably poses less of a hazard than the differentiation of deposits down the length of individual valleys. All of the valleys commence on high ground and extend to much lower elevations. It is probable that rainfall diminishes downvalley and that the degree to which tills lower down are weathered will not be directly proportional to their age.

C. Temporal Factors.

Temporal factors include the duration of site exposure to the agents tending to modify it, the specific time periods at which the site is exposed to those influences and the rate at which it responds. All three considerations are important if any attempt is to be made to estimate the age of a landform or deposit on the basis of PDM criteria.

Few post-depositional modifications are likely to be effected at rates which are consistent through time. Price (1980) found that most of the major changes to the morphology of moraines occurred in the first 100 years of their existence when meltwater runoff was abundant and there was little binding vegetation. If the climatic trend is towards amelioration then the rate of modification is likely to diminish greatly over time. However, if there is subsequent re-exposure that trend through time will be modified.

Climatic conditions change over time and hence the processes that give rise to PDM also change. Chemical weathering does not come to a standstill with the advent of cold climatic conditions (Birkeland, 1984). The higher CO_2 saturation equilibria of cold waters and the high CO_2 concentration in intergranular voids in snow and firn may aid chemical weathering and partly underlie the thicker weathering rinds that occur beneath snowpatches. On the other hand the onset of cold climate sees a diminution of vegetative productivity and hence the availability of biogenic CO_2. Subglacial precipitation of $CaCO_3$ testifies to the previous aggressivity of the cold waters involved (Corbel, 1957; Smith, 1969; Hallet, 1976).

Numerous attempts have been made to derive equations that describe the rate of weathering rind development over time. Using these equations attempts have been made to date specific glacial deposits (Cernohouz and Solc, 1966; Colman and Pierce, 1981). Such a curve has been developed by Caine (1983) to describe rind development on Tasmanian dolerite clasts.

Laboratory and field studies both suggest that the rate of chemical weathering diminishes over time as the build-up of residues impede the evacuation of solutes (Colman, 1981). While this is in all probability quite true, the situation may be complicated by Quaternary climatic change. If chemical weathering is enhanced under wetter (or warmer) interglacial conditions then viewing degrees of till weathering from a Holocene perspective might well give a false impression that weathering diminishes over time because the latest deposits will be weathered to a disproportionate extent. Although this issue may perhaps be intractable at present it is an important one. Some Tasmanian rind thicknesses exceed 250 mm. and using deposits presently being laid down and those of the late Last Glacial Stage as the only available calibration points, none of the previously published rind development equations provides an acceptable or realistic estimate of age, nor is it easy to develop one that can have real temporal meaning other than using linear or logarithmic time scales

4. Concluding Remarks.

A wide range of variables condition the degree of PDM of landforms and deposits. Seldom is the exact process by which each operates known nor its rate of operation or its impact. As a result the use of PDM as a dating tool demands extreme caution. Little faith can be placed in marginal differences between sites. The differences need to be of a substantial magnitude, and be demonstrable on the basis of several parameters for differentiation to be convincing.

Correlations are likely to be most secure where they are based upon similar relative sequences in different valleys rather than simply a similarity of numerical values for the measured parameters. This suggestion poses its own dangers as the glacial record is seldom identically preserved in all valleys. The dangers inherent in correlation by simply counting backwards might be overcome in due course through the development of satisfactory equations to describe the rates of modification or at least a more substantial body of evidence upon which to base qualitative judgements. The better alternative lies with evidence of a stratigraphic, radiometric or organic nature which is not always available. Such evidence has not been forthcoming in the study area and as a consequence morphostratigraphic and PDM techniques provide the only means for dating the glacial deposits.

PART B

FIELD EVIDENCE

STRUCTURE OF PART B.

The following chapters present the evidence that has been gathered during fieldwork in, and adjacent to, the upper Derwent River Valley. For convenience of description the area has been divided into a number of subsidiary catchments (Figure B1).

Chapters four to nine describe the evidence from tributary catchments within the present Derwent River drainage system. This represents the core area of the study that has been examined in detail. Chapters ten to twelve document the evidence obtained during reconnaissance investigations in neighbouring valleys that previously formed part of the drainage system of the Derwent Glacier. These valleys have not been studied in the same detail as the core area.

A map of the glacial geomorphology of each area is presented with each chapter, together with a second map that depicts the extent of the ice during the principal glacial events that have been recognised. The key to the geomorphological maps is presented in Figure B2.

Lithostratigraphic units have been recognised on the basis of the morphostratigraphic and postdepositional characteristics of the landforms and deposits. Their distribution is also depicted on the geomorphological maps. The derivation of these units is explained at greater length in section C of this thesis (chapter thirteen).

Integrated maps of the ice limits during each of the glaciations are also presented in section C (chapter fourteen).

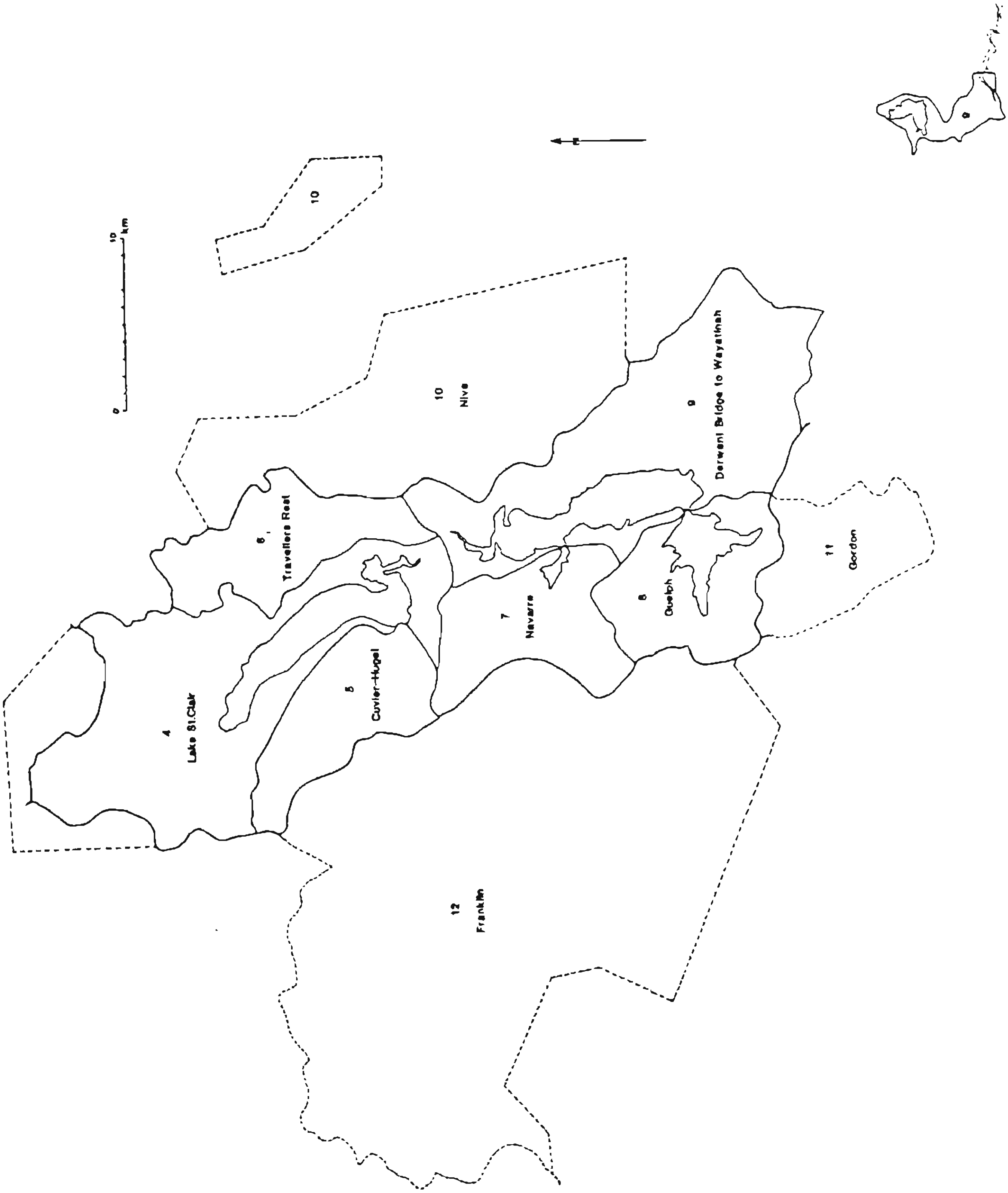


Figure B1 Location of tributary basins that form the core area of this study (solid boundaries) and adjacent basins in which reconnaisance investigations have been carried out (broken boundaries). The numbers indicate the chapter in which evidence from the basin is presented.

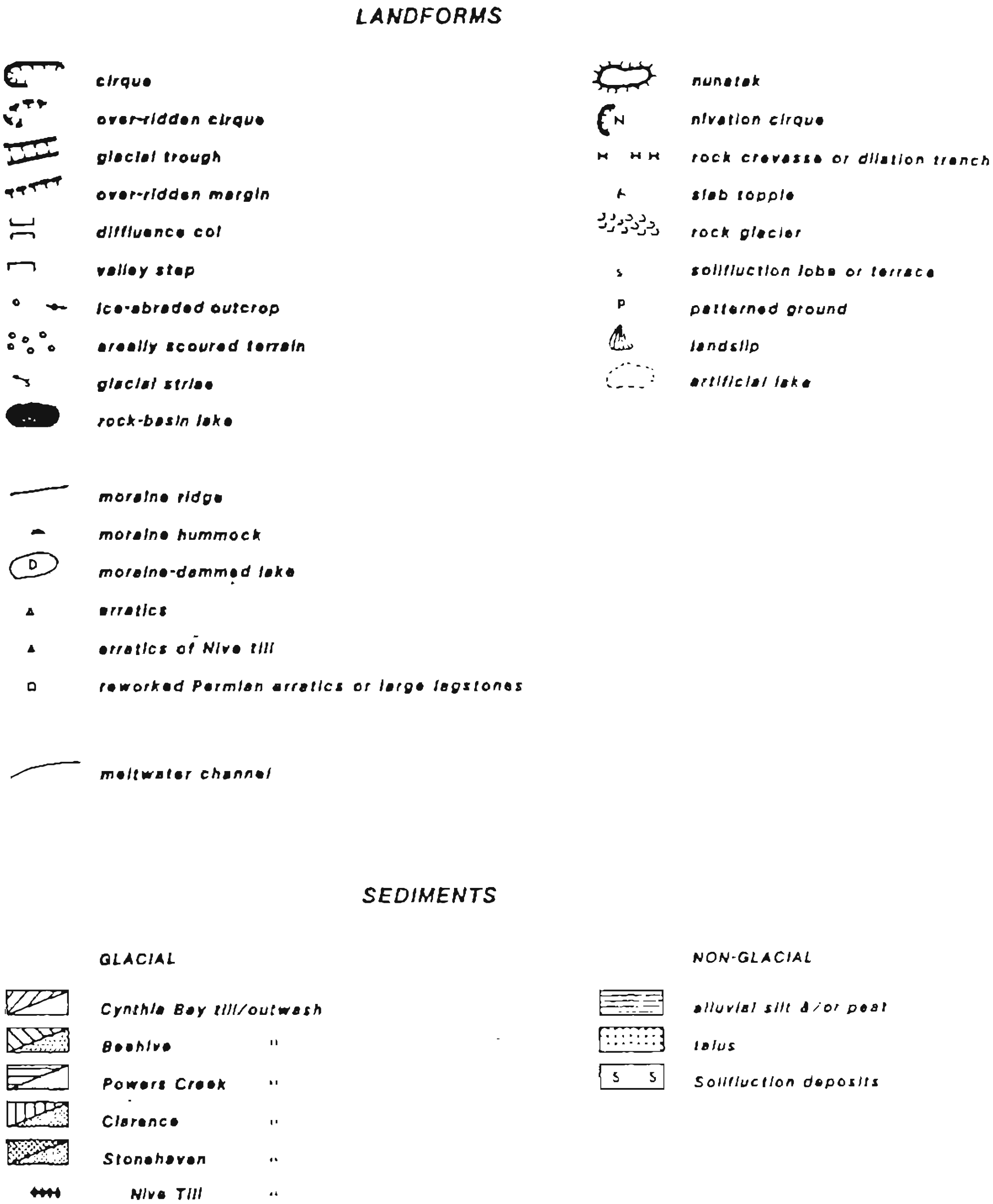


Figure B2 Key to the geomorphological maps (Figures 4.2, 5.2, 6.2, 7.2, 8.2, 9.2, 9.4, 10.2, 11.2, 12.2).

chapter four

THE LAKE ST. CLAIR AREA

"I will not dilate on the extreme beauty of the scenery as it might be considered out of place in an official Report .... the view from this point was beyond all description."

- George Frankland, 1835 Government Surveyor.

Lake St.Clair lies downstream of the convergence of the former Narcissus, Cephissus (Pine), Marion and Hamilton glaciers, all of which arose in the Du Cane Range. The lake marks the site at which the resulting Derwent Glacier was joined by ice that moved westwards from the Central Plateau, an ice abraded dolerite plain (Jennings and Ahmad, 1957) (Figure 4.1). This chapter explores the glacial geomorphology of the main catchment area of the Derwent Glacier, and examines the uppermost reaches of the adjacent Murchison, Mersey and Nive valleys in an effort to identify the glacial divides (Figure 4.2).

EROSIONAL MORPHOLOGY

A Glacial erosion:

A composite valley-head cirque 3 km. wide lies at the head of the Narcissus Valley. This cirque is floored by Permian sediments and its headwall is cut in Triassic sandstone that is overlain by columnar dolerite. The pronounced scallopping of the north-south aligned Geryon-Acropolis ridge that forms the western margin of the cirque indicates that this was a particularly important part of the Du Cane

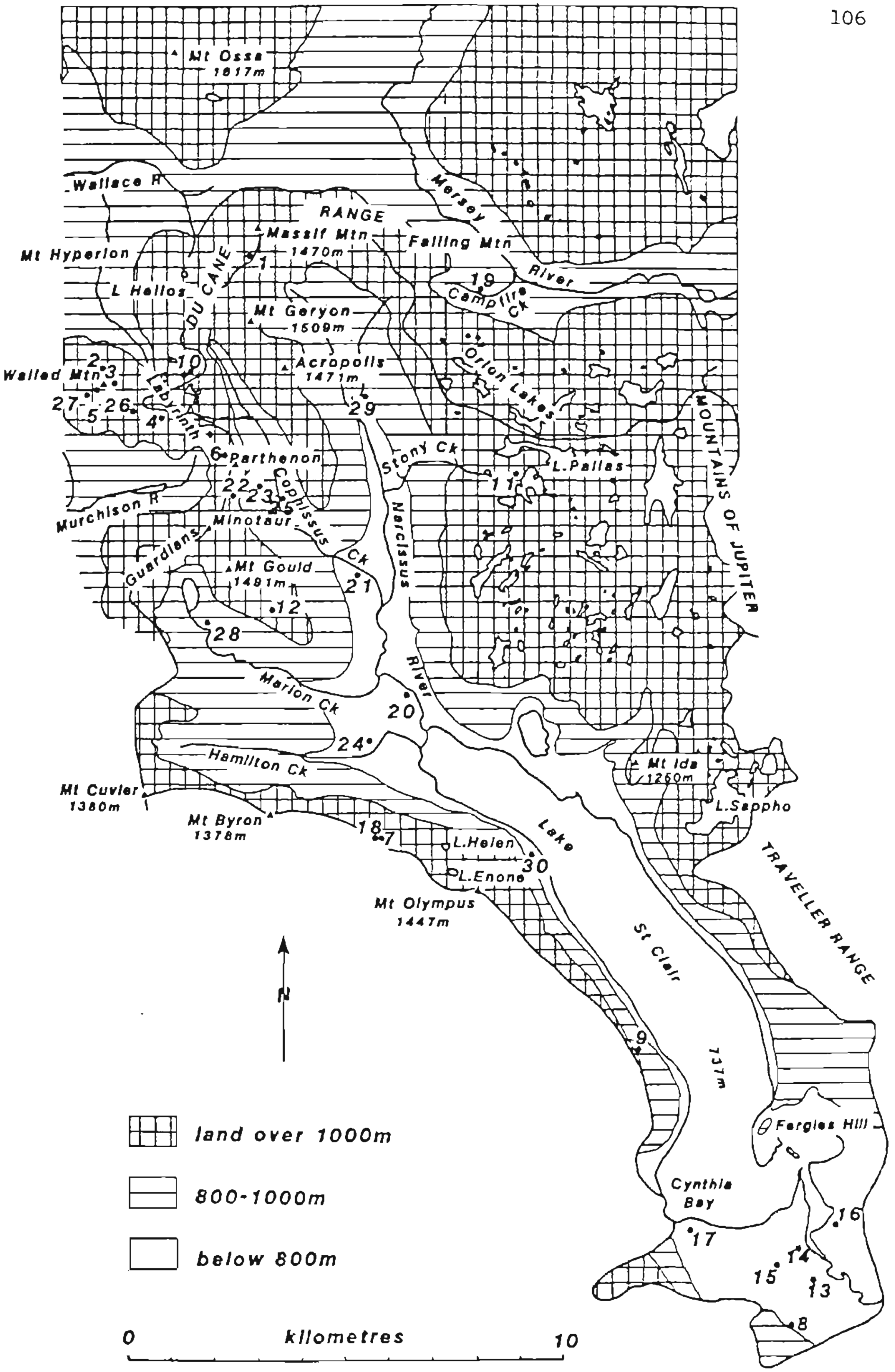


Figure 4.1 Locality map, Lake St. Clair area. Numbers refer to site locations in Tables 4.1 and 4.2.

Range snowfence (Figure 4.3).

Rock type has exerted a major control on cirque form. Whereas headward sapping of the Triassic sandstone beneath the dolerite columns has permitted the development of a steep 500 m headwall in the Narcissus Valley, the head of the Cephissus Valley cut only in massive dolerite is steep and mammilated rather than vertical. A steep arete occurs between the two valleys. The Marion and Hamilton cirques are also cut in Permo-Triassic rocks and again have vertical headwalls.

Each of these valley-head cirques has its floor at 900 - 950 m and a headwall 350 - 500 m high. Only in the Marion Cirque, where there is a small rock basin lake 13 m. deep (Derbyshire 1967), is there a clearly defined threshhold. Several factors have limited the degree of cirque development and conditioned the character of these cirques. These factors included the slowing of tributary ice flow by the main trunk glacier (Hamilton Cirque); over-riding of the headwall and later some downwind interception of snow (Cephissus Cirque); an unfavourably aligned snowfence (Marion Cirque); and an excessively low and broken valley head (Hamilton Cirque).

Structural geometry has played a major role in cirque location. Lake Helen (1190 m) and Lake Œnone (1210 m), the latter of which is a rock basin, occupy two prominent cirques on a lithological bench on the eastern flank of Mt. Olympus (Derbyshire, 1964, 1968) (Figure 4.4). A number of

smaller cirques are also present at about the same elevation on this mountain. Preglacial topography, in the form of benches on which snow could accumulate leeward of the snowfences, is likely to have been a major determinant of their location (Peterson 1969).

A series of cirques occurs on the slopes of the Du Cane Range at the head of the Murchison and Mersey valleys. They indicate that the range formed a major glacial divide (Figure 4.5). A shallow and totally enclosed cirque 200 m in diameter occurs on the summit of Massif Mountain. This has a distinct bedrock lip.

Many of the cirques have been over-ridden, particularly on the Labyrinth Plateau (Derbyshire, 1967). Cirques between Walled Mountain and Macs Mountain have been over-ridden by ice that was flowing to the east. Long Lake occupies a valley head cirque that was over-ridden by ice from the Labyrinth during a later phase of glaciation when the ice was less extensive and the Labyrinth ridge formed an important glacial divide between ice that flowed westwards into the Murchison Valley and eastwards into the Cephissus Valley. A cirque at 960 m on the southern end of Mt. Olympus has been over-ridden by ice that passed through a diffluence col from the Cuvier Valley. Valley head cirques that lie north and south of Mt. Ida have both been over-ridden by ice from the Central Plateau where almost all the cirque headwalls have also been over-ridden by ice which flowed generally southwards and southwestwards (Jennings and Ahmad, 1957).

Figure 4.5 Ice spilled northwards from the shallow Lake Helios cirque into the Wallace Valley. Glacial striae have been preserved beneath the surface of the lake but have been removed by chemical weathering elsewhere in the cirque.

Figure 4.6 Lake Pallas occupies one of many rock basins that have been eroded along structural lineaments in dolerite in the Traveller Range. This range forms the western margin of the Central Plateau and overlooks the Lake St. Clair trough.

Rock basin lakes are abundant on both the Central Plateau and the Labyrinth Plateau (Figure 4.6). In both these localities narrow elongate basins follow structural lineaments in the dolerite and are linked by a rectangular drainage system (Jennings and Ahmad, 1957). Some of the lakes in the Labyrinth are in excess of 10 m deep while others such as Lake Helios (Figure 4.5) are broad and shallow. Small rock basins formed transverse to the local flow of the ice occur 450 m above Long Lake on the western slopes of the Parthenon.

The most oustanding of the rock basin lakes is Lake St. Clair (igure 4.7). This piedmont lake occupies four rock basins on the floor of a glacial trough (Figure 4.8). Its maximum depth of 167 m (Derbyshire, 1971a) makes Lake St. Clair the deepest lake in Australia. Less than 50 m of this depth is due to impounding by end moraines. The deepest basin lies downstream from the entry point of the Ida glaciers from the Central Plateau. A fifth basin lies at the south-eastern extremity of Lake St. Clair and was developed jointly by the St. Clair Glacier and ice that descended from the Traveller Range. A rock bastion, Fergys Hill, has developed at this point. Sublacustrine benches along the lake margins may be of lithological origin. Derbyshire (1971a) suggests that 76.8% of the total lake volume of 1.69 km^3 is contained within the rock basins. The lake bed lies at a lower elevation than the bed of the Derwent River as far downstream as the limit of continuous drift 8 km south-east of Butlers Gorge.

Figure 4.7 Lake St. Clair and Mt. Olympus from Mt. Gould. Ice spilled into the lake along much of the plateau margin to the east. The prominent benches on Mt. Olympus are formed on Triassic sandstones. Ice from the Cephissus trough (left) over-rode the sandstone bench of the Gould Plateau (middleground). Alpine shrub vegetation re-established around the small tarn on the Gould Plateau before 7920 +- 250 BP (SUA 2080).

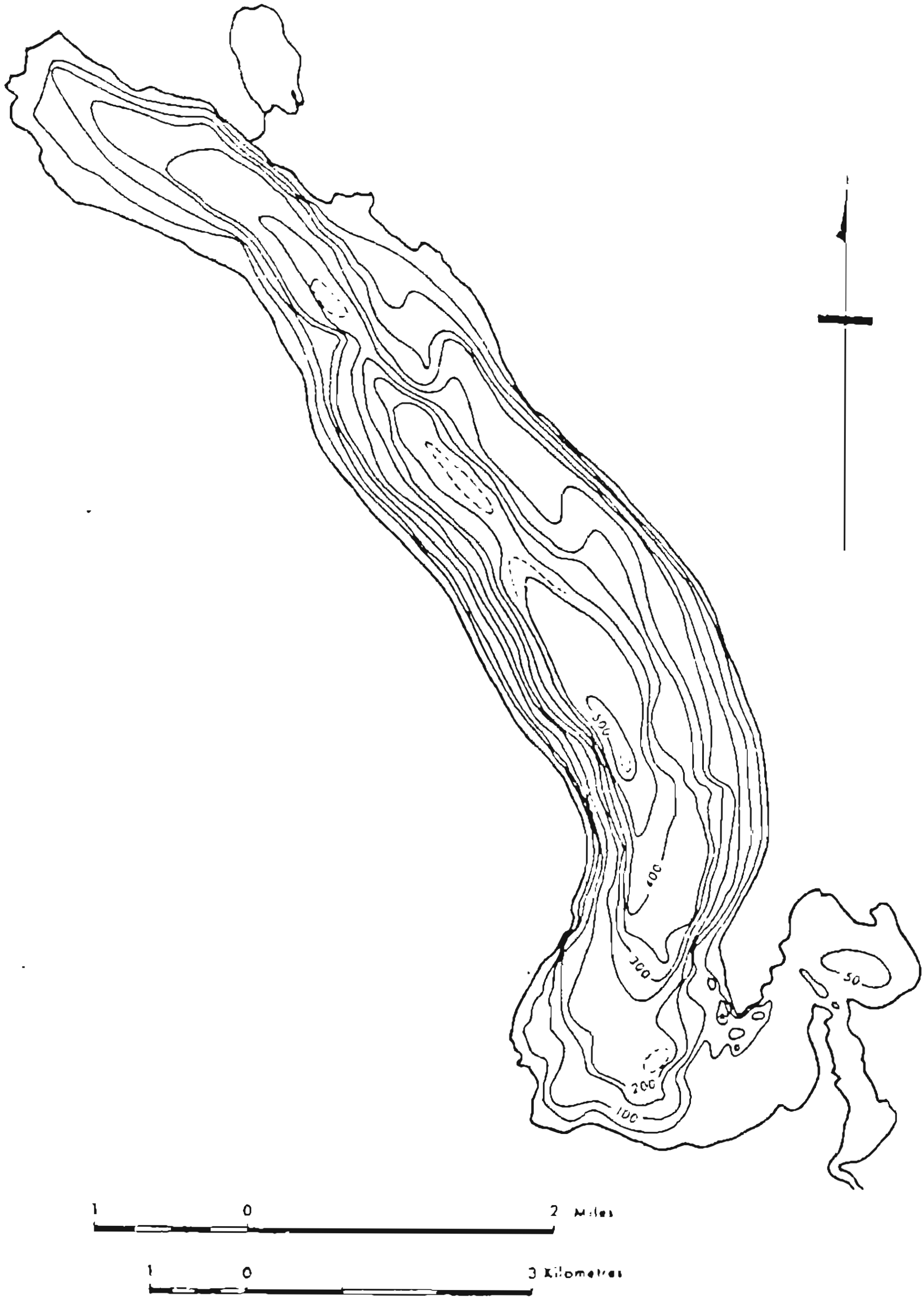


Figure 4.8 Bathymetric chart of Lake St. Clair (redrawn by Peterson & Missen, 1979, from bathymetry by Derbyshire, 1971a). The isobath interval is 50 ft. (15 m) and those shown dotted are at 225 ft., 265 ft., 425 ft. and 530 ft. (68 m , 80 , 130 m and 161 m).

The bedrock geometry and the consequent preglacial topography have clearly been responsible for determining the location of the glacial troughs. South of Mt. Ida the Lake St. Clair Trough is bounded by dolerite to the east and sub-horizontal Permo-Triassic rocks to the west. The base of the dolerite lies over 300 m above the lake on Mt. Olympus but descends below lake level at the foot of the Traveller Range. This suggests that the trough exploits a fault. The orientation of joint trends and structural lineaments in the dolerite of the Central Plateau (Jennings and Ahmad, 1957) is duplicated in the orientation of the tributary valleys upstream from Lake St. Clair which suggests that these valleys have been superimposed from the dolerite (Figure 2.2).

Rock type and valley gradient have conditioned the cross profile of the troughs. Broad U-shaped troughs are well developed in Permo-Triassic rocks. In strong contrast to this, narrower troughs cut in dolerite descend to the St. Clair Surface from the Central Plateau and at the head of the Cephissus Valley. The Lake St. Clair trough is strongly assymetric and has a steep western wall. The deepest part of the lake occupies a steep, straight trench which trends NNW - SSE close to the western shoreline. This is cut in Triassic rocks west of the St. Clair Fault. At this point a rock bastion, Fergys Hill, deflected the ice flow westward towards a re-entrant in Permo-Triassic rocks where the Cuvier Valley joins the St. Clair trough. The rock bastion probably owes its origin largely to interference with the flow of the Derwent Glacier by ice spilled over the margin of the Central Plateau. Therefore the most marked over-deepening of the trough may not reflect the amount of ice that descended from the Central Plateau so much as its gradient and energy, and its interference with the main ice flow southwards (cf. Derbyshire, 1971a).

Du Cane Gap is a transfluence col through which ice escaped westwards from the Mersey Valley into the head of the Narcissus trough (figure 4.3) (chapter thirteen). A further col occurs between Mt. Ossa and Mt. Massif. Ice flowed eastwards through this col into the Mersey Valley, which broadens appreciably downstream of this point. However, moraine ridges that are concave to the east were constructed east of the col during the most recent phase of glaciation and indicate that diffluence through the col did not occur at that time.. Cols between Mt.Olympus, Mt. Byron and Mt. Cuvier exhibit stoss and lee features which indicate that ice moved southwards through them from the Hamilton Valley into the Cuvier Valley. Similar evidence indicates that ice from the Cuvier moved into the St. Clair trough through a col on Mt. Olympus (figure 5.5). Derbyshire (1963) points out that ice would have flowed through the Olympus col in the opposite direction had it not been for the overdeepening of the St. Clair Trough.

Ice abraded bedrock indicates that the most recent glacial divide on the Central Plateau broadly coincided with the present divide between northward and southward flowing streams (Jennings and Ahmad, 1957). Some of the generally southward-moving plateau ice spilled into Lake St. Clair. Selective erosion along joints in the dolerite is common both here and on the Central Plateau. Stoss and lee surfaces on summits in the southern part of the Du Cane Range suggest that ice flowed from the upper Murchison into the Cephissus Valley (Figure 4.9). However, roches moutonées at lower levels north of the Parthenon suggest that at a later date ice flowed from the Labyrinth into the head of the Murchison Valley.

Smoothed bedrock on the summit of Walled Mountain suggests that it was over-ridden by ice when the glaciers were most extensive.. These summit rocks together with other abraded dolerite surfaces at 1250 m on the eastern ridge of Walled Mountain are exfoliating in slabs 15 - 20 cm thick (Figure 4.10). Glacial erosion may also have played a significant role in stripping overlying rocks from the lithological benches of Triassic sandstone developed on Mt. Gould and Mt. Olympus.

Glacial striae are generally poorly preserved due to the susceptibility of the rocks to chemical weathering. However, numerous striations on dolerite 1 m below the normal water suface in Lake Helios indicate that ice moved northwards into the valley of the Wallace River, a tributary of the Murchison. Striations trending towards 190^{o}- 205^{o} on sandstone bedrock just beneath the normal water surface in a small tarn of the Gould Plateau, together with plucking of nearby outcrops, indicates that the Cephissus Glacier flowed across the plateau from the north. The generally

Figure 4.9 The southern end of the Du Cane Range viewed from the Acropolis. All the summits except Mt. Gould (extreme left) have been over-ridden by ice that moved from the Murchison Valley (right) into the Cephissus Valley (left). A talus that was deposited in contact with the margin of the ice is just discernible at the break of slope below the summit block of Mt. Gould.

Figure 4.10 The summit rocks of Walled Mountain, west of the present Derwent - Murchison (Pieman) drainage divide, have been subject to considerable mechanical weathering since they were ice-smoothed. View eastwards towards the Labyrinth and Mt. Geryon.

southward flow of this ice across the Marion and Hamilton valleys towards the cols at the head of the Cuvier is confirmed by bedrock plucking and one lunate fracture. Lake Helios undoubtedly contained ice during the most recent glaciation, but the possibility that the Gould Plateau striae are earlier forms which have been exhumed must be kept open (cf. England et al., 1981).

B Glaciofluvial erosion:

Meltwater has played a major role in initiating or deepening many of the present stream channels in the Lake St. Clair area. Steep canyons incised in rock downstream of the higher cirques may have been initiated subglacially and have certainly been deepened as proglacial channels. Lateral meltwater flow along valley margins utilised lithological benches and also structural lineaments in the dolerite. As an example of the former, meltwater erosion has emphasised sandstone benches above Byron Gap and produced channels across the sandstone Gould Plateau. Deflection of meltwater along dolerite lineaments is evident above Long Lake, in the Labyrinth and on the Central Plateau. Some of the narrow elongate lake basins oriented parallel to the direction of ice flow on the Central Plateau may be the result of meltwater corassion by subglacial streams (Fairbridge 1968b).

The present outlet of the Derwent River from Lake St. Clair was probably initiated subglacially on the eastern margin of the St. Clair Glacier where confining ice pressures were lowest (Derbyshire, 1971a). It has subsequently developed as a proglacial channel cut through drift.

C Non-glacial erosion:

Because the major erosion surfaces of the Lake St. Clair area predate the glaciations (chapter 1) the suggestion that the principal valleys have been superimposed carries with it the corollary that the broad topography of the area is essentially the product of fluvial erosion.

Present day fluvial erosion is largely restricted to reworking of unconsolidated deposits. Some debris avalanching has occurred adjacent to a creek which descends steeply to the floor of Pine Valley south-east of the Parthenon, and other landslide scars and hollows are common in areas of steeper terrain on Mt. Olympus and elsewhere. Erosion runnels have developed on areas bared by nivation in the Narcissus Cirque and on Mt. Olympus. Small scale erosion and collapse of stream-banks is widespread, as exemplified along the course of Cephissus Creek in Pine Valley.

Solution pans have developed on the dolerite summit rocks on Walled Mountain (1411 m), and pans, runnels and sub-erratic pedestals are present on the Triassic sandstone benches of the Gould Plateau (1070 m) (Figure 4.11). The shallow and open cirque on the summit of Massif Mtn. forms an enclosed depression the size of a small football field. Drainage from this sinks into a pit 15 m in diameter and 8 m deep which appears to be focussed on a major joint intersection. While this may be due to very localised joint dilation there is little evidence of it. An alternative is that it is the product of solution processes acting upon autometasomatic alteration products in the joints (Hale, 1958; Spry and Hale, 1964).

The eastern ridge of the Acropolis and the summit of Mt. Gould consist of frost shattered dolerite columns that have not been over-ridden by ice at any time. Shattered rock that occurs on the Parthenon and Minotaur suggests that less than 50 m of ice passed through the adjacent cols during the most recent phase of glaciation.

Impressive free faces characterise much of the Lake St. Clair area, especially in the Du Cane Range and on Mt. Olympus. While many of these are the result of erosion by ice, some have resulted from the mechanical adjustment of rock materials to unloading or basal sapping. On a small scale such processes give rise to concentric shells such as are found on the summit rocks of Walled Mountain. Slopes formed on rocks with a high degree of structural anisotropy, particularly the vertically jointed dolerite, are prone to fail due to joint dilation once valley sides or cirque headwalls are steepened. The cambering and ultimate disintegration of columns due to this process is evident on Mt. Olympus (Clemes, 1925; Derbyshire, 1973) and elsewhere where it has effected considerable topographic change on a local scale. It is best described as slab-toppling (Caine, 1982).

The development of deep rock crevasses and dilation trenches which parallel the steepened slope is the first obvious sign of this process (Figure 15.12). Both are present above the Narcissus Cirque and on Mt. Olympus. The process is not confined to dolerite terrains as 'rock crevasses' and small gullies formed parallel to the valley side also occur at 1100m on the sandtone slopes east of the summit of Mt. Gould. The "Big Gun" in the Du Cane Range west of Massif Mtn. is a large leaning topple that has slipped downslope and is close to the point of outward collapse.

Toppling is most apparent in those areas where there has been glacial erosion of Permo-Triassic rocks that underlie the dolerite. Unloading is probably the principal cause but periglacial processes may play a role in dislodging columns (Banks, 1981; Derbyshire, 1973), while removal of alteration products in joints may also be significant in some cases (Hale, 1958). Some topples may be directly of glacial origin. Six metre high columns lean outwards at 30-50 degrees on the plateau margin just south of Mt. Hyperion and have clearly been pushed by a small plateau glacier that constructed a small blocky moraine against their foot. A similar process of pushing by a summit ice carapace may explain the presence of near horizontal columns that lie normal to the eastern margin of the Olympus Plateau. Here they are lodged between other columns that occur close to the edge and remain in a standing position (Figure 4.12).

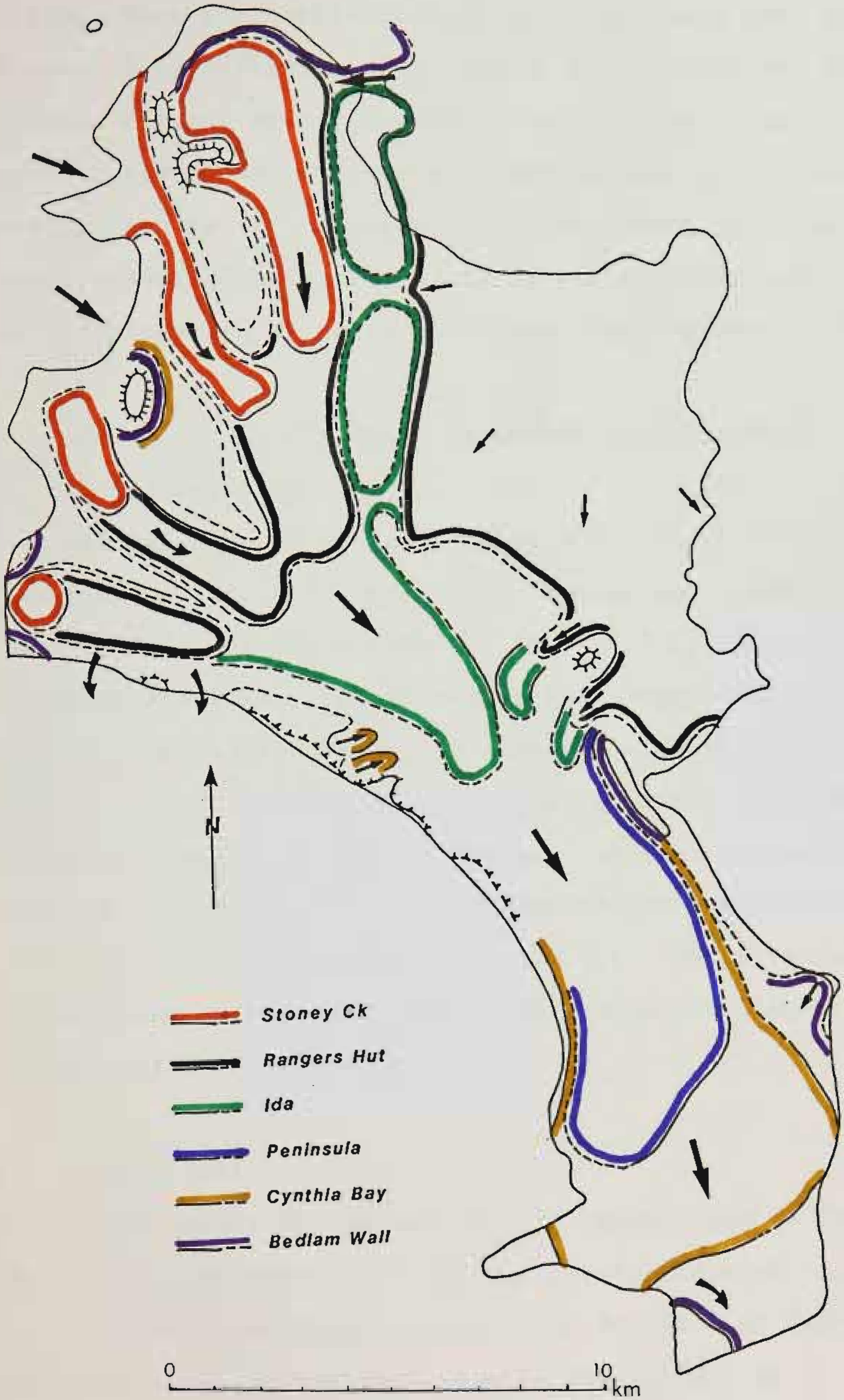
Occasional fresh scars on dolerite cliffs and freshly broken rocks at their foot suggest that only limited erosion by rockfall is presently occurring. A large fresh scar on

Figure 4.11 Pedestal developed on Triassic sandstone beneath a dolerite erratic on the Gould Plateau.

Figure 4.12 Toppled dolerite columns atop the eastern edge of the Mt. Olympus plateau lie normal to the plateau margin and are stacked horizontally between the buttresses.

overleaf

Figure 4.13 Principal ice limits in the Lake St. Clair area.


N
Stoney Ck
Rangers Hut
Ida
Peninsula
Cynthia Bay
Bedlam Wall
0
10
km

Falling Mountain delineates a recent collapse over an area of about 5 ha (Figure 4.3). Some steep gullies between dolerite columns on Mt. Geryon, Walled Mtn and elsewhere are probably the result of snow and rock avalanche. Snowpatch erosion occurs to a very limited extent in a number of areas, including the northern end of Mt. Olympus, and is far more effective on Triassic sandstones than on dolerite.

DEPOSITIONAL LANDFORMS AND SEDIMENTS

A Glacial Deposits

The Central Plateau adjacent to Lake St. Clair falls within a zone of predominantly erosional landforms (Jennings and Ahmad, 1957). Few moraines are to be found there, possibly because retreat was gradual rather than spasmodic (Davies, 1969). The steep walls of the St. Clair trough and the presence of Lake St. Clair mean that glacial deposits are accessible only in the tributary valleys and beyond the southern end of the lake. Lithological analyses of some of the tills are presented in Table 4.1. The principal end moraine complexes enable several phases in the glaciation to be identified (Figure 4.13).

(i) Bedlam Wall Moraine:

The steep flanks of the Bedlam Wall ridge south of Lake St. Clair have generally precluded the preservation of deposits but a lateral moraine extends along its foot at 830-840 m. At the northern end of the ridge it is overlain by 1 m of angular dolerite talus derived from a rock rib. The moraine can be traced southwards for nearly 2 km at which point it turns westwards against a rock spur and terminates. An outwash plain extends beyond the moraine and can be traced upvalley inside the moraine limit.

The Bedlam Wall Moraine marks a phase during which the Derwent Glacier terminated about 3 km. south of Lake St. Clair close to the site of the Derwent Bridge settlement. The moraine is well preserved but the till of which it is composed is weathered to a depth of at least 1.2 m. Small till knobs south of Mt Olympus; sandstone erratics at 1040 m. on the western side of the Parthenon; and a talus that formed in contact with an ice margin at 1260 m on the north-eastern shoulder of Mt. Gould (Figure 4.9) all appear to lie about 150 m above the most recent ice limits and may relate to this Bedlam Wall phase. The Bedlam Wall deposits generally lie inside the maximum ice limits suggested by the erosional topography but outside the succeeding Cynthia Bay moraines.

(ii) Cynthia Bay Moraines:

An impressive array of at least 25 end moraine ridges and latero-terminal moraines bounds the southern shoreline of Lake St. Clair. The southernmost of these moraines lies 1 km. from the lake shore. The moraines in this sequence are narrow and steep, none are more than 10 m high, and the spacing between them averages 50 m Some can be traced almost continuously for 3.6 km. They delineate the expanding foot of the Derwent piedmont glacier as it emerged from the confines of the St. Clair trough. Given the short duration of the Last Glaciation in Tasmania (chapter two) and the likelihood that retreat of the ice occurred very

Vicinity No.	16	14	[illegible]	17	21	22
		(> 15cm)	(< 15cm)			
%						
dolerite	96.0	92.0	31.0	58.0	50.0	91.0
quartz	4.0	6.0	68.0	12.0	5.0	4.5
Triassic sandstone	-	-	-	26.0	36.7	4.5
Permian & others	-	-	1.0	4.0	7.3	-

Table 4.1 Lithology of glacial deposits in the Lake St. Clair area. Site locations are indicated on figure 4.1.

rapidly, these end moraines and others of similar form that occur further upvalley appear to represent retreat cycles. The shortest likely cyclic event could be an annual fluctuation, but slightly longer periods could be involved.

His bathymetric survey of the southern end of the lake led Derbyshire (1971a) to argue that surficial deposits 4.6 m thick occurred on a bench south-east of the main terminal basin while deposits 1.8 - 5.4 m thick occurred in the southernmost basin of the main trough. His data suggest that the moraine barrage may be 40 - 45 m thick in the west, thinning to 24 m in the east. Recessional moraines occur to a depth of at least 30 m below the lake surface (Derbyshire 1971a).

The steep distal faces of the moraines suggest that they were formed by an active ice front. Steep end moraines of broadly equivalent age occur in some other locations, such as the northern cirque of Mt. Olympus where a steep moraine 3 - 5 m high has obliquely over-ridden an earlier cirque moraine 10 - 15 m high.

Basal till is exposed in many places around the shoreline of Cynthia Bay. It is generally tough and fissile with a greenish- grey coloured matrix. Individual clasts are facetted and striated clasts are common. Dolerite generally forms 70-80% of the clasts larger than small cobbles, with quartzite derived from the Permo-Triassic rocks accounting for the remainder (Table 4.1). Permo-Triassic rocks are more common in the pebble and granule grades.

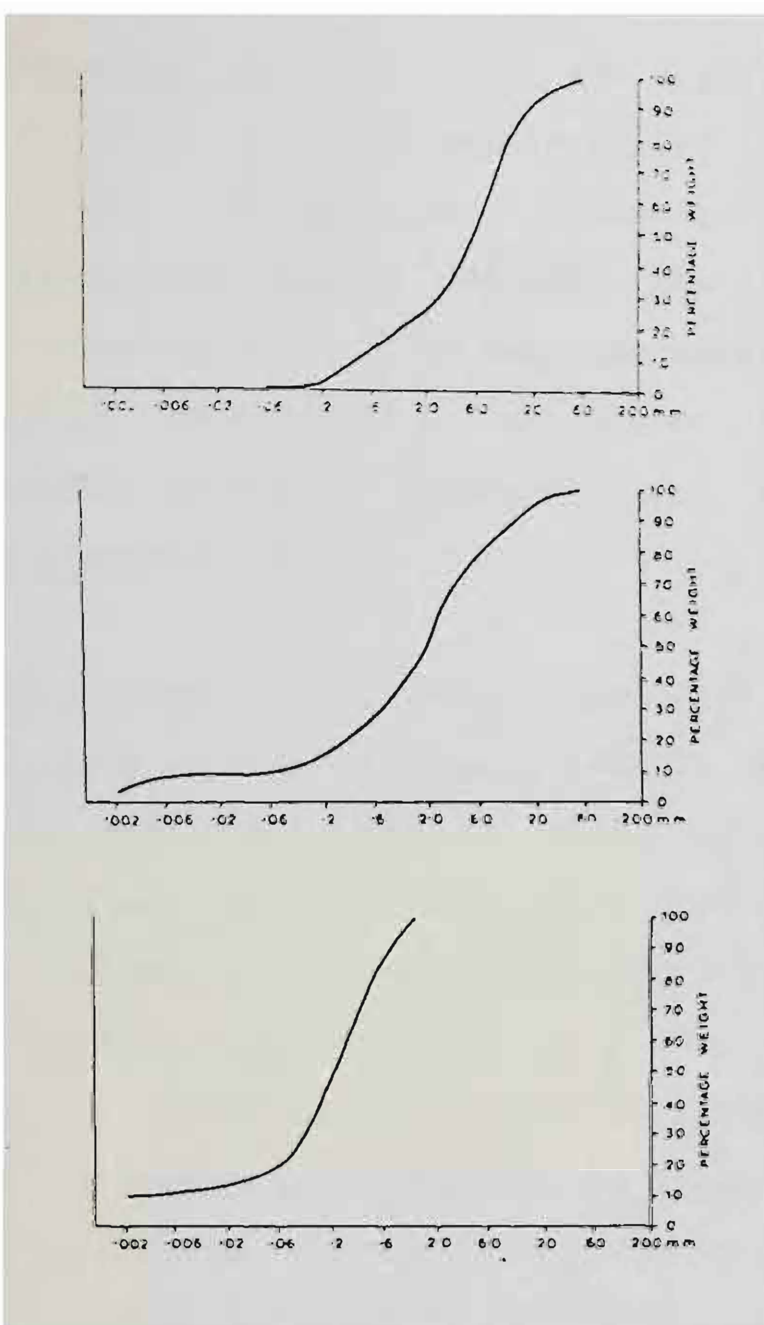


Figure 4.14 Grain-size analysis of some sediments from the Lake St. Clair area (after Derbyshire, 1967): (top) glaciofluvial sediment from near the Rangers Hut moraines in the lower Narcissus Valley; (middle) washed drift from the Cynthia Bay moraines; (bottom) glaciofluvial sediment from adjacent to the Cynthia Bay moraines.

Mechanical composition of some of the glacial and glaciofluvial deposits is presented in figure 4.14 (after Derbyshire, 1968a). Plagioclase feldspar, weathered dolerite mesostasis and quartz dominate the medium sand fraction, while X-ray diffraction of the clay suggests that it is virtually unweathered (Derbyshire, 1967) and represents deposits produced primarily by mechanical fracture during transport by ice.

Ice contact stratified drift has been exposed by quarrying and road construction in the end moraine (Figure 4.15) and one section has been described in detail by Derbyshire (1965). Beds of gravel, sand, massive silts and clay dip upstream, false bedding is undisturbed and balls of clay and sand that were probably transported in a frozen condition are also present. Slump structures are localised. These characteristics led Derbyshire to regard the deposit as the result of the slow melting of inert, debris-charged ice. He envisaged that extensive slumping and melting of the ice had been impeded by an insulating cover of drift and effective subsurface drainage through the adjacent outwash plain. The end moraine represents a steady state terminal position occupied for some time, the other moraines representing shorter time periods. The regular spacing of the moraines suggests consistent retreat of the ice front.

The regular spacing and size of the Cynthia Bay moraines recalls the form of De Geer moraines which form some distance behind an ice front that calves into water

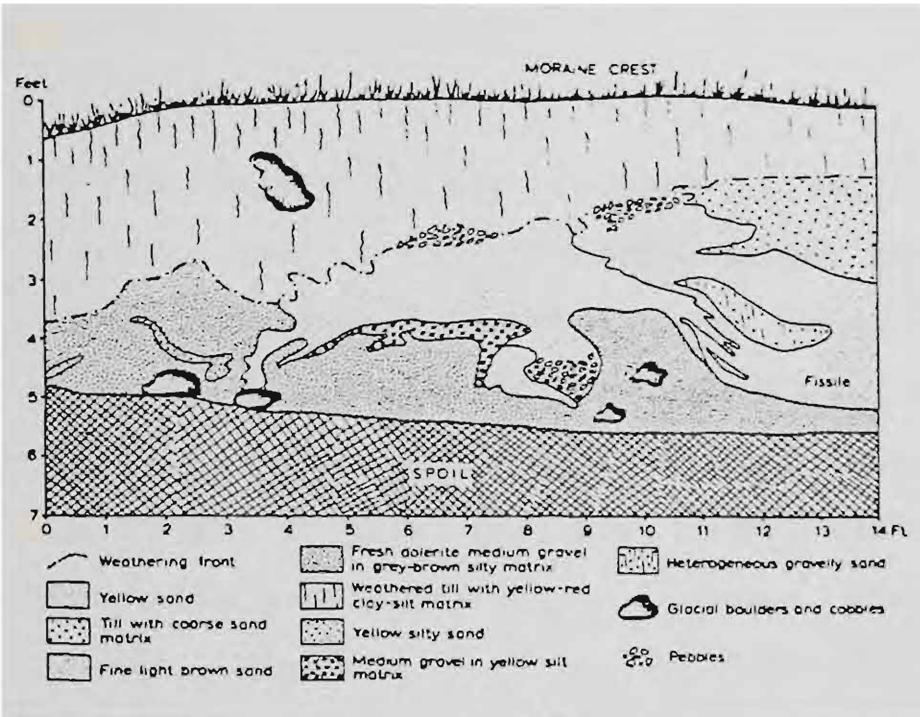


Figure 4.15 Section exposed in a road cutting through a recessional moraine of the Cynthia Bay phase, 400 m southeast of Cynthia Bay (after Derbyshire, 1967).

(Stromberg, 1965). However, while meltwater may previously have been impounded behind the Bedlam Wall Moraine further downstream a large lake would have been neccessary for the Cynthia Bay Moraines to have formed in this manner and there is no evidence for more than localised ponding of meltwater.

The development of the Cynthia Bay Moraines probably more closely parallels that of Thule-Baffin moraines (Embleton and King, 1975). Thule-Baffin moraines are cross valley landforms that have been described from the wasting margins of cold ice caps in the Thule area of Greenland (Weertman, 1961) and on Baffin Island (Andrews, 1963; Andrews and Smithson, 1966). The general absence of shear planes in sections exposed along the Cynthia Bay road and Derwent River suggests that the Cynthia Bay Moraines are not subglacial forms that have developed from the shearing of ice over the moraine crest in the general manner which has been proposed for Rogen moraines (Goldthwait, 1951; Derbyshire, 1965; Prest, 1968).

Weertman (1961) argues that Thule-Baffin moraines are formed where debris is brought to the ice surface in regelation ice that moves along flow lines rather than shear planes. This is consistent with the limited shearing at Cynthia Bay and is essentially the position adopted by Derbyshire (1965). The Cynthia Bay Moraines lack the alluvial fans that commonly form on the distal face of Thule-Baffin moraines (Goldthwait, 1951). The reason for this probably lies in differences in the thermal regime of the glaciers.

Thule-Baffin moraines are formed on the edge of cold ice sheets where there may only be a limited amount of basal debris entrained. Active temperate ice is likely to entrain more basal debris than will cold ice and upon reaching the surface this will form a thicker debris cover and will more effectively insulate the ice, thereby retarding melting and flowage.

Bouldery ablation till with angular dolerite clasts 2-3 m long in a loose and slightly oxidised matrix caps many of the moraines. Boulders which protrude through some of the intermorainal outwash probably represent supraglacial ablation deposits or lag boulders that could not be removed by meltwater.

The coarser grades of the outwash are dominated by dolerite, the finer grades by quartz. The dolerite cobbles are moderately well rounded and of moderate to high sphericity, but angularity increases in the finer grades, presumably due to water cushioning (Hatch et al., 1971; Blatt et al., 1972). The larger clasts occur in a loose textured matrix which is commonly bright yellowish brown (10 YR 7/6) in colour. Sections cut by the Derwent River reveal foreset beds, truncated bedding and current bedding indicative of seasonally torrential meltwater flow. Interbedded fine gravel, sand and silt lenses are common. Glaciolacustrine deposits consist entirely of massive clays either in the ice contact stratified drift or within intermorainal swales. No rhythymically bedded clays have been located.

The individual clasts in the moraines are little weathered and the soil profile can be best characterised as of A Cox Cu type (Birkeland, 1984). Nowhere do more than 30 cm of slope deposits overlie the Cynthia Bay till. X-ray diffraction led Derbyshire (1967) to suggest that some kaolinite was present in the till matrix but the peak on his diffractograms is very close to background and has not been duplicated by analyses during this study. All these characteristics suggest that the Cynthia Bay till is considerably younger than the Bedlam Wall till.

(iii) Peninsula Moraines:

The rock bastion of Fergys Hill that extends westwards into the southern end of Lake St. Clair is capped by a latero-terminal moraine that can be traced along the eastern side of the trough from 740 - 820 m A minor recurved segment of lateral moraine from 760 - 840 m. on the western side of the lake may correspond with the Peninsula Moraine, and, if so, indicates that the surface of the ice sloped eastwards at 25 - 30 m/km. Such assymetry of the ice surface would have been promoted by shading and snowfence effects, but it also suggests that little ice was entering the trough from the Central Plateau at this time.

The bathymetric chart suggests that the ice retreated initially to the western margin of the Derwent Basin and then to the southern margin of the deepest rock basin where 7 it became grounded.

(iv) Ida Moraines:

Discontinuous lateral moraines descend to Lake St. Clair from 800 m beneath Mt. Ida. These document the most recent occasion during which the Ida glaciers descended into the Lake St. Clair trough. The moraines that were formed on the southern side of the North Ida Glacier are sharply recurved southwards whereas lateral moraines of the Derwent Glacier occuring on its northern side appear to have been undisturbed by the Ida ice. This suggests that the largest of the glaciers to descend from the Central Plateau to Lake St. Clair was deflected southwards by the Derwent ice. This indicates that the valley glacier was not highly dependant upon the input of ice from the plateau.

Retreat of the main St. Clair glacier from the Ida limit must have occurred at a fairly uniform rate as there is little evidence of end moraines for 4 km upstream, after which many end moraines occur in two principal groups.

(v) Hamilton Valley moraines

A poorly developed interlobate complex of low moraines with an amplitude of less than 5 m occurs at the lower end of the Hamilton Valley. This complex indicates that the Hamilton Glacier was the next major tributary to withdraw from the main trunk glacier. Till with few definate moraine ridges occurs on the floor of the Hamilton Valley. Hummocky drift occurs in a few localities and suggests that the ice flow died. There is no evidence to indicate continued confluence of the Marion and Hamilton ice after the Hamilton and Narcissus glaciers had separated.

These moraines have an amplitude of less than 6 m. A fan of outwash occurs at the northern end of Lake St. Clair.

(vi) Rangers Hut moraines

An interlobate complex of low moraines occurs near the huts at the northern end of Lake St. Clair. A fan of outwash extends into the lake, and glaciofluvial deposits at least 4 m. thick also occur a short distance upstream from the huts. Small end moraine ridges and a limited area of hummocky moraines occur at the mouth of the Narcissus Valley. A sharp, sinuous ridge in the axis of the valley 2.5 km north of Lake St. Clair has the form of a small esker. The western part of the Rangers Hut complex contains moraines that are concave towards the Marion and Cephissus valleys. Hence, this interlobate complex documents the separation of the three glaciers.

Withdrawal of the Derwent Glacier to the Rangers Hut limit was at least partly a function of the continued shrinkage of the Central Plateau ice cap, which led to the decline of the Mersey Glacier, and the ending of diffluent flow from the Mersey through Du Cane Gap. A series of at least a dozen small end moraines was constructed at 1160 - 920 m east of Du Cane Gap as the Mersey Glacier retreated down the Campfire Creek Valley.

(vii) Marion Valley Moraines

Hummocky moraines occur in the lower and middle reaches of the Marion Valley upstream of the Rangers Hut Moraines. Nine end moraine ridges are present in the upper part of the valley, the innermost of which contains semi-circular or irregular shaped depressions on the ice proximal face. These are interpret ed as kettle holes that were formed during final decay of the ice. The greater development of end moraine ridges in the Marion Valley compared to the Hamilton Valley indicates that the Marion Glacier remained more vigorous than the Hailton Glacier during the final stages of glaciation. It would appear from this that the Guardians formed a far more effective snowfence and provided more adequate shading than did the ridges at the head of the Hamilton Valley.

(viii) Cephissus Valley Moraines

The Cephissus Glacier constructed numerous end moraines on rock ribs between the mouth of the valley and the Pine Valley Plains. The innermost of these again contains kettle holes. The Cephissus ice then died, possibly due to the loss of any significant ice imput from the Labyrinth. Two intermorainal outwash surfaces are overlain by 2 m of massive gleyed clays. These clays are due to meltwater ponding (Jennings, 1959). Intervening hummocky moraines indicate that deposition of the clays occurred in two stages (Derbyshire, 1963).

(vii) Stoney Creek Moraines

The descent of Stoney Creek from the Traveller Range to the Narcissus River is deflected southwards within a series of latero-terminal moraines with crests at 760-780 m. The Narcissus and Cephissus glaciers were no longer confluent at the time these moraines were constructed.

Retreat of the ice in the Narcissus Valley appears to have been fundamentally different in character to the pattern in the tributary valleys. Fluted drift with a mean relief of 2-4 m is abundant in the Narcissus Valley where it has been attributed to late glacial meltwater streams (Macleod et al., 1961); to a large glacier moving with vigorous rotational slip (Derbyshire, 1963); and to the presence of very thick ice (Peterson, 1969). Some apparent flutes can be traced back to small moraines that extend from the low divides between scallopped hollows that are inset into the main cirque. This is consistent with the proposition that the development of flutes results from a process similar to that involved in the development of crag and tail landforms. This view asserts that flute development is related to the migration of plastic till into cavities in the ice formed in the lee of obstructions at the glacier sole where low confining ice pressures apply (Galloway, 1956; Boulton, 1971). At least the largest of the Narcissus flutes have formed in this manner.

Unequivocal fluted drift occurs along about 4.5 km of the valley floor from 1040 - 850 m and indicates that the ice remained active even during its final retreat. This fluted drift contrasts with the moraines which contain kettle holes and the hummocky moraines which indicate that the ice flow died in the tributary valleys. The fluting originates in the head of the Narcissus Valley and does not extend towards Du Cane Gap. This suggests that the high Geryon-Acropolis snowfence rather than continued diffluence through Du Cane Gap lies behind the comparative longevity of the Narcissus Glacier

A further small area of fluted drift forms the southern shoreline of a lake several hundred metres north-east of Lake Sappho on the Central Plateau. This fluted drift indicates that the plateau ice cap also remained active during its retreat.

B Non-glacial deposits

(i) Slope deposits

Solifluction deposits mantle many of the slopes outside the Cynthia Bay ice limits but are scarce within those limits. They reach a depth of 2 m on the upper slopes of the Bedlam Wall ridge. These deposits consist of fragments of local bedrock or clasts up to 2m or more in size that have been reworked from till and occur in a silty clay matrix. They have moved over slopes of as little as 8° .

Slab toppling has produced extensive block slopes beneath many of the free faces. Such deposits are best developed on Mt. Olympus but also occur at Walled Mountain and on the western flank of Mt.Hyperion. They are most extensive beneath re-entrants in the clifflines. On the northeastern flank of Mt. Olympus large imbricate and joint-bounded blocks can be visually matched with one another. Block accumulations which appear to be the result of slab toppling are moderately common in steep terrain which has been eroded by glaciers.

Toppling is uncommon within the limits of the Bedlam Wall phase and is virtually absent within the Cynthia Bay ice limits. However, tongues of very large joint-bounded blocks extend steeply to the valley floor from broad block accumulations high on both sides of the upper Cephissus trough. These blocks are believed to represent topples that collapsed forward as the ice withdrew, together with some talus that was deposited against the ice margin. Another talus that was deposited in contact with the ice occurs at 1260 m on Mt. Gould.

Rockfall talus has accumulated in many localities. Mechanically shattered bedrock forms a mantle up to 1 m thick on the northern end of the Bedlam Wall ridge and is probably the result of frost wedging during the Cynthia Bay phase. Processes of rockfall from high dolerite clifflines commonly involve toppling to some extent, but smaller amounts of material are moved in each event, fragments of columns are widely dispersed and the prismatic form of the dolerite blocks is seldom preserved. Accumulations of this kind beneath the cliffs of Mt. Olympus lie at a gradient of 20-30 degrees. There is little freshly broken rock on the surface of the Mt. Olympus apron. Rockfall talus and some joint-bounded blocks dominate the lateral moraine north of Lake Helen but few rockfall deposits are present within the ice limit it defines (Figure 4.5).

The toppled blocks on the eastern wall of the Cephissus trough are overlapped on the eastern side of the trough by tongues of much smaller calibre rockfall and avalanche talus. Similar deposits also occur beneath the cliffs of Mt. Geryon in the Narcissus Cirque. A large rockfall deposit on the south-western slopes of Falling Mountain is associated with abrupt topple-like failure of a glacially oversteepened slope within the ice limits of the Cynthia Bay phase (Figure 4.3). The base of the collapse rests upon sediments underlying the dolerite which suggests that failure of the sediments has brought about cambering (Caine, 1982).

(ii) Alluvial, peat and dune deposits.

Up to 1.5 m of alluvial silt that contains fragments of charcoal has accumulated as overbank and levee deposits on the Pine Valley plains and forms a small delta where the Narcissus River discharges into the northern end of Lake St. Clair. The silts reflect a change from an environment in which vigorous meltwater streams carried mechanically weathered materials from an unstable landscape to an environment in which vegetation was present and the products of predominantly chemical weathering were transported by a less vigorous streamflow (Tricart and Cailleux, 1972). Organic-rich silt up to 1.8 m thick occurs in several rock basins in the Labyrinth, west of the lakes in the Helen and Enone cirques on Mt. Olympus. A core from the smaller of the tarns at 1070 m on the Gould Plateau revealed 85 cm of organic-rich silt that overlying minerogenic silt. Fragments of <u>Athrotaxis selaginoides</u> at the base of the organic-rich silt indicate that the ice had withdrawn and the landscape had stabilised sufficient for vegetation to develop prior to its deposition.

Minor paludal deposits have accumulated with fibrous peats on the button-grass plains. Glacial sands have been worked into beaches around parts of the shoreline of Lake St. Clair, most notably at the Frankland Beaches at the southern end of the lake. Dunes up to 4 m high have been constructed by aeolian action behind the southern beaches in response to northwesterly winds down the length of the lake during the Holocene.

DATING AND DISCUSSION

Radiocarbon assay of charcoal fragments obtained from 20 cm. above the base of 1.3 m of deltaic silt that overlies glaciofluvial sediments near the mouth of the Narcissus River indicates that the silt began to accumulate prior to 7650 +- 250 BP (SUA 2079). The fragments of *Athrotaxis selaginoides* at the base of the organic silt in the Gould Plateau tarn have been radiocarbon assayed to 7920 +- 250 BP (SUA 2080). Evidence from elsewhere in the state suggests that deglaciation was complete well prior to this time (Macphail and Peterson 1979).

The rockfall deposits at Falling Mountain lie within the ice limits of the Cynthia Bay phase and in some areas the ground is freshly disturbed. This indicates that the site remains unstable. Otherwise, apart from small fragments which have accumulated due to minor rock and snow avalanche activity around Mt. Geryon there is little evidence of sigificant rockfall today. Rockfall deposits are scarce within the Cynthia Bay ice limits, however a considerable amount of rockfall material has been reworked into moraines on Mt. Olympus. This suggests that the main phase of rockfall activity predates deglaciation.

The postdepositional data suggests that the glacial deposits as far downstream as the Cynthia Bay moraines do not differ significantly in age. All have highly immature A Cox Cu soil profiles (Birkeland, 1984). The average thickness of dolerite weathering rinds is less than 2 mm (table 4.2). The solifluction deposits occur almost exclusively outside the Cynthia Bay ice limits and are weathered to a comparable degree. This suggests that they are approximately equivalent in age.

Moraines that abut talus 1 km. south of Lake St. Clair and in the Lake Helen cirque delineate the maximum extent of ice during this glaciation. The Derwent Glacier extended 27 km from its headwall in the Du Cane Range. Fresh rock crevasses and ice abraded surfaces 300m. above the valley floor on the Gould Plateau indicate that the ice was 300 m. thick west of this point. Diffluent ice from the Derwent Glacier passed through two cols north-west of Mt. Olympus into the head of the Cuvier Valley until the end of the Peninsula phase. This ice surplus existed partly because the Derwent Glacier was supplemented by a difffluent lobe of the Mersey Glacier which flowed through Du Cane Gap at the head of the Narcissus Valley.

The regular spacing and limited evidence for shearing suggests that the moraines upstream of the Cynthia Bay

moraine[a] / vicinity no. / material[b]	BW 1 Bd	BW 2 Bd	BW 3 Bd	BW 4 Bd	BW 5 T	BW 6 T	BW 7 T	BW 8 T	BW 9 T	CB 10 Sd	CB 11 Sd	CB 12 Bss	CB 13 D	CB 14 T	CB 15 T
A. MORPHOSTRATIGRAPHY															
relative position	outside CB limit	margin CB	outside CB	outside CB	outside CB	above CB limit	over-ridden moraine	outside CB	above CB	inside 6	inside CB	inside CB	inside 6	=13	inside 14
crest/site elevation (m)	1350	1150	1410	1250	1280	1080	1280	830	970	1100	1210	1070	730	740	740
max. depth of local burial (by) (m)[c]	-	-	-	-	60.4	50.3	-	33.5	-	-	-	-	-	-	-
B. SUBSURFACE CLASTS															
dolerite rinds - max. (mm)	-	-	-	-	5.0	6.9	-	7.4	3.1	-	-	-	3.1	3.0	2.8
- min. (mm)	-	-	-	-	3.5	0.5	-	0.4	0.6	-	-	-	0.1	0.9	1.0
- mean (mm)	-	-	-	-	3.55	2.36	-	1.85	2.81	-	-	-	0.73	1.31	1.5
- SD (mm)	-	-	-	-	1.9	1.70	-	1.46	0.91	-	-	-	0.69	0.41	0.46
- hardness (1-5)	-	-	-	-	1-2	1-2	-	1	1-2	-	-	-	1	1	1
coherence of argillitic clasts	-	-	-	-	-	-	-	-	2	-	-	-	-	1	1
clast surface colours	-	-	-	-	7.5YR	7.5YR	-	7.5YR	7.5YR	-	-	-	10YR	7.5YR	10YR-7.5YR
C. SUBSURFACE MATRIX															
profile type	-	-	-	-	A Bt	A Bt	-	A Bt	A Bt	-	-	-	A Cox Cu	A Cox Cu	A Cox Cu
colour of B horizon	-	-	-	-	5YR	5YR	-	7.5YR-5YR	7.5YR	-	-	-	10YR	10YR-7.5YR	10YR
depth of B horizon (cm)	-	-	-	-	>100	>100	-	>100	>100	-	-	-	-	-	-
depth to lowest Cox (cm)	-	-	-	-	>100	>100	-	>100	>100	-	-	-	>150	>100	>80
pH at ~1m	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
organic at ~1m	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
clay size fraction mineralogy	-	-	-	-	-	-	-	Kaolinite[d] Gibbsite Quartz	-	-	-	-	-	montmor[d] illonite quartz	-
clay film development	-	-	-	-	2	1	-	2	2	-	-	-	0	1	1
coarse matrix: % >1cm	-	-	-	-	-	-	-	62	-	-	-	-	-	50	87
% 1-0.5cm	-	-	-	-	-	-	-	22	-	-	-	-	-	20	8
% <0.5cm	-	-	-	-	-	-	-	26	-	-	-	-	-	30	5
fine matrix: % sand	-	-	-	-	-	-	-	34.7	-	-	-	-	-	-	-
% silt	-	-	-	-	-	-	-	49.6	-	-	-	-	-	-	-
% clay	-	-	-	-	-	-	-	15.7	-	-	-	-	-	-	-
basal till lithification (1-5)	-	-	-	-	-	-	-	-	-	-	-	-	-	1	1
D. SURFACE MATERIALS															
no. of surface boulders/m²	-	-	-	-	-	-	-	-	-	-	-	-	-	3.8	-
% split clasts	-	-	-	-	-	-	-	-	30	-	-	-	-	1	-
surface pit depth (dolerite) - mean (mm)	-	-	-	-	-	-	-	≤2.0	-	-	-	-	-	<1.0	-
- max. (mm)	-	-	-	-	-	-	-	4.0	-	-	-	-	-	3.0	-
E. DEPOSITIONAL LANDFORMS															
moraine crest width (m)	-	-	-	-	-	-	-	-	8	-	-	-	-	9	17
proximal slope (°)	-	-	-	-	-	-	30	-	-	-	-	-	-	18	33
distal slope (°)	-	-	-	-	-	-	30	-	-	-	-	-	-	20	15
fluvial dissection (1-5)	-	-	-	-	-	1	0	1-2	2-2	-	-	-	-	0	0
F. BEDROCK LANDFORMS															
width of joint opening (cm)	-	-	-	-	-	-	-	-	-	0.5	0.3	-	-	-	-
depth of joint opening (cm)	-	-	30	25	-	-	-	-	-	-	0.8	-	-	-	-
width of solution pans (cm)	-	-	300-400	-	-	-	-	-	-	-	-	100	-	-	-
depth of solution pans (cm)	-	-	30	-	-	-	-	-	-	-	-	30	-	-	-
height of pedestals (cm)	-	-	-	-	-	-	-	-	-	-	-	5	-	-	-
presence of slab topples (1-5)	3	3	0	0	-	-	-	-	-	-	-	-	-	-	-
exfoliation slab thickness (cm)	-	-	15-20	15-20	-	-	-	-	-	-	-	-	-	-	-

moraine[a] / vicinity no. / material[b]	CB 16 T	CB 17 T	CS 18 T	I 19 T	RH 20 O	RH 21 T	RH 22 T	RH 23 T	RH 24 T	SC 25 O	SC 26 T	SC 27 T	SC 28 T	SC 29 T	- 30 T
A. MORPHOSTRATIGRAPHY															
relative position	=13	inside 15	overriding moraine	inside CB	inside 17	inside 20	inside 21	inside 22	inside 13	inside 22	inside CB	inside CB	inside RH	inside 20	inside CB
crest/site elevation (m)	760	735	1280	1200	760	785	1200	740	760	840	1135	1280	850	900	900
max. depth of local burial (by) (m)[c]	-	-	-	-	-	-	50.3	-	-	-	-	-	-	-	-
B. SUBSURFACE CLASTS															
dolerite rinds - max. (mm)	2.8	-	-	4.6	2.9	2.3	3.9	2.9	2.0	0.4	1.8	2.9	3.3	2.8	2.3
- min. (mm)	0.9	-	-	0.7	0.1	0.3	0.9	0.6	0.1	0.1	0.2	1.1	0.4	0.1	0.1
- mean (mm)	1.49	-	-	1.73	0.85	1.42	1.35	1.69	1.1	0.23	0.85	1.16	1.7	1.1	1.3
- SD (mm)	0.26	-	-	0.58	0.55	0.43	0.4	0.69	0.4	0.12	0.2	0.3	0.7	0.4	0.3
- hardness (1-5)	1	1	-	0.1	1	1	1	1	1	1	1	1	1	1	1
coherence of argillitic clasts	-	1	-	1	1	1	1	1	1	1	1	1	1	1	1
clast surface colours	-	7.5YR	-	7.5YR	10YR-7.5YR	10YR-7.5YR	7.5YR-5YR	7.5YR-5YR	7.5YR	10YR	7.5YR	7.5YR	5YR		7.5YR-10YR
C. SUBSURFACE MATRIX															
profile type	A Cox Cu	A Cox Cu	-	A Cox Cu	A Cox Cu	A Cox Cu	A Cox Cu	A Cox Cu	A Cox Cu	A Cox Cu	A Cox Cu	A Cox Cu	A Cox Cu	A Cox Cu	A Cox Cu
colour of B horizon	10YR	7.5YR-5YR	-	10YR-7.5YR	10YR-5YR	10YR-7.5YR	10YR	7.5YR-10YR	7.5YR	10-7.5YR	7.5YR-10YR	7.5YR-10YR	7.5YR	7.5YR	7.5YR-10YR
depth of B horizon (cm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
depth to lowest Cox (cm)	>120	>100	-	>60	>150	>30	>80	>130	>100	>100	>30	>50	>60	-	>30
pH at ~1m	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
organic at ~1m	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
clay size fraction mineralogy	quartz[d] kaolinite(?)	-	-	-	-	-	montmor-illonite quartz	-	-	-	-	-	-	-	-
clay film development	1	1	-	1	0	1	1	1	-	0	1	1	-	-	-
coarse matrix: % >1cm	-	36	-	-	-	-	-	-	-	-	-	-	-	-	-
% 1-0.5cm	-	32	-	-	-	-	-	-	-	-	-	-	-	-	-
% <0.5cm	-	12	-	-	-	-	-	-	-	-	-	-	-	-	-
fine matrix: % sand	-	-	-	-	41.8	-	-	-	-	-	34.8	-	-	-	-
% silt	-	-	-	-	42.2	-	-	-	-	-	38.1	-	-	-	-
% clay	-	-	-	-	16.2	-	-	-	-	-	27.1	-	-	-	-
basal till lithification (1-5)	-	1	-	-	1	1	-	-	-	-	-	-	-	-	-
D. SURFACE MATERIALS															
no. of surface boulders/m²	-	20.5	-	-	-	-	-	-	-	-	-	-	-	-	-
% split clasts	-	1	-	-	-	-	-	-	-	-	-	-	-	-	-
surface pit depth (dolerite) - mean (mm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
- max. (mm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
E. DEPOSITIONAL LANDFORMS															
moraine crest width (m)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
proximal slope (°)	-	-	40	-	-	-	-	-	-	-	-	-	-	-	-
distal slope (°)	-	-	10	-	-	-	-	-	-	-	-	-	-	-	-
fluvial dissection (1-5)	0	2	-	0-1	1-2	-	2	4	2-3	2	-	-	-	-	-
F. BEDROCK LANDFORMS															
width of joint opening (cm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
depth of joint opening (cm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
width of solution pans (cm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
depth of solution pans (cm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
height of pedestals (cm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
presence of slab topples (1-5)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
exfoliation slab thickness (cm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-

(a) BW= Bedlam Wall; CB= Cynthia Bay; I= Ida; RH= Rangers Hut; SC="Stoney Creek".

(b) B= bedrock (d= dolerite, ss= sandstone); T= till; O= outwash gravels.

(c) Refers to burial of the same depositional unit in the same vicinity but not at exact location where subsurface data were obtained. The latter have only been recorded from unburied sites. S= solifluction deposit

(d) after Dorbyshire (1967).

Table 4.2 Post-depositional modification of deposits, Lake St. Clair area. Site locations are indicated on figure 4.1.

terminal moraine are retreat features. However, insuffient sections are available to confirm this impression. The ice surface gradient declined eastwards near the southern end of the St. Clair trough which suggests that the glaciers that descended into the trough around Mt. Ida were of lesser significance as sources of ice than were the valley-head cirques. Indeed, only a comparatively small part of the ice cap is ever likely to have drained into the Derwent (cf. Derbyshire 1963). Ice retreat to the Rangers Hut limit was brought about largely through the decline of the Central Plateau ice cap which resulted in a halt to the passage of ice from the Mersey Valley into the Derwent.

Some slab topple deposits occur locally within the Cynthia Bay ice limits. In these cases toppling appears to have been a response to withdrawal of the supporting ice margin. However, most of the slab topple deposits at the foot of Walled Mountain and Mt. Hyperion lie outside the ice limits of the Cynthia Bay phase, during which time toppled blocks were reworked into moraines on Mt. Olympus. This indicates that the main phase of toppling predates the Cynthia Bay phase.

The clay-rich B horizon of the soil developed on the Bedlam Wall till is consistently at least three times thicker than in any of the subsequent tills. This indicates that the Bedlam Wall till is considerably older. On the other hand the dolerite weathering rinds are little thicker than in the subsequent tills (Table 4.2). During the Bedlam Wall phase ice extended well south of Lake St. Clair. A talus that accumulated in contact with the ice on the north-eastern shoulder of Mt. Gould at this time indicates that the Cephissus Glacier was at least 400 m. thick at this point.

Equivocal erosional evidence suggests that during a very intense phase of glaciation the summit of Walled Mountain (1410m) and many of the summits of the Du Cane Range were over-ridden by ice. Some summits appear to have been nunataks, including The Acropolis' eastern ridge and Mt. Gould. The glacial deposits generally appear to occur within previously ice-eroded landforms rather than to define the limits of the ice which was responsible for the erosion.

The development of moraines of Thule-Baffin type at Lake St. Clair reflects the slightly continental character of glaciation in this area. These moraines contrast starkly with the broader and more massive moraines that occur further down the Derwent Valley. No moraines of Thule-Baffin type have been recorded in the more maritime areas nearer the west coast. An extensive tract of fluted moraine in the upper Narcissus Valley indicates that the Narcissus Glacier remained active even during its final retreat. In contrast to this, the other tributaries valleys contain small end moraines and hummocky drift which suggests that the ice there became inactive and decayed *in situ*. This indicates that the high Geryon-Acropolis snowfence and the shading from the afternoon sun that it offerred were critical factors during the Stoney Creek phase. Erosional scallopping of this part of the Du Cane Range snowfence confirms its importance in this regard.

chapter five:
THE CUVIER AND HUGEL VALLEYS.

" In the Cuvier Valley I was struck.... by the similarity to the terminal moraine of a glacier presented by an enormous accumulation of boulders which chokes the lower end of the valley."

- Charles Gould, 1860 Government Geologist

The Cuvier River rises from Lake Petrarch, a shallow moraine dammed lake 1.5 km long that lies on the St. Clair Surface (Figure 5.1). Evidence from the Lake St. Clair area indicates that ice spilled into the head of the Cuvier Valley through two cols north of Lake Petrarch and that some of the Cuvier ice passed back into the Lake St. Clair trough through a col on Mt. Olympus (chapter five). The steeper Hugel Valley originates in cirques on the slopes of Mt Hugel (1397 m) and Mt. Rufus (1416 m). The Cuvier and Hugel rivers are confluent at Watersmeet near the southwestern extremity of Lake St. Clair. For convenience, the evidence from a small valley which adjoins the southern margin of the Hugel Valley is also presented here (Figure 5.2).

EROSIONAL LANDFORMS

A. Glacial erosion:

The two largest cirques occur at the heads of the Cuvier and Hugel valleys. They are incised into Permo-Triassic sediments but possess steep upper headwalls formed of

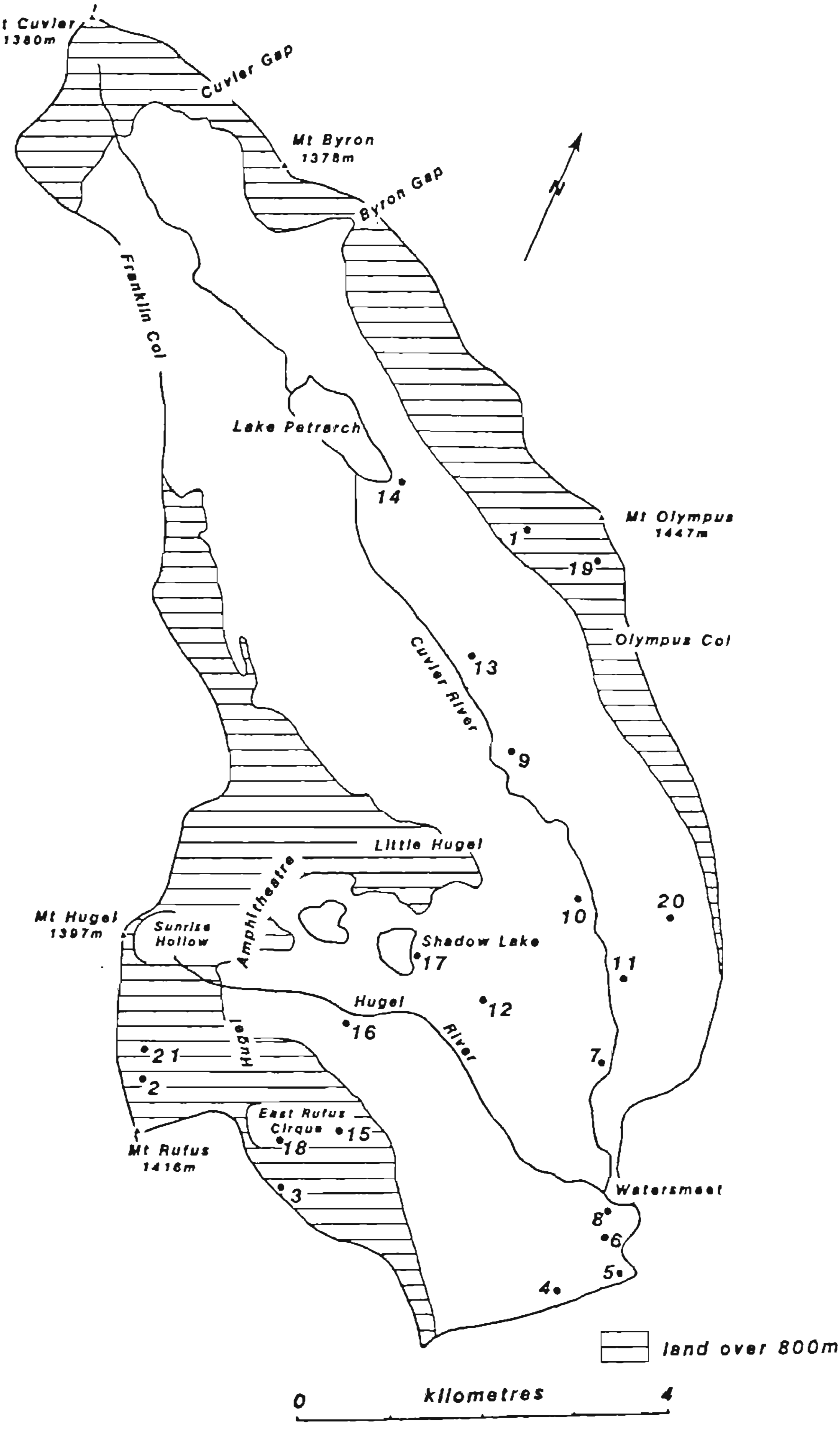


Figure 5.1 Locality map, Cuvier and Hugel valleys. Numbers refer to site locations in tables 5.1 and 5.2.

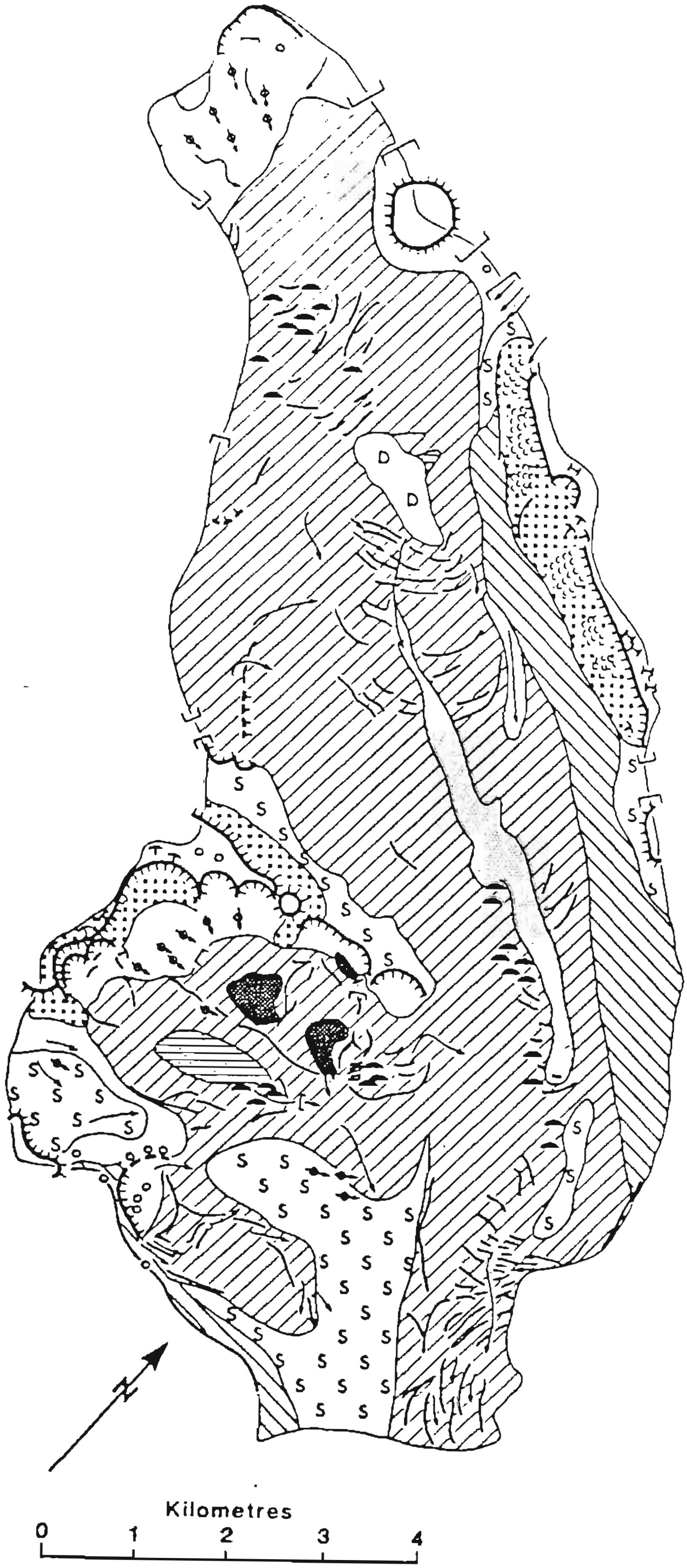


Figure 5.2 Glacial geomorphology of the Cuvier and Hugel valleys. See Figure B2 for key.

Figure 5.3 The glacier-scalloped eastern slopes of Mt. Rufus (left) and Mt. Hugel, viewed from the southern end of Lake St. Clair. The prominent cirque is Sunrise Hollow.

Figure 5.4 Sunrise Hollow, an immature cirque that is inset into the valley-head cirque at the head of the Hugel River.

Diffluence cols occur at the head of the Cuvier Valley between Mt. Cuvier and Mt. Byron (Cuvier Gap, 1010 m) and between Mt. Byron and Mt. Olympus (Byron Gap, 1010 m). Both of these appear to have exploited faults while stoss and lee features indicate that both conducted ice into the Cuvier valley from the north. West of Lake Petrarch lie the Alma (850 m) and Franklin (1030 m) cols through which ice flowed out of the Cuvier Valley. A further col 1.5 km south-east of Mt. Olympus also exploits a fault and is incised through the dolerite into the Triassic rocks (Figure 5.5). Ice smoothed bedrock on the walls of this col indicates that up to 100 m of ice flowed out of the Cuvier Valley into the Lake St. Clair trough.

The crest of the Little Hugel ridge is intensely shattered, probably by frost action. This suggests that it remained as a nunatak during at least the most recent phases of glaciation.

B. Glaciofluvial erosion:

Glacial meltwater channels have been formed in both drift and bedrock. Several bedrock meltwater channels occur in the Hugel Cirque, notably in the area below the Rufus-Hugel col. The Hugel River flows through a steep and rocky gorge below an end moraine at 930m.

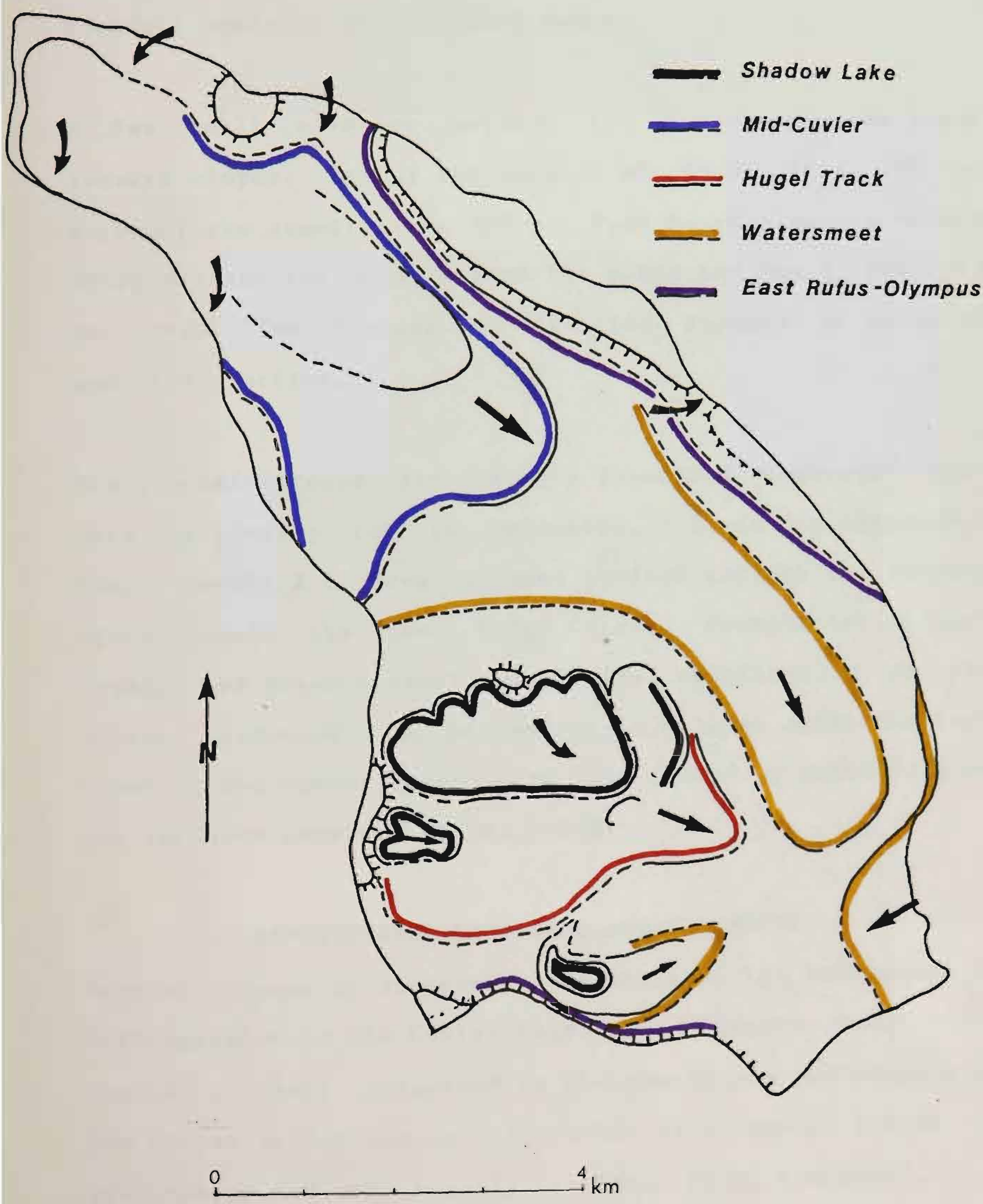
Near the head of the Cuvier Valley a thalweg meltwater channel up to 4 m deep and probably of subglacial origin occurs below Byron Gap. A series of Triassic sandstone benches on the north-western shoulder of Mt. Olympus have probably been exposed by meltwater erosion. These trend obliquely down towards the Cuvier, the largest of them probably exploiting a fault for part of its course. Many channels have been cut into drift south-east of Mt. Olympus. Others plunge downslope from the crest of the Mt. Olympus ridge towards the mouth of the Cuvier Valley. Large semi-circular depressions up to 50 m in diameter occur at the head of the largest of these. The meltwater responsible appears to have been generated from the Lake St.Clair side of the ridge. This indicates a glacial hydrological gradient from the Lake St. Clair trough to the lower Cuvier Valley and therefore indicates thicker ice and a higher ice surface in the Lake St. Clair trough at this point.

C. Non-glacial erosion.

Bold free faces are conspicuous features of the Cuvier-Hugel area. Many of these have been initiated by glacial action but have been maintained and sometimes enlarged by non-glacial processes, the most significant of which has been slab-toppling. Rock crevasses and dilation trenches up to 50 m wide parallel the margins of the Hugel Cirque and are also present at Mt. Olympus. These delineate large slabs that have not yet become totally detached from the hillside. The absence of freshly broken rock and disturbed ground indicates that these features are presently stable. Only incipient joint dilation is apparent above Sunrise Hollow which suggests that the main phase of toppling predates deglaciation of this cirque.

Figure 5.5 Ice flowed out of the Cuvier Valley into the Lake St. Clair trough via the Olympus col, viewed here from Little Hugel. The glacial tind beyond is Mt. Ida, which rises above the eastern shoreline of Lake St. Clair.

overleaf
Figure 5.6 Principal ice limits in the Cuvier and Hugel Valleys.


Shadow Lake
Mid-Cuvier
Hugel Track
Watersmeet
East Rufus-Olympus
N
0
4
km

Abundant erosion by rockfall has taken place in the Hugel cirques and beneath the clifflines on Mt. Olympus. Little freshly broken rock is present which suggests that little rockfall activity is occurring today.

A few small nivation hollows are present on some upper leeward slopes. One of the largest of these lies 150 m north of the summit of Mt.Rufus. This is an elongate hollow which extends for 30 m across the slope and has a headwall 8m high. The process of nivation appears to still be moderately active.

Postglacial streams have commonly inherited channels which were originally cut by meltwater. Since deglaciation a trench barely 2 m deep has been incised through the moraine which bounds the East Rufus Cirque. Postglacial fluvial erosion has brought about only limited modification of the former landscape. A series of 5 m high sandstone tors known as the Sphinx Stones have been formed by weathering on the northern shoulder of Mt. Rufus.

DEPOSITIONAL LANDFORMS AND SEDIMENTS

Several phases of successively diminished ice extent can be distinguished in the Cuvier-Hugel area (figure 5.6). The record is best preserved in the gently sloping terrain of the Cuvier Valley and is discernible to a lesser degree in the steeper and more heavily vegetated Hugel catchment.

(i) Olympus and East Rufus Moraines.

The steep walls of the Cuvier Valley have generally precluded the preservation of lateral moraines subsequent to the withdrawal of the supporting ice margin. This has resulted in extensive block-strewn slopes on the south-western flank of Mt. Olympus below 1100 m. These were referred to as the "lower talus unit" by Derbyshire (1973) who suggested that the boulders may represent rock glaciers which descended from Mt. Olympus. Derbyshire was prevented by the heavy vegetation from examining the deposits in detail but a severe wildfire has since occurred at this site. The position of this talus which accumulated in contact with the ice implies an ice thickness of about 160 m over Byron Gap, 260 m at Lake Petrarch and in excess of 100 m near the valley mouth. This would have been insufficient to have permitted the ice movement through the Olympus Col which must have occurred during an earlier phase. However, it was sufficient to maintain the flow of ice out of the Cuvier Valley through the Alma and Franklin cols.

The outermost and highest lateral moraine in the Hugel Valley is perched upon the ice-smoothed southern rim of the East Rufus Cirque. Originating at about 1370 m and standing over 100 m above the floor of the cirque this moraine can be traced eastwards for over 3 km. Although dominated by dolerite clasts the till contains a large proportion of hornfels derived from the sedimentary rocks at the base of the dolerite sill.

Fragments of a dense grey basal till are a particularly noteworthy component of this moraine. These occur as discrete clasts within the body of the moraine. The till consists of dolerite fragments of low to moderate sphericity and moderate angularity together with some quartz granules in a tough matrix which varies in colour from light grey (2.5 Y 6/1) to light yellow (2.5 Y 7/3). The highly lithified condition of this material must have been achieved prior to its incorporation as clasts in the moraine. This implies an earlier phase of glaciation and is consistent with the stratigraphic position of the Nive Till, a very similar deposit that is described in chapter ten.

The East Rufus Moraine implies the existance of a large glacier over 4 km wide and nearly 300 m thick which filled the entire Rufus-Hugel amphitheatre, with Little Hugel standing as a nunatak amid the confluent ice of the Hugel and Cuvier glaciers.

(ii) Watersmeet Moraines

A major complex of latero-terminal moraines of Thule-Baffin type (Embleton and King, 1975) was constructed by the expanding lobe of the St. Clair Glacier as it emerged from the glacial trough east of Mt. Olympus. Expansion of this lobe into the mouths of the Cuvier and Hugel valleys means that the alignment of the end moraines from those valleys parallels that of the moraines deposited by the St. Clair Glacier.

While it is difficult to discern any well defined moraines low down in the Hugel Valley a small series can be recognised in the Cuvier. An important feature here is a large moraine 38 m high with an irregular crest about 200 m broad. This probably originated in a medial position between the St Clair and Cuvier ice. Subsequent end moraines of the Cuvier Glacier are stacked against it. The descent of a meltwater channel from on top of the Mt. Olympus ridge to the foot of the most upstream of the Cuvier moraines implies that the ice in the St. Clair trough still stood at least 110 m higher than the floor of the Cuvier Valley subsequent to the withdrawal of the Cuvier ice from Watersmeet. A moraine which was subsequently constructed by the St. Clair Glacier now blocks the mouth of this channel.

The greater energy of the St. Clair Glacier at this point is reflected in the higher proportion of local sandstone clasts (26%) in the till deposited by the St. Clair Glacier compared to the tills deposited by the Cuvier ice despite the identical bedrock environments at Watersmeet (Table 5.1). About 95% of the clasts in the Cuvier till are dolerite which have been derived from further upvalley. Comminution of the Triassic rocks appears to be fairly rapid. While some slabs of sandstone have moved downslope and been carried upon the ice surface large sandstone clasts appear to be rare within the body of the moraines. This suggests that the lateral moraines are formed of material that was involved in a relatively large degree of lateral movement of comminuted englacial debris.

(iii) Moraines in the Hugel Valley

(a) Hugel Track Moraines

A prominant lateral moraine at 800-920 m to the south-east of Little Hugel was deposited upon a rock ridge which increasingly divided the Cuvier and Hugel glaciers as lowering of the ice surface proceeded. On the southern side of the valley a moraine can be traced from about 840 - 940 m. It remains undissected by stream erosion. Some of the ice in the East Rufus Cirque was probably deflected by a bedrock knob into the Hugel Glacier, which covered around 12 square kilometres at this time.

(b) Hut Moraines

A sharply defined lateral moraine extends downslope from the East Rufus Cirque at about 1140 m. At the time this moraine was constructed the East Rufus Glacier was independant of the main Hugel ice. The main part of the East Rufus Glacier was at least 1.5 km long and up to 80 m thick. That the ice in the main Hugel headwaters was much diminished is shown by end moraines at 920 m above a valley step along the Hugel River and also by a series of moraines which border the most easterly cirque in this series. Hummocky drift was subsequently deposited within the confines of these moraines, together with some silt deposits about 3 km east of Mt.Hugel.

(c) Shadow Lake Moraines

Shadow Lake, a small tarn on the lithological bench that floors the main Hugel amphitheatre, is impounded behind small end moraines that lie to its east and south, and also by hummocky moraine. Immediately to the east Forgotten Lake, a partial rock basin, lies behind a moraine capped rock ridge that had emerged to divide the ice body below Little Hugel from the Hugel Glacier. The most recent lateral moraine can be traced back to Sunrise Hollow. Abundant very angular blocks indicate that this moraine has a significant protalus component further upslope.

Proglacial gravels, sands and silts appear to have been laid down upon the bench beneath Sunrise Hollow at this time. The continued existence of Shadow and Forgotten lakes implies that no major phase of erosion that generated sediment took place on the headwalls after the lakes had been exposed by the retreating ice. The greater persistance of ice in this tributary position probably resulted from more adequate shading and a rather more favourable accumulation area on the snowfence than existed at Sunrise Hollow.

(iv) Moraines in the Cuvier Valley

Ground moraine covers the floor of the Cuvier Valley for about 5 km upstream from the Watersmeet moraines. Much of this terrain is thinly veneered by glaciofluvial deposits. East of the river lie poorly developed hummocky moraines with some kame and kettle features. Shallow sections that have been incised by creeks in this area, reveal waterlaid deposits among the till. A lack of distinct moraine ridges suggests that the glacier was retreating at a fairly uniform rate when these deposits were laid down.

(a) Middle Cuvier Moraines.

A major end moraine complex occurs about 2 km downstream from Lake Petrarch. This rises about 100 m above the valley floor and extends to within about 500 m of the lake. It stretches across almost the entire valley except where it has been incised by meltwater channels (Figure 5.7). The considerable bulk and general form of this moraine complex suggest that it was deposited over a considerable period of time when the ice front approximately maintained this position. Retreat to this point may have occurred in response to the severing of diffluent ice flow from the Cuvier and Byron cols. With the decline of the ice surface, diffluence of the Cuvier ice through the Franklin and Alma cols would have also ceased.

Dolerite is the dominant lithology in this till (Table 5.1). Outwash sand and gravel from this ice front has inundated swales and produced a veneer on the more downstream moraines. Occasional lag boulders occur on the plains. Dolerite dominates in all grades larger than granules. Beds and lenses of fine sand and gravel occur in the southern part of the valley.

(b) Lake Petrarch Moraines

A series of small moraines are stacked against the Middle Cuvier complex close to the southeastern end of Lake Petrarch. Outwash gravels occur along channels through the earlier moraines. The amplitudes of the moraine ridges are generally less than 15 m. Their surface form is fairly sharp with well preserved meltwater channels and a few

Vicinity No.	7	13	16	18
%				
dolerite	93.0	96.0	50.0	100.0
quartz	4.2	2.0	10.0	-
Triassic sandstone	2.8	2.0	29.0	-
Permian & others	-	-	11.0	-

Table 5.1 Lithology of glacial deposits in the Cuvier and Hugel valleys. Site locations are indicated on Figure 5.1.

probable kettle holes present. Shallow sections along the Cuvier River reveal ablation till with numerous angular boulders of dolerite and some sandstone underlain in places by basal till, as at the western end of the Lake Petrarch beach.

Upstream from Lake Petrarch the end moraines rapidly give way to ground moraine and hummocky moraine. The hummocky moraine often consists of sharply conical mounds up to 6 m. high and numerous kettle holes. Derbyshire (1963, 1967) has noted that the steepest slope angles are attained where the boulder content is high, and argued that this morphology is the product of slow, in situ decomposition of thin and inactive ice. Many of the clasts are angular to subangular and the mounds represent the inversion of deposits that were formed in former holes in and beneath the ice.

Upstream of a poorly defined end moraine about 1.2 km north-west of Lake Petrarch the hummocky moraine resumes with abundant waterlaid gravels, minor fault structures and some slumping in the melt-out till which is exposed along the banks of a creek south-west of Mt.Byron. Massive silty clay deposits more than 2m thick occur near the head of the Cuvier Valley.

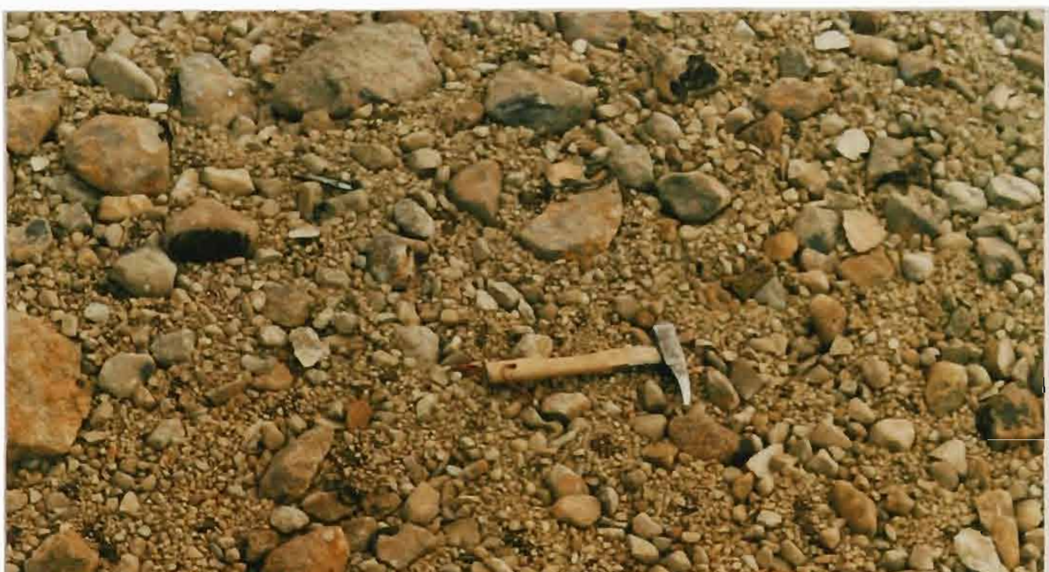
B. Non-glacial deposits.

(i) Slope deposits

Solifluction deposits are widespread on slopes that were not swept clear by ice during the most recent glacial phases. This material has moved down slopes of as little as 5^{o} but is presently inactive. Small solifluction terraces up to 3 m. wide with risers of 50 cm occur at about 1220 m on the East Rufus Moraine. These appear to be inactive despite the removal of much of the vegetation by a wildfire during the last decade. The till mantled slopes to the south have been smoothed, probably by solifluction and slopewash.

Periglacial nets up to 110 cm in diameter occur further east along this ridge and also appear inactive (Figure 5.8). However small nets about 20 cm in diameter within the East Rufus Cirque and small solifluction lobes up to one metre wide with risers of up to 15 cms on the northern slopes of Mt. Rufus both appear to be actively forming.

Lobes of blocks that extend down the eastern flank of Mt. Olympus (Figure 5.7) were recognised as rock glaciers by Derbyshire (1967, 1972). They are generally broader than they are long, and are equivalent to the lobate rock glaciers of Wahrhaftig and Cox (1959). The apparent absence of any remnant matrix in morphologically contiguous deposits supports Derbyshire's suggestion that interstitial ice rather than a saturated earthy matrix facilitated their flow. Derbyshire considered these to be ice-cemented rock glaciers in view of the general absence of any evidence of ice accumulation which might have nourished an ice core. In most cases their lateral transition into normal block slopes (White, 1981) and their position beneath a free face supports this interpretation. However, one of the rock glaciers has developed beneath a cirque. Hence, the possibility of an initial ice core must be kept open in this



case. White (1971) provides evidence that as an ice core decays movement may be maintained by interstitial ice until the characteristics of an ice cemented form (Potter, 1972) are assumed.

Slab toppling is the predominant mechanism of slope retreat in the high dolerite terrain and has produced extensive block slopes around the Mt. Olympus plateau. Toppling has also been involved in the accumulation of the talus beneath Little Hugel.

Rockfall talus occurs on the southwestern side of the Little Hugel ridge. Some large prismatic blocks are present but the deposit is dominated by smaller fragments. Only rarely can portions of columns that previously adjoined one another be identified. The position of this deposit suggests that it represents a major episode of rockfall that took place as the ice retreated and ceased to support the steep cirque headwall.

The protalus that forms a major component of the uppermost moraines at Sunrise Hollow is less well developed in the East Rufus Cirque where the backwall is less pronounced and less shattered. Little of this material appears to postdate deglaciation.

(ii) Alluvial, peat, dune and archaeological deposits.
Alluvial silts have accumulated to a depth of 1 m in depressions on the floor of the Cuvier Valley and on the benches in the Hugel Valley. Up to 1 m of organic-rich

silt overlies minerogenic silt beneath Sunrise Hollow. Postglacial fluvial sedimentation is also represented by alluvial fans that extend from Mt. Olympus onto the floor of the Cuvier Valley near Lake Petrarch and by gravels reworked from pre-existing glaciofluvial deposits to form shingle bars along streams.

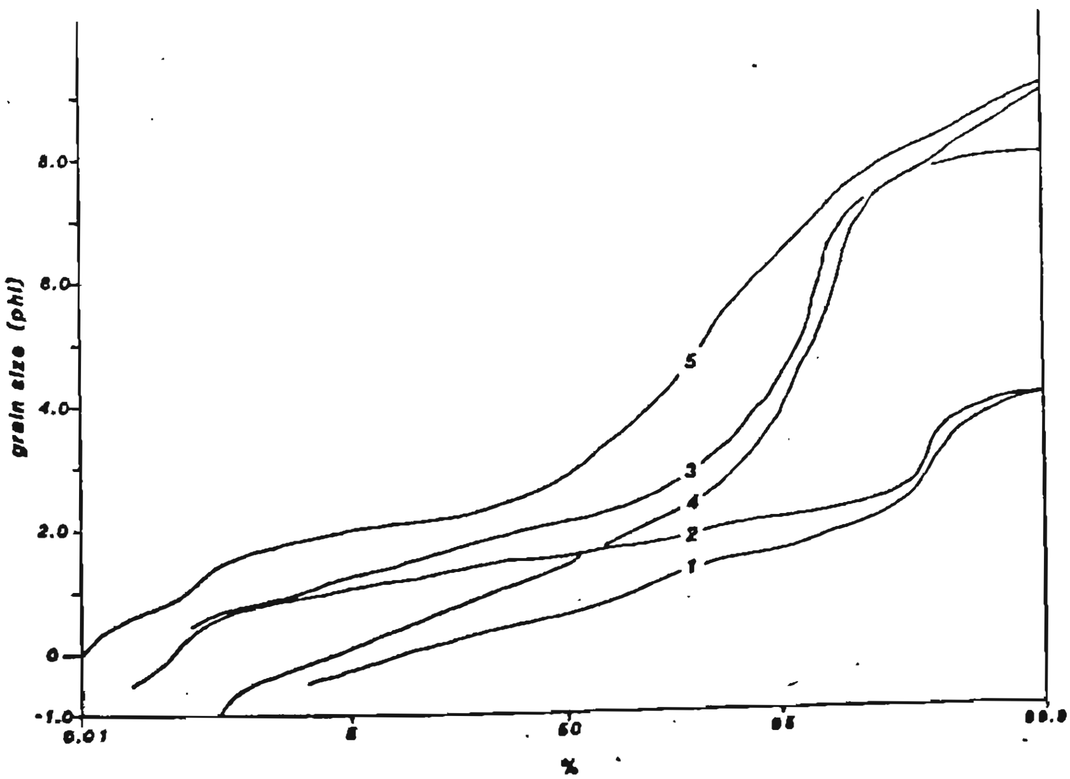
Fibrous peat has accumulated beneath the buttongrass plains that cover most of the Cuvier Valley floor and also occurs on the benches in the Hugel Valley.

The bed of the shallow Lake Petrarch consists of reworked outwash sand (Figure 5.9). An embankment of sand at the southeastern end of the lake has been blown onto the eroding margin of a small outwash plain (Figure 5.10).

A large prehistoric quarry site exists at an altitude of nearly 1400 m on the East Rufus Moraine. Clasts of calc-silicate hornfels have been gathered from the till and used in the manufacture of stone tools by Tasmanian aborigines. Thousands of tools and waste flakes are scattered across the surface of the moraine and adjacent rock ridge. Stone tools have also been found at lower elevations in the East Rufus Cirque, which may have served as a reasonably sheltered campsite for groups that visited the area to quarry the East Rufus Moraine. Isolated flakes have also been found lower down the Hugel Valley.

DATING AND DISCUSSION.

As in the Lake St. Clair basin the accumulation of peat upon


grain size (phi)
8.0
6.0
4.0
2.0
0
-1.0
0.01
5
50
95
99.9
%
5
3
4
2
1

organic-rich silts must postdate the retreat of the ice and the stabilisation of the landscape as a vegetation cover became established.

The protalus that forms the uppermost moraines at Sunrise Hollow and the rockfall and topple talus at Little Hugel both accumulated during the dying stages of glaciation. The greater shade offerred by the Little Hugel ridge would have permitted ice to linger there longer than was the case at Sunrise Hollow. The Little Hugel talus is not significantly attenuated along the former ice margin which suggests that the failure of this slope postdated the presence of any active ice.

Precise correlation of the phases in the Cuvier and Hugel valleys is difficult due to differences in topography and complications resulting from the diffluence of Lake St.Clair ice into the Cuvier. Hummocky moraines at the head of the Cuvier Valley indicate that the ice was stagnant. A more elevated and effective snowfence and better shade was available at the Hugel lakes than at the head of the Cuvier. Here small end moraines were constructed by active ice shortly prior to deglaciation. Hummocky moraine on the outermost part of the bench east of Shadow Lake suggests that debris-rich ice had earlier decayed in situ following a significant deficit in the glacier budget.

The massive Middle Cuvier end moraine complex represents a major phase during which the snout of the Cuvier Glacier approximately maintained its position for a lengthy period.

This complex probably reflects the re-adjustment of the Cuvier Glacier to its own accumulation resources after upstream supplementation from the St. Clair trough had ceased.

Downstream of the middle Cuvier Moraine lies an extensive tract of ground moraine that was laid down during a 5 km retreat. The failure to isolate a retreat of similar magnitude in the Hugel Valley may simply reflect individual glacier dynamics and in particular the much diminished supply of ice in the Cuvier Valley once diffluent flow through the Byron and Cuvier cols had ceased.

The Watersmeet Moraines form part of the extensive array of end moraines south of Lake St. Clair. Various lines of evidence that were presented in chapter four indicate that these end moraines represent the most recent glaciation of the St. Clair trough. Given that the Watersmeet Moraines form a benchmark at the mouth of the Cuvier Valley the other deposits further upstream on the valley floor must be of lesser age. Confluence of the Cuvier and Hugel ice during the Watersmeet phase has not been proven. However, any increase in the size of the Hugel Glacier is likely to have been matched by a higher ice limit on the edge of the St. Clair Glacier, leading to confluence of all three glaciers.

There is no evidence of large scale slab toppling within the ice limits of the Watersmeet phase which indicates that the main episode of toppling must have predated it.

Evidence of earlier glaciation is provided by the East Rufus and Olympus moraines. Most of the landscape must have been inundated by confluent ice masses at the time the East Rufus Moraine was constructed on the previously over-ridden arm of its cirque. No deposits can be definately attributed to this phase in the Cuvier Valley, although it is possible that the major talus on the flank of Mt. Rufus was deposited in contact with this ice. The glacier margin defined by this talus would have been insufficient to permit the passage of ice through the Olympus col at which time the Cuvier Glacier must have been 320 m thick. The rock glaciers extend below this highest ice limit and must therefore postdate this phase of glaciation. The incorporation in the rock glaciers of dolerite blocks that were derived by toppling indicates that a major phase of slab toppling followed retreat of this early glacier.

The use of post-depositional criteria (Table 5.2) as an aid to dating the earliest of the moraines is complicated by the reworking and collapse of the Olympus moraine and the uncertainty of its correlation with the East Rufus Moraine.

The moraines on the valley floors all possess A Cox Cu soil profiles. However, thicker dolerite weathering rinds occur in the Middle Cuvier complex than at either end of the valley. As this site appears no better drained than many others the only apparent site difference which might account for this is the presence of Nothofagus forest on the Middle Cuvier Moraine.

moraine[a]	OH	ER?	ER	WAT	WAT	MAT	WAT	WAT	MAT	WAT	WAT	HT	MC	PM	HH	SL	SL	SL	-	-	-
vicinity no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21
material[b]	T	B	T	T	T	T	T	T	T	T	T	T	T	T	T	T	T	T	R	S	S
A. MORPHOSTRATIGRAPHY																					
relative position	outside WAT	inside 3?	outside 17	outside 5	inside 4	inside 4	inside 4	inside 4	inside 4	inside 9	inside 10	outside 16	inside 11	inside 13	outside 16	inside 12	inside 12	inside 3		inside 10	insi 3
crest/site elevation(m)	1100	1240	1220	755	740	740	790	890	790	800	810	920	840	860	1180	930	960	1160	1140	880	1160
max.depth of burial (by)[c] (m)	-	51.5	-	-	-	-	-	-	-	-	80.4	-	-	-	-	-	-	-	-	-	-
B. SUBSURFACE CLASTS																					
dolerite rinds (mm)-max.	5.1	-	7.2	4.9	2.4	5.0	4.2	4.2	1.2	3.4	5.1	3.8	5.5	3.2	-	1.2	1.2	2.3	0.7	3.2	4.1
min.	3.5	-	3.2	0.7	0.6	0.3	0.2	0.5	1.2	1.0	0.9	0.8	1.8	0.8	-	0.6	0.4	0.6	0.5	0.1	0.3
mean	4.1	-	4.9	1.3	1.3	1.5	1.9	2.9	1.7	1.9	2.1	2.3	3.4	1.9	-	1.6	1.3	1.5	0.6	1.2	1.6
SD	0.6	-	1.0	0.9	0.5	1.2	1.6	0.9	0.3	0.5	0.9	1.1	0.9	0.9	-	0.6	0.5	0.5	0.1	0.7	0.8
Hardness (1-5)	1-2	-	2-4	1-2	1-2	1	2	1-2	1-2	2	2	1-2	2-3	1	-	1	1	1	1	1	1
clast surface colours	7.5YR	-	5YR	7.5YR	7.5YR	7.5YR	7.5-5YR	7.5YR	7.5YR	7.5YR	7.5-5YR	7.5YR	7.5YR	7.5YR	7.5YR	10-7.5YR	7.5YR	7.5YR	-	7.5YR	5-7.5
% unrecognizably weathered clasts	-	-	5	0	0	-	-	-	0	-	-	0	-	0	0	0	0	0	-	0	0
hornfels rinds (mm)-max.	-	-	4.2	-	-	-	-	-	-	-	-	-	-	-	-	-	-	2.2	-	-	-
min.	-	-	0.9	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.6	-	-	-
mean	-	-	2.0	-	-	-	-	-	-	-	-	-	-	-	-	-	-	1.7	-	-	-
SD	-	-	1.3	-	-	-	-	-	-	-	-	-	-	-	-	-	-	0.6	-	-	-
sandstone rinds(mm)-max.	-	-	4.9	-	-	-	-	-	-	-	3.3	-	-	-	-	-	-	-	-	-	-
min.	-	-	0.3	-	-	-	-	-	-	-	1.6	-	-	-	-	-	-	-	-	-	-
mean	-	-	1.0	-	-	-	-	-	-	-	2.7	-	-	-	-	-	-	-	-	-	-
SD	-	-	0.4	-	-	-	-	-	-	-	1.1	-	-	-	-	-	-	-	-	-	-
C. SUBSURFACE MATRIX																					
profile type	-	-	A Bt	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A BW Cox	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A Cox cu	-	A Cox cu	A Co
colour of B horizon	7.5YR	-	7.5-5YR	10YR	10YR	10YR	7.5-10YR	7.5YR	10YR	10YR	10YR	7.5YR	7.5YR	10YR	7.5YR	7.5YR	10YR	10YR	-	10YR	7.5
depth of B horizon(cm)	-	-	> 150	-	-	-	-	-	-	-	-	-	10	-	-	-	-	-	-	-	-
depth to lowest Cox(cm)	-	-	> 150	> 40	> 40	> 30	> 30	> 30	> 50	> 40	> 40	> 50	> 50	> 30	> 50	> 50	≥ 30	> 30	-	> 50	> 1
development of clay films (1-5)	-	-	3	1	1	1	1	1	1	1	1	1	1	1-2	1-2	1	1	1	1	1	1-
degree of basal till lithification(1-5)	-	-	2-3	-	-	-	-	-	-	-	1	-	1	-	-	1	-	-	-	-	-
clast socket staining(1-5)	-	-	3-4	1	1-2	1	1	1	1	1	1-2	1	1-2	1	1	1	1	1	-	-	-
notable component clasts	-	-	grey basal till	-	-	-	-	-	-	-	-	-	-	-	red basal till	-	-	-	-	-	-
D. SURFACE MATERIALS																					
No. of surface boulders (> 30cm/m^2)	3.5	-	1.4	-	-	-	-	-	-	-	-	-	-	-	2.0	-	-	2.4	-	-	-
% of split clasts	-	-	27	-	-	-	-	-	-	-	-	-	-	-	20	-	-	12	-	-	-
width of moraine crest(m)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	2-3	-	-	9	-	-	-
moraine slopes (°) - proximal	-	-	-	-	-	-	-	-	-	-	-	-	-	-	43	-	-	4,16	-	-	-
distal	-	-	-	-	-	-	-	-	-	-	-	-	-	-	48	-	-	32	-	-	-
degree of fluvial dissection (1-5)	-	-	-	1	1	1	1	1	1	1	1	1	1	1	1	1	1	-	-	-	-
max. diameter, sorted net(cm)	-	-	110&20	-	-	-	-	-	-	-	-	-	-	-	-	-	-	20	-	-	-
max.width, solifluction terrace tread(cm)	-	-	400	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
max.height, solifuction terrace riser (cm)	-	-	100	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
width of joint opening(mm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
depth of joint opening(mm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
height of tors (m)	-	5	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-

(a) OH= Olympus; ER= East Rufus; WAT= Watersmeet; HT= Hugel Track; MC= Middle Cuvier; PM= Petrarch; HH= Hut; SL= Shadow Lake.

(b) B= bedrock; T= till; S= solifluction deposit; R= rockfall talus.

(c) Refers to burial of the same depositional unit in the same vicinity but not at the exact location where subsurface data were obtained. The latter have only been recorded from unburied sites. S= solifluction deposit; O= outwash gravels.

Table 5.2 Post-depositional modification of deposits, Cuvier and Hugel valleys. Site locations are indicated on figure 5.1.

Only the Olympus and East Rufus moraines can be of greater age, both lying high on the valley walls. The consistently thicker rinds at these two sites suggests that a considerable age difference is involved. In addition, the chemically weathered sandstone tors known as the Sphinx Stones that lie on the northern shoulder of Mt. Rufus are located within the probable ice limits of the East Rufus phase and may not have survived nor had time to redevelop had the site been eroded by a glacier during the late Last Glaciation.

The presence of strongly lithified grey basal till as clasts within the East Rufus Moraine points towards a still earlier phase of glaciation.

Moraines of Thule-Baffin type are confined to the Watersmeet complex, which is consistent with their deposition by the St. Clair Glacier and by diffluent ice that flowed down the Cuvier Valley from the same ice source. The Cuvier Glacier retreated dramatically once the flow of ice into the valley head through the Cuvier and Byron gaps was severed. It subsequently constructed a massive end moraine and later deposited hummocky drift as the ice stagnated. Well formed end moraines indicate that the ice remained active high on Mt.Hugel where shading appears to have been important. Final deglaciation must have been rapid as the Hugel lakes were not filled by proglacial sediment.

chapter six
THE TRAVELLERS REST VALLEY

" Unfortunately we were unable to traverse the Traveller Plateau, but from its configuration as observed from Olympus I feel confident that signs of glaciation do not exist there.....it is evident that the glacial theory is of no use here."

- Officer (1894)

The Travellers Rest River rises from the south-western corner of the Central Plateau horst (Banks, 1973b) and descends steeply from the Lower Plateau Surface (known locally as the Traveller Plateau) to the St. Clair Surface where it joins the Derwent River (Figure 6.1).

Although some early workers argued that the Travellers Rest area had not been glaciated (eg. Officer, 1894) a number of workers have since shown the opposite to be the case. Erosional landforms, and in particular numerous rock basins on the Central Plateau east of Lake St. Clair, indicate that some of the plateau ice moved southwards into the present catchment of the Travellers Rest River (Clemes, 1925; Jennings and Ahmad, 1957). Clemes considered that two glaciers had moved westwards down the Travellers Rest Valley, one of these having originated on the plateau, the other having passed through a low col on the St. Clair Surface from the Clarence Valley to the east. Derbyshire (1967) argued instead that the glacial deposits on the St.

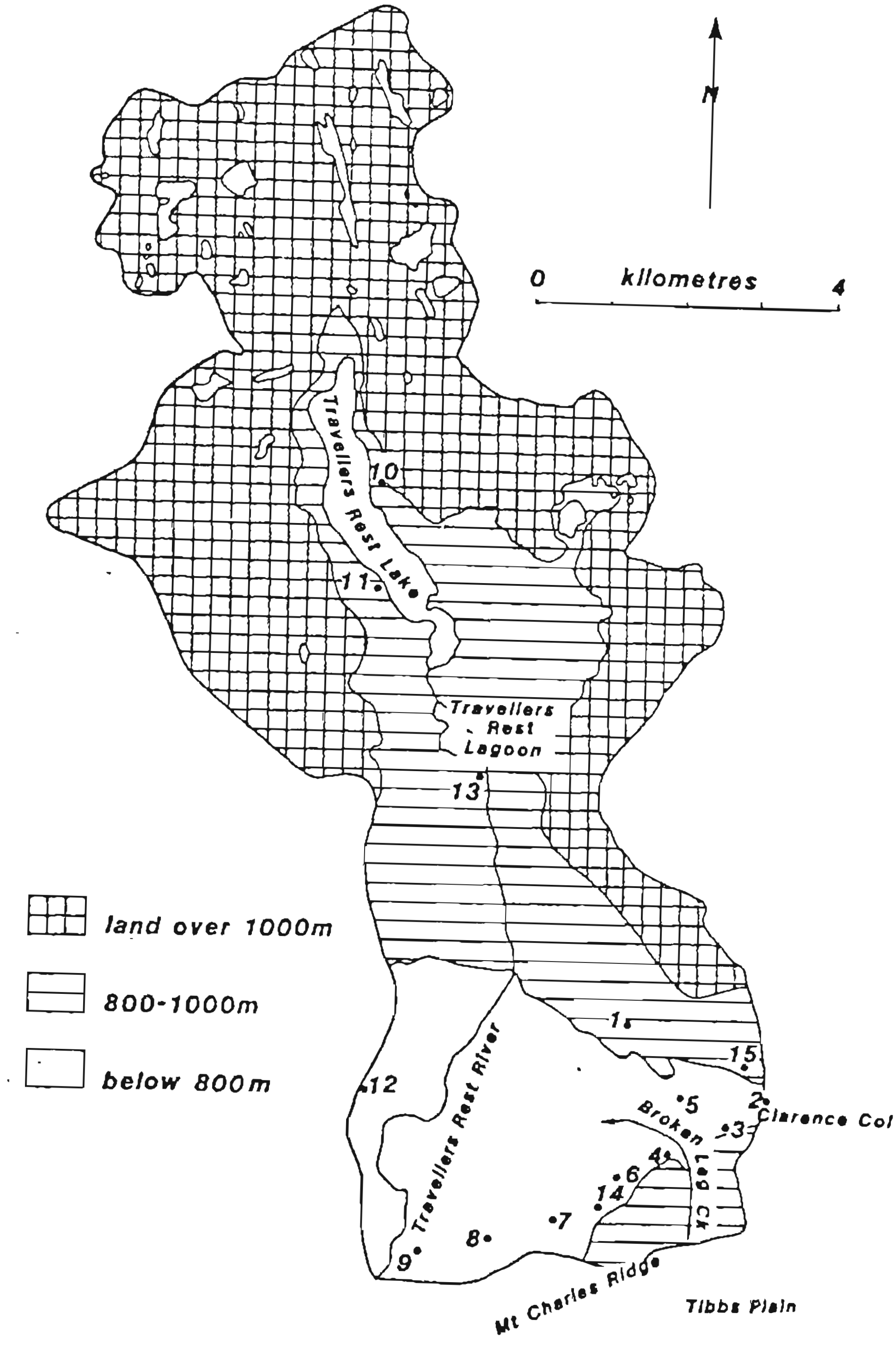


Figure 6.1 Locality map, Travellers Rest Valley. The numbers refer to site locations in tables 6.1 and 6.2

Clair Surface were deposited by a lobe of the Derwent Glacier which had expanded into the mouth of the Travellers Rest Valley. These issues are reassessed here on the basis of geomorphological mapping (figure 6.2).

EROSIONAL LANDFORMS

A Glacial erosion

Jennings and Ahmad (1957) recognised that the erosional landforms which dominate the western part of the Central Plateau give way in the south-west to a zone dominated by depositional landforms. The Travellers Rest River flows across this transition with the erosional zone occupying about 20% in the northern part of the catchment.

However the exhumation of abraded rock surfaces indicates that the depositional zone had also previously been subject to glacial erosion. Backwearing of the fault-line scarp on the southern margin of the Central Plateau has been assisted by the passage of ice down the Traveller Valley. Ice smoothed bedrock also occurs at many points along the Travellers Rest - Lake St. Clair divide overlooking Lake St. Clair and indicates that ice previously flowed directly over the scarp into Lake St. Clair. Erosional rounding and smoothing of the escarpment slopes south of the plateau and smoothing of the flanks of the Mt. Charles ridge further south suggests that almost the entire Travellers Rest catchment may have been affected by glacial erosion.

This area lacks any elevated ridges that could have been deeply eroded by cirque glaciers. The few cirque-like

LANDFORMS

cirque	nunatak
over-ridden cirque	nivation cirque
glacial trough	rock crevasse or dilation trench
over-ridden margin	slab topple
diffluence col	rock glacier
valley step	solifluction lobe or terrace
ice-abraded outcrop	patterned ground
areally scoured terrain	landslip
glacial striae	artificial lake
rock-basin lake	
moraine ridge	
moraine hummock	
moraine-dammed lake	
erratics	
erratics of Nive till	
reworked Permian erratics or large lagstones	
meltwater channel	

SEDIMENTS

GLACIAL	NON-GLACIAL
Cynthia Bay till/outwash	alluvial silt &/or peat
Beehive "	talus
Powers Creek "	Solifluction deposits
Clarence "	
Stonehaven "	
Nive Till "	

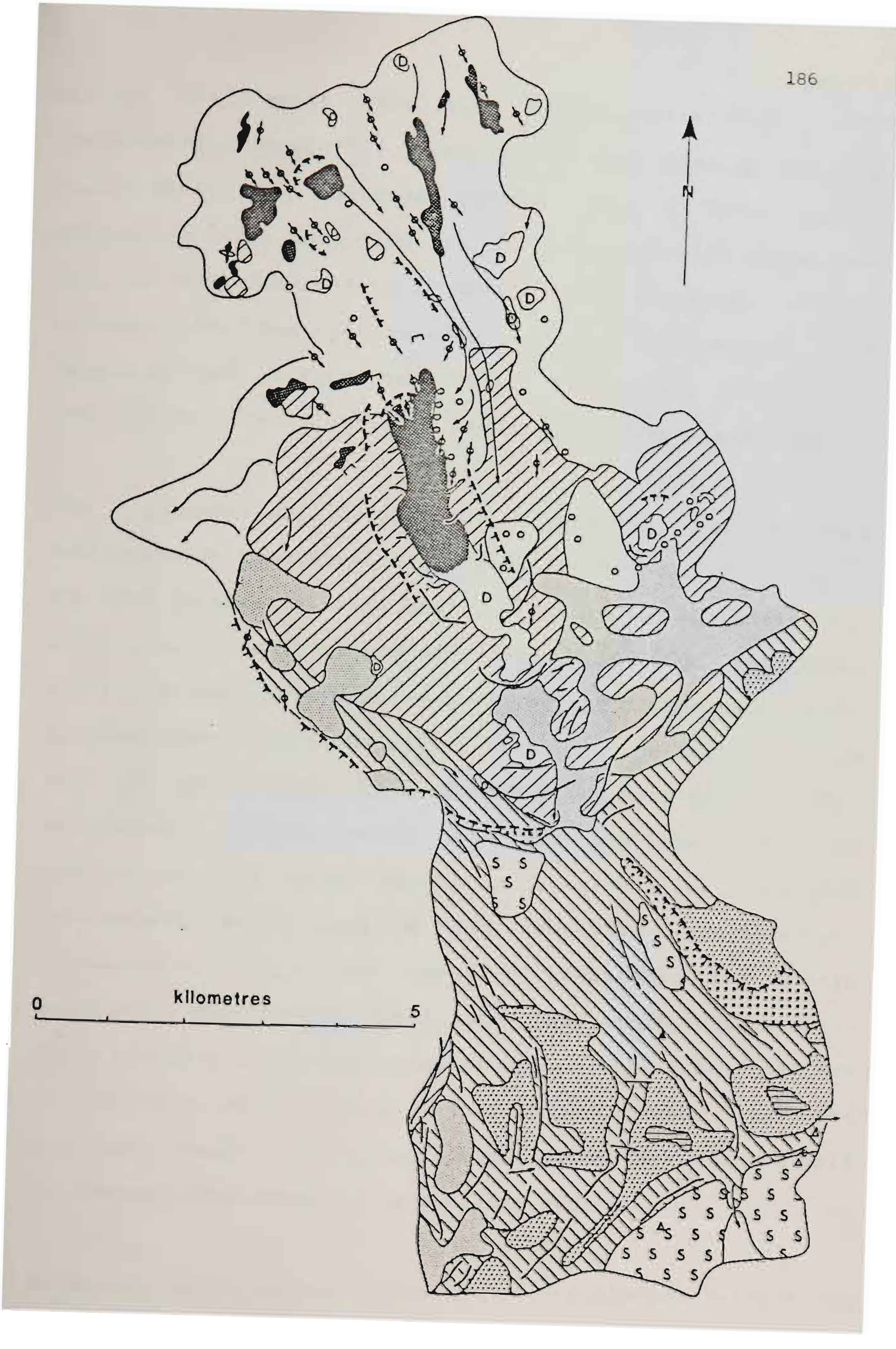


Figure 6.2 Glacial geomorphology of the Travellers Rest Valley. See Figure B2 for key.

hollows that exist have all been over-ridden. The outstanding topographic feature is the Central Plateau itself which has a subdued surface with a local relief generally less that 100 m. This low relative relief has kept the erosional energy of the plateau glaciers low and ensured that the low snowfences were overwhelmed. The result is that it forms part of an outstanding example of what Linton (1963) termed a "knock and lochan" landscape.

The predominant erosional landforms are roches mountonées,ice-smoothed ridges and rock basins. Plucking of the well jointed dolerite has occurred on the lee slopes and often also the side slopes of ridges (Jennings and Ahmad 1957). It indicates that most of the ice flowed southwards. At least forty lakes of varying size are present on this part of the plateau, most of them being rock basins. They are commonly rectilinear in form and are the result of erosion by ice which has been focussed along structural weaknesses. One or more of the shorelines are commonly ice-smoothed while the northern shorelines are almost invariably precipitous. The largest lake is Travellers Rest, which occupies a small glacial trough. It is 4 km long and its northern end is rock girt for 2 km. Moraines that are initially inset within the trough border the southern half of the lake and impound its southern end.

An exhumed area of both ice-smoothed and plucked bedrock in a col at 780 m on the Traveller-Clarence divide indicates that ice moved eastwards into the catchment of the Nive River (chapter 10). This is confirmed by four sets of striae (87, 90, 106 and 117 °) and supports the contention of Derbyshire (1971) that glacial deposits in this area were laid down by eastward moving ice which he considered to be a distributary lobe of the Derwent Glacier. However, this site lies 1 km beyond the maximum limits he proposed for that lobe and contradicts his deduction that the ice was contained by the divide.

B Glaciofluvial erosion.

Many linear meltwater channels have been cut along joints in the dolerite of the plateau. The low energy of the largely underfit present day streams on the plateau demands that these channels must have been cut by large volumes of meltwater under hydrostatic pressure due to an ice cover. Some very small ponds and lagoons, generally less than 10 m long, occupy shallow linear basins that were probably eroded by meltwater. These basins show little evidence of internal plucking and are incised below the level of adjacent ice abraded pavements. Almost all the meltwater channels of the plateau are cut in rock and were probably initiated as subglacial conduits (Sissons, 1960, 1961; Price, 1973).

The Travellers Rest River flows in a V-shaped proglacial meltwater channel that is cut into the floor of a shallow and very broad valley that descends from the plateau to the St. Clair Surface. An earlier marginal meltwater channel parallels the river from 880 - 800 m. Closer to the St. Clair Surface a marginal channel occurs between a latero-terminal moraine on the valley floor and the flank of the scarp. Other proglacial channels in this area have been cut through moraines, with each having a small outwash plain at its outlet. Broken Leg Creek. flows northwards via a poorly developed col gully (Sissons, 1961) through which meltwater previously passed southwards into the Nive catchment (chapter 10). The construction of Tibbs Plain provided a catchment at a sufficiently high elevation to reverse the direction of flow through the channel once the ice had receeded.

C Non-glacial erosion.

The dominant erosional landform of the area is the escarpment between the Lower Plateau Surface and the St. Clair Surface. This is probably a faultline scarp that was worn back in preglacial times by fluvial erosion and mass movement processes. The broad and shallow indentation in the plateau rim down which the Travellers Rest River flows delineates the area most severely eroded by ice and suggests that glacial erosion can have accounted for only a small proportion of the total erosion which has taken place (chapter thirteen).

A contrast exists between closed joints in ice-abraded dolerite bedrock on the present shoreline of Travellers Rest Lake and joints which have been etched open to widths of over 15 mm. on abraded surfaces outside the Cynthia Bay limits. As the edges are seldom fragmented the majority of this etching is probably the result of chemical weathering which would require a considerable period of time.

Small free faces produced by glacial plucking are widespread on the plateau but there are few free faces elsewhere. There is evidence for only very limited toppling on the plateau margin east of Travellers Rest Lagoon and little evidence of landslides. The modern surface therefore appears to be stable.

Human activities such as the construction of roads and excavation of small gravel pits have provided erosional landforms in their own right, while logging and burning on the escarpment slopes has promoted sheet erosion of the thin soils. Minor periglacial disturbance of exposed regolith has followed in the wake of some human activities and has prevented the re-establishment of a binding vegetation cover in some cases. Small polygonal nets are present at the southern end of Travellers Rest Lake (Jennings, 1956) and pipkrake forms each winter in road cuttings on the St. Clair Surface.

DEPOSITIONAL LANDFORMS AND SEDIMENTS

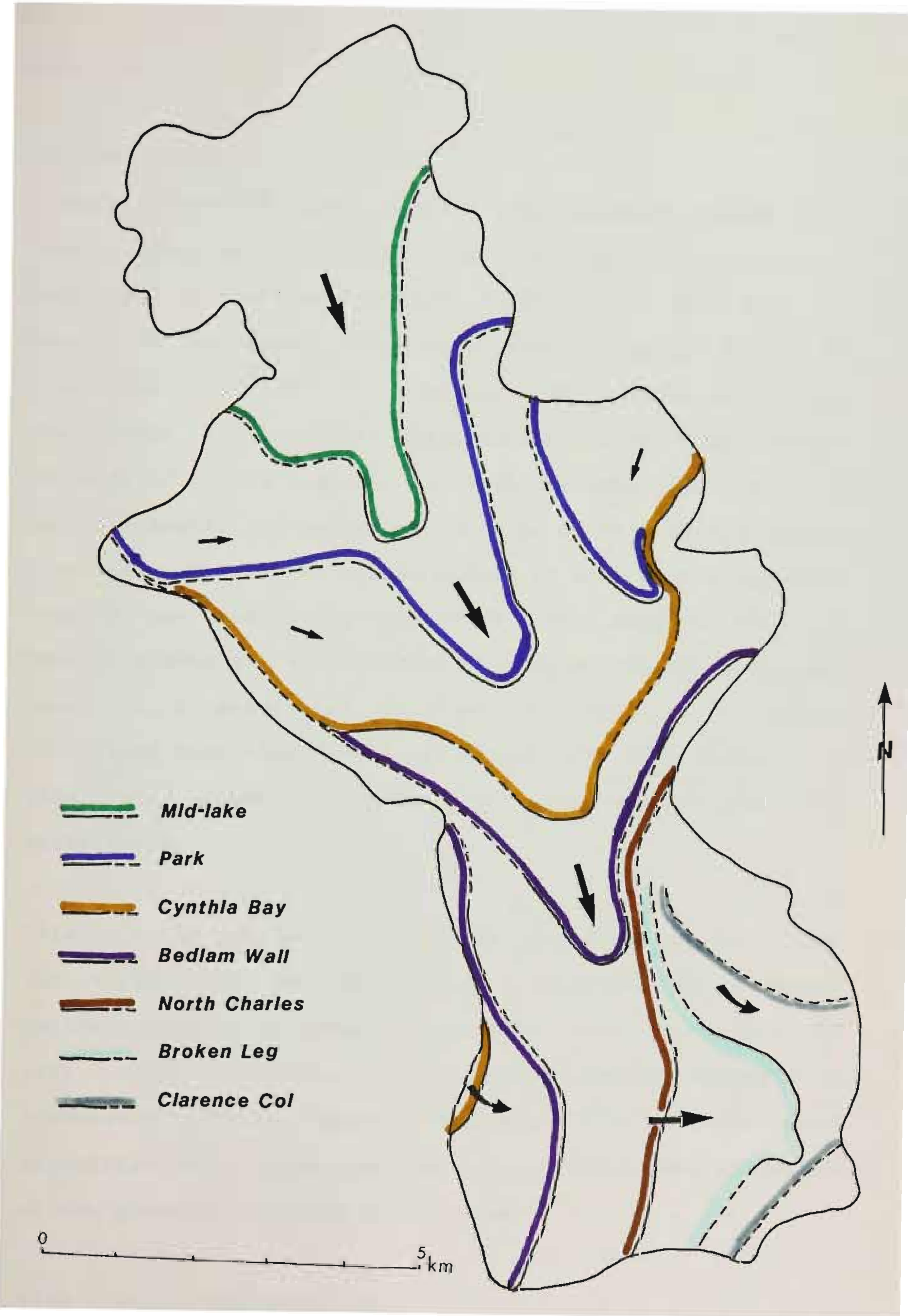
A Glacial deposits

Marked contrasts exist between the small moraines of the Travellers Rest Plateau and more massive moraines on the St. Clair Surface. The scarp slopes are mantled by bouldery ablation till which has commonly been reworked by slope processes. The tills may readily be differentiated on the basis of lithology into those that have emanated from the plateau and are composed entirely of dolerite, and those from Lake St. Clair which contain a variety of quartzitic erratics derived from the Permo-Triassic rocks (Table 6.1). Several phases of glaciation may be recognised (figure

Vicinity No.	4	6	10	13
%				
dolerite	82.0	67.0	100.0	100.0
quartz	16.0	8.0	-	-
Triassic sandstone	-	-	-	-
Permian & other	2.0	25.0	-	-

Table 6.1 Lithology of glacial deposits in the Travellers Rest Valley. Site locations are indicated on Figure 6.1.

Figure 6.3 Principal ice limits in the Travellers Rest Valley.



Mid-lake
Park
Cynthia Bay
Bedlam Wall
North Charles
Broken Leg
Clarence Col
0
5 km
N

6.3).

(i) Nive Till

A small outcrop of basal till has been exposed during the construction of a logging track at 820 m about 2.5 km north-west of the Clarence col (395 383). The till is exposed on the eastern flank of a probable bedrock knob. It is a dense grey basal till similar to the basal till clasts that occur in the East Rufus Moraine in the Hugel Valley (chapter 5). This deposit is best exposed in the Nive Valley (chapter 10) and is referred to in this thesis as the Nive Till. The larger clasts within it are generally well rounded and moderately spherical, but elongate and more angluar clasts are also present. Smaller angular pebbles occur in a sandy-clay matrix. Iron staining has seldom penetrated more than five centimetres into the surface of this very compact deposit, and the clasts are generally unweathered.

This till can only be fragmented by repeated hammer blows and borders on description as a tillite. It consists entirely of dolerite clasts which show that it cannot be part of the Permian tillite and strongly suggests a provenance on the Central Plateau. No moraines are associated with this deposit which is believed to be the oldest glaciial sediment in the area.

(ii) Clarence Col Moraines

Erratics of white quartzite occur in slope deposits at 780 m in the Clarence col and must have been derived from pre-existing till deposits. These erratics confirm the erosional evidence that ice passed through this col and that it was of westerly provenance.

Weathered till forms a lateral moraine immediately west of the col. The till extends to an elevation of at least 800 m on the northern side of the Mt. Charles ridge at which point it is overlain by slope deposits. Isolated zones of red basal till which has a blocky subangular pedal structure occur within this moraine. The less advanced lithification of this basal till suggests that it is much younger than the Nive Till. During this phase of glaciation abundant meltwater probably discharged eastwards via the Clarence River and southwards via the meltwater channel at Broken Leg Ck.

(iii) Broken Leg Moraines

A large latero-terminal moraine complex occurs 1.2 km west of the Clarence Col adjacent to Broken Leg Ck. Rock ridges that extend towards the valley floor at this point suggest that the complex may be rock cored. The moraine complex is convex towards the east, contains numerous quartzite clasts and appears to mark a major ice limit. The till is well weathered but cobbles and pebbles of sandstone and mudstone remain recognisable in some exposures. Meltwater was discharged eastwards and was impounded between the distal face of the moraine complex and the Clarence Col (figure 6.4). Dolerite predominates in the coarser fractions of the outwash. Excavation in the plain east of the moraine crest revealed over a metre of massive clays overlying outwash

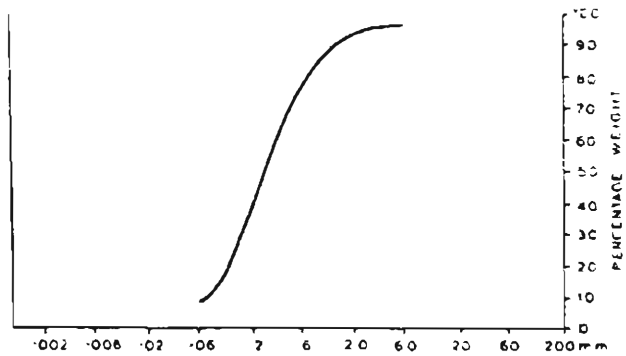


Figure 6.4 Cumulative curve of moderately well sorted glaciofluvial sediment from adjacent to the Broken Leg Moraine (after Derbyshire, 1967).

Figure 6.5 Cobbles on the southern beaches of Travellers Rest Lake have been reworked from glaciofluvial deposits and attest wave action due to northwesterly winds that have blown down the lake during the Holocene. The Traveller Range is visible in the background.

gravels.

A section close to the moraine face reveals 2 m. of current bedded sand and silt. Boulders become increasingly numerous towards the top of this section and some of the boulders disturb the bedding of the sands. This suggests deposition in an ice contact situation (Derbyshire, 1968a).

A section in the proximal face of the moraine at 770 m close to the Broken Leg meltwater channel reveals 1.5 m of basal till overlain by 2 m of rhythymically bedded glaciolacustrine sediments capped by 1 m of angular slope deposits. The glaciolacustrine sediments grade upwards from sandy beds at the base to clay. The fine "winter" layer is dominated by quartz while the coarser layer consists of plagioclase, pyroxene and quartzite, and if these sediments are true varves they would imply deposition over about 300 years (Derbyshire 1968a). The presence of dropstones indicates that the ice front must have remained fairly stable during this period.

(iv) North Charles Moraine

A low but broad end moraine extends across the plains north of Mt. Charles, 1.5 km. west of the Broken Leg complex. Its crest lies at about 750 m. and once again it may be rock cored. A probable extension of this moraine can be traced up the escarpment east of the Travellers Rest River to about 840 m. and appears to document the most recent occasion during which the Travellers Rest Glacier was confluent with the Derwent Glacier. A small outwash plain occurs between the North Charles and Broken Leg moraines. The till is moderately well weathered and some small coarse grained dolerite clasts have been decomposed into grus.

(v) Bedlam Wall Moraine

The eastern extension of the Bedlam Wall Moraine lies a short distance to the west of the Travellers Rest River and extends into the Derwent Bridge settlement. It can be traced discontinuously westward to the main Bedlam Wall Moraine in the St. Clair area. This moraine marks the limit of the Derwent Glacier following the retreat of the Travellers Rest ice from the St. Clair Surface to the plateau. The till is well weathered with the B horizon at least 150 cm thick and some argillitic clasts almost totally decomposed.

Remnants of a lateral moraine at 840 - 900 m on the scarp probably indicate the limit of the Travellers Rest Glacier at this time. Glaciofluvial sands and gravels were discharged down the Travellers Rest River onto the edge of the plain between the North Charles and Bedlam Wall moraines.

(vi) Cynthia Bay Moraines

The outermost of the Cynthia Bay Moraines can be traced eastwards from the Lake St. Clair area to the edge of the Travellers Rest catchment. This moraine ridge is narrow, less than 10 m high and topographically sharp. It forms part of the array of Thule-Baffin moraines that was described in chapter four. The till is little weathered. No ice from the Derwent entered the Travellers Rest area following construction of this moraine.

The corresponding moraine on the Traveller Plateau lies a few hundred metres south of Travellers Rest Lagoon. This correlation is based on the topographic freshness of the moraine and the limited weathering of the till. At least fourteen end moraines occur on the plateau. Basal till exposed just south of the lagoon is a dense, fissile, clayey sediment that contains moderately well rounded and unweathered dolerite clasts. Indurated ice contact glaciofluvial gravels occur on the edge of the plateau a short distance south of Travellers Rest Lagoon.

(vii) Park Moraine

The Park Moraine documents the withdrawal of the Travellers Rest Glacier to the southern end of Travellers Rest Lake. The main Travellers Rest ice separated from tributary ice that originated in a shallow valley head to the east. Low relief moraines downstream of Travellers Rest Lake were inundated by outwash during this phase, and Travellers Rest Lagoon was nearly filled by proglacial sediment.

(viii) Mid-lake Moraine

A low end moraine extends into the Travellers Rest Lake about half-way along its length. Echo-sounding has revealed the southern part of the lake to be generally less than 10 m deep. Upstream of the mid-lake moraine lies a deeper basin. The paucity of moraines further upstream suggests that the ice retreated from this position at a uniformly rapid rate. Glacial deposits are scarce north of this point.

B Non-glacial deposits

(i) Slope deposits

Almost the entire Travellers Rest catchment has been glaciated, with the exception only of the summit of Mt. Charles, Deposits of a non-glacial origin are therefore limited to those that have accumulated since the retreat of the ice.

Solifluction deposits are the most widespread nonglacial sediments. The angular dolerite clasts in these mantles have been derived from exposed bedrock on the southern scarp of the plateau and on the Mt. Charles ridge. In some cases tills that were deposited on the steep slopes have also provided a source of debris and it is commonly difficult to distinguish between reworked till and the periglacial deposits in these areas if erratic clasts are not present. Solifluction mantles do not occur within the ice limits of the Cynthia Bay phase but 1 - 3 m of slope deposits commonly overlie the older glacial deposits.

Present day periglacial activity is restricted to minor sorting of small calibre detritus in patches of bare regolith on the plateau (Jennings, 1956) and frost heaving by pipkrake development at lower elevations.

Rockfall talus has accumulated above 850 m beneath the dolerite clifflines on the crest of the Mt. Charles Ridge.

(ii) Alluvial, peat and beach deposits

Fluvial deposits include coarse gravels that have been reworked from outwash by the Travellers Rest River. Alluvial silts have been deposited to a maximum depth of 50 cm by Broken Leg Ck. and other streams that cross the outwash plains west of the Clarence Col. A variable thickness of minerogenic clay occurs beneath some of these silts and may be fluvial or niveofluvial rather than glaciofluvial. Up to 1 m of fibrous peat has accumulated on the surface of the buttongrass plains.

Beach gravels at the southern end of Travellers Rest Lake (Figure 6.5) have been reworked from the glacial sediments by wave action brought about by northwesterly winds that have blown down the length of the lake during the Holocene.

DATING AND DISCUSSION

The outstanding erosional legacy of glaciation in the Travellers Rest catchment is the upland of the Traveller Plateau with its many rock basins. The Midlake and Park Moraines that occur on the southwestern edge of this ice abraded plain are not represented on the St. Clair Surface within the Travellers Rest catchment. The surface gradient of the Traveller Glacier was low at this time and it is probable that little erosion was achieved. The possible extension of the glacial trough that contains Travellers Rest Lake was filled by glacial and glaciofluvial deposits.

The morphological relationships suggest that the outermost moraines on the Traveller Plateau were constructed at the same time as the Cynthia Bay moraines on the St. Clair Surface

The Bedlam Wall, North Charles, Broken Leg and Clarence Col moraines mark stages in the retreat of the Traveller lobe of the Derwent Glacier. The Clarence Col moraines and associated erratics indicate that initially the ice passed eastwards through the Clarence Col into the Clarence River area (chapter ten).

The Nive Till occurs beneath the till of the Broken Leg phase and must therefore predate it. Although it occurs within the limits of a distributary lobe of the Derwent Glacier that extended eastwards through the Clarence Col the Nive Till contains only dolerite clasts. This suggests that it was deposited by ice which originated on the Traveller Plateau. This indicates that either the Traveller Glacier was deflected eastwards on the northeastern margin of the Derwent lobe, or that it was locally more vigorous than the Derwent Glacier at the time.

The erosional morphology of the area hints at the occurrence of an early phase of glaciation during which ice covered almost all the Traveller Plateau, scarp slopes and St. Clair Surface, and may have overtopped much of the Mt. Charles ridge. No depositional evidence has been obtained which supports this proposition. Glacial erosion seems likely to have accounted for only a relatively small proportion of the total retreat of the scarp which is essentially a preglacial landform.

Postdepositional data are presented in Table 6.2. The Midlake, Park and Cynthia Bay deposits are similar and represent the most recent glaciation of the area. Weathered clasts are found to a depth of no more than 50 cm in the Cynthia Bay moraines where A Cox Cu soil profiles (Birkeland 1984) are the norm. The average thickness of dolerite weathering rinds in the Cynthia Bay Till is less than 1.8 mm.

Slope deposits that occur outside the Cynthia Bay limits and overlie the earlier tills are comparably weathered to the Cynthia Bay Till. This suggests that the slope deposits are of approximately the same age as the Cynthia Bay Till and that they are periglacial in origin.

Dolerite weathering rinds in the Bedlam Wall, North Charles, Broken Leg and Clarence Col moraines are twice the thickness of those in the later tills. The totally decomposed pebbles of coarse grained dolerite and argillitic clasts that can be found in these moraines contrast with minimal weathering in the Cynthia Bay Till. There is little difference in the degree to which the Clarence Col, Broken Leg, North Charles and Bedlam Wall deposits have been modified by postdepositional processes. They are believed to represent successively diminishing phases of the same glaciation.

The dense grey basal Nive Till, exposed on the logging track 2.5 km north-west of the Clarence Col, is a highly

moraines[a]	CC	CC	CC	CC	BL	BL	BL	NC	BW	CB	CB	CB	CB	CB	CB
vicinity no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
material[b]	T	B	T	T	T	T	T	T	T	B	O	T	T	S	S
A. MORPHOSTRATIGRAPHY															
relative position	outside 5	outside 4	outside 5	outside 5	inside 4	outside 8	outside 8	inside 7	inside 7	inside 9	inside 9	inside 9	inside 11	inside 8	
crest/site elevation(m)	820	780	800	795	790	780	760	730	740	945	950	740	950	795	
max. depth of burial (by)(cm)[c]	200	-	100(S)	150(S)	150(S)	100(S)	100(S)	-	-	20(S)	-	-	-	-	-
B. SUBSURFACE CLASTS															
dolerite rinds - max.(mm)	-	-	6.4	5.4	5.6	9.8	4.2	3.9	5.6	-	1.4	2.3	1.75	1.9	2.2
- min.	-	-	1.8	2.1	2.2	1.1	1.5	1.7	1.3	-	0.3	0.6	0.32	0.8	0.4
- mean	-	-	3.5	2.9	3.4	3.2	2.8	2.5	2.7	-	0.71	1.8	1.2	1.1	1.8
- SD	-	-	1.3	0.8	1.0	1.3	0.7	0.6	1.1	-	0.39	0.69	0.2	0.1	0.8
- hardness(1-5)	-	-	2	2	2	2-3	2	1	2	-	1	1	1	1	1-2
coherence of argillites	-	-	3	3	2	3	2	3-4	3	-	0	1	0-1	1-2	1-2
clast surface colours	-	-	5YR	5YR	5YR	5YR	7.5-5YR	5-7.5YR	5YR	-	10YR	7.5YR	10YR	5YR	5YR
C. SUBSURFACE MATRIX															
profile type	A Bt	-	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt	-	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A Bt
colour of B horizon	-	-	5YR	5YR	7.5YR	5YR	7.5YR	7.5YR	7.5YR	-	10YR	10YR	10-7.5YR	7.5-5YR	7.5-5YR
depth of B horizon(cm)	-	-	> 150	> 300	> 150	> 300	> 200	> 150	> 150	-	-	-	-	-	-
depth to lowest Cox	-	-	> 150	> 300	> 150	> 300	> 200	> 150	> 150	-	-	-	-	-	-
pH at ~1m	-	-	-	-	-	-	-	-	-	-	> 50	> 80	60?	50?	?
clay size fraction										-	-	-	-	-	-
mineralogy	-	-	-	-	-	-	-	quartz kaolinite	-	-	-	-	-	-	-
development of clay films	-	-	4	4	3	4	3	3	3	-	1	1	1	1	-
organic at ~100cm	-	-	-	-	-	-	-	2.5	-	-	-	-	-	-	-
degree of lithification of basal till	5	-	2	3	2	3	3	3	2	-	-	0	1	1	-
clast socket staining	-	-	-	4	4	3	4	3	4	-	-	0	1	1	-
coarse matrix: % > 1cm	-	-	-	-	-	61.8	50.0	-	-	-	53.0	-	-	-	-
%1-0.5cm	-	-	-	-	-	17.6	12.5	-	-	-	9.0	-	-	-	-
% < 0.5cm	-	-	-	-	-	20.6	37.5	-	-	-	38.0	-	-	-	-
E. DEPOSITIONAL LANDFORMS															
moraine crest width(m)	-	-	-	-	70-120	-	-	-	-	-	-	12	10	-	-
moraine slopes - proximal	-	-	-	-	3-9	-	-	-	-	-	-	20	15	-	-
distal	-	-	-	-	5-13	-	-	-	-	-	-	20	25	-	-
degree of fluvial dissection	-	-	-	-	3	1	2	3	3	-	-	0	0	1	-
F. ERODED BEDROCK LANDFORMS															
width of joint openings (mm)	-	12	-	-	-	-	-	-	-	1-2	-	-	-	-	-
depth of joint openings (mm)	-	35+	-	-	-	-	-	-	-	< 2	-	-	-	-	-
height of tors (m)	-	3	-	-	-	-	-	-	-	-	-	-	-	-	-

(a) CC= Clarence Col; BL= Broken Leg; NC= North Charles; BW= Bedlam Wall; CB= Cynthia Bay.

(b) T= till; B= bedrock; O= outwash gravels; S= solifluction deposit.

(c) Refers to burial of the same depositional unit in the same vicinity but not at the exact location where subsurface data were obtained. The latter have only been recorded from unburied sites. S= solifluction deposit.

Table 6.2 Post-depositional modification of deposits, Travellers Rest Valley. Site locations are indicated on figure 6.1.

distinctive sediment and is so well lithified as to suggest that it is considerably older than the red basal till that overlies it.

The Cynthia Bay Moraines on the St. Clair Surface are of Thule-Baffin type and mark shortlived, perhaps annual positions of an ice front that was undergoing steady retreat. The more massive end moraines that occur outside the ice limits of the Cynthia Bay phase document long-lived positions of the ice margin. Rhythymically bedded clays in the Broken Leg Moraine suggest that the margin of the Derwent Glacier may have been maintained in the same position for centuries. However, this need not imply that the glacier was in a steady state throughout this time. Muller (1958) has recorded that the Khumbu Glacier on Mt. Everest thinned by 70 m between 1930 and 1956 but that its snout did not retreat during this period. Hence, a glacier may lose mass without retreating. The greater bulk of the glacial deposits outside the Cythia Bay ice limits compared to the deposits upstream of Cynthia Bay is therefore probably due to the older glacial events having been of greater duration.

chapter seven
THE NAVARRE VALLEY

The Cynthia Bay Moraines that form the southern shoreline of Lake St. Clair (chapter 4) stand on the north-eastern margin of the Navarre basin at the upstream end of a broad glacial trough (figure 7.1). The Derwent Trough, which extends further to the south, is a broad extension of the St. Clair glacial trough. At the southern end of the Derwent Trough lies a subdued moraine and outwash tract where four ice tongues previously coalesced on the St. Clair Surface between Mt. Rufus and the King William Range (Figure 7.2).

EROSIONAL LANDFORMS

A Glacial erosion:

The highest cirque in the area lies at 1220 m and contains the Gingerbread Hut, south-east of Mt. Rufus. This cirque is shallow and broad and is overlooked to the north by the East Rufus Moraine (chapter five). Over-ridden cirques and valley steps occur at 1020 -1040 m in the same valley, while another over-ridden cirque lies 3 km east of the summit.

The principal cirque in this part of the King William Range is located on the eastern slopes of Mt. Pitt. It has a headwall 180 m high but no well defined threshhold.

The most striking erosional landform is the Derwent Trough between Mt. Rufus and Bedlam Wall. The eastern wall of this

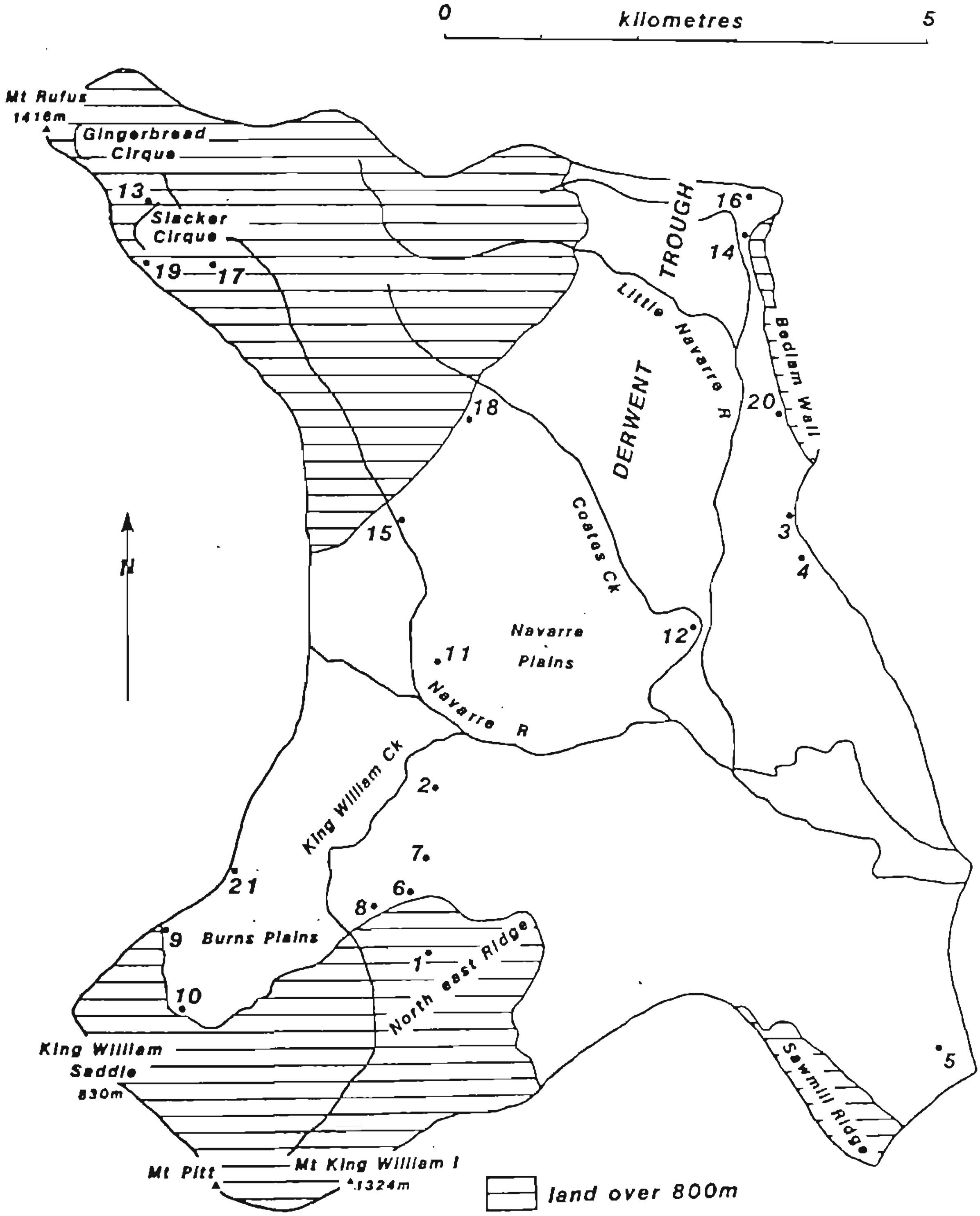


Figure 7.1 Locality map, Navarre Valley. Numbers refer to site locations in tables 7.1 and 7.2.

LANDFORMS

cirque
over-ridden cirque
glacial trough
over-ridden margin
diffluence col
valley step
ice-abraded outcrop
areally scoured terrain
glacial striae
rock-basin lake

moraine ridge
moraine hummock
moraine-dammed lake
erratics
erratics of Nive till
reworked Permian erratics or large lagstones

meltwater channel

nunatak
nivation cirque
rock crevasse or dilation trench
slab topple
rock glacier
solifluction lobe or terrace
patterned ground
landslip
artificial lake

SEDIMENTS

GLACIAL

Cynthia Bay till/outwash
Beehive "
Powers Creek "
Clarence "
Stonehaven "
Nive Till "

NON-GLACIAL

alluvial silt &/or peat
talus
Solifluction deposits

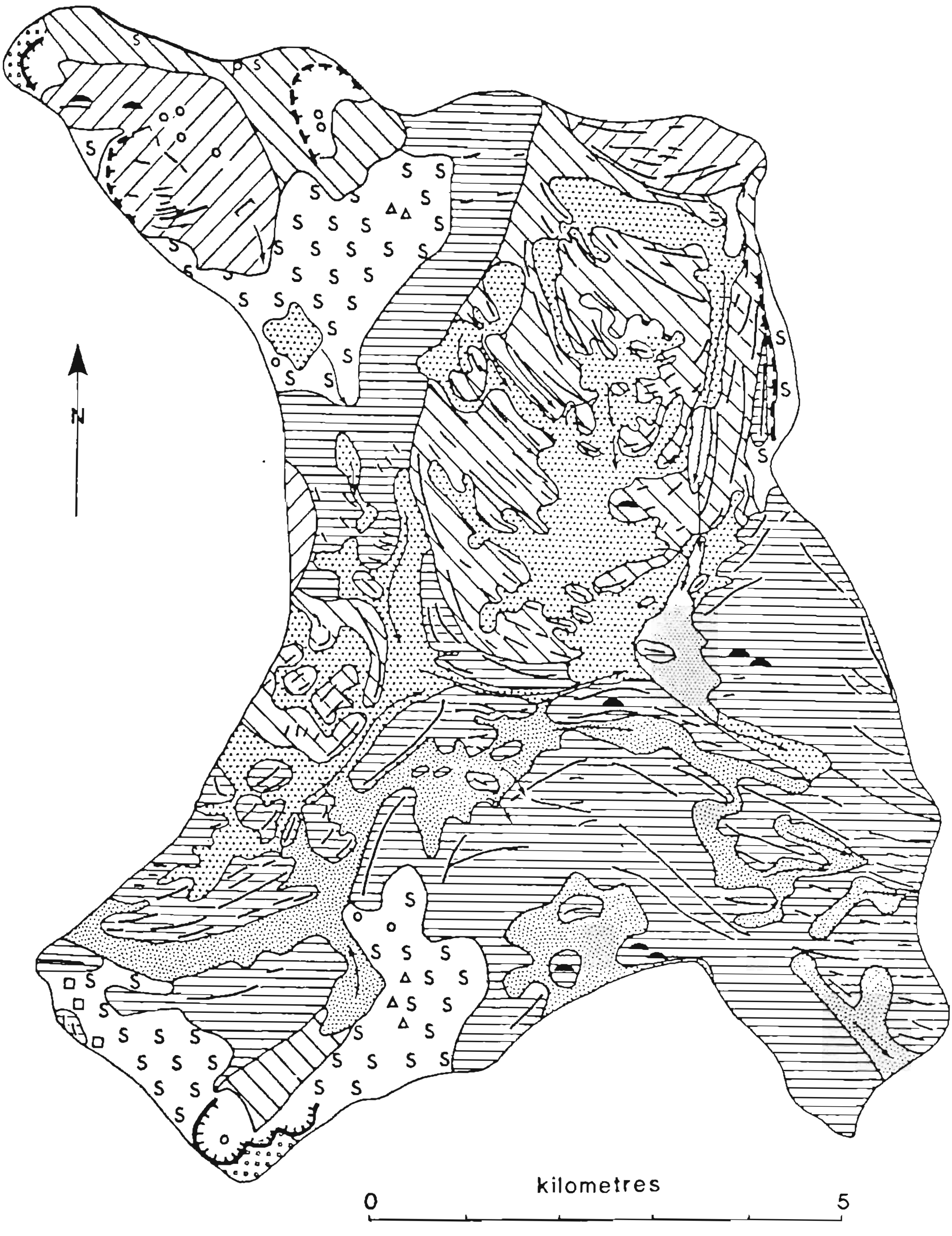


Figure 7.2 Glacial geomorphology of the Navarre Valley. See Figure B2 for key.

trough is considerably steeper than that to the west because it served to deflect the St. Clair piedmont glacier westwards onto the Navarre Plains and hence was subjected to severe truncation. In addition, the strength of the dolerite at Bedlam Wall permits the maintenance of a steep free face which contrasts with the gentler slopes formed on the weaker Permo-Triassic rocks which form the western wall of the trough. The western wall was also over-ridden by ice generated on Mt. Rufus. A smaller glacial trough is present beneath Mt. Pitt in the King William Range

King William Saddle is a shallow diffluence col on the edge of the St. Clair Surface. Derbyshire (1967) considered that ice from the Derwent Glacier escaped westwards through this saddle into the Franklin River system via the Surprise Valley. However, the pattern of moraines south of Lake Dixon in the Franklin Valley indicates that the ice that flowed through King William Saddle came from the Franlin and not the Derwent.

Ice-abraded rock surfaces indicate that the arms of the shallow valleys on Mt. Rufus and also those of the trough below the Mt. Pitt cirque have been over-ridden. Bedrock at 950 m on the eastern slopes of Mt. Rufus has also been smoothed and plucked by ice that flowed southeastwards. If this smoothing was caused by the Derwent Glacier rather than ice from Mt. Rufus it would imply that the Derwent Glacier was at least 200 m thick at this point. The northern slopes of the King William Range give an impression of having been ice smoothed below about 880 m while a spur that descends north-eastwards from the range appears to have been smoothed below 900 m. The summit of the Bedlam Wall ridge (880 m) is also broadly rounded, although shattered tors up to 5 m high occur at the highest point. These tors indicate that the summit was not over-ridden by glaciers during the last glaciation.

B Glaciofluvial erosion:

Meltwater channels are incised in bedrock downstream from the high cirques, but the most striking examples of glaciofluvial erosion occur on the low terrain between the King William Range and Mt. Rufus. As the Derwent Glacier retreated upvalley this meltwater drainage slipped progressively down the nose of the King William I ridge and cut through a moraine belt at the foot of the ridge. The Navarre River now flows through this channel. Further east the Little Navarre River flows southwards close to the foot of Bedlam Wall and the likely position of a former basal meltwater conduit. Streams such as Coates Ck. meander through moraine belts via former proglacial meltwater routes.

C Non glacial erosion

Only limited slab toppling is evident in the Navarre basin due to the relatively low relief and low altitude of the dolerite hills. The low relief has meant that the dolerite has generally been over-ridden and smoothed rather than steepened, and the low altitude has meant that even those areas which were steepened have not been subject to major periglacial activity subsequently. Free faces formed on

Permo-Triassic rocks on Mt. Rufus are the result of initial glacial steepening followed by spring sapping. Minor solutional sculpturing of sandstone pavements in the form of pans and runnels is abundant on the slopes of Mt. Rufus and to a lesser extent Mt. King William I (Figure 7.3).

A few small longitudinal nivation hollows (French, 1976) up to 15 m wide and several tens of metres long are present high on Mt. Rufus.

Fluvial erosion has largely been limited to some deepening of pre-existing channels formed in unconsolidated sediments and the stripping of exposed soil surfaces. It has been greatly aided by human activities where the regolith beneath the surface peat has been exposed by bull-dozing, a striking example of this being the track to the former fire-spotting tower on Mt. King William I. Two canals have been excavated for hydro-electric purposes. Both of these have used the courses of former meltwater channels that discharged from the margin of the Franklin Glacier into the Navarre basin.

DEPOSITIONAL MORPHOLOGY AND SEDIMENTS

A Glacial deposits

The depositional morphology of the Navarre basin is dominated by an extensive array of end moraines on the lower terrain. These have been deposited by the Derwent Glacier which entered the area from the north-east, and by the Franklin Glacier which breached the present Franklin - Navarre divide from the north-west. The Gingerbread Glacier which descended the Navarre Valley from Mt. Rufus and the

Pitt Glacier from the King William Range also descended to low altitudes.

The Derwent Trough has been infilled by till and outwash, the thickness of which is unknown. Early estimates of 120 - 150 m. (Clemes, 1925) and 220 m (Lewis, 1939) seem excessive in the context of surrounding bedrock morphology ; a more recent estimate of 45 m (Derbyshire, 1971a) cannot be confirmed without geophysical investigations which were not feasible during the present study.

Resolution of the former ice flow directions is sometimes difficult: the morphology of some of the downvalley moraines has been modified by meltwater erosion; some low moraines have been inundated by later outwash; sediments suitable for fabric analysis are poorly exposed; and the similar bedrock terrain over which both the Derwent and Franklin glaciers passed means that provenance of the till deposits is difficult to determine on the basis of lithological evidence (Table 7.1). Several major ice limits have been identified (Figure 7.4)

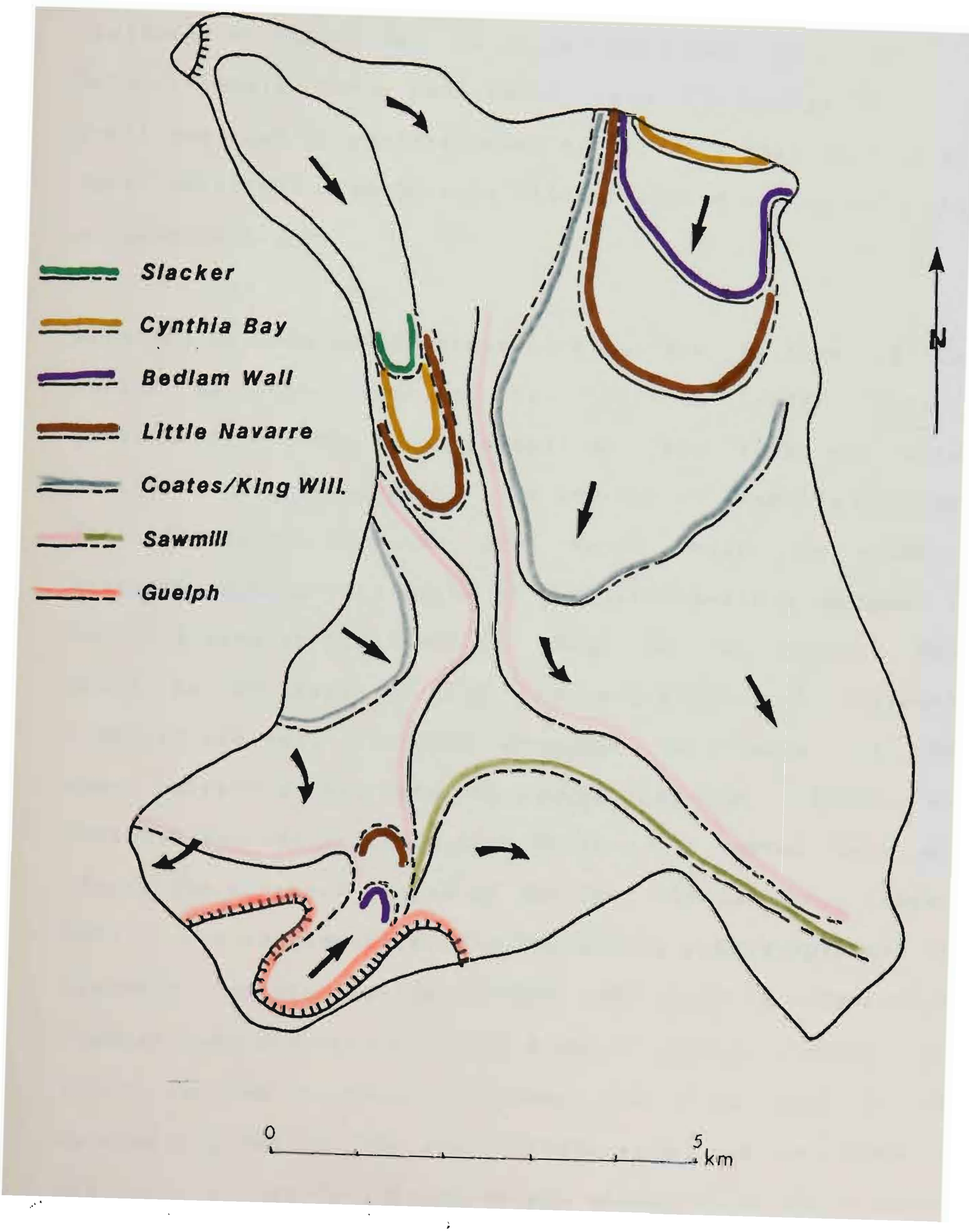
(i) Guelph moraines:

The Guelph moraines occur on the northern and north-eastern slopes of the King William Range and extend southwards into the Guelph basin. These poorly defined and degraded moraines occur at a higher elevation on the north-eastern side of the range than they do further west (900 m. vs 860 m) due to the massive discharge of ice from Lake St. Clair. They extend onto a ridge of Permian sandstone and dolerite

Figure 7.3 Solution pans formed in Triassic sandstone on the southern flanks of Mt. Rufus.

overleaf

Figure 7.4 Principal ice limits in the Navarre Valley.


Slacker
Cynthia Bay
Bedlam Wall
Little Navarre
Coates/King Will.
Sawmill
Guelph
N
0
5
km

rocks west of Lake King William. Ground moraine and hummocky moraine occur north-east of King William Saddle and on the Guelph-Navarre divide. The moraine morphology implies that the Derwent and Franklin glaciers were confluent at this time. It is unlikely that ice from the Derwent would have penetrated into the Guelph Valley as local ice that accumulated east of Mt. King William I would have deflected the Derwent Glacier around the eastern side of Sawmill Ridge.

Erratics of Owen Conglomerate occur on the surface of the Guelph moraines south of King William Saddle. This is problematic because no outcrops of this rock are known further upstream in either the Derwent or Franklin valleys. Their deposition from the west would imply the eastward movement across the grain of the landscape for at least 14 km. of a massive ice sheet at least 500 m thick. This would be contrary to all the other geomorphic evidence (chapter twelve). The most probable explanation is that these erratics are pre Cainozoic in age. Tillite and striated bedrock indicate that Permian ice moved eastwards across the northern slopes of the King William Range (Ahmad, 1957). The conglomerate erratics on the Guelph moraines are probably bedouin boulders that were carried eastwards by Permian ice, reworked from the Permian glacial deposits and later carried a short distance westwards again by late Cainozoic glaciers. Permian tillite crops out just east of the saddle, and Owen Conglomerate occurs as erratics within it (Ahmad, 1957).

The provenance of a series of quartzitic erratics at 950 m on the eastern slopes of Mt. Rufus is uncertain, but if they were deposited by the Derwent Glacier an ice thickness of at least 200 m is implied.

(ii) Sawmill Ridge moraines:

This moraine complex is wrapped around the rock ridge west of Lake King William.These moraines extend discontinuously around the northern slopes of the King William Range towards King William Saddle. As the ice receded and the Derwent and Franklin glaciers ceased to be confluent a broad interlobate moraine complex was laid down across the plains between Mt Rufus and Mt. King William I. It would appear from this that little if any ice from the Derwent Valley could ever have escaped westwards through King William Saddle (cf. Derbyshire, 1963). Instead, the Derwent Glacier was forced to flow southwards along the eastern side of Sawmill Ridge by the Franklin and Guelph glaciers. The tills of the interlobate complex are locally enriched in Permo-Triassic rocks which suggests that it is constructed upon a rock ridge (Table 7.1)

The end moraines on the southern shore of the Navarre arm of Lake King William contain meltout tills and some bedded gravels. These occasionally dip upstream with the ice proximal surface of the moraine plastered by red basal till that has a blocky subangular pedal structure. This suggests that the ice margin was active and advanced locally over ice marginal sediments that had already been deposited.

Vicinity No.	5	7	11	17	18a	18b	19	21
%					> (cobbles)	< (cobbles)		
dolerite	69.6	56.2	95.0	30.0	90.0	27.8	28.0	28.0
quartz	26.1	22.3	5.0	60.0	10.0	16.7	14.0	49.0
Triassic sandstone		15.2?	-	5.0	-	55.5?	-	6.0
mudstone	4.3	7.3	-	5.0	-	-	51.0	17.0
other			-		-	-	7.0	-
provenance:	Derwent	Franklin?	Derwent	Rufus	Rufus	Rufus	Rufus	Franklin

Table 7.1 Lithology of glacial deposits in the Navarre area. Site locations are indicated on figure 7.1.

Meltwater was later impounded between the snout of the Franklin Glacier and the north-eastern spur of Mt. King William I. An exposure at 810 m to the east of King William Creek reveals rhythymites that have been glacio-tectonically deformed. They are stiffly pliable to brittle, lightly oxidised, and fragment into small cubes when dried out (Figure 7.5).

A sharp crested lateral moraine bounds the eastern margin of the Mt. Pitt trough from 1050 - 830 m. This has been constructed upon ice - smoothed bedrock by the glacier that flowed down the Pitt Valley.

(iii) Coates Creek - King William Creek moraines:
A low moraine loop extends across the Derwent Trough 4 km upstream from the Sawmill limit. This moraine forms part of the interlobate complex south-east of Mt. Rufus. A further moraine ridge that was deposited by the Franklin Glacier on the western side of the interlobate complex north of King William Creek is probably of equivalent age. Once the Derwent Glacier had withdrawn to the Coates Ck. limit, meltwater from the south-eastern margin of the Franklin Glacier discharged via the channel now occupied by the lower reaches of the Navarre River. This also carried meltwater from the Gingerbread Glacier on Mt. Rufus, and from the Pitt Glacier which may have briefly extended onto the plains after the Franklin ice withdrew.

These moraines consist largely of meltout and ablation tills, and in some cases are constructed upon bedrock

Figure 7.5 Weathered rhythmites in the Sawmill Ridge moraine complex east of King William Creek.

Figure 7.6 Large glacial boulder of Triassic sandstone in the Slacker Cirque, Mt. Rufus. Note broad solution runnels that have developed down its flanks. Solution pans up to 50 cm. wide occur on top of the boulder, which probably dates from the Cynthia Bay phase.

ridges. Red basal till occurs in the moraines near Coates Ck. Angular to sub-angular clasts in much of the ground moraine suggest that it is primarily of supraglacial origin. Angular slabs of Triassic sandstone from Mt. Rufus that are up to 2 - 5 m long occur on the surface of the interlobate complex. Hummocky moraine occurs north-west of the Navarre arm of Lake King William on gently sloping terrain that appears to lie on the downvalley side of a bedrock ridge. This suggests slow <u>in situ</u> melting of stagnant ice which may have become detached from the glacier front as it retreated.

(iii) Little Navarre moraines:

A discontinuous and gullied end moraine extends across the valley floor between Bedlam Wall and Mt. Rufus about 2.5 km upstream from the Coates Ck. limit. Few sections are available through the slopes below the Bedlam Wall ridge. However, angular surface boulders suggest that local slope deposits that were deposited against the ice margin form a significant component of these moraines. A broad outwash plain occurs downvalley of the Little Navarre Moraines. Protruding large erratics suggest that the glaciofluvial gravels only thinly veneer the ground moraine in places. Meltwater escaped via at least three channels through the Coates Ck. moraines (Figure 7.2).

The Franklin Glacier appears to have retreated steadily after it withdrew from the King William Creek limit, but the considerable bulk of some of the moraines suggests that the ice front remained active. The moraine remnants on Burns

Plains are generally broader than they are wide due to their having been dissected by meltwater. The Gingerbread Glacier probably still extended to below 800 m. (Figure 7.6). The less favourably situated northeastward facing Pitt Glacier had retreated well into its trough, but continued to discharge some fluvioglacial sediment onto the south-eastern corner of Burns Plains.

(iv) Bedlam Wall moraine:

A subdued end moraine ridge extends most of the way across the Derwent Trough west of Bedlam Wall. Its crest lies at about 760 m. A small outwash plain is impounded between it and the Little Navarre moraines. Meltwater from this limit discharged via Coates Ck. and the Little Navarre River.

(v) Cynthia Bay moraines:

The sequence of end moraines that forms the southern shoreline of Lake St. Clair extends into the mouth of the Derwent Trough. These moraines have an amplitude of around 10 m. and successive crests tend to occur 50 - 70 m. apart. The highest crests are at 760 m. The outermost Cynthia Bay moraines document the most recent occasion on which meltwater from the Lake St. Clair trough discharged into the present Navarre basin. The principal meltwater drainage route at this time was via the Little Navarre River.

The till is only slightly weathered and at the northern end of the Bedlam Wall Ridge is overlain by less than 50 cm of slope deposits derived from the earlier tills that flank the ridge.

The Gingerbread Glacier extended no lower than about 900 m while there is little evidence for any active ice remaining in the Pitt Cirque at this time.

(vi) Slacker Moraines:

Small lateral moraines and associated hummocky drift occur in the over-ridden Slacker Cirque (grid ref. (8013)266 345) on the south-eastern slopes of Mt. Rufus, and a slightly weathered lateral moraine occurs on the southern arm of the higher Gingerbread Cirque (263 355). The lower cirque is the more shaded of the two and the fairly abundant moraines here contrast with the soliflucted northern slopes of the Gingerbread Cirque. This suggests that ice in the Slacker Cirque remained active longer than that in the more open Gingerbread Cirque. The cirque 3 km east of the summit that had previously been over-ridden by ice from the Gingerbread Cirque was by now ice free.

B Non - glacial deposits:

(i) Slope deposits

Solifluction deposits are the most widespread non-glacial sediments in the Navarre basin. Boulders 3-4 m in length occur in these deposits which have moved over slopes of as little as 5° in a silty-clay matrix. Many of these deposits include spheroidally weathered boulders that have been derived from bedrock weathering profiles. Up to 2 m of solifluction deposit overlies the lateral moraine of the Bedlam Wall phase along the foot of the Bedlam Wall ridge.

The summit block of Mt. Rufus consists of a severely shattered remnant of Jurassic dolerite from which block slopes extend a short distance downslope. These terminate in ice-smoothed rock ridges. The summit rubble is probably the combined result of glacial sapping, dilation and intense frost action upon a nunatak. However, such block accumulations are likely to have a complex history and may be the result of freeze-thaw action upon bedrock that was deeply weathered during the Tertiary (Caine, 1966). Limited protalus occurs in the Gingerbread Cirque.

Rockfall talus is present on Mt. King William I where fresh angular fragments overlie much larger cambered blocks that have surface weathering rinds of 1.4 - 2.0 mm.

(ii) Alluvial, peat and archaeological deposits

Alluvial deposits are limited to accumulations of silt in overbank situations and local topographic depressions. These seldom exceed 1.5 m in depth on the Navarre Plains and are generally less than 1 m thick in the swales between the Cynthia Bay Moraines and in the Gingerbread Cirque. Fibrous peats have accumulated beneath the buttongrass plains.

Thousands of aboriginal stone flakes are scattered on the East Rufus moraine (Figure 7.7). The clasts from which they have been struck were quarried from the till. Other archaeological sites have been found adjacent to the lower Navarre River (Figure 13.14). Because aborigines obviously ranged widely over this area the possibility exists that

Figure 7.7 Aboriginal stone tools have been struck from clasts of calc-silicate hornfels in the East Rufus Moraine. This prehistoric quarry is the largest archaeological site that has been recorded in the study area. It lies on an exposed and often windy ridge at an elevation of nearly 1400 m.

human carriage of clasts for the purpose of stone tool manufacture might be misinterpreted as a reflection of ice flow patterns. In an attempt to minimise this risk several factors have been taken into account before clasts have been accepted as glacial rather than manuports. Clasts have only been interpreted as glacial where there was no evidence of human modification such as multiple fracturing, bulbs of percussion or retouch. However, large clasts have generally been accepted as glacial erratics even if working or associated waste flakes are present, in the belief that the aborigines are unlikely to have carried them far when suitable materials are generally widespread.

3. DATING AND DISCUSSION

The deposition of organic-rich silt and the accumulation of peat postdates climatic amelioration and stabilisation of the landscape. In the Gingerbread Cirque such silt overlies soliflucted till at the foot of the East Rufus Moraine. The solifluction terraces on this moraine do not appear to be forming at present despite the limited vegetation cover. This indicates that these terraces did not form under the present climate. The archaeological deposits on the East Rufus Moraine overlap the solifluction terraces but no stone flakes are exposed in sections that have been cut through the terraces by running water. This suggests that quarrying of the moraine by prehistoric aborigines postdates formation of the terraces.

The Slacker Moraines indicate that the ice in the Slacker Cirque remained active during deglaciation. Solifluction mantles over 1 m. thick occur in the middle reaches of the Navarre Valley. These suggest that the Navarre Glacier probably extended no lower than 900 m during the Cynthia Bay phase. A succession of small moraine ridges formed south of Lake St. Clair at this time.

The Bedlam Wall Moraine is the innermost of the style of massive and broad moraines that occur downstream of Lake St. Clair in the Derwent Trough. The Bedlam Wall, Little Navarre and Coates Creek-King William Creek moraines all mark significant retreats of the Derwent Glacier, but the mass of material deposited suggests that the ice mass remained active.

Morphological criteria suggest that two major lobes of ice were confluent in the middle of the present Navarre basin. One of these was the Derwent Glacier which entered the area from the north-east. The other was the Franklin Glacier which entered from the north-west. The Franklin lobe prevented the Derwent Glacier from extending westwards through King William Saddle and was itself forced in that direction by Derwent ice.

The Derwent and Franklin piedmont lobes remained confluent until the latter part of the Sawmill Ridge phase, during which time the ice front had been active. They subsequently retreated broadly in phase with one another due to the fact that diffluence through the col at the head of the Franklin meant that they were both effectively draining the same ice source. An interlobate moraine complex was constructed as they retreated. The Pitt Glacier had been dammed by the Franklin Glacier during the Sawmill phase but with the withdrawal of the Franklin ice may have briefly spilled out onto the margin of Burns Plains.

Erosional evidence suggests that at its maximum the Derwent Glacier was at least 200 m. thick east of Mt. Rufus, and may have over-ridden the Bedlam Wall ridge during an early phase of glaciation. Rounding of the slopes on the King William Range, if of glacial origin, suggests that at least 50 m. of ice passed westwards through King William Saddle and that the ice north-east of Mt. King William I may have been 160 m. thick.

The morphological evidence for the Bedlam Wall ridge having been over-ridden by ice is not supported by depositional evidence. However, despite the apparent absence of erratics the massive extent of the Derwent Glacier at its maximum (see chapters 9 and 12) makes it almost inevitable that over-riding by ice did take place. Probably this occurred most recently during the Guelph phase when the ice reached equivalent elevations on the King William Range 7 km further downstream in a much broader part of the valley.

This morphological evidence is confirmed by the pattern of post-depositional modification of the glacial deposits (Table 7.2). Soils that have developed on the Slacker and Cynthia Bay moraines have shallow A Cox Cu type profiles. In contrast, textural B horizons over 1 m. thick occur on the Bedlam Wall Moraine. This indicates that the Cynthia

moraines[a]	G	S	S	S	S	S	S	S	CKW	CKW	CKW	CKW	CKW	CKW	BW	BW	CB	CB	CB	CB	C
vicinity no.	1	2	3	4	5	6	7	8	9	10	11	12	21	13	14	15	16	17	18	19	20
material[b]	T	T	T	T	T(O?)	T	T	T	T	T	ST?ST	T	T	B	T	T	T	T	O	T&B	B
A. MORPHOSTRATIGRAPHY																					
relative position	outside 2	outside 3	outside 4	outside 5*	inside 5	inside 3	outside 9	outside 9	outside 21	outside 21	outside 12	outside BW	inside 8	inside 8?	inside CKW	outside CB	inside 15	inside 21	inside 21	inside 15	outside 6
crest/site elevation(m)	830	765	742	760	780	770	770	760	760	760	750	730	-	1130	730	770	760	795	755	1080	880
max.depth of burial (m) (by)[c]	2(S)	-	-	-	-	1(S)	0.5(S)	1.5(S)	1(S)	1.5(S)	2(ST?)	-	-	-	2(S)	-	0.5(S)	1(S)	-	-	-
B. SUBSURFACE CLASTS																					
dolerite rinds(mm)-max.	16.9	11.7	9.3	10.4	5.9	8.0	7.8	8.2	6.6	6.9	6.7	6.2	7.1	-	4.2	5.6	2.8	2.4	1.6	1.7	-
min.	3.4	2.1	3.1	3.0	2.8	1.4	2.7	2.4	2.2	2.9	0.6	0.7	0.9	-	0.5	0.9	0.7	0.9	0.1	0.3	-
mean	8.06	7.03	6.26	6.59	4.09	4.58	4.22	4.95	4.65	4.34	3.75	2.53	3.8	-	2.9	2.4	1.50	1.5	0.66	1.06	-
SD	3.88	3.35	2.03	2.2	0.96	1.49	1.46	1.54	1.14	1.12	1.99	1.23	2.07	-	0.9	1.06	0.46	1.04	0.5	0.39	-
hardness	3	2	2	2	2	2	2	2	2	2	2	2/1	-	-	2/1	1	1	1	1	1	-
coherence of argillitic clasts(1-5)	4	4	5	-	2/3	2/3	2/3	3	3	2	4	3	-	-	3	2	2	2	2	1	-
clast surface colours	5YR	5YR	7.5-5YR	7.5-5YR	7.5-5YR	5YR	5YR	7.5-5YR	7.5YR	7.5-5YR	5YR	5-7.5YR	-	-	10-5YR	10-7.5YR	10-7.5YR	10-7.5YR	10-7.5YR	10-7.5YR	-
C. SUBSURFACE MATRIX																					
profile type	A Bt	A Bt	A Bt	A Bt	A Bt	A BT	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt		A Bt	A Bt	A Cox cu	A Cox cu	A Cox cu	A Cox cu	
colour of B horizon	7.5-5YR	7.5-5YR	5YR	7.5-5YR	7.5YR	7.5YR	7.5YR	7.5YR	7.5-5YR	7.5-5YR	7.5-5YR	-	-	-	7.5-5YR	10-7.5YR	10-7.5YR	10YR	10YR	10-7.5YR	-
depth of B horizon (cm)	>300	>300	>200	>200	>100	>300	>300	>200	>200	>200	>300	>200	-	-	>100	> 50	<30	<50	<20	<30	-
depth to lowest Cox (cm)	>300	>300	>200	>200	>100	>300	>300	>200	>200	>200	>300	>200	-	-	>100	> 50	-	-	-	-	-
pH at ~1m	-	-	-	-	-	-	-	-	-	-	4.8	-	-	-	-	-	-	4.9	-	-	-
clay films (1-5)	4	4	2	2	2	3	3	3	2	3	2	3	-	-	2	2	1	1	1	1	-
coarse matrix:% >1cm	-	-	56	-	-	-	70	-	-	-	-	50	-	-	-	-	-	-	-	-	-
%1-0.5cm	-	-	15	-	-	-	16	-	-	-	-	32	-	-	-	-	-	-	-	-	-
%<0.5cm	-	-	29	-	-	-	14	-	-	-	-	18	-	-	-	-	-	-	-	-	-
basal till lithification (1-5)	3	2	3	3	3	3	2	3	-	2	-	3	-	-	-	-	1	1	-	-	-
clast socket staining(1-5)	4	3	3	4	4	3	4	3	2	3	3	3	-	-	2	2	1	1	2	1	-
D. SURFACE MATERIALS																					
% split clasts	-	-	-	-	-	-	-	-	-	-	-	13.6	-	-	30	-	3.0	1.0	2.0	<10.0	-
E. DEPOSITIONAL LANDFORMS																					
moraine crest width (m)	-	23	-	-	-	-	-	-	-	-	-	60	-	-	10	-	10-20	-	-	2-5	-
proximal slope (°)	-	8	-	-	-	-	-	-	-	-	-	4-10	-	-	17	-	17-25	-	-	-	-
distal slope (°)	-	12	-	-	-	-	-	-	-	-	12,17,26	5-15	-	-	-	-	17-25	-	-	-	-
dissection (1-5)	-	4	3	3	3	2	2	2	2	2	2	2	-	-	2	2	1	1	1	1	-
F. BEDROCK LANDFORMS																					
max.depth of solution pans(mm)	-	-	-	-	-	-	-	-	-	-	-	-	-	400	-	-	-	-	-	300	
max.width of solution pans(mm)	-	-	-	-	-	-	-	-	-	-	-	-	-	500	-	-	-	-	-	500	
max.height to tors(m)	-	-	-	-	-	-	-	-	-	-	-	-		2	-	-	-	-	-		8

(a) G= Guelph; S= Sawmill; CKW= Coates/King William; BW= Bedlam Wall; CB= Cynthia Bay.

(b) T= till; O= outwash gravels; ST= soliflucted till; S= solifluction deposit; B= bedrock.

(c) Refers to burial of the same depositional unit in the same vicinity but not at the exact location where subsurface data were collected. The latter have only been recorded from unburied sites. S= solifluction deposit.

Table 7.2 Post-depositional modification of deposits, Navarre area. Site locations are indicated on Figure 7.1.

Bay and Bedlam Wall moraines differ considerably in age. Because sections through the moraines are generally shallow it has not been possible to compare the thickness of the soil profile that has developed on the Bedlam Wall Moraine with the profiles that have developed on the earlier moraines.

There is a general systematic increase in the thickness of the dolerite weathering rinds. Maximum rind thickness in the Cynthia Bay Moraines is 6.0 mm but rinds up to 16.9 mm thick are present in the Guelph Moraines. However, rind thickness increases in a progressive rather than stepwise fashion. The significant age difference between the Cynthia Bay and Bedlam Wall moraines that is suggested by the comparative degrees of soil development is not reflected by the weathering rinds. Weathering rinds in the Sawmill and Guelph moraines are generally thicker than any that were recorded in the Travellers Rest basin. This suggests that the Sawmill and Guelph moraines are older.

Solifluction deposits less than 50 cm thick have accumulated within the Cynthia Bay ice limits. In contrast, up to 2 m of solifluction deposit overlies the Bedlam Wall and earlier moraines.

Solution pans formed on sandstone benches on Mt. Rufus also suggest that the Cynthia Bay and Bedlam Wall phases differ significantly in age. The western arm of the Slacker Cirque was probably ice-smoothed during the Cynthia Bay phase and solution pans up to 30 cm deep have developed subsequently

(Figure 7.5). Because pans may deepen only slowly once an insulating floor of insoluble residue accumulates, pan width may be a more useful measure of age. Pans within the Cynthia Bay limits are seldom more than 30 cm wide while those on the slopes above commonly reach 150 cm in width.

Further evidence of the age difference between the Bedlam Wall and Cynthia Bay tills is available at the Slacker Cirque on Mt. Rufus. Here sandstone tors up to 2m high occur in an area that undoubtedly was over-ridden by ice during the Coates Ck. phase but which probably escaped active erosion by ice during the Cynthia Bay phase. The presence of large tors on top of the Bedlam Wall ridge suggests that a considerable period of time has elapsed since the crest was over-ridden during the Guelph phase. The Guelph phase is therefore likely to be of considerable age. The absence of erratics may be due to the midstream location of Bedlam Walls with respect to medial moraines on the Derwent Glacier, or the occurrence of multiple episodes of periglacial slope instability since ice last over-rode the ridge.

In summary, post-depositional characteristics suggest that the Coates Ck./ King William Ck. till represents a phase of glaciation that was separated from the Cynthia Bay phase by a period of prolonged chemical weathering. The Bedlam Wall moraines bear greater similarity to the Coates Ck. deposits than to the Cynthia Bay tills.

The massive form of the oldest end moraines suggests that they document positions of the ice margin that were maintained for lengthy periods. Bedded till in the Sawmill moraines dips upstream and red basal till has been plastered upon it. Rhythymically bedded clays east of King William Creek have been glaciotectonically deformed. These characteristics indicate that the ice margin was active. In contrast, the Cynthia Bay Moraines are narrow ridges that extend across the valley and document shortlived positions of the ice margin. Hummocky moraines in the Gingerbread Cirque indicate that the ice in this shallow and open cirque became inactive and decayed *in situ.*

chapter eight

THE GUELPH VALLEY

The Guelph River drains eastwards from the northern part of the King William Range (Figure 8.1). It originally joined the Derwent River immediately upstream of Butlers Gorge, but its lower reaches and the glacial deposits on the valley floor have been inundated following the construction of the Clark Dam and the filling of Lake King William. A bathymetric chart of Lake King William (Peterson and Missen, 1979) enables some interpretation of the topography on the lake bed (Figure 8.2). Ice from the eastern margin of the Derwent Glacier coalesced in the Guelph Valley with glaciers which arose on the slopes of the King William Range. The glacial geomorphology of the area is presented in Figure 8.3.

EROSIONAL LANDFORMS

A Glacial Erosion

The summit of the King William I massif is an ice abraded dolerite plateau. Elevated crests on the plateau have been little modified by glacial erosion which suggests that only a thin carapace of ice was present. Stoss and lee orientations on bedrock indicate a predominantly eastward movement of the plateau ice towards the Guelph Valley, but there are also local indications of movement westward into the Surprise Valley (chapter twelve).

The leeward eastern slopes of the King William Range have

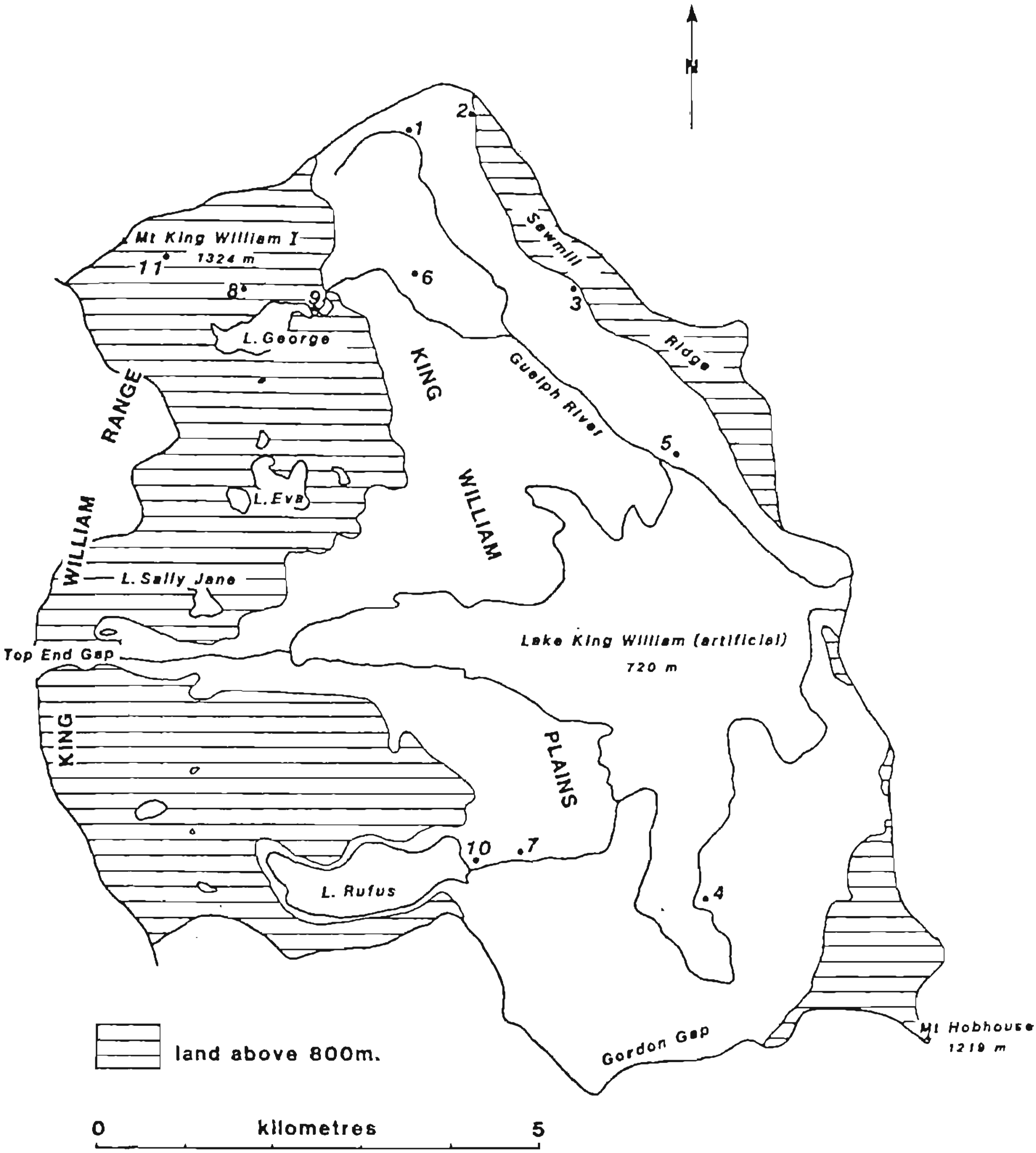


Figure 8.1 Locality map, Guelph River Valley. Numbers refer to site locations in tables 8.1 and 8.2.

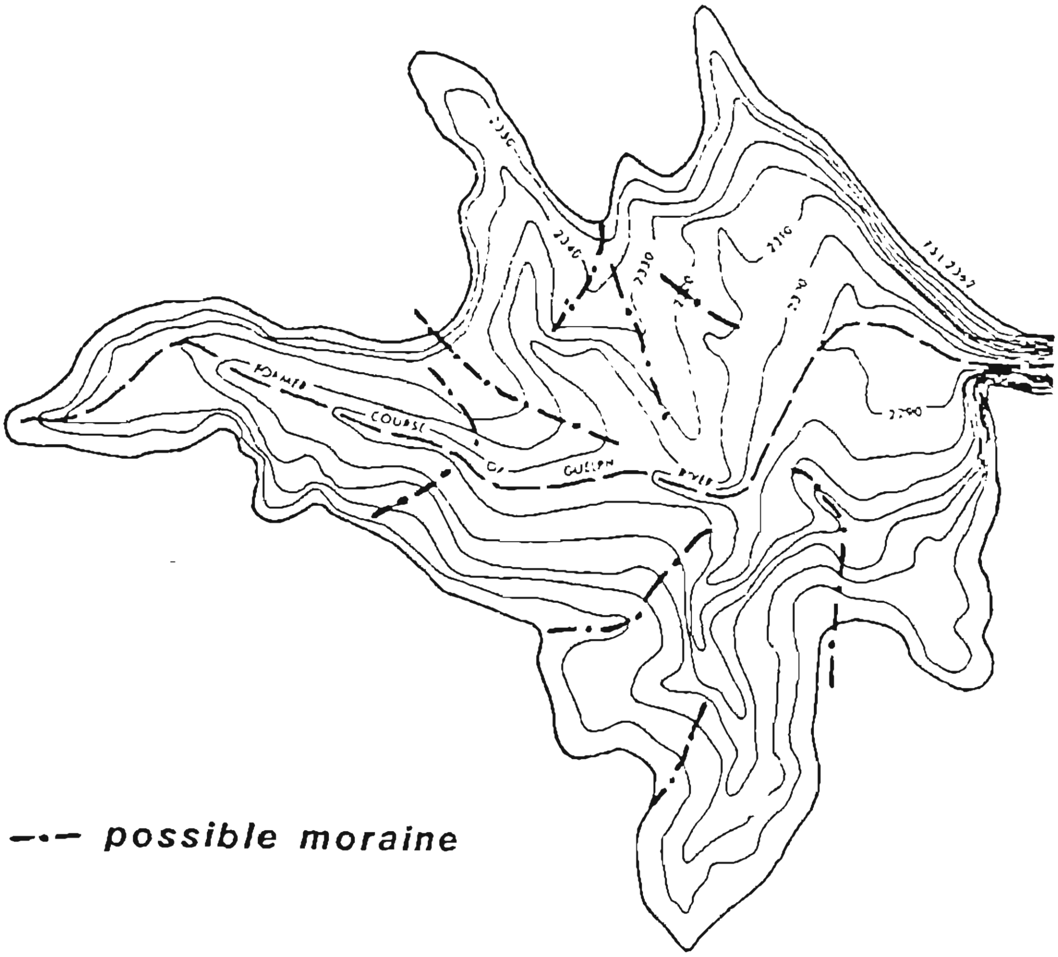


Figure 8.2 Bathymetric chart of the Guelph Basin of Lake King William (after Peterson and Missen, 1979). Possible moraines on the floor of the lake have been indicated.

been markedly steepened by intense glacial erosion and in particular by the excavation of numerous cirques. The mean orientation of these cirques is 89^{o}, and this primarily reflects the north-south orientation of the range. Most of the cirques lie at valley-heads and are cut in Permian sediments at the foot of columnar dolerite clifflines. A low threshold is common. A number of rock basin lakes occur, often bounded by a single end moraine as exemplified at Lake Eva. This contrasts with the general absence of end moraines at other cirques in the study area where no threshold is present.

Three glacial troughs are present. The northernmost of these occurs at the head of the Guelph River and contains Lake George (Figure 8.4). Lake George is a rock basin that has been extended by the deposition of a moraine barrage at its eastern end. Bathymetric surveying by Derbyshire (1967) reveals that the deepest part of Lake George is oriented about 30 degrees more northerly than the main basin (Figure 8.5). This reflects some deflection of the erosion by a structural lineament.

The Top End Valley is a further trough that extends from the head of the Surprise River (chapter twelve) into the Guelph Valley (Figure 8.6). This narrow defile is over 300 m deep and separates the Mt. King William I massif from the rest of the King William Range to the south. Shallow valleys that extend downslope of cirques on both walls of the Top End Gap have been deflected eastwards as a result of an ice stream which flowed eastwards through the Gap. It is clear from

LANDFORMS

- cirque
- over-ridden cirque
- glacial trough
- over-ridden margin
- diffluence col
- valley step
- ice-abraded outcrop
- areally scoured terrain
- glacial striae
- rock-basin lake

- moraine ridge
- moraine hummock
- moraine-dammed lake
- erratics
- erratics of Nive till
- reworked Permian erratics or large lagstones

- meltwater channel

- nunatak
- nivation cirque
- rock crevasse or dilation trench
- slab topple
- rock glacier
- solifluction lobe or terrace
- patterned ground
- landslip
- artificial lake

SEDIMENTS

GLACIAL

- Cynthia Bay till/outwash
- Beehive "
- Powers Creek "
- Clarence "
- Stonehaven "
- Nive Till "

NON-GLACIAL

- alluvial silt &/or peat
- talus
- Solifluction deposits

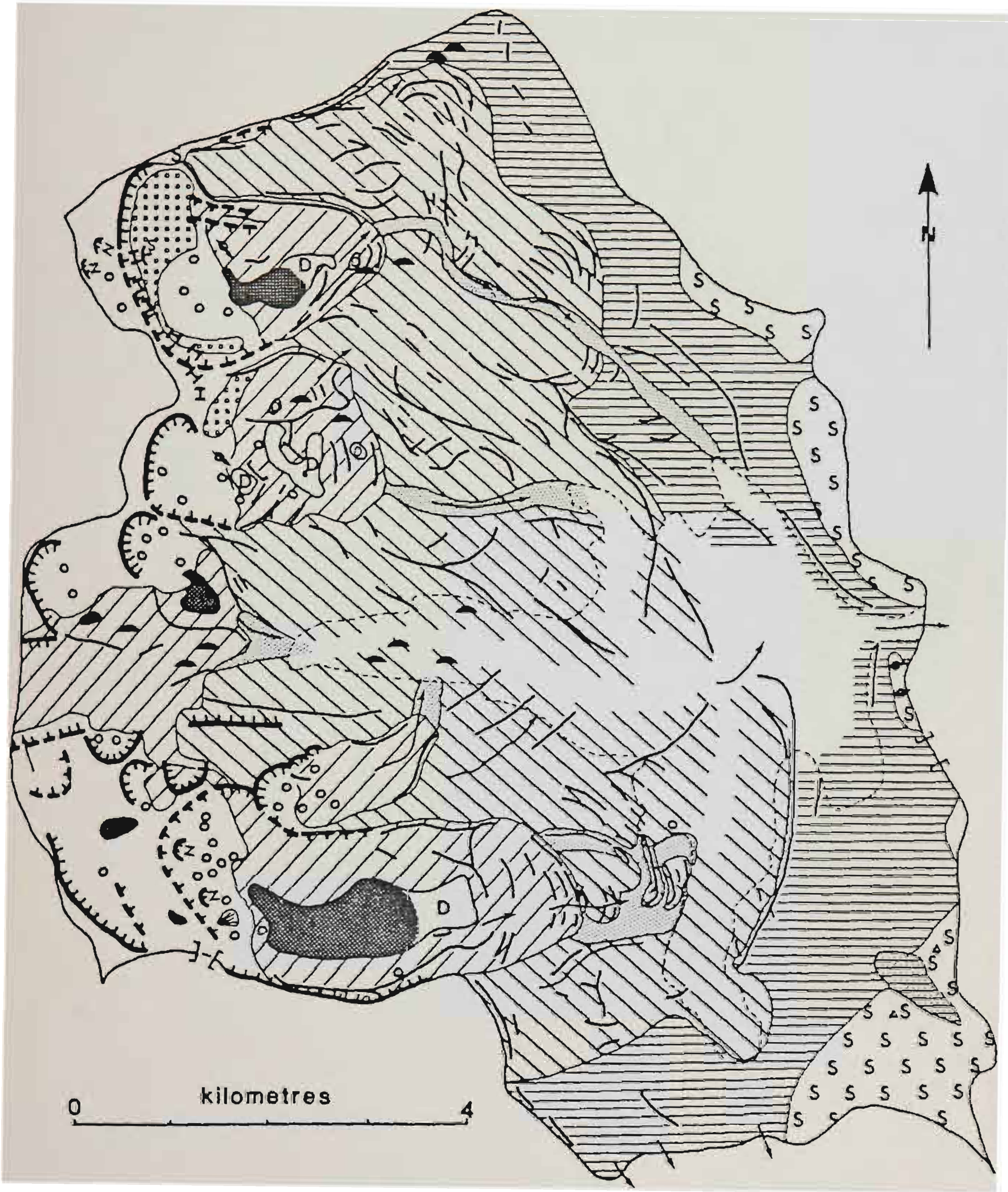
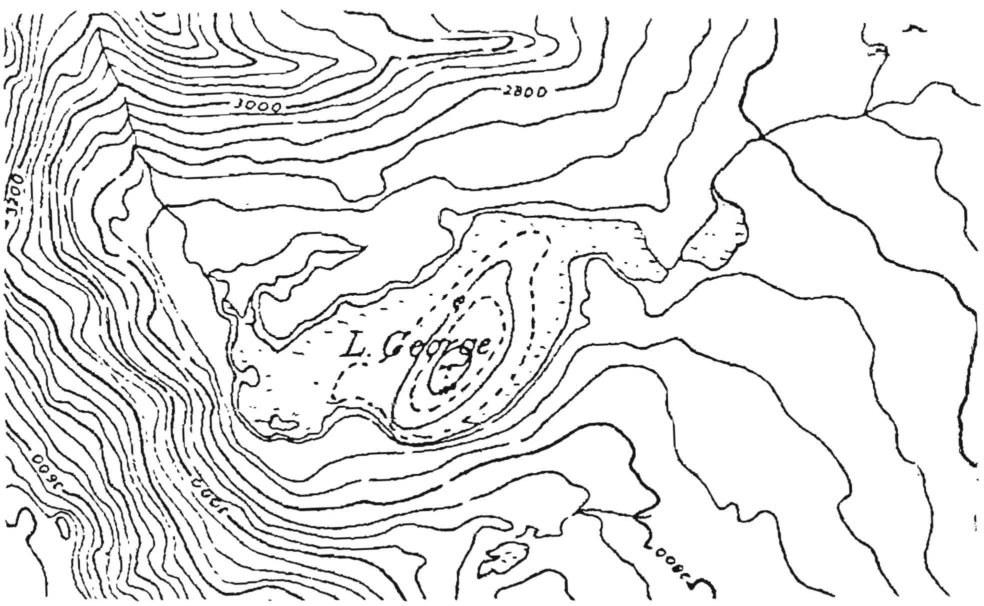


Figure 8.3 Glacial geomorphology of the Guelph River Valley. See Figure B2 for key.

Figure 8.4 Lake George (foreground) and Lake King William from the summit of Mt. King William I. Note the pronounced lateral moraine on the northern (left) side of Lake George.

Figure 8.5 Bathymetric chart of Lake George. Interval between the solid isobaths is 50ft. (15 m.); broken isobaths are intermediate (25 ft.) (after Derbyshire, 1967).


3000
2800
L. George

this that a considerable quantity of ice that accumulated in the lee of the Loddon Range at the head of the Surprise Valley passed through the Top End Gap and supplemented the glaciers that arose in the King William Range. Some of this ice also flowed WNW down the Surprise Valley.

Lake Rufus lies in a well developed glacial trough 3 km long and up to 1 km wide. The plateau margin at its head is severely ice abraded and glacial erosion has left a small "Half Dome" hill (Derbyshire, 1967). The lake lies in a rock basin and has been extended by the construction of a moraine barrage. It is about 80 m deep (Figure 8.7) (Derbyshire, 1967) and the trough walls rise 200 m on either side of it. Excavation of this trough has involved the removal of over 60 million cubic metres of rock. In some ways the Lake Rufus trough is the most striking glacial landform in the study area. Ice smoothed bedrock on the arms of the trough near its head indicates that up to 300 m of ice accumulated here.

In each case the troughs terminate abruptly at the level of the St. Clair Surface where piedmont ice lobes developed.

The Top End Gap is a diffluence col by means of which ice flowed eastwards from the head of the Surprise Valley into the Guelph basin. A smaller diffluence col occurs south-west of Lake Rufus and a number of valley steps up to 80 m. high are present on the eastern slopes of the King William Range.

Ice abraded bedrock is present at the foot of the range. Derbyshire (1967) reported striae near the south-western shoreline of Lake King William that indicated ice movement at 80^{o} - 90^{o} from the Lake Rufus area towards Butlers Gorge. Ice abraded bedrock near the crest of Sawmill Ridge has been frost shattered and has a less fresh appearance than the abraded rocks that occur on the King William Range. The direction of ice movement over Sawmill Ridge is uncertain from the erosional evidence. The bathymetric chart of Lake King William reveals a series of ridges that are aligned north-south on the lake floor upstream of the lower Guelph Gorge, and a steep knoll at the gorge outlet. Abraded rock ridges that are aligned parallel to these drowned ridges occur above lake level. This suggests that many of the ridges beneath the lake that have this alignment are bedrock rather than moraines (Figure 8.2).

B Glaciofluvial erosion

Meltwater channels have been cut in rock along structural lineaments in the dolerite of the King William I Plateau. Other meltwater channels plunge steeply downslope into the valley heads. The lower Guelph Gorge between Mt. Hobhouse and Sawmill Ridge has been deepened by glaciofluvial erosion.

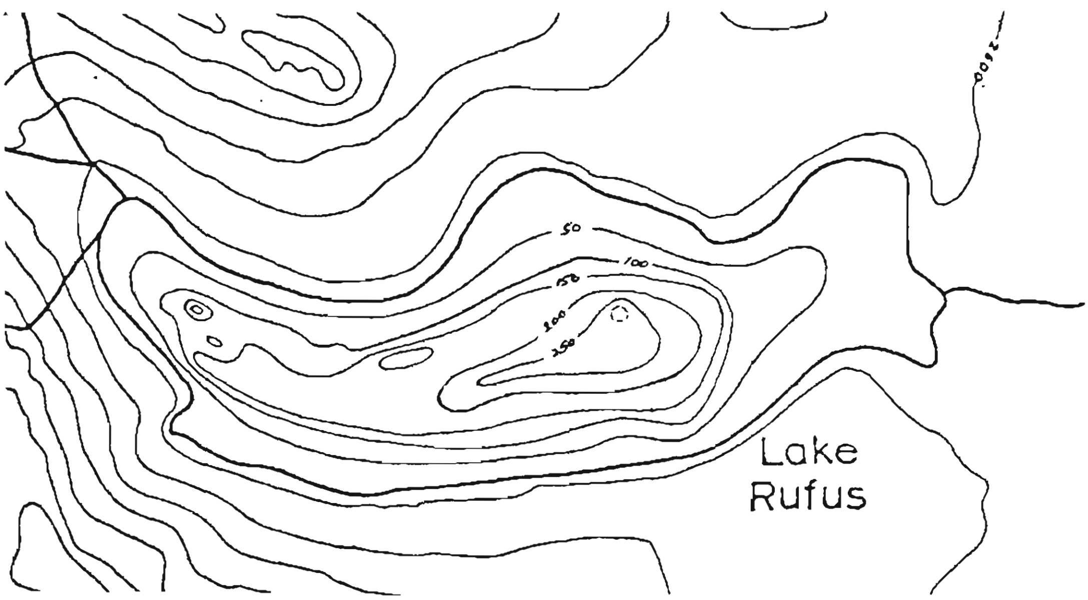
The outermost moraines on the King William Plains have been heavily dissected by meltwater to an extent that often complicates the interpretation of their original morphology. Shallow channels that plunge downslope from the Guelph-Navarre divide between Mt. King William I and Sawmill

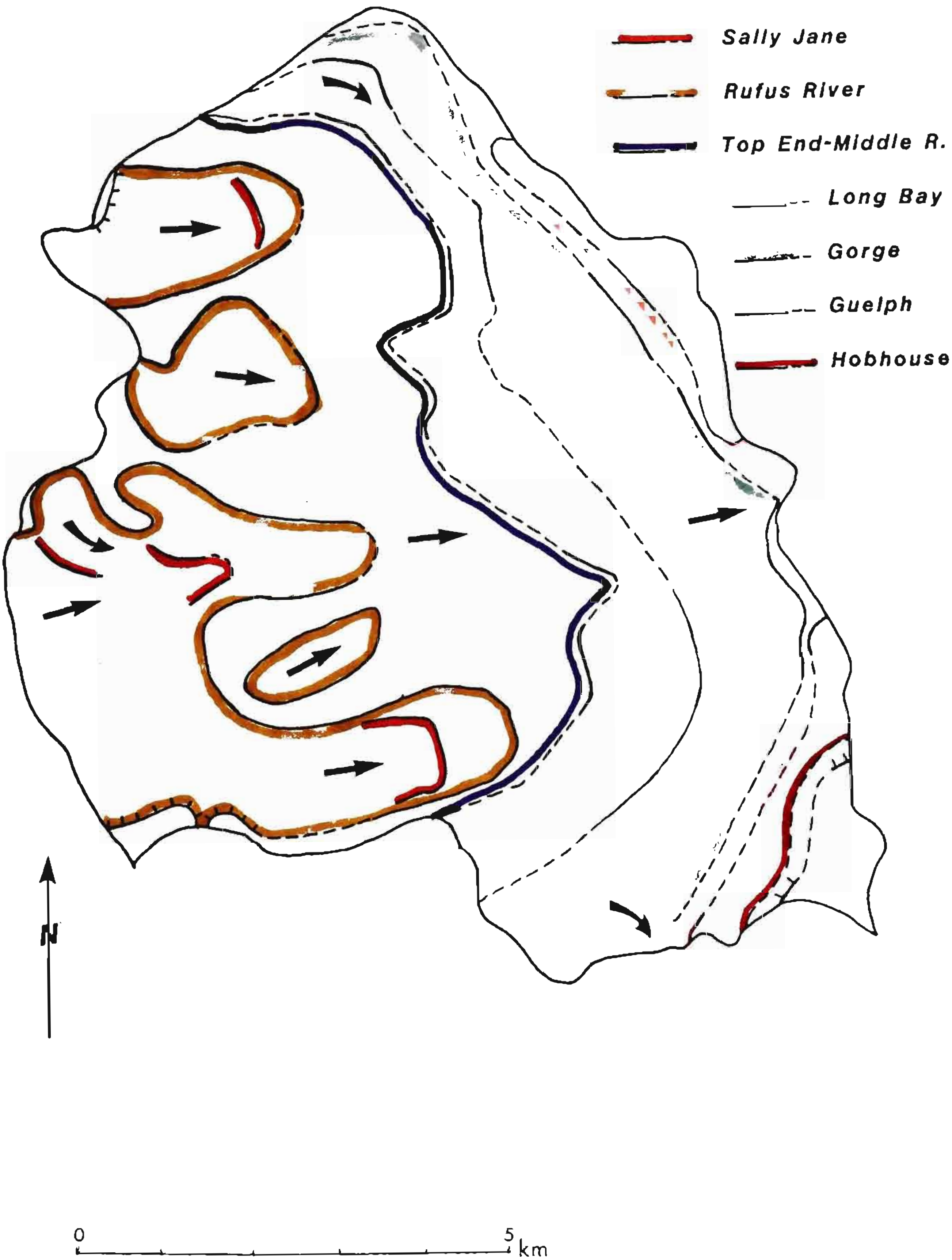
Figure 8.6 The Top End Gap separates the Mt. King William I massif from the southern part of the King William Range. Lake Sally Jane lies on its northern margin and is impounded by a moraine.. The view is from the Mt. King William I plateau.

Figure 8.7 Bathymetric chart of Lake Rufus. Interval between the solid isobaths is 50 ft. (15 m); broken isobaths are intermediate (25 ft.). (after Derbyshire, 1967).

overleaf

Figure 8.8 Principal ice limits in the Guelph River Valley.


50
100
150
200
250
Lake
Rufus


Sally Jane
Rufus River
Top End-Middle R.
Long Bay
Gorge
Guelph
Hobhouse
N
0
5
km

Ridge were cut in drift by meltwater from the retreating Derwent Glacier (Figure 8.3).

C Non-glacial erosion

The broad topography of the Guelph Valley is the result of fluvial erosion that has been controlled by structural lineaments. The floor of the valley has been eroded into Permian and Triassic sediments west of a fault that trends north-south along the edge of the resistant dolerite spine of Sawmill Ridge (Gulline et al., 1963. Permian and Triassic sediments crop out to over 1000 m altitude north of Lake George but ice smoothed dolerite is present on the south-western shoreline. This suggests that the valley which previously existed here may have exploited a fault.

The Top End Gap is developed along a structural lineament which extends east-west through the King William Range and can be traced eastwards to the lower Guelph Gorge. This lineament is of uncertain origin but its orientation parallels that of other troughs and ridges on the eastern flank of the King William Range which suggests that it is a major joint or fault in the dolerite. It previously conducted the main branch of the Guelph River eastward through the range until headward erosion by the cirques of the Loddon Range led to the capture of its headwaters by the Surprise (Figure 15.8a).

A partly detached mass of dolerite at the head of the Lake George trough demonstrates the cambering of slab topples on oversteepened dolerite slopes (Figure 15.13). The margin of the King William I Plateau is characterised by rock crevasses and dilation trenches for a distance of several hundred metres south of this point. Some of these trenches are up to 100 m wide and contain small semi-permanent ponds.

Rockfall has been a comparatively minor agent of scarp retreat at the head of some of the troughs, most notably above Lake Rufus (Derbyshire, 1967) and on the eastern slopes of Mt. King William I. Very limited rockfall has also occurred from some of the steeper residuals on the King William Plateaux and from some partly detached slab topples.

Nivation hollows are well developed on the King William plateaux. The largest are seldom more than 20 m wide with a backwall height of about 8 m. Some of these hollows have developed on solifluction deposits that accumulated above the level of the plateau ice carapace. Therefore, they must postdate the solifluction deposits.

DEPOSITIONAL LANDFORMS AND SEDIMENTS

A Glacial Deposits

(i) Hobhouse Erratics

Quartzite erratics that occur among dolerite slope deposits indicate that ice reached 850 m on the northern slopes of Mt. Hobhouse (Figure 8.8). In view of the large volume of ice that was generated on the King William Range it is reasonable to interpret these erratics as deposits of the Guelph Glacier rather than the Derwent Glacier, deposited at a time when the two glaciers were confluent and the Derwent ice would have extended far downstream of Butlers Gorge (chapter nine). Part of this Guelph ice would have flowed eastwards from the Lake Montgomery Cirque (315 160) which now lies in the Gordon catchment (chapters eleven and thirteen). The lithology of the clasts in the till exhibits considerable local variation (Table 8.1).

(ii) Guelph Moraines

Degraded moraine remnants occur from lake level to 800 m on the southern end of Sawmill Ridge and to a similar elevation on the foot of the ridge that extends towards Sawmill Ridge from Mt. Hobhouse. These moraines represent part of a moraine complex that occurs in the Derwent Valley immediately north of Butlers Gorge. This documents an event during which the Guelph and Derwent glaciers ceased to be confluent and retreated into their respective valleys (chapter nine). Till deposits within the area described in this chapter imply that ice may also have passed eastwards into the Derwent via two minor cols on the northern end of the Hobhouse Range.

Quartzite erratics at 790 m on the Guelph-Gordon divide west of Mt. Hobhouse probably date from this period. They indicate that diffluent ice from the Guelph would have been able to over-run the divide and spill into the upper Gordon Valley via the Gordon Gap (chapter eleven).

(iii) Gorge Moraines

Till that is plastered against the western side of the

Vicinity No.	1	3	4	5	7	8	9
%							
dolerite	60.0	37.5	54.0	68.9	38.5	92.0	52.0
quartz	32.0	47.5	29.0	26.8	46.6	4.0	38.5
quartz schist	-	-	-	-	10.1	-	-
argillites	8.0?	15.0?	15.0?	4.3?	4.8	4.0?	9.5
other	-	-	-	-	-	-	-

Table 8.1 Lithology of glacial deposits in the Guelph area.

Site locations are indicated on Figure 8.1.

Sawmill - Mt. Hobhouse ridge appears to mark a distinct phase during which the Guelph Glacier was trapped behind the rock bar at the Guelph Gorge. A small outwash plain was constructed against the inner margin of the earlier moraine at the gorge outlet. The steep flanks of the ridge have militated against the preservation of deposits, but some highly degraded moraines occur on more gentle terrain to the south of Lake King William.

Degraded moraines also extend onto the Guelph-Gordon divide and it is probable that a thin lobe of ice about 1 km wide passed southwards through the Gordon Gap. A till mantle below about 800 m indicates that the northern end of Sawmill Ridge was overridden. Hummocky drift occurs in the far north west of the Guelph area within this ice limit.

(iv) Long Bay Moraines

The bathymetric chart indicates that a probable major end moraine extends northwards across the bed of Lake King William from a degraded moraine on the eastern shoreline of Long Bay (Figure 8.2). The moraines north of Lake King William have been severely dissected by meltwater, but probable continuations of the Long Bay Moraines can be traced eastwards to the Guelph River and northwards along the western flank of Sawmill Ridge.

The moraines along the middle reaches of the Guelph River are arcuate to the north and were deposited by the Lake George Glacier. The ice was most extensive to the south where the Sally Jane, Top End and Rufus glaciers remained confluent, but ice no longer passed southwards through the Gordon Gap into the upper Gordon River Valley.

(v) Top End - Middle River Complex

Moraines deposited during this phase imply that the Rufus Glacier barely reached the present shoreline of Lake King William and was discrete from the broad confluent lobes that lay to the north. This phase was also noteworthy for the construction of a major end moraine over 20 m high and 400 m broad at Lake Montgomery. This now forms part of the Guelph-Gordon divide and no ice entered the Guelph Valley from this source following its construction.

The most extensive ice body at this time remained the Top End Glacier which still extended 5 km eastwards of the Guelph-Surprise divide. This was the last occasion during which ice accumulated in the shallow cirque north of Lake George. Hummocky drift well inside this ice limit suggests that retreat may have been rapid and left stranded masses of debris-rich ice to decompose _in situ_ in a number of localities.

(vi) Rufus River Moraines

A complex of topographically fresh and little dissected moraines commences about 1 km downstream of Lake Rufus and extends westwards towards the lake shoreline. Fresh lateral moraines occur 2 km east of the Top End Gap and other fresh moraines occur around Lake Eva and to the north and east of Lake George. The lateral moraine on the northern side of Lake George has been constructed within the limits of a degraded outer moraine which now has little topographic expression.

Many of the lateral moraines lie within the ice-abraded arms of the glacial troughs and cirques, while the terminal moraines indicate that the ice did not extend any significant distance beyond the break of slope at the foot of the range. The glaciers responsible for the construction of these moraines were generally very short and steep. The northern lateral moraine of the George Glacier suggests an ice surface gradient of as much as 160 m/km. Small outwash plains were constructed amid the older degraded moraines on the King William Plains.

The topographic freshness and limited dissection of these moraines suggests that they are equivalent in age to the Cynthia Bay Moraines in the Lake St. Clair area (chapter four).

(vii) Sally Jane Complex

Latero-terminal moraines just east of the Top End Gap on the slopes of the King William Range impound a number of small lakes, most notably Lake Sally Jane (Figure 8.6). These moraines appear disproportionately large given the limited extent of the ice. Older lateral moraine materials deposited by the Top End Glacier have probably been partly reworked but it is nonetheless clear from the size of the moraine that the Sally Jane Glacier was very active. Protalus has also contibuted to the moraines at Lake Sally Jane, quite angular blocks being present just inside the crest.

The Sally Jane Complex is represented elsewhere by the innermost of the cirque moraines that bound the rock basins. The Top End Glacier had by this time declined from being the largest glacier in the area to virtually the smallest. The cause of this decline lay in the continued retreat of the Surprise Glacier which no longer had a sufficient surplus to direct ice eastwards through the Top End Gap. Little sediment has accumulated in the lakes upstream of the Sally Jane Complex which suggests that deglaciation was probably fairly rapid.

B Non-glacial deposits

(i) Slope deposits

Solifluction mantles are once again the most widespread non-glacial deposits. These mantles are up to 3 m. thick on the northern flank of Mt. Hobhouse where they overlie weathered till and incorporate occasional quartzite erratics. Argillite-rich mantles over 1 m thick cloak the upper western flank of Sawmill Ridge.

Few fully detached topples occur on the King William Range, where most of the plateau margin above the cirques has been over-ridden by ice during the most recent phases of glaciation. Small solifluction terraces are developed on the partly detached topple above Lake George. Bare treads about 2 m wide are separated by risers 30 - 50 cm high which support small shrubs. Some primary frost sorting has occurred on the treads. Small polygonal nets up to 30 cm in diameter occur at about 1100 m elevation north-east of Mt. King William I on a surface devegetated during the last decade. Subnival pavements of closely fitted stones (White, 1972) floor some of the nivation cirques on the King William I plateau. They have probably resulted from block movement due to the weight and creep of the snow-pack and frost heaving (Troll, 1958; White, 1972).

Rockfall talus has accumulated beneath dolerite clifflines at high elevations on the King William Range and Mt. Hobhouse. A considerable thickness of rockfall talus has accumulated beneath the summit cliffs of Mt. King William I. Most of this lies above 1100 m which suggests that the upper surface of the George Glacier lay below this level at the time the talus accumulated. A small protalus rampart is present in this locality.

(ii) Alluvial, peat and archaeological deposits

Swales between moraines on the King William Plains have been filled by up to 1 m of alluvial sand and silt. Organic silt has accumulated to depths of 30 cm in ponds that have formed in dilation trenches on the margin on the King William I plateau. The silt reaches 60 cm in thickness in the lagoon at the eastern end of Lake George. Fibrous peat up to 80 cm. thick has accumulated beneath the buttongrass plains.

Several archaeological sites (Figure 15.12) record the presence of prehistoric aborigines who have modified the form of some surface clasts in the process of manufacturing stone tools.

3 DATING AND DISCUSSION

The youngest deposits are organic-rich silts that have accumulated inside the ice limits of the Sally Jane phase at the head of Lake George and in dilation trenches on the edge of the King William I plateau. Less than 30 cm of silt has accumulated on the floors of nivation hollows on the King William Range. Silt has also been deposited in swales between moraines on the King William Plains. Archaeological deposits have been found on the surface of the silt on the King William Plains. They have also been found on moraines from which the peat has been stripped.

Small solifluction terraces have developed inside the ice limits of the Rufus River phase on the partly detached slab topple that overlooks Lake George. Thick solifluction mantles only occur outside the ice limits of the Rufus River phase and must therefore predate the close of that phase.

A small ridge of protalus was constructed beneath the cliffs of Mt. King William I on an earlier talus that probably accumulated in contact with the ice margin during Rufus River time and locally slipped downslope as the ice retreated. The protalus must postdate the Rufus River phase. Protalus also forms a significant part of the cirque moraine at Lake Sally Jane.

The crest of the partly detached topple above Lake George has been smoothed by glacial erosion. This erosion could not have been accomplished without collapsing the topple if it had already been partly detached. Therefore, the toppling must postdate the Rufus River phase and probably occurred in response to the retreat of the supporting ice margin.

During the Rufus River phase the glaciers were short and steep. The Rufus River Moraines occur inside the ice limits of the Top End-Middle River phase. The glaciers formed separate piedmont lobes on the King William Plains. Outwash sand and gravel extends from the ice limits of the Rufus River phase along meltwater channels that were eroded through the Top End-Middle River and earlier Long Bay moraine sequences.

The Long Bay Moraines define a major ice limit at a time when the King William Range glaciers formed a confluent piedmont apron. During the Gorge phase the Guelph Glacier was lodged behind the rock bar at the Guelph Gorge. The Derwent and Guelph ice was still confluent at this time around the northern end of Sawmill Ridge.

The glacial sediments that were deposited during the Guelph phase lie outside the ice limits of the Gorge phase. They indicate a more extensive ice cover that inundated much of Sawmill Ridge. Diffluent ice from the Guelph Valley also spilled through the Gordon Gap into the upper Gordon Valley.

At its maximum the Guelph Glacier probably over-rode Sawmill

Ridge and reached an altitude of 850 m on the northern slopes of Mt. Hobhouse. A tongue of ice at least 50 m thick flowed southward into the head of the Gordon River Valley.

Postdepositional characteristics of the landforms and deposits is presented in Table 8.2. The Sally Jane and Rufus River moraines are topographically fresh. The Top End-Middle River, Long Bay and Gorge moraines are rounded in form and have been extensively dissected by stream erosion and give an impression of greater age. Erratics high on the slopes of Mt. Hobhouse are the sole remaining depositional legacy of the Hobhouse phase.

Shallow A Cox Cu type soil profiles have developed on the Sally Jane and Rufus River moraines. However, the Cox horizon reaches a depth of at least 1.5 m in the Top End-Middle River moraines. The soil profiles on the earlier moraines are characterised by thick textural B horizons.

The mean thickness of weathering rinds formed on dolerite clasts in the Rufus River moraines is less than 2 mm and maximum rind thickness is less than 5 mm. This is half the thickness of the weathering rinds in the moraines of the Guelph phase.

These characteristics suggest that the Rufus River moraines are of equivalent age to the Cynthia Bay moraines at Lake St. Clair, and that the earlier deposits are considerably older. The Hobhouse erratics suggest a phase of glaciation

moraine[a]	G	GG	GG	LB	TEM	TEM	RR	RR	SJ	-	-
vicinity no.	1	2	3	4	5	6	7	8	9	10	11
material[b]	T	T	T	T	O	T	T	T	T	S	S
A. MORPHOSTRATIGRAPHY											
relative position	outside	outside	inside	inside	inside	inside	inside	inside	inside	inside	inside
	2	3	2	3	3	5	4	TEM	8	3	9
crest/site elevation (m)	770	770	760	740	740	780	760	960	810	730	1200
max. depth of burial (by) (m)[c]	-	-	-	S2	S1	-	-	-	-	-	-
B. SUBSURFACE CLASTS											
dolerite rinds (mm) - max.	8.2	8.0	8.8	-	2.7	-	-	4.3	4.2	1.9	1.2
min.	1.9	1.0	4.0	-	1.2	-	-	0.3	0.9	0.4	0.3
mean	4.7	4.5	6.1	-	1.8	-	-	1.3	1.7	0.8	0.8
SD	1.8	1.5	1.7	-	0.5	-	-	0.8	0.6	0.4	0.3
hardness (1-5)	3	3-4	2-4	3	1-2	1-3	1-2	1	1-2	1	1
coherence of argillites (1-5)	5	3-5	4-5	4	2	4	1	2	1	1	1
clast surface colours	5YR	7.5YR	5YR	5YR	5YR	7.5-5YR	10YR	10YR	10YR	-	-
C. SUBSURFACE MATRIX											
profile type	A Bt	A Bt	A Bt	A Bt	A Cox cu	A Bt	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A Cox cu
colour of B horizon	5YR	7.5YR	7.5-5YR	7.5YR	7.5YR	7.5-5YR	10YR	7.5-10YR	10YR	10YR	7.5YR
depth of B horizon (cm)	>150	>100	>150	>150	-	>150	-	-	-	-	-
depth of lowest Cox (cm)	>150	>100	>150	>150	>150	>50	<50	<80	-	-	-
development of clay films (1-5)	4	3-4	3	4	-	3-4	1	1	1	1	1
fine matrix: % sand	-	-	29.2	-	-	-	-	-	13.7	27.8	-
% silt	-	-	37.1	-	-	-	-	-	48.8	40.0	-
% clay	-	-	33.7	-	-	-	-	-	37.5	32.2	-
degree of lithification of basal till	-	-	3	3	-	3	-	1	-	1	-
clast socket staining	4	3-4	2-4	3	-	2-3	-	1	-	1	-
D. SURFACE MATERIALS											
no. of surface boulders ($/m^2$)	-	-	-	-	-	-	-	-	-	-	-
% split clasts	20	-	-	23	-	16	3	-	-	-	-
E. DEPOSITIONAL LANDFORMS											
moraine crest width (m)	-	-	-	15	-	25	5	5	-	-	-
moraine slopes (°) - proximal	-	-	-	<5	-	8	20	25	-	-	-
- distal	-	-	-	<5	-	10	20	15	-	-	-
degree of fluvial dissection (1-5)	3	3-4	3-4	4	-	4	2	1	2	-	-

(a) G= Guelph; GG= Gorge; LB= Long Bay; TEM= Top End-Middle River; RR= Rufus River; SJ= Sally Jane.

(b) T= till; O= outwash gravels; S= solifluction deposit.

(c) Refers to burial of the same depositional unit in the same vicinity but not at the exact location where subsurface data were collected. The latter have only been recorded from unburied sites. S= solifluction deposit.

Table 8.2 Post-depositional modification of deposits, Guelph area. Site locations are indicated on Figure 8.1.

that is of still greater age.

chapter nine

DERWENT BRIDGE TO WAYATINAH

After the Derwent Glacier emerged from its glacial trough at Lake St. Clair it spread out upon the St. Clair Surface as a piedmont lobe. Glacial deposits floor the Derwent Valley almost continuously for 20 km. beyond Lake St. Clair to the headwaters of Mossy Marsh Creek, which was regarded by Derbyshire (1963, 1967) as the downstream limit reached by the glacier. Near Mossy Marsh Creek the Derwent River descends from the St. Clair Surface via a steep-sided gorge that offers few prospects for the preservation of glacial deposits (Figure 9.1).

This chapter documents further exposures of probable glacial drift that have been located well downstream of Mossy Marsh Creek near the lower end of the Derwent Gorge (Kiernan, 1983a) (Figure 9.2). It also re-examines the deposits upstream of Mossy Marsh where outwash plains are inset into the moraines on the valley floor. The artificial Lake King William, created for hydro-electric purposes, has flooded more than 40 km^2 of the valley floor in an area that would otherwise appear most promising for obtaining evidence on the age of the glacial deposits. Information on this critical part of the study area is therefore limited to the exposures above the lake shore and a geological report prepared prior to filling of the lake which briefly touches upon the Pleistocene deposits. (Prider, 1948). A bathymetric chart (Peterson and Missen, 1979) also provides

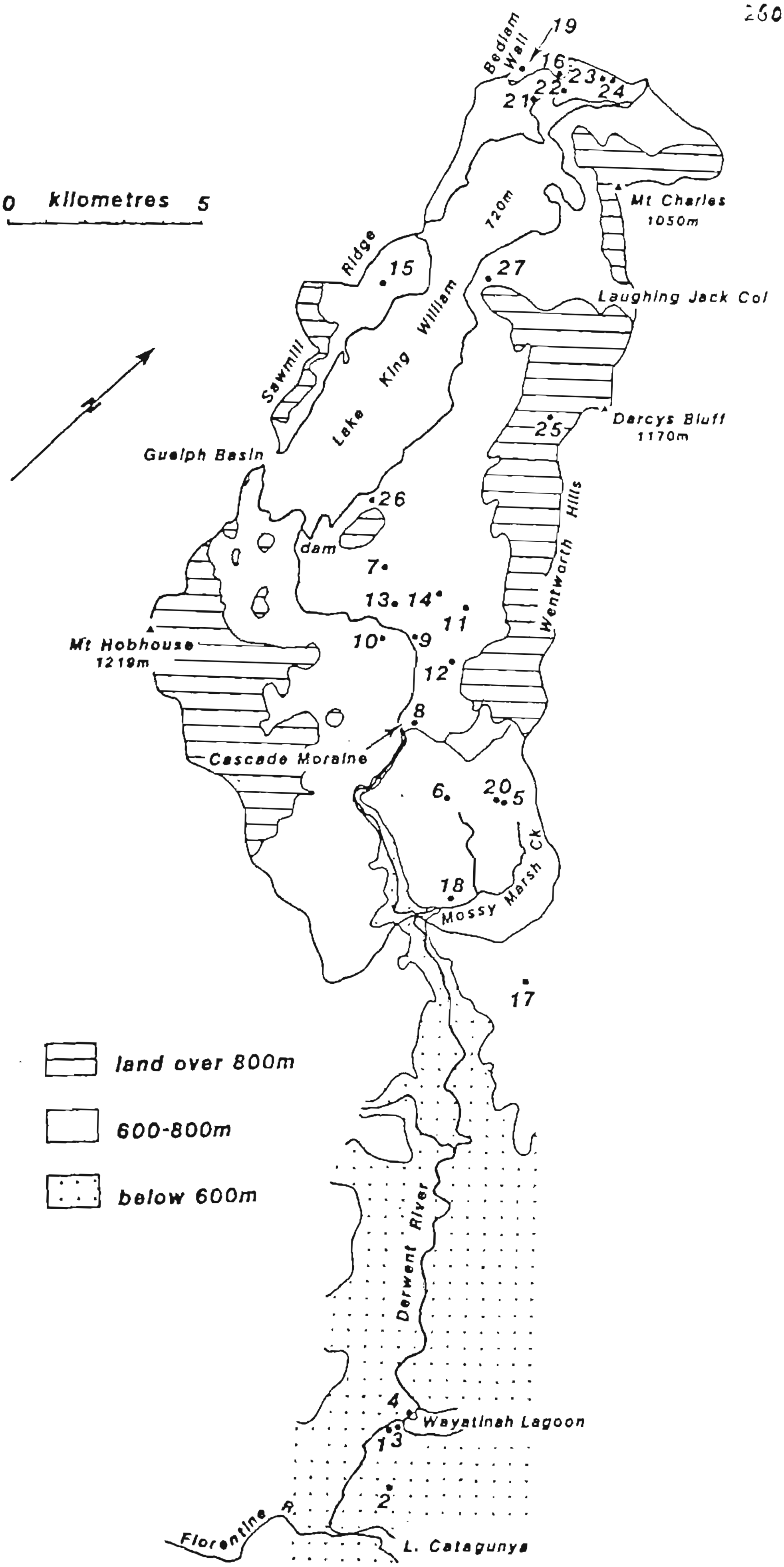


Figure 9.1 Locality map, Derwent Bridge to Wayatinah. Numbers refer to site locations in tables 9.1 and 9.2.

LANDFORMS

cirque

over-ridden cirque

glacial trough

over-ridden margin

diffluence col

valley step

ice-abraded outcrop

areally scoured terrain

glacial striae

rock-basin lake

moraine ridge

moraine hummock

moraine-dammed lake

erratics

erratics of Niva till

reworked Permian erratics or large lagstones

meltwater channel

nunatak

nivation cirque

rock crevasse or dilation trench

slab topple

rock glacier

solifluction lobe or terrace

patterned ground

landslip

artificial lake

SEDIMENTS

GLACIAL

Cynthia Bay till/outwash

Beehive "

Powers Creek "

Clarence "

Stonehaven "

Niva Till "

NON-GLACIAL

alluvial silt &/or peat

talus

Solifluction deposits

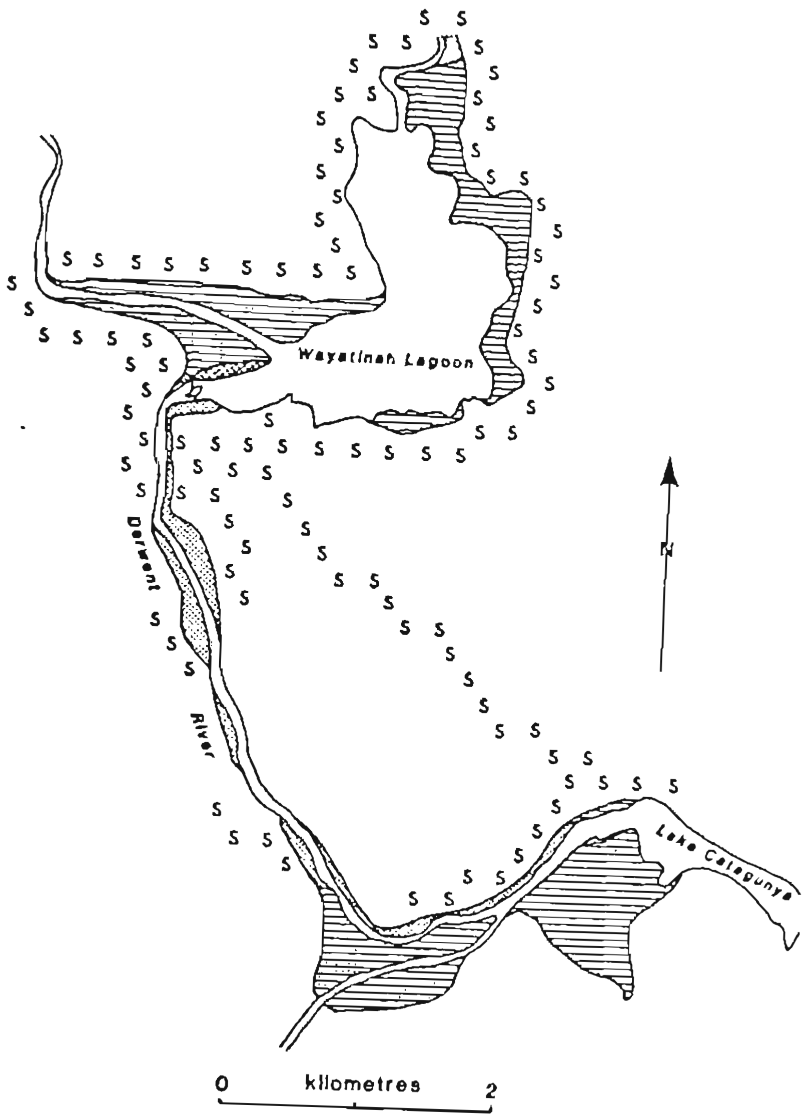


Figure 9.2 Glacial geomorphology of the Wayatinah area. See figure B2 for key.

some morphological evidence (Figure 9.3).

The distribution of the erosional landforms and sediments is presented in Figure 9.4.

EROSIONAL LANDFORMS

A Glacial Erosion

The only potential snowfence in this area is the Hobhouse Range. Shadowed by the more elevated King William Range which lies immediately upwind, the Hobhouse Range was regarded as unglaciated by Derbyshire (1967) and as the site only of nivation by Peterson (1969). Equivocal erosional evidence suggests that the Hobhouse Range may have been lightly touched by glacial erosion above 960 m where some rock surfaces appear to have been abraded by glaciers. At least four cirque-like hollows occur at the head of small streams which drain into the Derwent River. However, no glacial deposits have been recognised on the upper slopes of the range and glaciation cannot be proven.

A diffluence col, the Laughing Jack col, occurs to the east of Lake King William between Mt. Charles and the Wentworth Hills. Ice from the Derwent Valley passed eastwards through this col into the Nive River catchment.

Many small knolls on the floor of the Derwent Valley have been abraded by ice that flowed generally south-eastwards. Spurs that project westwards from the Wentworth Hills have been shorn off by the Derwent Glacier. A broken fault-bounded ridge of dolerite that is oriented north-south

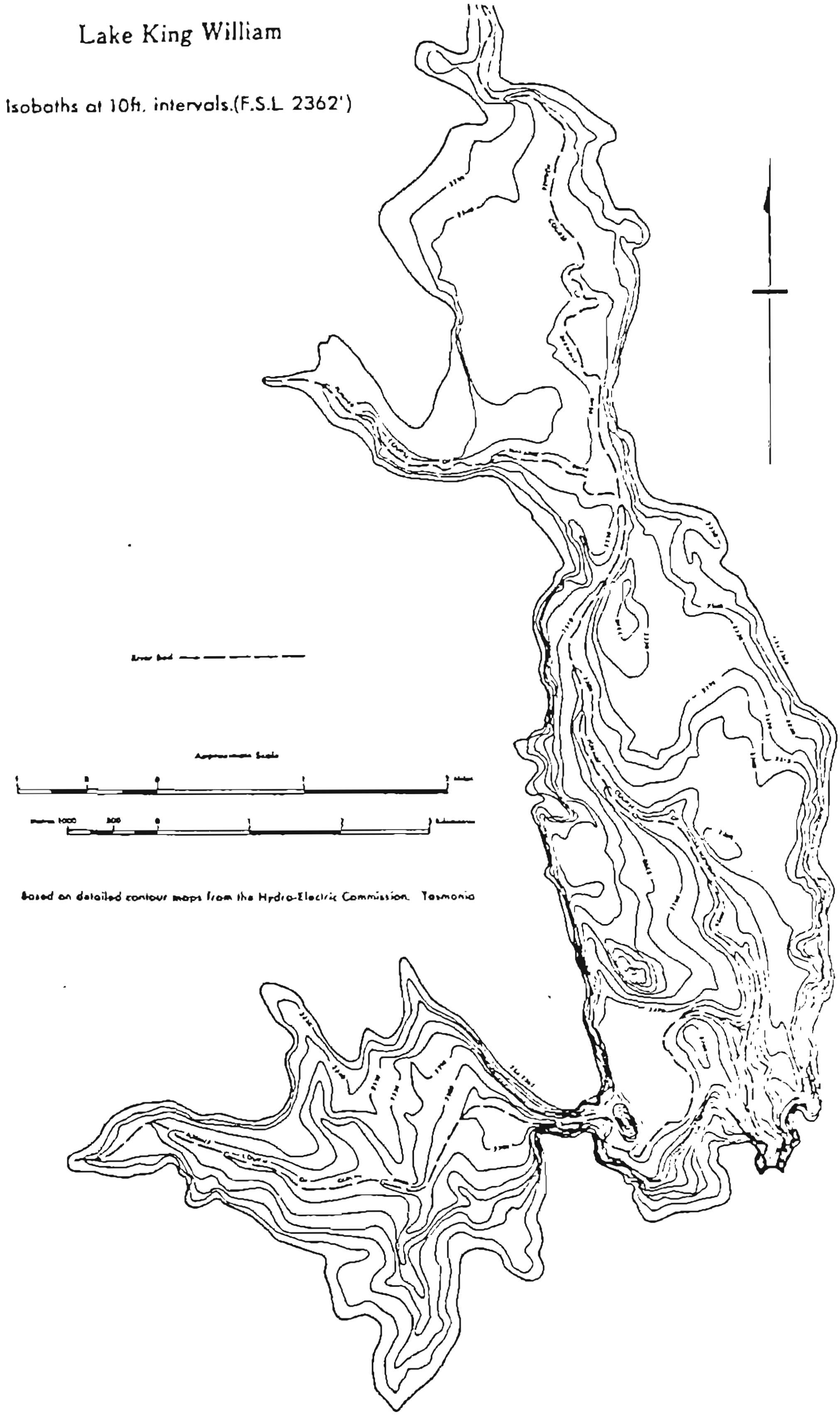


Figure 9.3 Bathymetric chart of Lake King William (after Peterson and Missen, 1979).

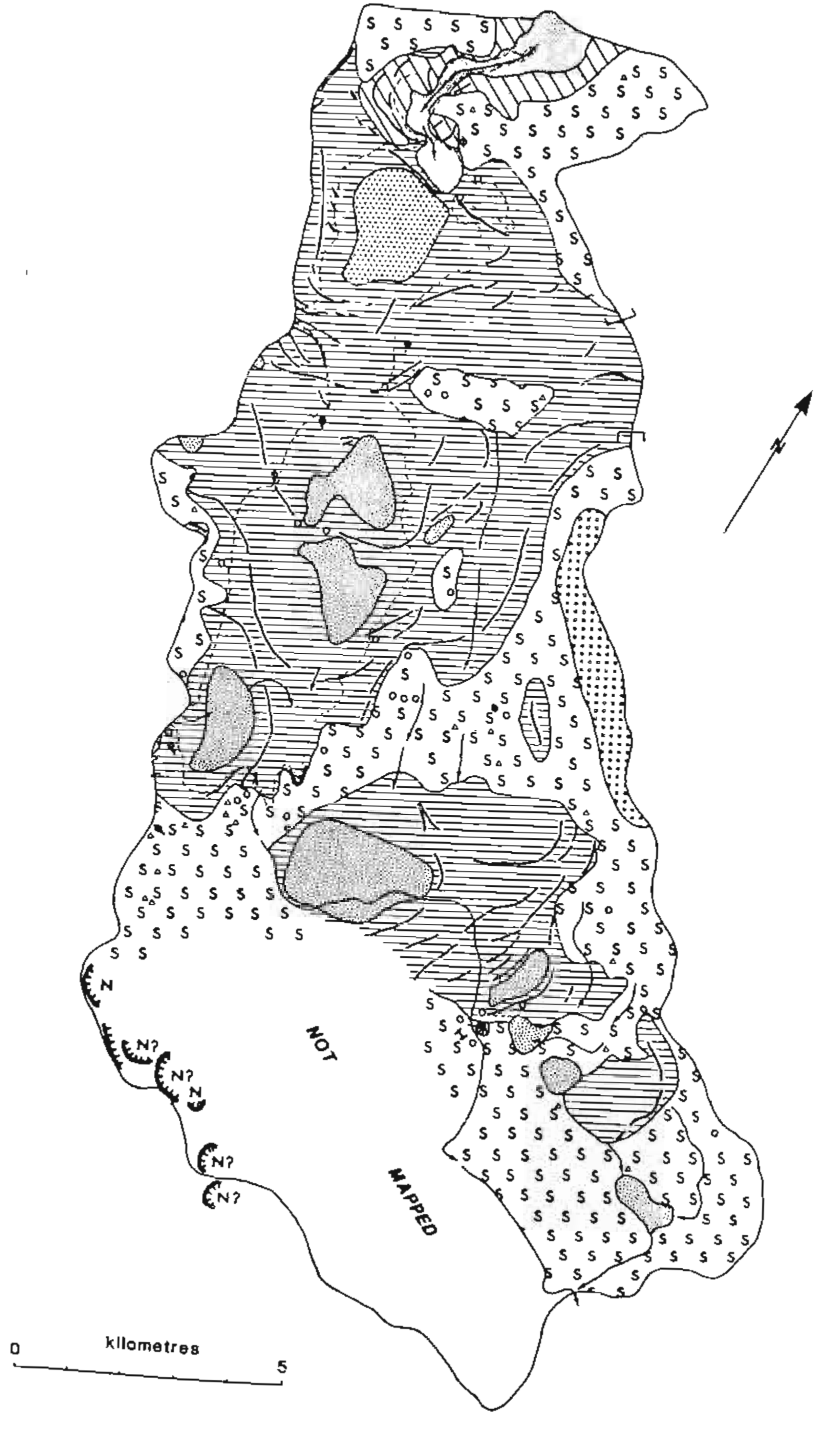


Figure 9.4 Glacial geomorphology, Derwent Bridge to Mossy Marsh Creek. See Figure B2 for key.

and that rises 130 m above the southeastern shoreline of Lake King William has been over-ridden. In part this probably occurred due to the deflection of the Derwent Glacier eastwards by the Guelph Glacier. This ice then descended to erode a broad valley in Triassic sandstone that crops out to the east of the fault. The crest of Sawmill Ridge which forms the western shoreline of Lake King William also appears to be ice smoothed.

At least 5 m of saprolite formed from dolerite is overlain by the Cascade Moraine 5.6 km south-east of Lake King William. The upper part of the saprolite has been disturbed during the construction of this moraine. This suggests that it was not frozen when the moraine was constructed (cf. Hale, 1958). The site occurs close to the entrance to the Derwent Gorge where active glacial erosion may have been anticipated if the ice had passed through the gorge. However, the saprolite has not been removed despite the fact that probable glacial deposits suggest the Derwent Glacier previously extended 16 km. beyond this point.

B Glaciofluvial erosion

A short gorge 15 m deep has been cut by the Derwent River south of the Derwent Bridge settlement. This was probably initiated as a subglacial meltwater channel, the location of which was governed by low confining ice pressures on the eastern margin of the Derwent Glacier (Derbyshire, 1971a).

Marginal meltwater channels are cut in rock and drift about 20 m above the level of Lake King William on the western slopes of the Wentworth Hills. Examination of the bathymetric chart of Lake King William reveal many probable proglacial channels that have been cut in drift on the lake bed (figures 9.3 and 9.4). Small col gulleys cross the dolerite ridge on the south eastern shoreline.

Butlers Gorge, site of the 67 m high Clark Dam which impounds Lake King William, has been deepened as a subglacial and later a proglacial meltwater route. The Derwent River was forced to the eastern side of its valley immediately upstream of Butlers Gorge by high ice pressures that were associated with the entry of a major tributary glacier from the Guelph Valley. Following the retreat of the ice the river was maintained in this position by the end moraine that was deposited during the Guelph phase (chapter eight).

Other meltwater channels south-east of Butlers Gorge still conduct small streams around the distal faces of terminal moraines. Small cascades in the bed of the Derwent River a short distance downstream of the Cascade Moraine probably owe their existence to increased hydrostatic pressure of meltwater beneath a glacier. This suggests that the Derwent Glacier extended downstream of the Cascade Moraine. Meltwater has probably contributed significantly to deepening the Derwent Gorge upstream of Wayatinah.

C Non-glacial Erosion

Some small free faces that overlook the Derwent Gorge appear to be the result of slab toppling, while toppling is also responsible for the development of some small faces near the crest of the Wentworth Hills. Smaller scale rockfall has contributed significantly to the creation of free faces on Mt. Hobhouse but there are few fresh scars. In many cases a lichen cover extends between partly detached blocks and the faces from which they were derived, which indicates that little rockfall is occurring under present day conditions.

A number of shallow hollows that are bounded by small rock scarps occur on the Hobhouse Range and are probably nivation cirques (Peterson, 1969). Frost processes are probably responsible for the shattering of previously ice smoothed surfaces on the Hobhouse Range and on the floor of the Derwent Valley. These outcrops have also been subject to some etching of joints by chemical weathering.

Present day erosion by frost processes is restricted to the development of pipkrake in locations where the regolith has previously been bared. At about 680 m on the southern end of the Wentworth Hills small gaps up to 1 cm wide have been observed to form around clasts in slope deposits each winter. Ice pushes the finer matrix away from the clast due to differences in the thermal conductivity of the two materials (Washburn, 1979).

Lake King William occupies a small graben (Gulline et al., 1963) and hence the orientation of this portion of the river is the result of fluvial erosion that has been broadly controlled by faulting. Downstream of Butlers Gorge the river appears to follow joint orientations in the dolerite and faults that trend NNW - SSE (Jennings, 1955).

Present day fluvial erosion appears limited to the stripping of exposed regolith and the reworking of unconsolidated deposits from channel banks.

Human activity has impinged significantly upon geomorphic processes through the construction of dams, pipelines and roads which have changed the face of the landscape and altered stream regimes. The Derwent River is now virtually dry between Butlers Gorge and Wayatinah. The failure of a canal bank 5.8 km south-east of Butlers Gorge in the early 1970s caused a large landslip with a vertical range of about 60 m. Slippage occurred along a plane which has a mean slope of 30^{o}-40^{o} and the slide was 50 - 70 m wide. Its broad spoon-shaped headwall is formed in at least 10 m of saprolite.

DEPOSITIONAL LANDFORMS AND SEDIMENTS

A Glacial Deposits

(i) Wayatinah Moraine.

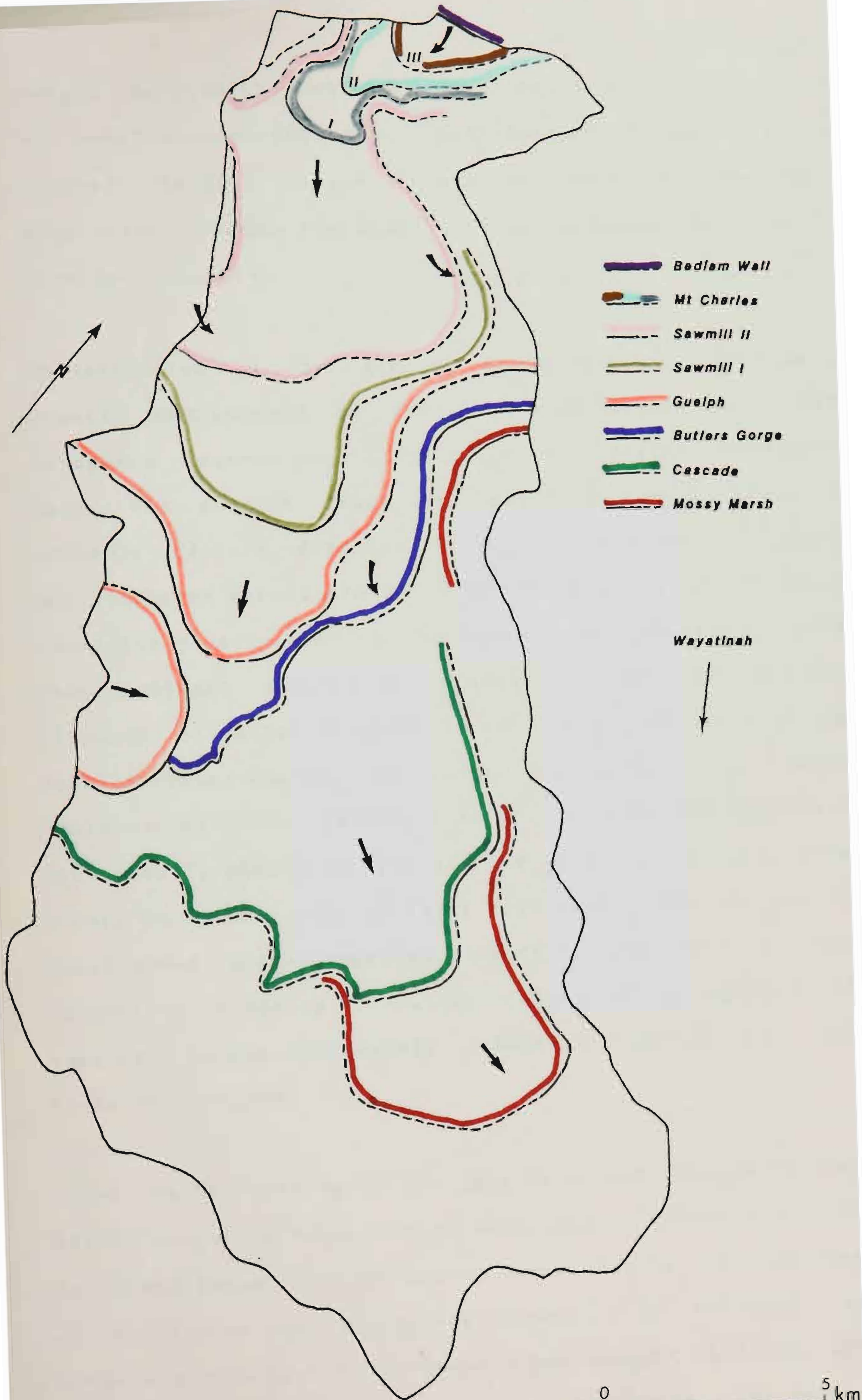
Sections exposed along the banks of the Nive River near its confluence with the Derwent, and for a short distance along the Derwent either side of the confluence reveal several metres of massive washed boulders up to 5 m long (Figure 9.5). The size and rounded form of these boulders strongly suggests glacial transport to this site (Kiernan, 1983a). The boulders overlie up to 1 m of angular dolerite slope deposits that in turn overlie weathered bedrock but which include few preweathered clasts. Bedded gravels of fluvial

Figure 9.5 Massive boulder deposits beneath fluvial sands and gravels at the confluence of the Nive and Derwent rivers are probably of glacial origin. These deposits lie nearly 70 km. from the headwall of the Derwent Glacier in the Du Cane Range. The wall of the section is 10-15m. high.

Figure 9.6 Partly flooded deposit 200m north of the section illustrated in Figure 9.5. Rotted dolerite cobbles and very large boulders occur in a sandy clay or silty clay matrix.

overleaf

Figure 9.7 Principal ice limits in the Derwent Valley downstream of Derwent Bridge.


III
II
I
N
Bedlam Wall
Mt Charles
Sawmill II
Sawmill I
Guelph
Butlers Gorge
Cascade
Mossy Marsh
Wayatinah
0
5
km

origin overlie the boulder deposit and are in turn overlain by solifluction deposits. Solifluction deposits 1 km further south contain clasts of sandstone, mudstone, quartzite, dolerite and basalt. Some of these clasts may be reworked erratics.

Investigation of the glacial deposits in this area is greatly complicated by the Wayatinah Lagoon and Lake Catagunya hydro-electric storages but further fieldwork during the present study has revealed the presence of probable glacial deposits. Probable till that is compact but weathered occurs along the shoreline of Wayatinah Lagoon immediately adjacent to the boulder deposit (Figure 9.6). This sediment appears to consist entirely of dolerite although it is not possible to confirm the identity of some totally rotted clasts. The clasts range from sub-rounded boulders up to 2m in length to cobbles that are rounded or occasionally platey in form and may be quite angular. These occur in a sandy clay or silty clay matrix that is sparsely distributed in some sections. Some of the boulders have weathering rinds up to 70.0 mm thick. It is highly likely that this is glacial deposit although its origin may prove to be contentious.

Prior to the filling of the Lake Catagunya storage Jennings (1955) mapped an extensive accumulation of dolerite boulders in a weathered dolerite matrix downstream of the confluence of the Derwent and Florentine rivers. He believed that these deposits were river gravels and thought that they were Tertiary in age, presumably due to their degree of weathering. His map extends northwards to the Nive River where the boulder deposit described earlier is mapped as part of the same unit. It is possible that the now flooded downstream gravels are also of glacial origin but it has not been possible to confirm this.

This site lies 16 km downstream of the maximum limit of the Derwent Glacier that was recognised by Derbyshire (1963, 1967), and lies 27 km downstream of the nearest glacial deposits in the Nive Valley. The intervening terrain in both valleys is mantled by thick slope deposits. As the Derwent Glacier is likely to have always been much larger than the Nive Glacier it seems the more likely of the two glaciers to have reached Wayatinah. If glaciers transported the large boulders down either valley then some large boulders should also occur in the lower parts of the valleys. Similar large boulders are present in the Derwent Gorge, but they are absent from the lower end of the Nive Gorge.

There are no other likely sources for the boulders at Wayatinah. Cirques and rock basins indicate that glaciers developed on the elevated Wylds Craig - Mt. Shakespeare massif but Wayatinah lies 20 km from these cirques along the shortest practicable route via the Florentine Valley, and the arrival of the till from this direction seems highly improbable. Some ice may have accumulated on the north-east slopes of Mt. Shakespeare only 4 km from Wayatinah but the potential snowfence lies at only 800 m Cirque-like hollows occur at 700 m at the head of the Counsel River on the southern part of the Hobhouse Range and at 800 m on the northeastern slopes of Mt. Shakespeare. Neither of these is likely to have been the source of extensive glaciers that could have reached Wayatinah.

It is therefore argued that at its maximum extent the Derwent Glacier probably reached at least 70 km from its headwall in the Du Cane Range, and descended from the St. Clair Surface to an elevation of as low as 210 m (Figure 9.7). Little evidence is available regarding the thickness of the ice, but quartzite erratics exposed among the roots of windthrown trees at 860 m on the northern slopes of Mt. Hobhouse and the erosional morphology at a similar altitude on the eastern slopes of the Hobhouse Range suggest at least 180 m of ice filled the Derwent Valley at that point.

Time has not permitted exploration for glacial deposits downstream of Wayatinah where the search is increasingly comlicated not only by a near continuous string of artificial lakes but also the possible entry into the Derwent Valley of glacial deposits from eastward flowing tributaries.

(ii) Mossy Marsh Moraines

A low moraine with its crest at 680 - 710 m is plastered against a rock bar 9 km south-east of Butlers Gorge in the headwaters of Mossy Marsh Creek. A small outwash plain lies downstream of the moraine. An exposure of meltout till at the eastern end of the moraine reveals that all lithologies except quartz are deeply rotted (Figure 9.8, tables 9.1 and

Figure 9.8 Weathered till in the headwaters of Mossy Marsh Creek.

Figure 9.9 The Cascade Moraine has been constructed upon deeply weathered dolerite bedrock downstream from Butlers Gorge.

9.2). A fabric diagram by Derbyshire (1967) indicates that this till was deposited by ice that flowed in a south-easterly direction and this is compatible with the morphological evidence. The maximum thickness of the dolerite weathering rinds in the till is 14.2 mm. The till is overlain by solifluction deposits in which the maximum weathering rind thickness is 3.6 mm.

The overlying slope deposits make it difficult to identify the limits of the till on the valley walls but the presence of erratics at 790 m south-west of Butlers Gorge, and at 850 m in the Laughing Jack Col suggests a minimum ice surface gradient of 14 - 16 m./km. This gradient implies that the Bedlam Walls ridge south of Lake St. Clair must have been over-ridden, despite the apparent absence of erratics and the thick slope deposits that occur on its crest (chapter six).

(iii) Cascade Moraine

An end moraine complex with its most downstream crest at about 670 - 690 m occurs 5.6 km south-east of Butlers Gorge and marks the downstream limit of continuous drift in the Derwent Valley. A small cascade and meltwater gorge occur in the bed of the Derwent adjacent to the downstream crest. Part of this moraine complex has been constructed upon deep saprolite (figure 9.9).

The till on the eastern side of the valley contains a higher proportion of Triassic sandstone than occurs to the west where the till consists almost entirely of dolerite (Table

9.1). A fragment of highly lithified grey basal till (Nive Till: chapter 12) was found among recent landslide debris that extends from the distal face of the Cascade moraine into the Derwent River. As has previously been argued in chapter four, this degree of lithification must have been achieved prior to the incorporation of the clasts into subsequent moraines. The presence of Jurassic dolerite in this basal till clast proves that it is not a fragment of Permian tillite. This indicates that the Derwent Valley was glaciated well prior to the construction of the Cascade Moraines.

The Cascade Moraines mark a phase during which the Derwent Glacier was initially impounded behind a rock ridge that extends southwestwards across the valley, and that the glacier then commenced a slow retreat. Undulating ground moraine and fluvioglacial sediments occur upstream of this point.

(iv) Butlers Gorge Moraine

A partly drowned latero-terminal moraine is plastered against the rock ridges that form the southern shoreline of Lake King William. It can be traced upstream towards the Wentworth Hills. A thin outwash plain lies at 660 m with its apex 1 km downstream from the Clark Dam. This moraine marks a phase during which the Derwent Glacier was locked behind the rock bars and was probably still confluent with ice that flowed eastwards through the Guelph Gorge from the King William Range (chapter eight). The glacial deposits are rather thin and it is probable that the phase was

Vicinity No.	2	5a	5b	7	8a	8b	12	13a	13b	14	15	26	27
%		(> 10cm)	(< 10cm)		(> 10cm)	(< 10cm)		(> 10cm)	(< 10cm)				
dolerite	52.2	70.0	40.9	99.2	66.7	40.0	64.8	98.5	66.7	100.0	84.1	85.2	82.5
quartz	25.1	23.0	22.7	0.5	26.7	60.0	1.9	1.5	31.1	-	12.5	11.7	11.0
argillites	6.3?	7.0	25.3?	0.3	3.5	-	-	-	-	-	3.0?	2.1?	6.2?
basalt	11.4	-	-	-	3.1	-	33.3	-	2.2	-	-	-	-
other	5.0	-	11.1	-	-	-	-	-	-	-	0.4	1.0	0.3

Table 9.1 Lithology of glacial deposits in the area between Derwent Bridge and Wayatinah. Site locations are indicated on Figure 9.1.

short-lived.

(v) Guelph Moraines

During the Guelph phase the Derwent Glacier stood a short distance upstream of the Guelph - Derwent confluence. A pronounced ridge on the lake floor that is indicated on the bathymetric chart corresponds with the position of a large moraine reported by Prider (1948). Its crest lies at 687 -690 m and a probable small outwash plain lies immediately downstream. The ridge is continued above the lake where a moraine can be traced westwards onto Sawmill Ridge (Tasmap 8113: 368 220). The morphological evidence indicates that this is an end moraine which was constructed by the Guelph Glacier following its separation from the Derwent Glacier. The northern limb of the moraine is slightly concave to the north-east which suggests that part of it was deposited by the Derwent Glacier. This interlobate moraine therefore seems to indicate the separation of the two glaciers. A small outwash plain was later constructed by meltwater from the Guelph Glacier on the upstream side of this moraine.

The moraine ridge that was constructed by the Derwent Glacier defines an ice limit that would have been sufficient to enable the Derwent Glacier to overtop the northern end of Sawmill Ridge and extend into the upper Guelph Valley. It is therefore equivalent to the Guelph phase in the Navarre Valley (chapter seven) and in the Guelph Valley (chapter eight). Lateral moraines to the east indicate that ice may no longer have continued to flow out of the Derwent Valley through the Laughing Jack Col.

Few sections are available through that part of the complex which remains above lake level, but Prider (1948) records the presence of 2 m boulders of Precambrian quartz schist on the valley floor 800 m north of Butlers Gorge. Prider argued that these implied movement of ice from the north-west. Quartz schist bedrock is certainly exposed in the upper Franklin Valley (chapter twelve) but the extension of the Franklin ice south-eastwards to Butlers Gorge is contrary to the morphological evidence in the Navarre Valley. This indicates that during a later phase the Derwent Glacier prevented ice from the Franklin Glacier passing eastwards of the King William Range (chapter seven), a situation which is likely to have also applied during earlier phases of mutually greater ice extent.

The quartz schist erratics at Butlers Gorge could have been reworked from Permian glacial deposits in the same manner as the bedouin boulders at King William Saddle (chapter seven). Quartz schist forms a significant proportion of the clasts in the Lake Rufus end moraine complex where they have been derived from the Permian sediments that were deposited immediately to the east of a large area of quartz schist terrain (chapter eight). It would have been a simple matter for erratics to have been carried towards Butlers Gorge from this source. Alternatively, these quartz schist erratics may have been carried from source rocks west of the King William Range by the Top End Glacier which was a tributary of the Guelph Glacier (chapter eight).

The survival of a schist-rich moraine on the floor of the Derwent Valley confirms that the Derwent Glacier had retreated upstream of this locality, enabling the Guelph Glacier to extend through the Guelph Gorge.

(vi) Sawmill Ridge Moraines

A complex of low end moraines that can be traced upslope from the shores of Lake King William continue the trend of prominent ridges on the lake floor. These moraines form the Sawmill Ridge complex in the Navarre area, and at Lake King William can be readily differentiated into two phases on morphological grounds, both of which are represented by an end moraine and outwash plain.

During the Sawmill I phase the Derwent Glacier constructed a major latero-terminal moraine between the mouth of the Laughing Jack Col in the east and Sawmill Ridge in the west. A probable outwash plain occurs on the lake bed at 696 - 699 m. With the continued decline of the Derwent Glacier, diffluent flow through Laughing Jack col had ceased by this time. Bedlam Wall would have emerged through the ice surface to thereafter divide the Derwent Glacier into its western (Navarre) and eastern (Traveller) lobes.

During the Sawmill II phase a massive end moraine over 200 m. broad was constructed immediately south of Navarre Arm. Shallow sections on the southern side of the arm reveal red basal till plastered against the moraine face, which suggests that the ice front was active. A probable outwash plain occurs on the lake floor at 705 - 708 m. After this phase confluence of the two lobes of the Derwent Glacier south of the Bedlam Wall ridge ceased. A medial moraine was deposited south of Bedlam Wall. As the majority of the Derwent Glacier passed to the west of Bedlam Wall the ice at the head of the present Lake King William diminished rapidly after this time.

(vii) Mt. Charles Moraine

A small terminal moraine descends from 735 m on the Bedlam Wall ridge into the northern end of Lake King William. It was constructed by a thin body of ice which passed through the gap between Bedlam Wall and Mt. Charles from the eastern lobe of the Derwent Glacier. Low lake levels in the summer of 1983 - 84 revealed an extensive outwash plain immediately downstream of this moraine at 708 - 711 m. A typical section in the riverbank revealed 1 - 1.5 m of rounded dolerite cobbles up to 20 cm long. These are overlain by poorly sorted coarse cobbles with some clasts up to 50 cm in size in a sparse matrix of sand. Current bedded sands and steeply foreset cobble beds suggest torrential flow. They contrast with the imbricate structure of the present day point bar deposits that have been deposited by less torrential flows. This section suggests that some readvance may have occurred.

Upstream from this moraine lies a further plain formed of outwash gravel and sand. Massive clay indicates that a small meltwater lake occurred at this site (Derbyshire, 1971a). Two smaller moraines (Mt.Charles II and III) were constructed between the Mt. Charles Moraine and Derwent

Bridge as the ice retreated.

(viii) Bedlam Wall Moraine

The northernmost moraine in the area encompassed by this chapter can be traced discontinuously eastwards into the Travellers Rest Valley and also northwards into the Lake St. Clair area. It has been described in chapters four and five. A later small outwash plain has partly inundated the Bedlam Wall Moraine around Derwent Bridge and can be traced northwards to the Cynthia Bay moraines (chapter four).

B Non-glacial deposits

(i) Slope deposits

Solifluction deposits mantle the slopes of the Wentworth Hills and the northern end of the Hobhouse Range. Slope deposits up to 1.5 m thick have been located within all the ice limits described in this chapter, but have not been found to overlie the Cynthia Bay outwash at the northern end of Lake King William. At least 1 m of weathered slope deposits overlie the till in the headwaters of Mossy Marsh Ck.

These solifluction deposits can usually be differentiated from bouldery till because the till stones tend to be more rounded. In other cases corestones that have been derived from deeply weathered dolerite are incorporated into the slope mantles. The progressive incorporation of corestones into slope deposits is well seen in a section on the eastern bank of the Derwent 300 m downstream of the Cascade Moraine. Such deposits can sometimes be differentiated from reworked glacial clasts because the latter are generally less spherical and the till usually contains erratics.

Mantles of mechanically shattered clasts occur on the flanks of the Wentworth Hills, Mt. Hobhouse and the Bedlam Wall ridge. Most of these mantles occur beneath free faces and have accumulated as rockfall talus. Block slopes on the upper levels of the Wentworth Hills a few hundred metres northwest of Darcys Bluff have been reworked into small block streams. Minor accumulations of spalled dolerite shells occur in some locations, notably west of the weir downstream of Butlers Gorge. These are probably the result of repeated wildfires.

(ii) Alluvial, peat and archaeological deposits.

Up to 2 m of yellowish coloured sand and silt with occasional small pebbles overlies the outwash gravel at the northern end of Lake King William and has accumulated as an overbank deposit. Tributary streams from the Wentworth Hills and Mt. Hobhouse have generally deposited only sand and silt upon the Pleistocene gravel. The sand is locally overlain by peat south of Derwent Bridge. The bed deposits of the Derwent River consist of reworked outwash.

Up to 6 m of debris which ranges in calibre from fine clay to boulders 3 m in diameter has accumulated on the semi-dry bed of the Derwent River 6 km south-east of Butlers Gorge, following a landslide caused by the failure of a hydro-electric diversion canal. The abundant water associated with deposition of this material is reflected in small bodies of waterlaid gravel and cobbles together with cross-bedded sand and fine clay. The bulk of this material was derived from deeply weathered dolerite and slope deposits, but one or two clasts of grey basal till (Nive Till) have also been located among the debris 20 m above river level.

Prehistoric archaeological deposits have been found close to the Cascade moraine and west of the Laughing Jack col. Prehistoric human activity has probably resulted in some redistribution of rock materials suited to the manufacture of stone tools. This trend has continued on a far larger scale with the advent of roadbuilding and hydro-electric construction activities.

DATING AND DISCUSSION

Shallow organic-rich silt that overlies overbank sand south of Derwent Bridge and occurs in depressions on the buttongrass plains is the youngest sediment in this area. The earlier sand has been reworked from outwash deposits. The deposition of similar sand at Laughing Jack Lagoon in the adjacent Nive River basin ceased shortly after 1540 +- 60 BP (SUA 1958) (chapter ten).

Solifluction deposits occur throughout the area and overlap all the till deposits, which they must therefore postdate.

The Bedlam Wall Moraine lies upstream of the Mt. Charles Moraine. The outwash deposits that extend downstream from the Mt. Charles Moraine partly inundate the earlier glacial deposits that occur further downstream. The Mt. Charles, Sawmill Ridge, Guelph and Butlers Gorge end moraines and their associated outwash plains record earlier and more extensive ice limits. Apart from the steep rock slopes immediately downstream of the Clark Dam, glacial deposits extend continuously between Butlers Gorge and the Cascade Moraines. The presence of clasts of Nive Till in the Cascade Moraines indicates that an early glaciation took place in the Derwent Valley but Nive Till has not been found <u>in situ</u> elsewhere in the area.

Weathered solifluction deposits occur between the Cascade and Mossy Marsh moraines. Little moraine morphology remains apparent at the Mossy Marsh ice limit. No moraines have been identified on the St. Clair Surface downstream of Mossy Marsh Creek but probable glacial deposits are exposed a further 16 km downstream at an altitude of 210m.

The presence of deeply weathered bedrock between the Mossy Marsh and Cascade moraines, and beneath the slope deposits and glaciofluvial gravels at Wayatinah points to a protracted period of warm humid climate prior to the initial glaciation of the Derwent Valley.

The maximum limits of the Derwent Glacier remain uncertain. The deposits that occur downstream of the Florentine-Derwent confluence have largely been inundated in the filling of the Lake Catagunya hydro-electric storage. Other storages occur at close intervals further downstream and have flooded the most likely sites for the preservation of any glacial deposits. Unfortunately the associated excavations are largely confined to steep damsites where deposits are unlikely to have been preserved.

Postdepositional characteristics of the landforms and deposits are presented in Table 9.2.

Deep soil profiles occur on all the glacial deposits. The slope deposits that overlie moraines at the foot of the Wentworth Hills generally have shallower profiles in which weathered clasts are confined to the top metre. Solifluction deposits in excess of 1 m thick overlie the Mossy Marsh Moraine. Solifluction deposits up to 4 m thick occur between Mossy Marsh and Wayatinah. This suggests that the till at Wayatinah may be ancient.

The difference in the mean thickness of the weathering rinds in the glacial deposits between the Sawmill Ridge moraines and Wayatinah is not significantly different at one standard deviation. This suggests that they do not differ significantly in age. However, the maximum weathering rinds of 5.2 mm that occur in the Mt. Charles and Bedlam Wall moraines contrast with rinds as thick as 14.2 mm in the glacial deposits at Mossy Marsh Creek. The weathering rinds in the possible drift at Wayatinah are up to 70.0 mm thick.

While the moraines are well preserved upstream of the Cascade Moraine, the Mossy Marsh deposits are degraded. No moraines occur between Mossy Marsh and Wayatinah which

moraines[a]	-	-	WY	WY	HH	HH	C	C	C	C	C	C	C	C	BG	C	S	S	S	S	S	S	MC	MC	BW	BW	-
vicinity no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	26	15	16	17	18	19	20	21	21	22	23	24	25
material[b]	B	S	O	T	T	T	T	T	F	T	F	T	F	F	T	T	T	T	T	S	T	T	T	T	T	T	S
A. MORPHOSTRATIGRAPHY																											
relative position	outside 4	overlies 1	overlies 2	inside 3	inside 4	outside 3	inside 5	inside 7	inside 8	inside 8	inside 77	inside 77	inside 7	outside 13	inside 7	inside 41	inside 14	outside 3	outside 5	outside 15	inside 1	inside 15	inside 16	inside 13	inside 13	inside 13	inside 14
crest/site elevation(m)	370	335	350	331	700	700	370	680	630	640	530	700	670	560	360	730	735	420	640	850	700	760	740	730	715	815	1010
max. depth of burial (ky) (m)[c]	[illegible]	05	[illegible]	31	[illegible]	51.5	25	-	-	-	51	51	-	-	55	51	53	-	-	-	-	-	450	51.5	41.5	41.5	-
B. SUBSURFACE CLASTS																											
dolerite rinds-max. (mm)	-	8.1	0.6	20.0	14.2	13.8	11.0	7.5	8.1	11.1	6.0	12.1	7.5	13.2	12.3	-	14.2	boulder 8.2	6.1	-	3.8	8.6	-	5.0	5.7	5.2	3.2
min.	-	0.5	0.1	3.2	2.1	6.0	1.0	0.3	1.1	4.5	1.8	1.8	1.5	3.3	2.1	-	3.3	1.0	1.7	-	1.3	2.8	-	1.4	0.1	2.2	0.8
mean	-	2.3	0.3	13.6	8.8	9.5	7.9	2.2	4.0	7.6	3.3	4.5	4.0	6.6	4.3	-	6.6	4.1	3.1	-	2.5	5.1	-	2.9	2.3	3.5	1.3
SD	-	1.5	0.1	10.2	3.6	3.0	2.4	2.1	3.4	2.3	0.7	3.5	1.6	3.4	1.9	-	2.6	2.2	0.1	-	0.5	2.6	-	1.0	0.5	1.7	0.6
hardness(1-5)	-	2	1	3-4	2-3	1-4	2-4	2-4	1-4	3-3	2-4	3-4	2-4	1-4	1-3	-	3-4	1-3	3	-	2	2	-	2	1-2	3	2-2
coherence of argillites	-	-	2	-	4-2	4	4	4	4-5	4	4	4	4	4-5	-	-	4	4	4-5	-	4	2	-	4	4	3-4	1
dolerite clast surface colours	5YR	5YR	7.5-10YR	10-5YR	5YR	5YR	7.5-5YR	5YR	5YR	5YR	7.5YR	5YR	7.5-5YR	5YR	5YR	5YR	5YR	7.5-5YR	5YR	7.5-5YR	7.5YR	5YR	5YR	7.5YR	7.5YR	5YR	7.5YR
% of totally unrecognizable clasts	-	-	< 5	20	-	10	40	33	-	-	-	13	-	30	-	-	20	-	-	-	-	-	9	-	5	-	0
C. SUBSURFACE MATRIX																											
colour of B horizon	5YR	-	-	7.5-5YR	7.5-5YR	7.5-5YR	5YR	7.5YR	7.5YR	7.5YR	7.5YR	7.5YR	7.5YR	7.5YR	5YR	7.5-5YR	7.5-5YR	5YR	7.5YR	5YR	7.5-5YR	7.5-5YR	7.5-5YR	7.5YR	5YR	7.5YR	10YR
depth of B horizon(cm)	35	-	-	> 100	> 300	> 300	< 400	> 200	> 300	> 300	> 400	>300	< 150	> 300	> 100	> 50	> 300	> 50	>130	>200	< 50	> 80	>100	<100	> 150	> 100	> 70
depth of CaX horizon(cm)	-	-	-	> 100	> 300	> 300	> 400	> 200	> 300	> 200	> 400	>300	> 150	> 300	> 100	> 50	> 200	< 50	<130	>200	> 50	> 80	<100	<100	> 150	> 100	> 70
clay size fraction mineralogy	-	-	-	-	kaolinite [illegible]	-	-	-	-	-	-	quartz kaolinite	-	-	-	-	-	-	-	-	-	-	-	-	quartz kaolinite	quartz	-
development of clay films (1-5)	5	-	-	4-5	5	5-3	3	3	3	3	3	3	3-4	3-4	5	2-1	3-4	3	4	3-4	1-2	2	3	3	1-2	3-4	1
coarse matrix: % > 1cm	-	-	-	-	32	-	73	-	-	-	-	-	-	61	-	-	-	-	-	-	-	-	43	-	-	-	-
% 1-0.5cm	-	-	-	-	17	-	20	-	-	-	-	-	-	13	-	-	-	-	-	-	-	-	11	-	-	-	-
% < 0.5cm	-	-	-	-	31	-	7	-	-	-	-	-	-	26	-	-	-	-	-	-	-	-	36	-	-	-	-
fine matrix: %sand	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	60.3	-	-	81.3	-	-	-	63.6
%silt	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	20.5	-	-	1.8	-	-	-	20.1
%clay	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	19.2	-	-	16.9	-	-	-	16.3
organic at ~100cm (%)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	6.8	5.0	-	-	13.8	-	-	-	-
degree of lithification of basal till (1-5)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
clast socket staining (1-5)	-	-	-	4-5	4-5	3-5	3-5	1-3	3	3-5	2-4	3-4	2-4	4-5	3-4	-	4-5	3-4	2-4	2-5	2-4	-	3	3	3	3	3
D. SURFACE MATERIALS																											
% split clasts	-	-	-	-	-	-	-	-	-	-	-	-	-	-	32.0	23.0	-	-	-	-	-	-	-	-	-	-	1
E. DEPOSITIONAL LANDFORMS																											
moraine crest width(m)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	50	-	-	-	-	-	-	-	-	-	-	-
moraine slopes - proximal (°)	-	-	-	-	<7	-	-	-	<10	-	-	-	-	-	-	<10	-	-	-	-	-	-	-	-	-	-	-
- distal (°)	-	-	-	-	<3	-	-	-	<10	-	-	-	-	5,15	-	<10	-	-	-	-	-	-	-	-	-	-	-
degree of fluvial dissection(1-3)	-	-	-	-	4-5	3	-	-	2-3	2	-	-	-	3-4	2	2-4	-	-	3	-	-	4	-	-	3	3	-
F. ERODED BEDROCK																											
width of joint opening (mm)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	10	-	-	-	-	-	-	-
depth of joint opening(mm)	60	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	>70	-	-	-	-	-	-	-
height of tors (m)	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	5	-	-	-	-	-	-	-

(a) WY= Wayatinah; HH= Honey Marsh; C= Cascades; BG= Butlers Gorge; G= Guelph; S= Sawmill; MC= Mt. Charles; BW= Bedlam Wall.

(b) B= bedrock; T= till; S= solifluction deposit.

(c) Refers to burial of the same depositional unit in the same vicinity but not at the exact location where subsurface data were collected. The latter have only been recorded from unburied sites. O= outwash gravels; T= till; S= solifluction deposit; F= fluvial gravels; A= alluvial sand and silt.

(d) after Derbyshire (1963).

Table 9.2 Post-depositional modification of deposits between Derwent Bridge and Wayatinah. Site locations are indicated on Figure 8.1.

suggests that the glacial sediments in this area may have been reworked during a phase or phases of periglacial activity that did not affect the moraines further upstream. Weathering rinds on dolerite clasts in the solifluction deposit that overlie till and glaciofluvial sediment upstream of Mossy Marsh reach a thickness of 3.6 mm. Those that occur in the slope deposits that overlie the till at Wayatinah reach 6.1 mm. This suggests that the solifluction deposits at Wayatinah may be significantly older than those at Mossy Marsh Creek. This suggests in turn that the till beneath the deposits at Wayatinah may be older than that near Mossy Marsh.