LATE CAINOZOIC GLACIATION

AND

MOUNTAIN GEOMORPHOLOGY

IN THE

CENTRAL HIGHLANDS OF TASMANIA.

by Kevin Kiernan

VOLUME
TWO

chapter ten

THE WESTERN NIVE VALLEY

A large proportion of the Central Plateau Ice Cap drained southeastwards into the headwaters of the Nive River (Jennings and Ahmad, 1957). The westernmost tributary of the Nive, known as the Clarence River, rises from Clarence Lagoon, a moraine-dammed lake on the edge of the plateau 5 km. southeast of Travellers Rest Lake. From there it descends steeply down onto the St. Clair Surface (Figure 10.1). Ice accumulated on the plateau at the head of the Clarence Valley (Clemes, 1925).

Two distributary lobes of the Derwent Glacier extended eastwards across the St. Clair Surface at the foot of the scarp that forms the southern margin on the plateau. One of these passed through the Clarence Col between the foot of the scarp and Mt. Charles (chapter six), the other through the Laughing Jack Col between Mt. Charles and the Wentworth Hills further to the south (Derbyshire, 1967) (chapter nine). Small valley glaciers also developed on the eastern slopes of the Wentworth Hills. It has previously been argued that ice did not accumulate on this range due to its unfavourable NW - SE alignment and the fact that it lay within the rainshadow of the King William Range (Derbyshire, 1967).

This chapter details the evidence left by the ice masses that were confluent with the Derwent Glacier in this western

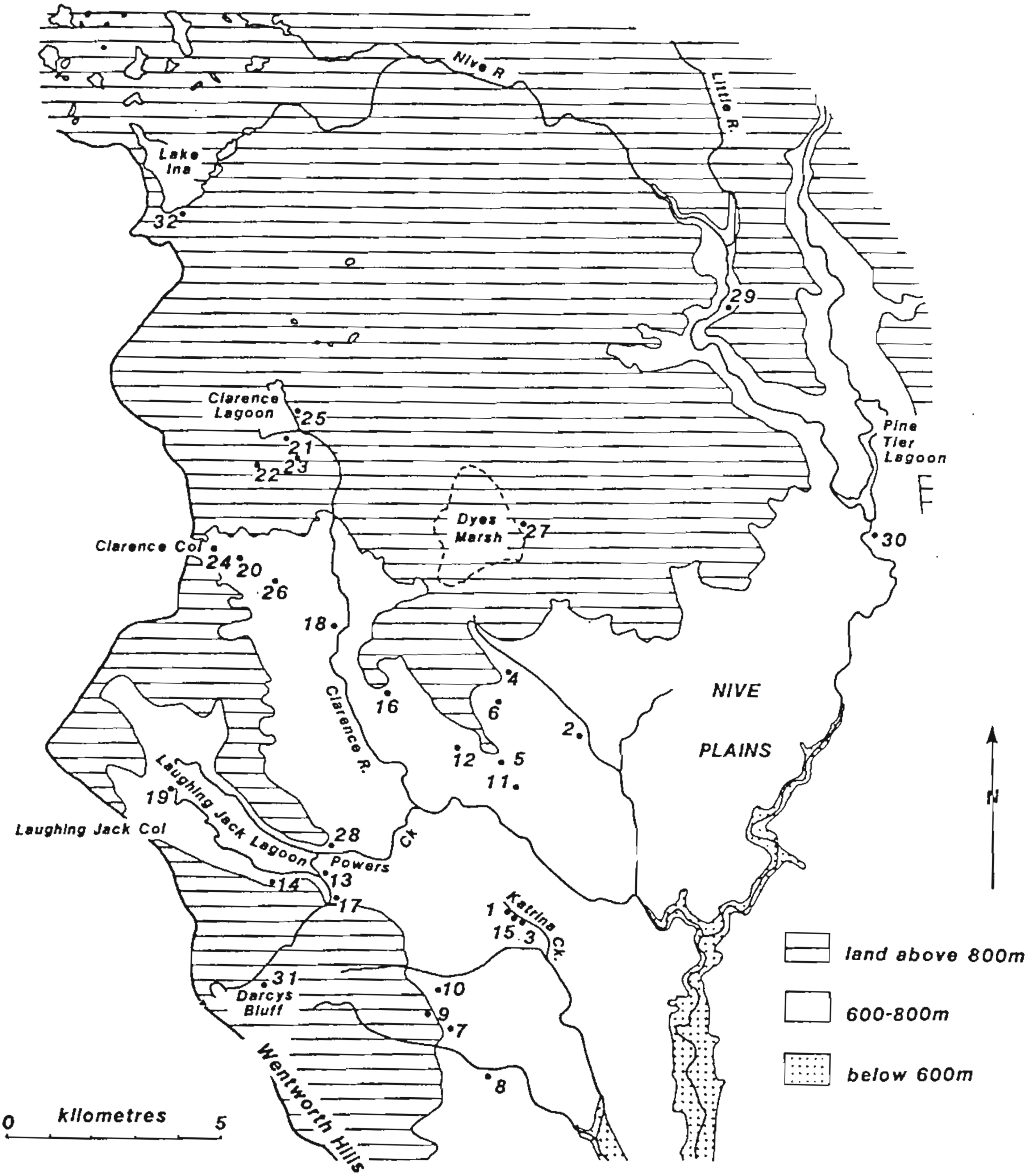

Figure 10.1 Locality map, western part of the Nive Valley. Numbers refer to site locations in tables 10.1 and 10.2.

part of the Nive catchment. Reconnaissance exploration along the main branch of the Nive further to the east has confirmed that the glacial sediments in the Clarence Valley and Wentworth Hills were deposited by local glaciers. They were not deposited by the main Nive Glacier which did not extend this far downstream. The distribution of glacial landforms and deposits in this area is presented in Figure 10.2. and the principal ice limits are presented in Figure 10.3.

EROSIONAL MORPHOLOGY

A Glacial Erosion

That portion of the Nive catchment which has been investigated during this study is characterised by weak glacial erosion of dolerite. The most widespread erosional landforms are ice-abraded rock surfaces. Freshly ice smoothed rock occurs north of Lake Ina on the plateau. Stoss-side smoothing and lee-side plucking indicates that the plateau ice moved generally south-eastwards, but some of the ice north-west of Lake Ina crossed a low divide and passed southwestwards into the Travellers Rest catchment. The abraded bedrock surfaces north of Lake Ina are fresher than those further south on the plateau and elsewhere within this area where considerable joint etching has taken place.

Abraded bedrock surfaces are common on the St.Clair Surface up to 4 km east of Laughing Jack Lagoon, where they are associated with till deposits. A brief reconnaissance along the main branch of the Nive between Pine Tier Lagoon and the Nive - Little Nive confluence found outcrops with a very

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 trenc
- 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

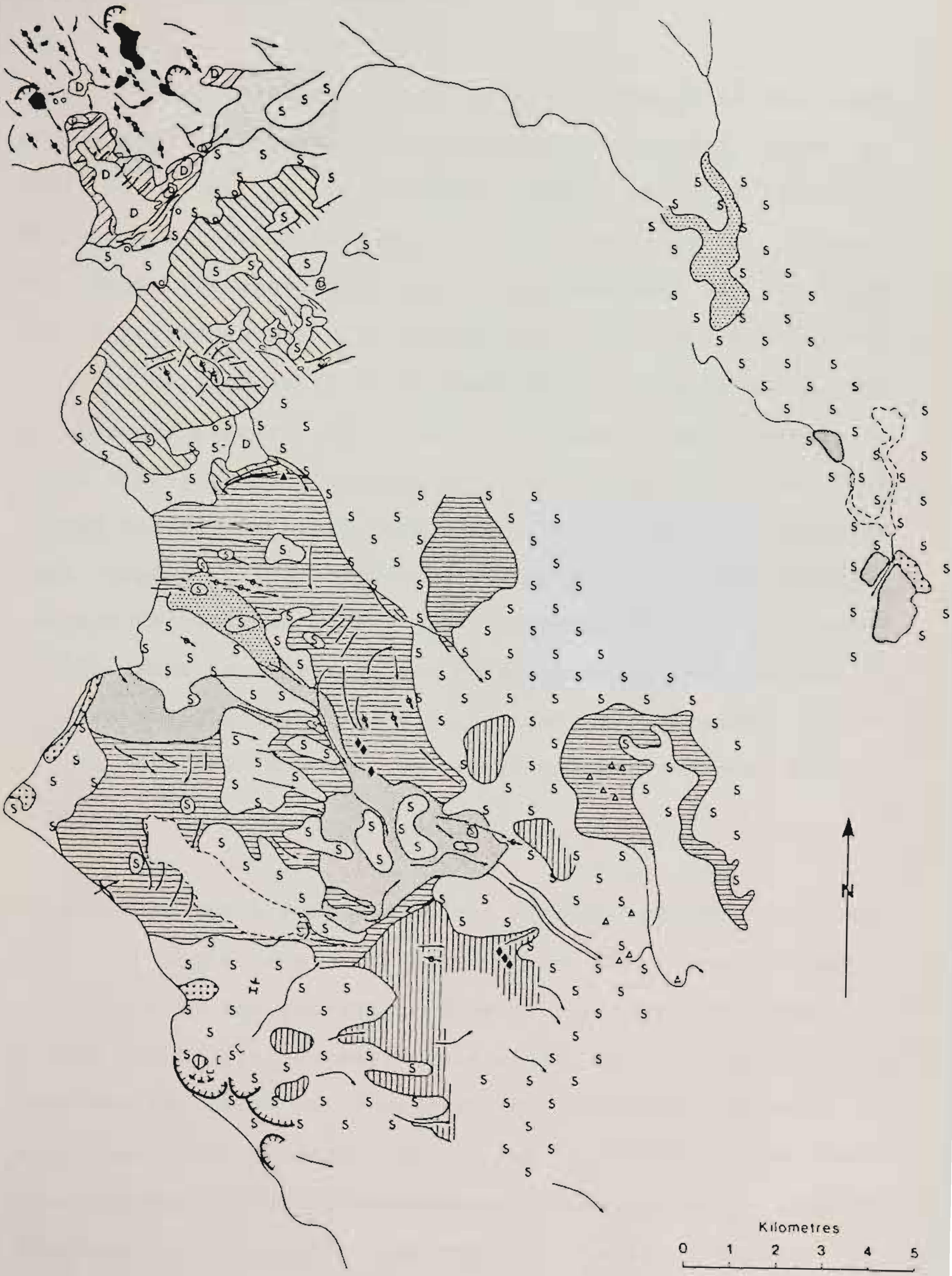

Figure 10.2 Glacial geomorphology of the western part of the Nive Valley. See figure B2 for key.

similar form but these could not with certainty be attributed to glacial erosion.

Only weakly developed cirques are to be found in this part of the Nive catchment. Three over-ridden cirques occur at 1040 -1060 m on the Central Plateau north of Lake Ina. Three poorly formed cirques occur at the head of valleys that drain the eastern slopes of the Wentworth Hills. These were regarded as nivation features by Derbyshire (1967) but are now reinterpreted as cirques on the basis of their size and because glacial deposits have been found downslope. Three valley steps occur along the course of the stream that drains the Darcys Bluff tarn into the artificial Laughing Jack Lagoon. The absence of end moraines in these shallow cirques is consistent with the observation by Derbyshire (1967) that in this area end moraines are seldom present in cirques that lack a threshold. A further possible cirque occurs at 890 m. on the south-eastern slopes of Mt. Charles.

A number of small rock basins are present on the plateau, most of which are elongated NW - SE along structural weaknesses in the dolerite. Some of the lakes have bays and inlets that are oriented almost at right-angles to this. Dyes Marsh, 4.5 km southeast of Clarence Lagoon, is surrounded by bedrock shores and is drained by two small channels that lie at its southeastern and southern margins. Although it gives an impression of being a shallow rock basin that has been filled by sediment there are no glacial deposits to support this suggestion.

The Clarence Col (790 m) and Laughing Jack Col (800 m) are diffluence cols by means of which distributary lobes of the Derwent Glacier extended into the Nive catchment. A further small col that has been cut across the Mt. Charles Ridge north-west of Tibbs Plain now conducts Broken Leg Ck. northwards into the Travellers Rest catchment (chapter six).

B Glaciofluvial erosion

Meltwater channels on the Central Plateau range from steep proglacial outlets which plunge down the scarp from the Clarence Lagoon area to rectilinear joint-controlled networks in the heavily eroded lake country that were probably initiated as subglacial conduits. A col gulley occurs at Broken Leg Creek. Meltwater scalloping of the bedrock is striking to the east of the Clarence Col.

Other meltwater channels have been cut in bedrock downslope from the Wentworth Hills cirques and at the eastern end of Laughing Jack Lagoon, where one has become the site of a small hydro-electric dam. The easternmost of the channels that plunge downslope from Dyes Marsh towards the Nive Plains may have been deepened by meltwater that discharged from the margin of the Derwent Glacier when its Traveller-Clarence lobe was lodged behind a rock bar southwest of the marsh. Further to the east the Nive River Gorge in the hill country that neighbours Pine Tier Lagoon has been deepened by meltwater. The major gorges of the Nive downstream of the Clarence River have probably also been deepened by meltwater from the eastern margin of the Derwent Glacier and from the glaciers that arose on the Wentworth Hills.

C Non-glacial erosion

The southeastward drainage of this part of the Central Plateau is the result of the tilting of the rocks to a gradient of about 1 in 130 in that direction some considerable time prior to 26 Ma BP when basalt was deposited in a pre-existing valley further east on the Central Plateau (Banks, 1973b). The more precise orientation of the Nive gorges broadly parallels the course of the Derwent and reflects structural lineaments in the dolerite. The drainage eastwards from the Clarence Col and Laughing Jack Col is strikingly parallel which suggests that both drainage lines have been guided by jointing in the dolerite, the Laughing Jack drainage having been superimposed onto Triassic bedrock.

Outpourings of basalt on the Nive Plains appear to have filled a small valley between the scarp and the Nive River, and led to the development of a minor twin stream system. As the youngest K/Ar assay that has been obtained on the basalts of the Central Plateau is 21.8 Ma BP (Sutherland _et al._, 1973) and basalt crops out close to the foot of the scarp it is evident that the rate of scarp retreat has been very slow (chapter thirteen). The scarp becomes progressively less steep towards the east where only limited slab toppling is in evidence. Localised rock crevasses and dilation trenches occur in dolerite bedrock at about 970 m on the northeastern ridge of Darcys Bluff and small slab topples occur immediately downslope.

Outcrops of dolerite at high elevations have been severely fractured by physical weathering. Partly detached fragments that occur near the top of some taller outcrops are highly angular, in strong contrast to spalling shells that occur near the base of outcrops and are probably the result of repeated fires. At the downstream end of Laughing Jack Lagoon a bedrock ridge is ice-smoothed to an elevation which approximates that of an adjacent moraine crest. Above this level joints in the dolerite have been deeply etched by weathering. This etching therefore provides an approximate geomorphic measure of the time interval between deposition of the moraine and the deposition of earlier tills further to the east by ice that completely over-rode the ridge.

Human activity is the major control upon present day erosion. Numerous erosional landforms are the direct result of developmental activities, including road-cuttings, quarries and canals. Timber harvesting and burning in the Dyes Marsh area has promoted severe sheet erosion of thin soils on the slopes of the scarp (figure 15.13). Once the regolith has been bared it is open to disturbance by minor periglacial processes. Road-cuttings are commonly disturbed by pipkrake formation in winter. Differences between the thermal conductivities of dolerite clasts and the surrounding matrix in tills and clay-rich slope mantles permits pipkrake to develop on the downslope side of clasts exposed in road-cuttings. This forces the matrix downwards and the clast migrates downslope into the resulting void once the ice melts.

DEPOSITIONAL MORPHOLOGY AND SEDIMENTS

(i) Nive Till

The earliest glacial sediment in the Nive catchment is a highly compact grey basal till that contains numerous rounded clasts of dolerite in a dense matrix. The absence of any weathering rinds on dolerite clasts in this sediment and the lack of significant iron staining of its matrix suggests that it is watertight. This material is exposed in three sections in which the stratigraphic sequence is identical. The first of these lies at 730 m adjacent to Katrina Ck. about 1.8 km south of the Clarence weir. In this section 1.5 m of Nive Till is exposed, and no more than the upper 50 cm of the sediment shows any iron staining. This thin weathered zone is overlain in turn by a thin sheet of heavily weathered ablation till. The Nive Till is dominated by dolerite clasts but the ablation till contains occasional erratics of quartz and quartz schist. This suggests that the ablation till may not have been derived from the same source area as the basal till. The other sections occur at 725 m on either side of the Lyell Highway 1.2 km north of Laughing Jack Marsh (Figure 10.4).

The Nive Till is a tough and strongly lithified sediment. It is identical in almost all respects to the basal till that occurs as clasts in the East Rufus Moraine (chapter five), beneath weathered till in the Travellers Rest Valley (chapter six) and in landslide debris derived from the

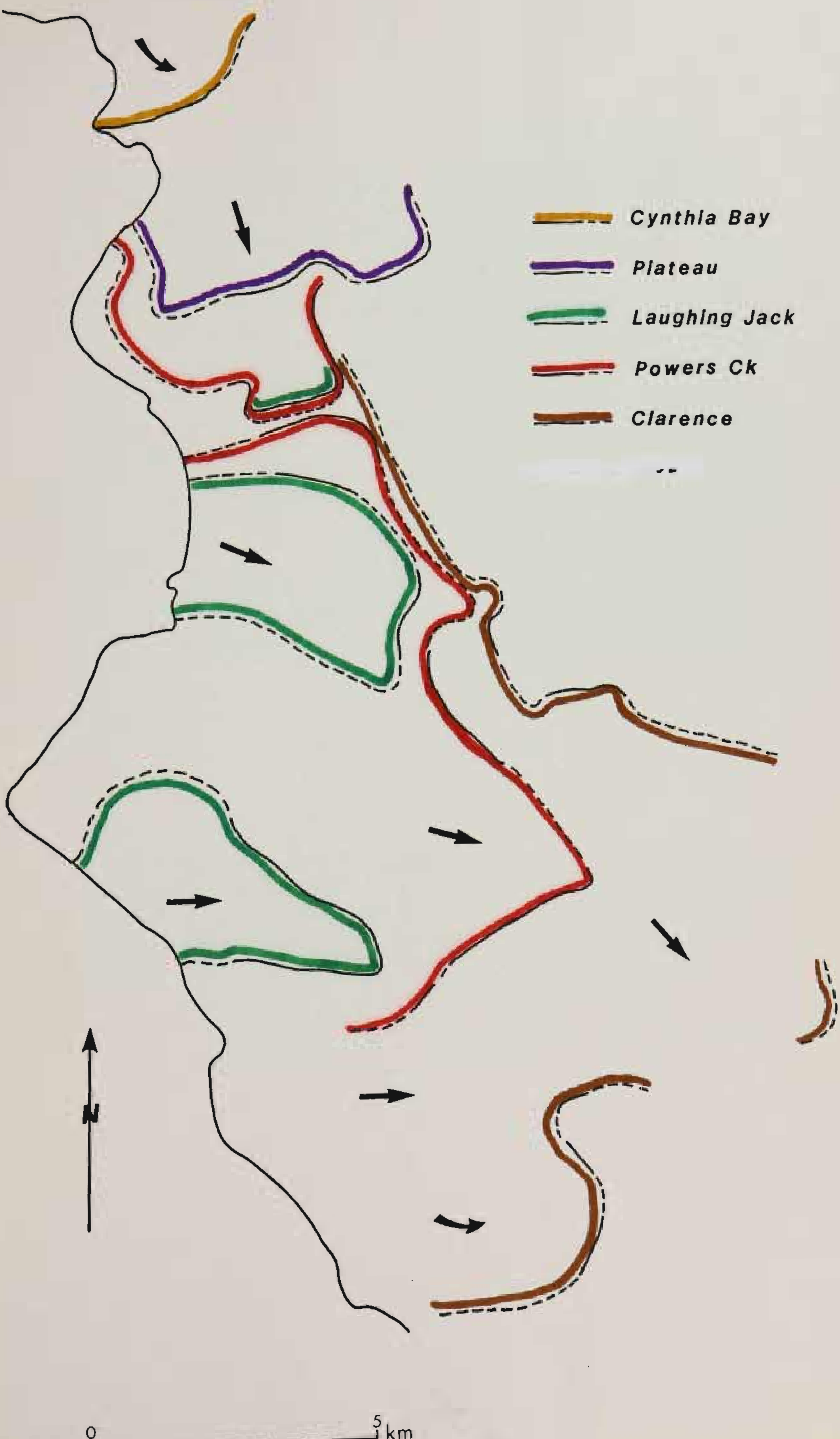
Cynthia Bay
Plateau
Laughing Jack
Powers Ck
Clarence
N
0
5 km

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Figure 10.3 Principal ice limits in the western part of the Nive Valley.

Figure 10.4 Highly lithified grey basal till (Nive Till) overlain by weathered till of the Laughing Jack phase, 1.2 km north of Laughing Jack Marsh.

Figure 10.5 Weathered till in the Laughing Jack moraines at the eastern end of Laughing Jack Lagoon.

Cascade Moraines downstream of Butlers Gorge (chapter nine).

The total domination of the Nive Till in this area by dolerite clasts (Table 10.1) indicates that it probably originated on the Central Plateau. The discovery of fragments of this till among severely shattered boulders at 960 m. close to the edge of the scarp at Clarence Lagoon is consistent with this interpretation. As ice from the Derwent Glacier appears to have subsequently passed across all three sites where the Nive Till occurs in situ it seems likely that the Nive Till predates the full development of the Clarence and Laughing Jack cols. The extent of the ice in the Clarence Valley during the glaciation responsible for its deposition cannot be ascertained from the limited exposures that are available. The fact that the three sites that have been found all lie reasonably close to the easternmost ice limit in the area probably reflects no more than that the subsequent till covers are thinner in such a location. However, no Nive Till has been found to date beyond the later ice limits.

(ii) Clarence Moraines

The most easterly of the glacial deposits are scattered erratics up to 2 m in size that protrude through glaciofluvial deposits on the Nive Plains. Deeply weathered till occurs at 770 m on the scarp slopes SSE of Dyes Marsh. This till is far more deeply weathered than any of the glacial deposits that have been located on the St. Clair Surface within the Derwent River Valley. This suggests that

Vicinity No.	13	14	17	21	22	23	25	32	33
%									
dolerite	64.0	79.5	75.0	93.0	95.0	94.0	100.0	100.0	100.0
quartz	26.0	-	12.5	7.0	4.0	6.0	-	-	-
argillites	-	12.8	12.5	-	1.0	-	-	-	-
basalt	-	-	-	-	-	-	-	-	-

Table 10.1 Lithology of glacial deposits in the western Nive Valley. Site locations are indicated on Figure 10.1.

it is considerably older.

Weathered red basal till that has a blocky sub-angular pedal structure is exposed in a shallow section on the west bank of the southernmost outlet from Dyes Marsh. It is overlain in this section by loosely textured ablation till in which dolerite clasts have weathering rinds up to 23.0 mm thick. The presence of Permo-Triassic clasts suggests that this till was deposited by ice that over-ran the rock bar to the west. The presence of weathered till at 800 m. adjacent to Dunnys Ck. on the south side of the Clarence Valley suggests that the Clarence Glacier may have been 100 m thick between these two points.

The easternmost limit reached by ice from the Derwent remains uncertain as no definite end moraine appears to have been deposited and the area has been inundated by glaciofluvial deposits. Till indicated on the Glacial Map of Tasmania (Derbyshire et al., 1965) as occurring on the south bank of the Clarence River 2 km. from the Nive has been reinterpreted on figure 10.2 as a periglacial slope deposit. However, boulders of up to 3 m in the bed and banks of the Clarence a further 1 km east of this point hint that the Derwent Glacier may have almost reached the Nive River.

As ice from the Derwent was sufficiently thick to over-run the rock bar at 800 m south of Dyes Marsh the east-west gradient of the ice surface would not need to have been very great for it to have reached almost to the top of the scarp further west, to the south of Clarence Lagoon. The erosional morphology of the scarp here hints that it has been almost buried by ice (chapter six). Large dolerite boulders occur in many localities along its face. Often these were rounded but have been fragmented by postdepositional weathering. These boulders extend to at least 960 m on the crest of the scarp south of Clarence Lagoon and fragments of red basal till have been found among the rock debris at 900 m in this area.

Deeply weathered till also occurs at 800 m further south on the Wentworth Hills between Dunnys Ck. and Wentworth Ck. A further 2.5 km to the southwest weathering rinds are exfoliating in sheets 4 - 5 mm. thick from dolerite boulders that have been bull-dozed during the construction of a logging track across a degraded moraine. These more southerly deposits are dominated by dolerite clasts which suggests that they were deposited by glaciers that descended from the Wentworth Hills. The presence of till further upslope on the flanks of this range confirms that glaciers developed here. They descended to within 5 km of the Nive River.

A brief reconnaissance along the main branch of the Nive River revealed a major accumulation of large rounded boulders up to 1 m. in diameter adjacent to the gorge that has been dammed to form Pine Tier Lagoon. This boulder deposit is 4 m thick and is best exposed on the western side of the river. It is interpreted as glaciofluvial sediment that was discharged through the gorge from ice that was trapped upstream of the rock bar at this point. The deposit becomes coarser downwards, sand comprising only about 5 % of the sediment at the base but about 30 % towards the top. Permian argillites that occur in the deposit are weathered. This site lies 10 km north-east of the Clarence Moraines south of Dyes Marsh. A reconnaissance investigation at several points along the Nive between Pine Tier Lagoon and the mouth of the Clarence River has failed to locate any similar deposits suggestive of ice south of this point.

(iii) Powers Creek Moraines

The rock spur that was over-ridden by the ice that deposited the Clarence Moraines south of Dyes Marsh extends southwards to within 1 km of the Clarence River immediately north of its confluence with Powers Creek. The Powers Creek Moraines were constructed against this spur at a time when the distributary lobe of the Derwent Glacier was lodged behind it. The southern margin of the ice would have lain against a rounded rocky knoll 3 km east of Laughing Jack Lagoon. Weathered ground moraine at 800 m elevation 1 km southeast of Laughing Jack Lagoon may represent the same glacial event.

No appreciable end moraine exists but the contrast between the very patchy distribution of till east of Powers Creek and the virtually continuous till cover that exists west of this point suggests that an important ice limit existed here. Soliflucted till that is plastered against the dolerite hills and lag boulders that occur on the plains indicate that ice from the Derwent still extended 10 km. into the Nive catchment at this time. An outwash plain extends downstream from this ice limit. The till is similar in composition to that which forms the Clarence Moraines. The basal facies are red in colour and have a blocky subangular pedal structure. The ablation till is loosely structured and deeeply weathered.

(iv) Laughing Jack Moraines

The Laughing Jack Moraines document a phase during which the ice that passed through the Clarence and Laughing Jack cols was no longer confluent. The Clarence lobe probably reached the northern end of Laughing Jack Marsh, which is an outwash plain. Degraded lateral moraines extend towards the Clarence River which has washed out the terminal segment. The Laughing Jack lobe was locked behind a rock bar at the downstream end of Laughing Jack Lagoon (Figure 10.5). It constructed a low end moraine in a re-entrant on the southern side of the bar, and discharged meltwater through the Powers Ck. Gorge to its north. The extent of the outwash plain at Laughing Jack Marsh suggests that this was an important ice limit.

A section exposed in a gravel pit at 725 m. 1.2 km. north of Laughing Jack Marsh reveals over a metre of weathered till, the upper part of which has locally been disturbed by secondary movement (Figure 10.4). Dolerite clasts in the ablation till have weathering rinds up to 17.0 mm. thick. The till matrix is bright reddish brown in colour (5YR5/8). Weathering rinds on dolerite clasts in the till at Laughing

Jack Lagoon are equally thick, but the colour of the matrix is rather lighter (7.5YR4/6) due to a greater admixture of local Triassic bedrock. Tibbs Plain is composed of outwash gravels and sands that were discharged southwards through the Broken Leg Col.

The bouldery terminal moraine that bounds the southern shoreline of Clarence Lagoon may date from around this time. This moraine is about 15 m high. The lagoon appears to be totally depositional in origin (Jennings and Ahmad 1957). Initiation of the moraine barrage that impounds Clarence Lagoon probably resulted from stemming of the southward flow of plateau ice by the Derwent Glacier and its lateral moraines which reached this level during the Clarence phase. A fossil shoreline occurs 1 m above the present water level at the marshy northern end of the lagoon and is the result of erosional deepening of the lagoon outlet.

The erosional morphology at the head of the Clarence River is not strongly developed and the thin ice body which existed here subsequent to the Clarence phase is likely to have had little capacity for erosion. The surrounding shores have been degraded by periglacial slope processes and the lagoon now appears to be quite shallow.

(v) Plateau Moraines

A series of small moraines occurs 1 km north of Clarence Lagoon. These moraines are constructed amidst bedrock ridges. They can be traced towards some small lakes that have been almost completely filled by sediment 1.8 km northeast of Clarence Lagoon.

The ice would have remained confluent with the Travellers Rest Glacier southwest of Lake Ina when these moraines were constructed. However, once the ice had retreated upvalley of the rock ridge to the north meltwater would have been discharged southwestwards into the Travellers Rest River and northeastwards into the Little Nive River. The survival of Clarence Lagoon is probably attributable to the fact that all the proglacial sediment was deflected into these alternative meltwater routes after this time.

Lightly weathered outwash gravels that are overlain by fresh slope deposits on the main branch of the Nive River downstream of the Little River are unlikely to be older than the plateau phase.

(vii) Cynthia Bay Moraines

A further ice limit is defined by moraines that were constructed against a rock ridge close to the southern shoreline of Lake Ina. The till can also be traced westwards to the Cynthia Bay limit in the upper Travellers Rest Valley. The fresh form of the moraines and the negligible weathering of the till suggests that they are equivalent in age to the Cynthia Bay Moraines at Lake St. Clair.

These moraines mark the northeastern margin of the zone of predominant glacial erosion that was identified on the Central Plateau by Jennings and Ahmad (1957). While Lake

Figure 10.6 Frost shattered rubbles overlie weathered till on the northern side of Laughing Jack Lagoon.

Figure 10.7 Gilgai on the shoreline of Laughing Jack Lagoon.

Ina is impounded behind moraines, steep rocky shorelines towards its northern end suggest that an erosional basin could be present.

A few small moraine ridges appear to be submerged on the floor of the lake and an ice-pushed rampart is present at its far northern end. Further north lies a myriad of small rock basins and ice abraded knolls, with only a few very small moraines in evidence. However, the glacial geomorphology of the plateau north and east of Lake Ina has not been investigated in detail during the preparation of this thesis.

B. Non-glacial deposits

(i) Slope deposits

Solifluction mantles that contain angular dolerite clasts are widespread throughout the area. They are no more than 30 cm thick within the Cynthia Bay ice limits on the Central Plateau, but beyond those limits commonly attain a thickness of 2 m or more. Remobilised till and clay-rich mantles extend into Clarence Lagoon from the adjacent slopes. They reach 3m in thickness southeast of Dyes Marsh where they overlie till of the Clarence phase. Slope mantles of this sort overlap all the glacial deposits on the St. Clair Surface. Similar deposits are abundant in the vicinity of the Pine-Nive confluence and they overlie the ice-proximal gravels at the mouth of the Pine Tier Lagoon Gorge.

On the Wentworth Hills the solifluction mantles are up to 2

m. thick and indicate that glaciers did not develop there during the Cynthia Bay phase. The very shallow tarn near the crest of the Wentworth Hills has been impounded behind slope deposits that have spread onto the valley floor from the southwest. The slope deposits may be reworked till. The absence of any identifiable moraine morphology on the top of the range suggests that the glaciation was ancient.

Accumulations of angular dolerite rubbles that are probably the result of frost shattering are locally abundant. Highly angular rubbles of this nature occur along the northeastern shoreline of Laughing Jack Lagoon for a kilometre beyond its outlet (Figure 10.6). Similar rubbles occur at the foot of free faces on the edge of the plateau escarpment, beneath clifflines on Mt. Charles and on the Wentworth Hills.

(ii) Alluvial, peat and archaeological deposits.

At the western end of Laughing Jack Lagoon up to 1 m of silt overlies weathered till and is in turn overlain by up to 50 cm of organic silts. A number of devegetated mounds up to 30 cm high and 2 m wide occur on the north-western shoreline of the lagoon (Figure 10.7). These exhibit cracked clay summits and are interpreted as gilgai (Leeper, 1964). The bed of the lagoon at this point is frequently exposed during drawdown of the storage. They are interpreted as gilgai (Figure 10.6).

Up to 1 m of dark brown alluvium that has a columnar structure overlies glaciofluvial gravels at 790 m on the Nive River 2 km downstream from the Little River. This alluvium overlies lightly weathered slope deposits a short distance to the south-east.

Organic-rich silts have accumulated to a depth of 1.25 m on the northern margin of Darcys Tarn, where they overlie minerogenic sands. Sandy silts have been laid down as overbank deposits by many streams in the area.

Derbyshire (1967) reported the presence of 17 ft (5.2 m) of peat in Dyes Marsh. However, systematic coring of the marsh by E.A. Colhoun has revealed a maximum sediment depth of 4.5 m, with organic deposits occurring to a depth of only 1.3 m. This suggests that the sediment depth in Derbyshire (1967) is misprinted and should read 1.7 ft. or 1 ft. 7 ins. (E.A.Colhoun, pers. comm.). Thick <u>Sphagnum</u> occurs a few hundred metres east of Darcys Tarn in the Wentworth Hills but the depth and age of this deposit is unknown. Fibrous peats that have accumulated on the buttongrass plains seldom exceed 1 m. in thickness.

Numerous scatters of prehistoric stone artefacts together with isolated flakes and possible manuports (figure 10.8) have been exposed by wave action upon the peats around the shores of Laughing Jack Lagoon. The majority of these artefacts have been manufactured from calc-silicate hornfels while others are of quartz. They are concentrated around till exposures which suggests that the raw material rocks were obtained from the glacial deposits.

Cosgrove (1984) has recorded other sites of unknown age

Figure 10.8 A manuport at Laughing Jack Lagoon. Numerous scatters of prehistoric aboriginal stone tools are exposed around the shores of the lagoon.

along the Little River and adjacent reaches of the upper Nive. It is to be anticipated that favourable lithologies have probably been concentrated at local campsites by the aborigines. Larger scale redistribution of rock materials by humans is associated with recent road-building and hydro-electric construction activities.

No artefacts have been found at depth in the silts that overlie the till despite the accessibility of numerous sections that have been cut by small streams, but hundreds have been found on the surface of the silt.

DATING AND DISCUSSION

The glacial deposits in the western part of the Nive catchment have been superimposed upon a more ancient landscape that is likely to have been subject to only very slow rates of erosional modification over the past 21.8 ma.

Driftwood from the base of the organic silt at the eastern end of Laughing Jack Lagoon has been radiocarbon assayed at 810 +- 60 BP (SUA 1957). Driftwood from the upper 20 cm of the underlying minerogenic silt has given a radiocarbon result of 1540 +- 60 BP (SUA 1958). Deposition of the organic-rich silt clearly occurred during the late Holocene. Hundreds of stone tools have been found on the surface of the organic silt but none have been found within it. This indicates that the majority of the artefacts postdate 810 +- 60 BP (SUA 1957).

Derbyshire (1967) obtained a radiocarbon assay of 4930 +-

180 BP (Gak 784) on humified peat from beneath Sphagnum in Dyes Marsh. This indicates that peat has been forming for about 5 ka.

Solifluction deposits are widespread outside the ice limits of the Cynthia Bay phase but are absent inside those limits. This indicates that the solifluction occurred prior to the retreat of the Central Plateau Ice Cap from the ice limit at Lake Ina. Solifluction deposits overlie the Plateau moraines and also the earlier moraines near Clarence Lagoon. These latter moraines may have been deposited during the Laughing Jack phase, but ice would have been able to persist at high altitudes on the Central Plateau long after the distributary lobes of the Derwent Glacier had withdrawn westwards across the St. Clair Surface.

The Clarence and Laughing Jack lobes were confluent during the Powers Creek phase. The glacial deposits of the Powers Creek phase lie well inside the ice limits of the Clarence phase and therefore must be younger. The nearly continuous cover of glacial deposits inside the ice limits of the Powers Creek phase contrasts with the patchy distribution of the Clarence till. This may be due to the Clarence phase having been shortlived, but may also be the result of the Clarence till having been more extensively reworked by periglacial and other processes than have the deposits of the Powers Creek phase. This might imply that the Clarence till is significantly older than the Powers Creek till. During Clarence time the Derwent Glacier was probably over 160 m. thick south of Clarence Lagoon. Glaciers also arose at this time on the eastern slopes of the Wentworth Hills.

Glacial erosion and the burial of glacial deposits by outwash gravels and periglacial slope mantles has prevented identification of the easternmost ice limits in this part of the Nive River catchment. Isolated exposures on the floor of the Clarence Valley indicate that the Derwent Glacier extended eastwards to within 5 km of the Nive River, while lateral moraine remnants suggest that this ice may have been 100 m thick only 7 km from the Nive.

Postdepositional characteristics of the deposits in this area are presented in Table 10.2.

The shallow soil profiles that have developed on the moraines of the Cynthia Bay phase at Lake Ina are of A Cox Cu type. Reconnaissance excavations to less than 50 cm depth in the moraines on the plateau south of Lake Ina revealed that a textural B horizon was present. The Bt soil horizon in the moraines at the eastern end of Laughing Jack Lagoon is at least 2 m thick. The maximum depth of the soil profile in the moraines further to the east has not been ascertained. However, solifluction deposits within the ice limits of the Clarence phase north of the Clarence River have a Bt horizon at least 3 m thick.

Weathering rinds on dolerite clasts also reveal systematic variation. Weathering rinds in the moraines at Lake Ina reach a maximum thickness of 2.1 mm, rinds in the Laughing Jack Moraines reach 17.5 mm and those in the Clarence

moraines[a]	N	–	C	C	C	C	C	C	C	C	C	–	PC	PC	PC	PC	PC	PC	PC
vicinity no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19
material[b]	T	B	T	T	T	T	T	T	T	T	ST	B	T	T	T	T	T	T	T
A. MORPHOSTRATIGRAPHY																			
relative position	underlies 3	inside CL?	overlies 1	outside 5	outside PC	inside 4	outside PC	outside PC	outside PC	outside PC	outside PC	inside C	inside C	= 13	inside C	inside C	inside 13	inside 16	inside 17
crest/site elevation(m)	720	720	720	730	720	730	800	800	780	780	720	715	765	760	720	720	780	770	780
max.depth of local burial(by)(m)[c]	T1.5,S1	–	B1.3	B2	B1.1	B1.5	B2.1	B2	S1	S2	–	–	B1.5	B1.2	S30	B1.5	B1.3	S1	–
B. SUBSURFACE CLASTS																			
dolerite rinde(mm)-max.	–	–	–	23.1	14.6	–	37.0	23.0	12.3	12.8	19.6	8.0	10.2	8.0	3.4	17.5	9.0	10.1	15.0
-min.	–	–	–	8.6	6.8	–	13.0	9.1	6.0	4.0	9.4	4.0	1.2	0.1	1.0	1.0	3.0	3.1	4.0
-mean	–	–	–	12.7	9.8	–	23.9	15.4	9.0	7.9	14.2	5.9	5.1	1.6	2.1	8.3	4.7	5.0	8.0
- SD	–	–	–	3.8	1.9	–	5.3	3.6	1.8	2.3	3.6	0.7	2.0	1.2	0.8	5.6	1.8	1.8	3.2
- hardness (1-5)	1	–	2-3	3-4	3-4	–	5	5	4	3-4	4	4	3-4	3	3-4	4-5	3-4	3	3
clast surface colours	2.5Y	–	5YR	5YR	5YR	5YR	5YR	7.5YR	5YR	7.5-5YR	5YR	5YR	5YR	5YR	5YR	5YR	7.5-5YR	7.5YR	7.5-5YR
%unrecognizably weathered clasts	0	–	–	10	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–
coherence of argillite clasts	1	–	–	4-5	4-5	5	–	–	–	–	–	–	5	4	–	–	4-5	4-5	–
hornfels rinde(mm)-max.	–	–	–	–	–	–	–	–	–	–	–	–	–	5.2	–	–	–	–	–
-min.	–	–	–	–	–	–	–	–	–	–	–	–	–	0.9	–	–	–	–	–
-mean	–	–	–	–	–	–	–	–	–	–	–	–	–	3.4	–	–	–	–	–
- SD	–	–	–	–	–	–	–	–	–	–	–	–	–	1.9	–	–	–	–	–
C. SUBSURFACE MATRIX																			
profile type	A Cox cu	A Bt	–	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt	A Bt	ABt	A BT	A Bt	A Bt	A Bt
colour of B horizon	–	–	5YR	5YR	5YR	5YR	5YR	5YR	5YR	5YR	5YR	7.5-5YR	7.5YR	7.5YR	5YR	5YR	7.5YR	7.5YR	7.5YR
depth of B horizon(cm)	–	> 300	> 1.5	> 1	> 1.5	> 2	> 150	> 200	> 100	> 100	> 200	> 100	> 300	> 100	> 100	> 200	> 100	> 200	> 100
depth to lowest Cox(cm)	20	> 300	> 1.5	> 1	> 1.5	> 2	> 150	> 200	> 100	> 100	> 200	> 100	> 300	> 100	> 100	> 200	> 100	> 200	> 100
pH at ~100cm	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–
% organic at ~100cm	–	–	–	–	–	5.6	–	–	–	–	–	–	–	–	–	5.2/5.6	–	–	–
clay size fraction mineralogy	–	–	chlorite quartz	–	–	–	–	–	–	–	–	–	–	–	–	Gibbsite[d] quartz	–	–	–
clay film development(1-5)	1	4	4	4-5	4-5	5	5	4	3-4	3	3-4	–	3	4	5	5	4	3-4	4
grain size < cobbles:																			
% > 1 cm	–	–	–	–	–	32.4	–	–	–	–	36.0	–	–	–	–	33.8/11.6	–	–	–
% 1cm-0.5cm	–	–	–	–	–	3.9	–	–	–	–	11.0	–	–	–	–	13.5/8.1	–	–	–
% < 0.5 cm	–	–	–	–	–	63.7	–	–	–	–	53.0	–	–	–	–	32.7/70.3	–	–	–
grain size < gravel:																			
% sand	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–
% silt	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–
% clay	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–
degree of basal till lithification(1-5)	5	–	3	3	3	–	–	–	–	–	–	–	–	–	3	3	–	–	–
clast socket staining	1	4-5	4-5	4-5	4	5	5	5	5	4	3	–	–	3	5	5	–	3-4	–
D. SURFACE MATERIALS																			
no.of surface boulders ($/m^2$)	–	–	7	–	–	4	–	–	–	–	–	–	–	–	–	–	–	–	11
% split clasts	–	–	50	–	–	36	–	–	–	–	–	–	–	24/40	–	–	–	–	3
clast surface colours	–	5YR	5YR	5YR	5YR	5YR	5YR	5YR	5YR	5YR	5YR	5YR	7.5YR	7.5-5YR	–	–	–	–	–
depth of clast pitting	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–	–
E. DEPOSITIONAL LANDFORMS																			
local moraine relief(m)	–	–	<1	–	<2	–	–	3	–	–	–	–	–	–	2	–	–	–	8
width of moraine crest(m)	–	–	<3	–	15	–	–	10	–	–	–	–	–	–	<3	–	–	–	5
proximal slope angle(°)	–	–	<3	–	<5	–	–	<10	–	–	–	–	–	–	–	–	–	–	–
distal slope angle(°)	–	–	<3	–	<5	–	–	15	–	–	–	–	–	–	–	–	–	–	10-12
fluvial dissection(1-5)	–	–	4	–	3?	3	–	4?	4	4?	–	–	3-4	–	–	–	–	–	2

(a) N= Nive; C= Clarence; PC= Powers Creek; PL= Plateau; CB= Cynthia Bay.

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

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

(d) after Derbyshire (1967).

Table 10.2 Post-depositional modification of deposits in the western Nive Valley. Site locations are indicated on Figure 10.1.

moraine[a] / vicinity no. / material[b]	PC 20 T	PC 21 T	PC 22 T	PC 23 T	PL? 24 T	PL 25 ST	PL? 26 O	CB 27 S	CB 28 S	CB 29 S	CB 30 S	CB 31 S	CB 32 T	CB 33 F
A. MORPHOSTRATIGRAPHY														
relative position	inside	outside	outside	outside	inside	inside	inside	outside	outside	inside	outside	inside	inside	outside
	18	25	25	25	20	21	24	all	all	30	all	7	25	30
crest/site elevation(m)	710	900	880	880	720	980	760	860	790	780	680	1080	1020	580
max.depth of local burial(by)(m)[c]	B1.3	S2?	S2?	S	-	-	S1.2	-	-	-	A0.4	-	-	A0.3
B. SUBSURFACE CLASTS														
dolerite rinds(mm)-max.	11.0	-	-	-	6.8	2.1	2.5	2.2	1.4	1.4	1.6	2.6	2.1	0.2
-min.	2.2	-	-	-	2.8	0.8	0.1	0.2	0.1	0.3	0.1	0.5	0.2	0.1
-mean	7.0	-	-	-	3.8	1.0	1.2	1.8	0.4	0.9	0.6	1.2	1.3	0.1
- SD	2.3	-	-	-	1.5	0.4	1.0	0.8	0.3	0.3	0.3	0.4	0.4	0.02
- hardness (1-5)	-	2	3	2	2-3	1-2	2-3	1	1	1	2	1	1	1
clast surface colours	7.5YR	5YR	5YR	5YR	7.5-5YR	7.5YR	7.5YR	10YR	10YR	10YR	10YR	10-7.5YR	7.5YR	7.5YR
%unrecognizably weathered clasts	-	-	-	-	-	-	-	-	-	-	-	-	-	-
coherence of argillite clasts	-	-	-	-	-	-	-	-	-	-	-	-	-	-
hornfels rinds(mm)-max.	-	-	-	-	-	-	-	-	-	-	-	-	-	-
-min.	-	-	-	-	--	-	-	-	-	-	-	-	-	-
-mean	-	-	-	-	-	-	-	-	-	-	-	-	-	-
- SD	-	-	-	-	-	-	-	-	-	-	-	-	-	-
C. SUBSURFACE MATRIX														
profile type	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	A Cox cu	A Cox cu
colour of B horizon	7.5-5YR	7.5YR	-	-	7.5YR	7.5YR	7.5YR	10YR	10YR	7.5-10YR	10YR	5YR	10YR	7.5YR
depth of B horizon(cm)	-	-	-	-	> 100	-	> 0.8	-	-	-	-	-	-	-
depth to lowest Cox(cm)	-	-	-	-	> 100	-	> 0.8	-	-	-	-	-	-	-
pH at ~100cm	-	-	-	-	-	-	-	-	-	-	-	-	-	-
% organic at ~100cm	-	-	-	-	-	-	-	5.9	-	-	-	-	-	-
clay size fraction mineralogy	-	-	-	-	-	-	-	-	-	-	-	-	-	-
clay film development(1-5)														
grain size < cobbles:	4	-	-	-	3-4	2	-	-	-	-	-	-	1	-
% > 1 cm	-	-	-	-	-	-	-	-	-	-	-	-	-	-
% 1cm-0.5cm	-	-	-	-	-	-	-	-	-	-	-	-	-	-
% < 0.5 cm	-	-	-	-	-	-	-	-	-	-	-	-	-	-
grain size < gravel:														
% sand	-	-	-	-	-	-	46.4	-	-	13.3	-	15.1	-	-
% silt	-	-	-	-	-	-	34.2	-	-	50.0	-	64.7	-	-
% clay	-	-	-	-	-	-	19.4	-	-	36.7	-	20.2	-	-
degree of basal till lithification(1-5)	-	-	5	-	-	-	-	-	-	-	-	-	-	-
clast socket staining	-	-	4	-	4	3-4	3	-	-	-	-	-	1	-
D. SURFACE MATERIALS														
no.of surface boulders (/m²)	-	30	10	23	-	4	-	-	-	-	-	-	22	-
% split clasts	-	30.0	15	26	-	2	-	-	-	-	-	-	3.0	-
clast surface colours	-	-	-	-	7.5-5YR	7.5YR	7.5YR	-	-	-	-	5YR	7.5YR	-
depth of clast pitting	-	-	-	-	-	-	-	-	-	-	-	-	-	-
E. DEPOSITIONAL LANDFORMS														
local moraine relief(m)	-	-	-	-	-	10	-	-	-	-	-	-	8	-
width of moraine crest(m)	-	-	-	-	-	7	-	-	-	-	-	-	3	-
proximal slope angle(°)	-	-	-	-	-	15	-	-	-	-	-	-	16	-
distal slope angle(°)	-	-	-	-	-	23	-	-	-	-	-	-	11	-
fluvial dissection(1-5)	4	-	-	-	3-4	2	-	-	-	-	-	-	1-2	-

Moraines are as thick as 37.0 mm.

The moraines that impound Lake Ina are fresh in form, only slightly weathered, and are undoubtedly equivalent in age to the Cynthia Bay Moraines at Lake St. Clair. The moraines at Clarence Lagoon are much more degraded. Little moraine morphology is to be found east of Laughing Jack Lagoon.

The Nive Till is much more highly lithified than the red basal till that occurs inside the ice limits of the Clarence phase. This suggests that the Nive Till is much older.

These variations in the postdepositional modification of the deposits are consistent with the morphological relationships between the moraines and outwash terraces. The dramatic change in the character of the soil profiles downstream of Lake Ina suggests that the moraines further south on the plateau are considerably older. If a maximum dolerite weathering rind thickness of only 5.1 mm has been achieved since retreat from the ice limit at Lake Ina it is difficult to see how rinds of up to 17.5 mm at Laughing Jack Lagoon and nearly double that figure in tills at the foot of the Wentworth Hills could have developed during the same glaciation. The highly lithified condition of the Nive Till suggests that it may considerably predate all the later deposits.

chapter eleven
THE UPPER GORDON VALLEY

The King William Range has probably been more severely eroded by glaciers than any other snowfence in the study area (Figure 11.1). Ice that accumulated in the northern part of the range flowed down the Guelph Valley (chapter eight) to join the Derwent Glacier at Butlers Gorge. The low divide between the Guelph and the headwaters of the Gordon River to the south (described in this thesis as the Gordon Gap) was over-ridden at times by a distributary lobe of the Guelph Glacier. This extended southwards into the Gordon Valley where it merged with local glaciers that formed in the southern part of the King William Range.

This chapter records a reconnaissance of the glacial geomorphology of the upper Gordon Valley (Figure 11.2). The object was to confirm impressions gained in the Guelph Valley that ice must have passed through the Gordon Gap. It was hoped to establish the amount of ice involved, and to assess the relative contributions of the Guelph lobe and the local glaciers to any trunk glacier that may have formed in the Gordon. These objectives have been only partly achieved due to the dense vegetation that made access very difficult. A thorough assessment would require much more time than was available.

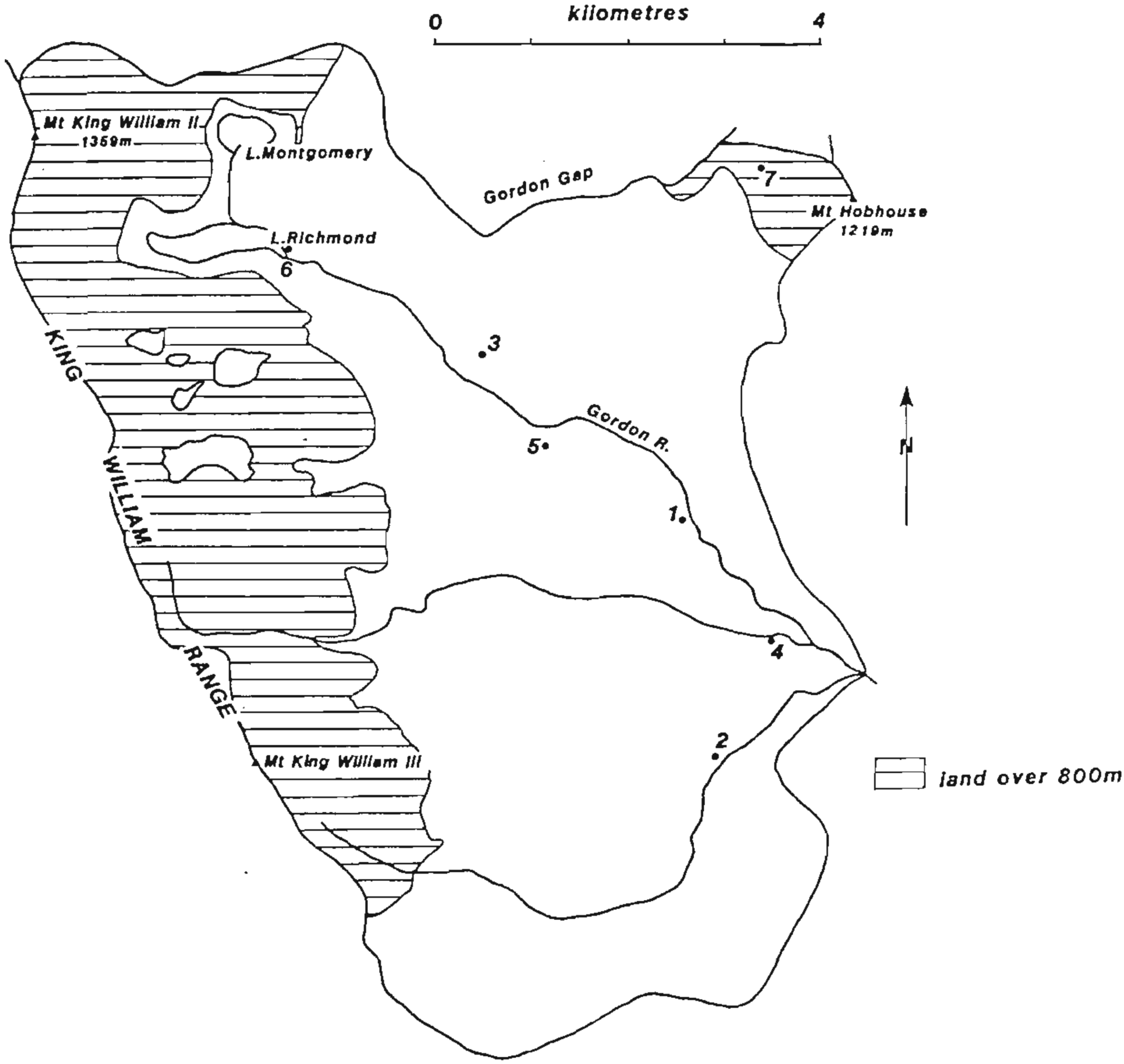


Figure 11.1 Locality map. upper Gordon River Valley. Numbers refer to site locations in tables 11.1 and 11.2.

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

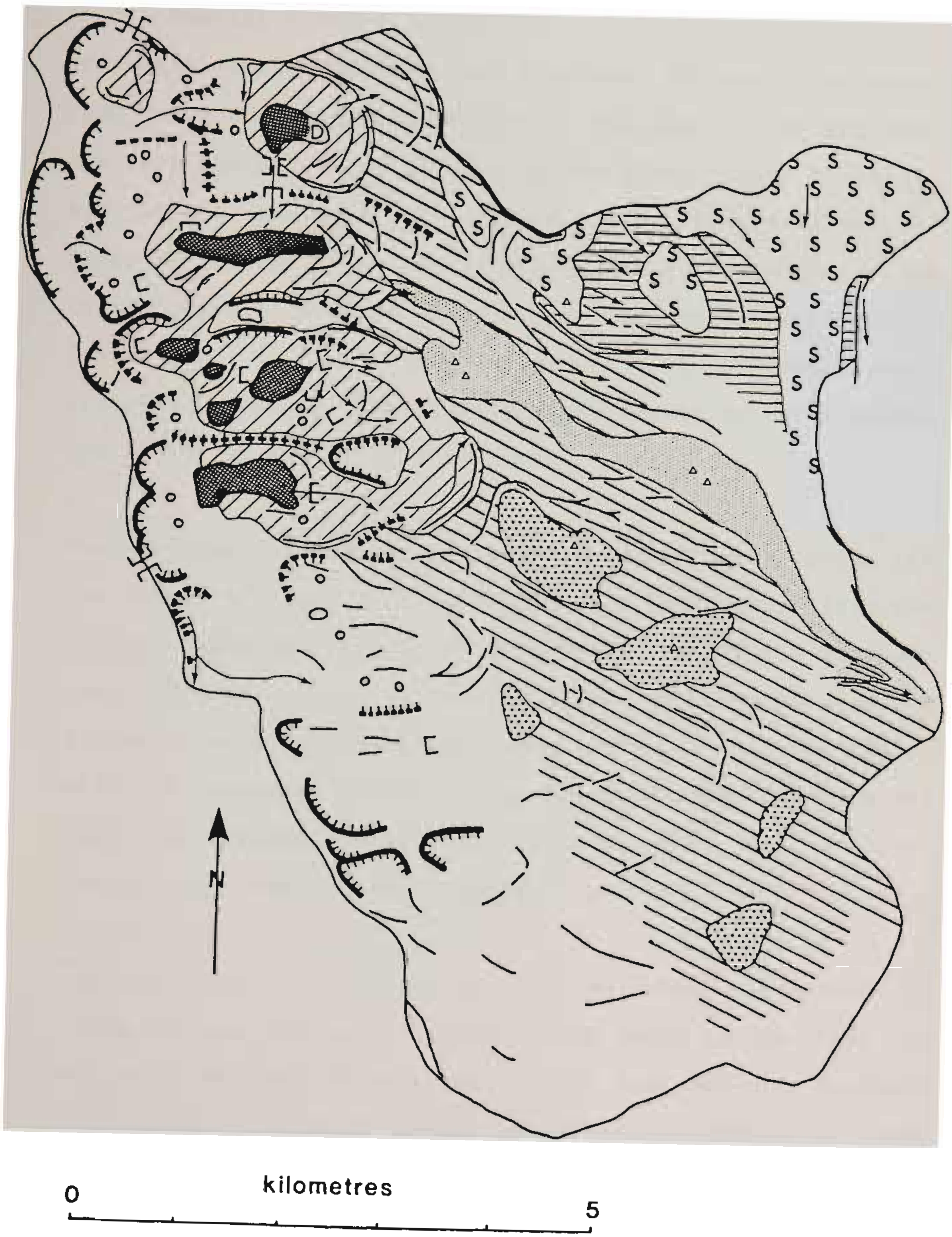


Figure 11.2 Glacial geomorphology of the upper Gordon River Valley. See Figure B2 for key.

EROSIONAL LANDFORMS

A Glacial Erosion

Broad and shallow cirques have been cut in dolerite above 1050 m on the King William Range. One such cirque has been cut into the head of a glacial trough above Lake Richmond, and others surround the summit of Mt. King William III. Large lateral moraines indicate that glaciers extended from these cirques to low levels in some cases (Derbyshire, 1968a). Ice-smoothed rock on the arms of many cirques indicates that glaciers previously extended beyond the limits of the lateral moraines.

The principal cirques in the upper Gordon Valley occur at the head of glacial troughs that have been eroded into the eastern flank of the King William Range. As in the northern part of the range the troughs are cut into Permian and Triassic rocks, and sapping at the foot of the overlying sill of columnar dolerite has produced steep and imposing headwalls (Figure. 11.3). Valley steps up to 160 m high occur within some of the troughs.

Most of these troughs are occupied by lakes, the largest of which is Lake Richmond, a narrow rock basin 1.4 km long and up to 33 m deep (Derbyshire, 1968a) that has been extended by the construction of a moraine barrage at its eastern end. The threshold of the Lake Richmond trough lies a few hundred metres west of the break of slope at the foot of the range. Half a dozen smaller lakes are developed on the range, most of these probably being rock basins. Lake Montgomery, the small moraine-bounded lake north of Lake Richmond, is 25m deep (Derbyshire, 1967) and is probably also a rock basin that has been extended by the construction of a moraine barrage (Figure 11.4).

Small diffluence cols occur between some of these cirques. Headwall crests have frequently been over-ridden. In most cases these over-ridden cirques occur adjacent to wider parts of the King William Plateau where ice abraded rock is widespread. The ice carapace that over-rode the cirques was probably thin as it did not abrade the higher rock crests (Peterson, 1968).

Little ice-abraded rock is exposed on the floor of the Gordon Valley where thick drift is generally present. Micro-erosion features are uncommon due to the susceptability of most of the rock to postglacial weathering. One set of striae trending towards 135° was found at 580 m (site 4 on Figure 11.1) adjacent to the creek that drains the small southernmost lake on the range. These striae indicate that a valley glacier moved southeastwards across this site.

B Glaciofluvial Erosion

Numerous meltwater channels that have been cut in rock link the lakes on the slopes of the King William Range, and other smaller channels occur on the broader plateaux. Most of these channels exploit joints in the dolerite.

Proglacial channels have been cut in rock beyond the threshhold of the glacial troughs. A series of steep subglacial channels plunge downslope into the Gordon Valley from the Guelph-Gordon divide. They developed at a point where ice pressure was relieved in a zone of extending flow, and appear to be the equivalent of the "chutes" of Mannerfelt (1945).

Many channels have been cut in drift at the foot of the King William Range. The majority of these are the result of proglcial meltwater erosion. They postdate the retreat of the Guelph ice and were cut at a time when local glaciers terminated not far from their cirques.

C Non-glacial Erosion.

The upper Gordon River Valley trends NW-SE from the Guelph-Gordon divide. Although it is eroded in Permo-Triassic rocks it parallels the Derwent River Valley to the east which is largely controlled by joints and faults in the dolerite (chapter nine). This suggests that the upper Gordon Valley has been superimposed from the original dolerite cover rocks that still occur on the crests of Mt. Hobhouse and the King William Range

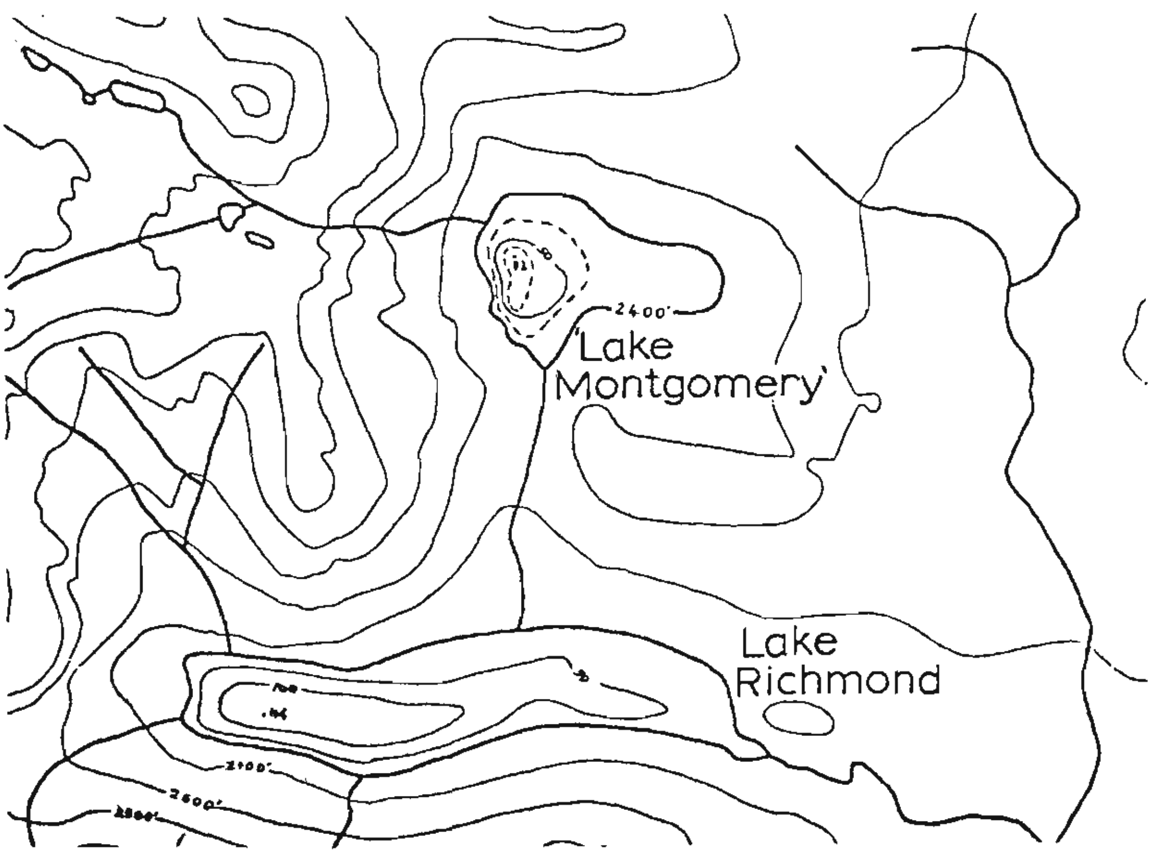
The Guelph-Gordon divide is strongly asymmetric, with a gentle gradient to the north and a steeper slope down into the Gordon. This is probably the result of the dolerite rock bar at the lower end of the Guelph Valley having impeded base level lowering by that river. No such impediment existed in the Gordon Valley where the river was able to cut more readily into the Permian rocks. The southern slopes of the divide have been further steepened by

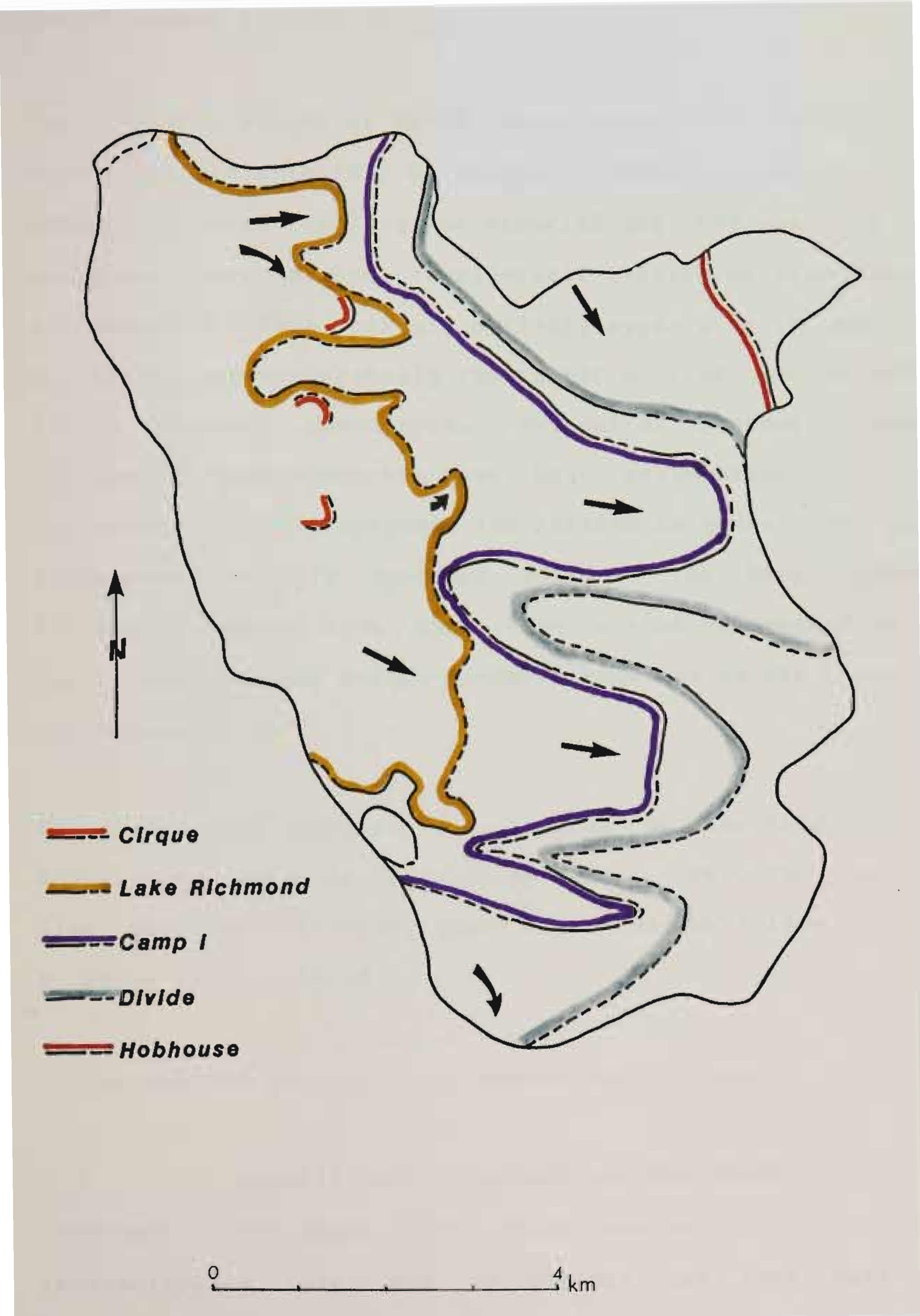
Figure 11.3 Lake Richmond, source of the Gordon River. The steep trough-end has been over-ridden by ice from the plateau south of Mt. King William II.

Figure 11.4 Bathymetric charts of Lake Richmond and Lake Montgomery. Isobaths are in feet. (after Derbyshire, 1967).

overleaf

Figure 11.5 Principal ice limits in the upper Gordon River Valley.


2400'
'Lake
Montgomery'
Lake
Richmond
2400'
2600'
2800'


N
Cirque
Lake Richmond
Camp I
Divide
Hobhouse
0
4 km

erosion on the margin of the glacier that flowed out of the Lake Richmond trough.

The western slopes of the Hobhouse Range above 800 m have been fashioned primarily by fluvial erosion. A very limited amount of slab toppling has occurred near the crest of the range but there has been considerable erosion of free faces by rockfall. This rockfall activity appears to be inactive at present and was probably the result of frost action under colder climatic conditions. Mechanical weathering under periglacial conditions has also been responsible for the disintegration of previously ice-abraded eminences that rise above more recently smoothed rock on the King William Plateau. Joints have also been etched by weathering at lower levels on the Guelph-Gordon divide and on the floor of the Gordon Valley.

Small nivation cirques are present on the King William Range. The source of Tasmania's largest river, the Gordon, lies amid the late-lying snows of a nivation hollow at 1320 m above Lake Richmond.

Little erosion appears to be occurring at present.

DEPOSITIONAL LANDFORMS AND SEDIMENTS

Fieldwork in the upper Gordon Valley has only been at the reconnaissance level and no attempt has been made to comprehensively map all the depositional landforms and sediments at the foot of the King William Range. The ice limits in this part of the upper Gordon Valley have been identified mainly by air photo interpretation with some field verification (figure 11.5).

(i) Hobhouse Moraines

The uppermost lateral moraine on the eastern side of the Gordon Valley can be traced SSE from 730 - 630 m on the southwestern slopes of Mt. Hobhouse. It marks the passage of at least 100 m of ice southwards through the Gordon Gap from the Guelph Glacier. Quartzite erratics occur up to 850 m on the eastern wall of the Gap. The moraine crest suggests that the ice surface sloped southwards at a gradient of about 70 m/km. If this moraine represents the maximum phase of glaciation in the upper Gordon, it implies that the surface of the Gordon Glacier was strongly asymmetric with the eastern margin of the ice lying at least 300 m lower than the upper limit of the ice that overflowed the principal cirques and troughs of the King William Range 7 km to the west. A gradient in this direction is consistent with probable snowfence and shading effects.

The maximum ice limits further downvalley are difficult to discern on the air photos due to dense vegetation and the possibility that some moraine-like ridges are rock, perhaps only thinly mantled with till. A pronounced ridge 3.5 km south of Mt. Hobhouse is probably a moraine and if so it suggests that at least 80 m of ice covered the valley floor at this point. This seems surprisingly little for the maximum phase of a glacier from such a major snowfence as the King William Range, particularly in view of the emerging picture of very extensive Pleistocene ice in the Tasmanian

Central Highlands. It is probable, therefore, that the maximum ice limits lay higher on the steep slopes of the Hobhouse Range.

A moraine-like ridge at 800 - 720 m below the cirque on the southern flank of Mt. King William III lies on a rock bench and forms the divide between the Gordon and Denison rivers. While it is probable that some ice and meltwater spilled into the upper Denison near this point most would have flowed into the Gordon. Fieldwork at a glacial limit downstream of this cirque has revealed till deposits that are no more than moderately weathered in comparison to the till that defines the maximum ice limit in most of the other valleys examined during this study. This compounds the impression that the Gordon Glacier must have extended further than indicated here.

Despite the fact that the maximum limit of the ice does not appear to have been located, the fieldwork has been sufficient to reveal that all the ice bodies that arose from the King William Range were at one time confluent both with one another and with the lobe of the Guelph Glacier that extended into the valley. It is also clear that the majority of the ice in the Gordon Valley was generated locally.

(ii) Divide Drift

The Guelph-Gordon divide is mantled by thin till and erratics transported by ice that spilled over it from the north. No moraine morphology remains on the steep upper slopes south of the divide. Two lateral moraines were constructed across its lower slopes on the margin of the Gordon Glacier. The uppermost of these was probably initiated in an interlobate position between the Guelph and Gordon glaciers. It parallels the Gordon River about 1 km to its north at 720 - 640 m This moraine contains a moderately high proportion of quartz schist clasts that may either have come from the Guelph Valley or been reworked from the Permian sediments at Lake Richmond.

The limited distribution of the Divide Drift suggests that it represents the final occasion on which ice spilled southwards through the Gordon Gap.

(iii) Camp I Moraines

The crest of the inner lateral moraine south of the Gordon Gap lies at 680 - 720 m (Tasmap grid ref. 8113-345 135). A probable continuation of this moraine terminates 5.5 km south-east of the Gordon Gap at about 600 m. elevation, but its crest has not yet been continuously traversed. These moraines appear to represent the next major phase after construction of the Hobhouse Moraines. Pronounced lateral moraines 3 - 5 km north-east of Mt. King William III (Figure 11.6) indicate that two separate ice lobes extended from the range at this time, while another smaller glacier also arose from the southernmost cirque on Mt. King William III.

The till that forms the lateral moraine 2.5 km downstream from Lake Richmond may be up to 20 m thick but no sections

Figure 11.6 The upper Gordon River Valley from Mt. Hobhouse. The crest of the Guelph-Gordon divide is marked by the transition from buttongrass plains to forest. Moraines exten d eastwards across the floor of the Gordon Valley from the cirques of the King William Range.

are available that permit certain confirmation that the ridge is not partly cored by rock. A few quartzite clasts in the till exhibit nail-head striae. The till becomes increasingly dominated by dolerite clasts as one progresses down the valley (Table 11.1). This results from the rapid comminution of the argillites. Localised exceptions to this general pattern are common. Comminution till (Dreimanis, 1976) is exposed at 580 m on the south bank of the creek that drains the southernmost lake. This till contains abundant angular fragments of dark siltstone amid rounded clasts of dolerite.

About 7 km south-east of Lake Richmond the moraine morphology is poorly defined, heavily dissected by meltwater and overlapped by a broad outwash terrace. The till here contains thick iron pans. The glaciofluvial sediments thinly mantle older drift and glacial boulders that protrude through the surface have been subject to considerable chemical weathering. A few lenses of rhythmically bedded silts were located in a shallow exposure at 620 m. about 4.8 km ESE of Mt. King William III.

(iv) Lake Richmond Moraines

Small end moraines occur a few hundred metres from the eastern end of Lake Richmond and some of the other small lakes that occur further south in the King William Range. The Lake Richmond moraine can be traced to a lateral moraine on the northeastern side of the lake. It is morphologically fresh, the till is little weathered and it is probably equivalent to the Cynthia Bay Moraines at Lake St. Clair.

Vicinity No.	1	2	5	6	9	10
%						
dolerite	35.0	41.0	30.0	17.0	40.0	70.0
quartz	48.0	58.0	52.0	28.0	30.0	25.0
argillites	2.0	1.0	8.0	45.0	25.0	5.0
black schist	15.0	-	10.0	-	-	-
other	-	-	-	10.0	5.0	-

Table 11.1 Lithology of glacial deposits in the upper Gordon Valley. Site locations are indicated on Figure 11.1.

A high proportion of local argillites and quartzite erratics reworked from them is evident in the till at Lake Richmond. This contrasts strongly with the more highly comminuted tills further downvalley (Table 11.2).

The Lake Richmond Moraines record a glacial event during which ice was generally confined above 680 m and seldom extended beyond the cirques and glacial troughs where it accumulated. The moraine to the east of Lake Montgomery has been constructed upon a low recessional moraine of the Guelph Glacier and is oriented at right angles to it.

Glaciofluvial sediments that can be traced back to the Lake Richmond Moraines are inset between the Camp I Moraines. They generally appear to mantle only thinly older ground moraine. Sections cut by the Gordon River 4 km southeast of Lake Richmond reveal up to 4 m of generally compact glaciofluvial gravels that contain moderately developed iron pans and are thinly veneered by peat. In some exposures the fine fraction has been washed out of the top few centimetres of the deposit. Most of the clasts are less than 20 cm in size. A few atypical clasts of up to 50 cm have probably been reworked from pre-existing till deposits. Smaller outwash plains extend downvalley from the southern cirques and infill intermorainal swales. These sediments are invariably dominated by dolerite and quartzite clasts. The end moraine and outwash terrace are indicative of a major stage limit.

(v) Cirque Moraines

Small terminal moraines form the present eastern shoreline of Lake Richmond and also bound the shores of lakes further south. Those moraines that stand upon well defined threshholds are generally well preserved. No significant outwash morphology can be traced back specifically to the cirque moraines.

This till is at least 15 m thick near the outlet to Lake Richmond. Slabs of Fern Tree Mudstone (Permian) up to 1 m long occur on the surface of the moraine. These slabs are highly angular despite the incompetence of this rock and must therefore be supraglacial in origin. This suggests considerable rockfall from the walls of the trough while the glacier was present.

The survival of the lakes without their being filled by glacial and glaciofluvial sediment suggests that final retreat of the ice occurred rapidly.

B Non-glacial Deposits

(i) Slope deposits

Solifluction deposits that contain abundant angular dolerite clasts cloak the western flank of Mt. Hobhouse but are only sparsely distributed on the King William Range. At least 4 m of poorly sorted rubble in a silty clay matrix are exposed in a creek on the western flank of Mt. Hobhouse. Lesser accumulations, that are probably no more than 2 m thick, are present on the southern slopes of the Guelph-Gordon divide. They incorporate erratics of dolerite and extend into some of the meltwater chutes.

A few large joint-bounded blocks of dolerite that occur beneath the free faces high on the King William Range are the result of slab toppling. They tend to occur above the ice limits recorded for the Lake Richmond phase.

Some large accumulations of rockfall talus occur in joint-controlled gullies between mechanically shattered butresses on Mt. Hobhouse. Deposits of angular talus beneath the cirque headwalls on the King William Range are generally confined to the foot of rock chutes. They probably represent avalanche and rockfall deposits.

(ii) Alluvial and peat deposits

Alluvial sediments are confined to overbank silts up to 50 cm thick that have been deposited by the principal streams, and to organic-rich silts that have accumulated locally in ponds and swales. Fibrous peats up to 1 m thick have accumulated beneath the buttongrass plains.

3 DATING AND DISCUSSION

Accumulations of peat and deposits of alluvial silt occur within the most recent ice limits at Lake Richmond and must therefore postdate deglaciation.

The cirque moraines represent a retreat from the Lake Richmond limit and document considerable rockfall activity, in contrast to the general stability of the free faces today. Solifluction deposits are virtually absent within the ice limits of the Lake Richmond phase but are widespread outside those limits. This indicates that the main phase of solifluction predates deglaciation. The end moraine and outwash terrace that were deposited during the Lake Richmond phase suggests that it represents the ice limit during a major stage.

The Camp I Moraines enclose the outwash plain that was deposited during the Lake Richmond phase. The broad outwash terrace 7 km southeast of Lake Richmond records another important ice limit. During this phase no ice from the Guelph Valley passed through the Gordon Gap. The glaciers that arose from the King William Range within the upper Gordon Valley occurred as two main lobes and one smaller glacier.

The Divide Drift and Hobhouse Moraines record earlier phases during which the ice cover was more extensive. The lateral moraine on the western slopes of Mt. Hobhouse was constructed by a 100 m thick distributary lobe of the Guelph Glacier that extended southwards through the Gordon Gap. In the Gordon Valley the ice from the Guelph was confluent with local ice.

Smoothing of the arms of the principal cirques and troughs of the King William Range by glacial erosion indicates that during the time of greatest ice cover the terrain between the cirques was over-ridden. Ice accumulated to depths of up to 300 m on the eastern side of the King William Range.

moraines[a]	D?	D?	D	CI	CI	LR	-
vicinity no.	1	2	3	4	5	6	7
material[b]	O	O	T	T	O	T	S
A. MORPHOSTRATIGRAPHY							
relative position	outside 2	inside 1	outside 6	inside 2	outside 6	inside 3&5	outside 3
crest/site elevation	600	640	740	640	630	710	850
max.depth of burial(by)(m)[c]	-	-	52	-	-	-	-
B. SUBSURFACE CLASTS							
dolerite rinds (mm)-max.	2.1	1.9	6.1	3.1	-	1.4	2.3
min.	0.9	0.8	2.9	1.6	-	0.6	0.7
mean	1.4	1.5	3.9	2.3	-	1.2	1.3
SD	0.7	0.8	1.5	1.3	-	0.9	0.6
hardness (1-5)	1-2	1-2	2-4	1-2	1	1	1-2
clast surface colour	7.5YR	7.5YR	5YR	5YR	5YR	7.5YR	7.5YR
% unrecognizably weathered clasts	-	-	20	7	-	0	-
C. SUBSURFACE MATRIX							
profile type	A Cox cu	A Cox cu	A Bt	A Bt	A Cox cu	A Cox cu	A Cox cu
colour of B horizon	7.5YR	7.5YR	7.5YR-5YR	7.5YR	7.5YR	10YR-7.5YR	7.5-5YR
depth of B horizon (cm)	-	-	> 200	> 100	-	-	-
depth to lowest Cox(cm)	> 100	> 100	> 200	> 100	> 70	50	> 70
development of clay films (1-5)	-	-	3-4	3-4	-	1	-
clast socket staining	-	1-2	3-4	3-4	-	1	-
D. SURFACE MATERIALS							
% split clasts	-	-	20	16	0	2	-
E. DEPOSITIONAL LANDFORMS							
moraine crest width (m)	-	-	-	12	-	6	-
moraine slopes (%)							
- proximal	-	-	-	< 5	-	15	-
- distal	-	-	-	< 5	-	10	-
degree of fluvial dissection(1-5)	-	-	-	4	-	1	-

(a) D= Divide; CI= Camp I; LR= Lake Richmond.

(b) T= till; O= outwash gravels.

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

Table 11.2 Post-depositional modification of deposits in the upper Gordon River Valley. Site locations are indicated on Figure 11.1.

The available data concerning postdepositional modification of the landforms and deposits is presented in table 11.2.

The A Cox Cu type soil profiles that have developed on the moraines of the Lake Richmond phase contrast with deeper soil profiles on the older moraines that have B horizons at least 1 m. thick.

Weathering rinds on dolerite clasts in the Lake Richmond Moraines are no more than 1.4 mm thick. This is less than half the thickness of those in the Camp I Moraines.

The morphology of the Lake Richmond moraines is well preserved, but the Camp I moraines are degraded and partly buried by outwash gravels and sands.

These characteristics suggest that the Camp I Moraines are significantly older than the Lake Richmond moraines. Insufficient postdepositional data has been obtained to permit adequate comparison of the Camp I moraines with the earlier moraines that occur higher on the walls of the valley.

More work is required in the Gordon River Valley. The fairly limited weathering of the glacial deposits 11 km downstream from the valley head contrasts with the heavy weathering characteristic of the ice limit in most of the other valleys studied. The King William Range is a major snowfence and it is probable that a significantly larger glacier than that envisaged here previously existed in the Gordon Valley.

chapter twelve
THE FRANKLIN VALLEY

" The valley of Lake Dixon is, par excellence, the ideal of a perfect glacial valley. No-one, however ignorant of glacial action, could in this neighbourhood gaze upon those beautiful scooped, or rather abraded, lakes or tarns..(many with islets).., the snow-white polished, billowy, and cascade-like roches moutonées...together with the tumbled moraines...without being impressed that its singularly characteristic features must have been produced by the slow rasping flow of an ancient river of ice."

– Johnston (1894a)

Study of the glacial landforms and deposits in the Cuvier Valley indicated that diffluent ice spilled into the head of the Franklin Valley and into the head of its tributary the Alma (chapter 5). Further south the Derwent and Franklin glaciers became confluent in the present Navarre basin, east of King William Saddle (chapter 7) where some of the end moraine belts can be traced westwards into the upper Franklin Valley. Further south again, ice from the Surprise Valley flowed eastwards through the Top End Gap to join the Guelph Glacier, a tributary of the Derwent (chapters eight and nine)(Figure 12.1).

This close relationship between the Derwent and Franklin glaciers coupled with the difficulties imposed by the flooding of Lake King William in a part of the Derwent

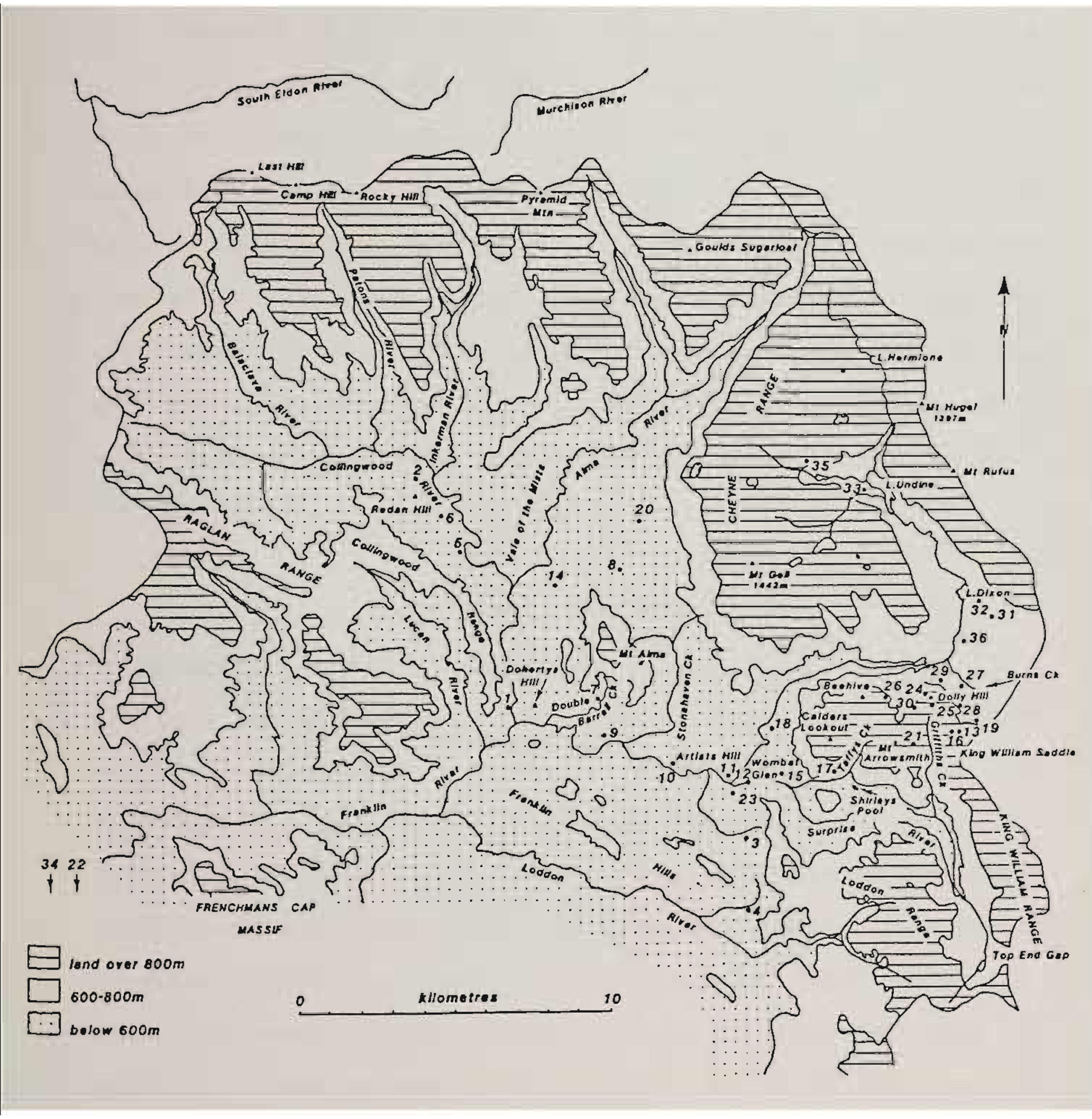


Figure 12.1 Locality map, Franklin River Valley. The numbers refer to site locations in tables 12.1 and 12.2.

Valley likely to be critical to this study together prompted the extension of fieldwork on a reconnaissance level into the Franklin River Valley. Exploration of this valley enables a number of important interpretations of the evidence. Unfortunately it has not been possible to explore these fully in this thesis, but they serve as clues for future research. Fieldwork in the Franklin Valley was limited to the Franklin and Alma glaciers and other glaciers that were confluent with them. No attempt was made to map all the lesser ice limits in the Franklin River Basin once each tributary glacier separated from the Franklin Glacier (Figure 12.2).

EROSIONAL MORPHOLOGY

A Glacial Erosion

Within a few kilometres of its descent from the St. Clair Surface the upper Franklin River is joined by two tributaries, the Surprise and Collingwood rivers. The Collingwood River results from the confluence of the Alma, Inkerman, Patons, Balaclava and Cardigan rivers. Erosional evidence indicates that ice accumulated independantly in each of these valleys, and also in the Franklin and Surprise valleys. Cols at the head of the Franklin and Alma valleys fed diffluent lobes of the Derwent Glacier into the Franklin River Basin. King William Saddle fed part of the upper Franklin ice into the Surprise Valley. Further south glacial erosion at the head of the Surprise has captured an increasing proportion of the ice that previously flowed through the Top End Gap into the Guelph Valley (chapters eight and fifteen).

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

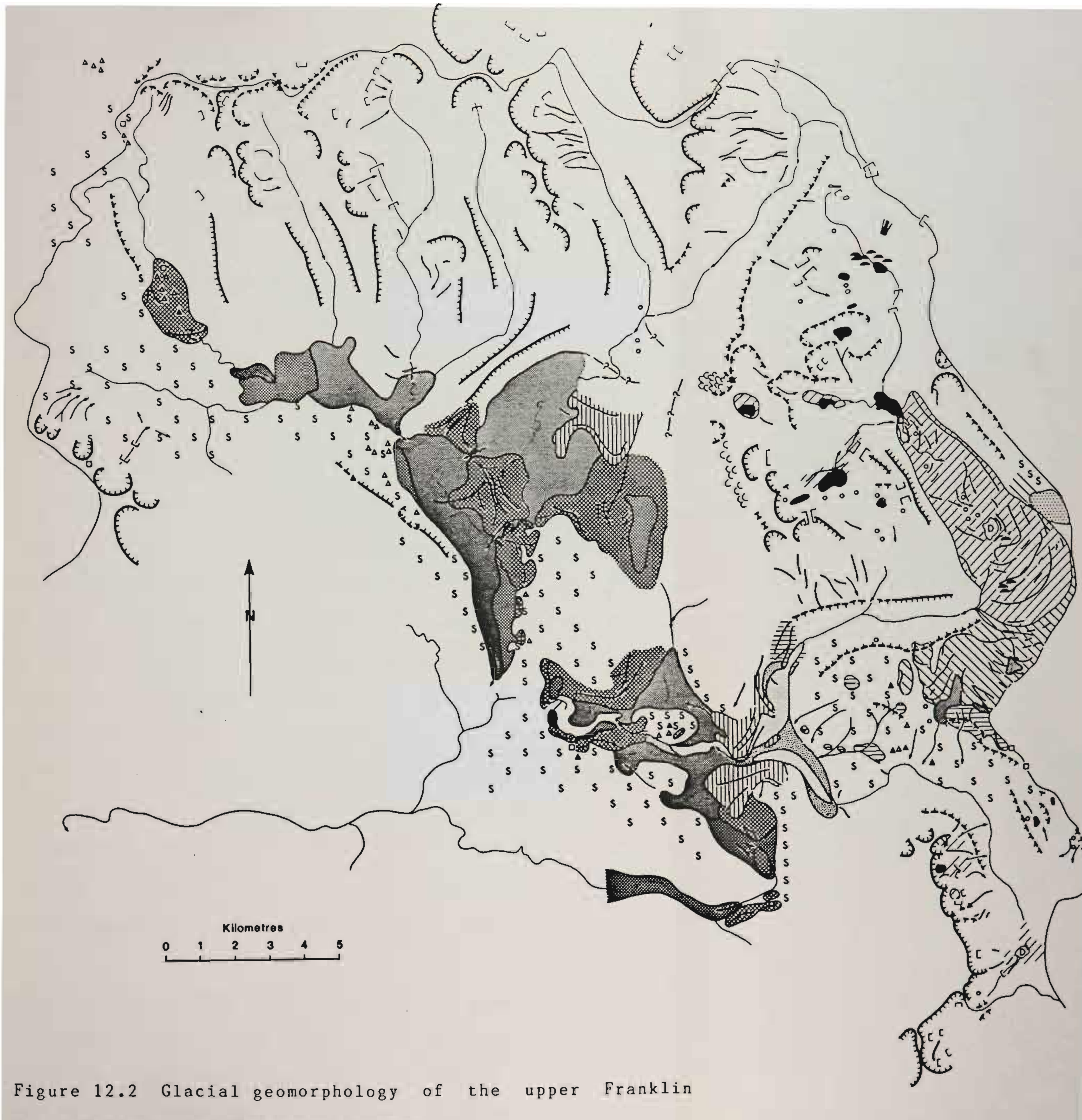


Figure 12.2 Glacial geomorphology of the upper Franklin River Valley. See figure B2 for key.

Shallow over-ridden cirques form a "glacial stairway" in the lee of the Cheyne Range snowfence (Figure 12.3) and owe their form to both piedmont and plateau ice sheets (Derbyshire 1963, 1968a). The cirques on the southern end of the range have not been over-ridden. The Surprise River rises from one of eight cirques in the lee of the Loddon Range snowfence, most of which are poorly developed. On the opposite side of the Surprise Valley one or two very immature cirques have developed on the windward western slopes of Mt. King William I.

A single cirque-like hollow occurs on the western slopes of the less elevated Loddon Range that overlook the Loddon River. This river joins the Franklin downstream of the Collingwood. The hollow lies on the south-western flank of an 890 m outlier of the range. However, till that occurs at 560 m beaneath it was probably deposited by distributary lobe of the Franklin Glacier. A reconnaissance beneath other valley heads of similar aspect and higher elevation (over 1000 m) further south in the headwaters of the Loddon River failed to locate any further till or outwash. If ice accumulated on the western slopes of the Loddon Range the glacial deposits may since have been buried. If true, this would suggest a considerable age for the glaciation.

Many of the cirques in the Collingwood catchment are poorly developed. Most have been initiated on lithological benches. The western branch of the Alma River rises from a series of cirques in the Pyramid Mountain area. Other

Figure 12.3 Cirques on Mt.Gell in the Cheyne Range, viewed from Mt. Hugel. Lake Undine is in the middleground. Part of the shattered dolerite mass that forms the summit of Mt. Hugel is visible in the foreground.

Figure 12.4 The glacial trough of the Surprise Valley, viewed from the southeastern slopes of Mt. Arrowsmith. The King William Range is to the left and Mt. Ronald Cross, a part of the Loddon Range, lies to the right. Mt. King William II is in the background.

cirques are present on Goulds Sugarloaf. The Inkerman, Patons and Balaclava valleys all originate from valley-head cirques in the Last Hill - Camp Hill - Rocky Hill area. The headwalls of some of these cirques appear to have been ice-smoothed, which Peterson (1969) attributed to their having been invaded by an ice sheet. Peterson considered that westward movement of such an ice sheet had inhibited the flow of ice from these cirques and from others on the Raglan Range and thereby retarded their development. This issue is explored more fully in chapter fourteen.

Rock basin lakes are confined to the dolerite mountains, notably in the Cheyne Range. Lake Undine in the upper Franklin Valley is moraine-dammed but steep rock walls partway along its length suggest that at least part of it may be a rock basin.

Most of the valleys exhibit a broadly U-shaped cross profile in their upper reaches where they cut through Permo-Triassic rocks. The troughs are less symetric and poorly developed as they cross Palaeozoic rocks and merge with one another further downstream.

Ice-smoothed bedrock is widespread in the upper reaches of the valleys. Further downstream it has commonly been modified by mechanical weathering except where it lies buried beneath glacial or periglacial deposits. An ice abraded surface of Permo-Carboniferous age has been exhumed upstream of Lake Dixon (Banks, in Derbyshire, 1966) and modified by late Cainozoic ice. Ice-smoothed bedrock at 918

m. suggests that Mt. Arrowsmith was over-ridden at the maximum phase. Evidence of glacial erosion in the lower Collingwood Valley is limited to the truncation of some spurs and a possible ancient cirque that lies on the northern flank of the Collingwood Range within the maximum limits of the Collingwood Valley Glacier.

B Glaciofluvial erosion

Small meltwater channels link many of the ice abraded hollows on the eastern flank of the Cheyne Range. Channels incised in both rock and drift occur upstream of Lake Dixon. About 2 km downstream from this lake the Franklin enters a major meltwater gorge 5 km long and up to 200 m deep. Further channels descend into this from cirques on the southern slopes of Mt. Gell and from north of Dolly Hill.

In the Surprise Valley a series of meltwater channels have been cut in rock at about 1050 m across the western flank of the King William Range. These are interpretted as marginal meltwater channels that imply an ice thickness of 500 m in this part of the valley. Other channels descend steeply from King William Saddle, the slopes of Mt. Arrowsmith and from Calders Lookout. A narrow pass between Mt. Arrowsmith and McKays Peak may be a meltwater channel. A series of shallow ponds in this area, the largest of which is Shirleys Pool, were regarded by Gulline (1965) as karst lakes. However, their position in relation to the meltwater channels suggests that at least of these ponds some may occupy depressions eroded by meltwater.

Meltwater has scallopped bedrock near the western margin of the Carbonate Ck. plain and a prominent meltwater channel extends southwards towards the Loddon River. Many of the valleys to the west have been deepened by meltwater erosion. A broad depression between Artists Hill and the foothills of the Cheyne Range is now drift mantled but may have been initiated by meltwater that flowed between the Alma and Stonehaven valleys.

The Collingwood River flows through a narrow channel 800 m. southeast of Redan Hill, where the river is incised in resistant quartz schists. This is anomalous as more easily eroded pelitic schists parallel the foot of the range less than 200 m to the north-east and underlie most of the lower Collingwood Plain. This course may be due to the river may having been held against the foot of the Collingwood Range by glacial ice or deposits. The Collingwod has cut down 10 m. into the bedrock floor of the channel since the ice retreated and now lies 60 m below its highest outwash terraces.

C Non-glacial erosion

Slab toppling is not as widespread in the Franklin Valley as it is further to the east because only a few of the mountains in this part of the study area bear residual caps of dolerite. Incipient toppling is evident on the eastern slopes of the Cheyne Range above Lake Undine, but toppling is more advanced around the dolerite cliffs that form the summit block of Mt. Gell. Here the dolerite has been extensively shattered by mechanical weathering, in the same manner as the residual dolerite rubbles that form the summits of Mt. Hugel (Figure 12.3) and Mt. Rufus (chapter five).

Ice-abraded rock on the shoulders of Mt. Arrowsmith has also been shattered by mechanical weathering. Shattered tors overlook the Taffys Ck. meltwater channel and must have developed since ice last passed through the saddle between Calders Lookout and Mt. Arrowsmith. Much of the eroded rock around the Collingwood - Franklin junction has been shattered.

Small nivation hollows are present on higher peaks, notably in the Cheyne Range. Kirkpatrick (1984) reports other snow-patches on Rocky Hill and Pyramid Mountain. Pipkrake has been observed on road cuttings down to the lowest levels investigated (400 m).

Fluvial erosion does not appear to have markedly altered the glacial landscape but has created more striking features outside the glaciated areas. Massive gorges such as that downstream of the Collingwood-Franklin confluence cut across the structural grain of the landscape and are probably the result of superimposition from earlier dolerite cover rocks (Davies, 1965). Smaller ravines and gulleys are the product of a long period of fluvial dissection, possibly aided by snow-meltwaters that acted selectively upon less resistant rocks and structures. Apart from some failure of steep slopes formed on unconsolidated materials, particularly after fire, present day fluvial erosion appears very minor and is limited to the incision and reworking of pre-existing sediments on valley floors, and the erosion of exposed soil surfaces. Microscale weathering and erosion features include pans and other solution features that have developed on sandstone benches on Mt. Rufus (eg. at Tasmap grid ref. 8013- 263 348). These lie within the area that was eroded by the the Navarre Glacier.

Some karst development has also taken place in the study area. The outwash plain along Carbonate Creek is pock-marked by sinkholes that have developed since these gravels were deposited. Several largely inactive stream-sinks occur along the margin of the Mt. Ronald Cross plateau close to the former margin of the Surprise Glacier. Glacial erosion of the plateau margin and joint dilation following withdrawal of the valley ice, or less probably dissolution of the dolomite by marginal meltwaters, may have contributed to the position in which the stream-sinks developed. An enclosed depression 500 m in diameter and at least 50 m deep occurs on the crest of a spur on the steep western flank of Mt. Gell. An inverted treeline may be at least partially due to clays that floor the depression and impede drainage (Figure 12.5). This depression may have formed where meltwaters were ponded against the margin of the Alma Glacier by a lateral moraine.

DEPOSITIONAL MORPHOLOGY AND SEDIMENTS.

A Glacial Deposits

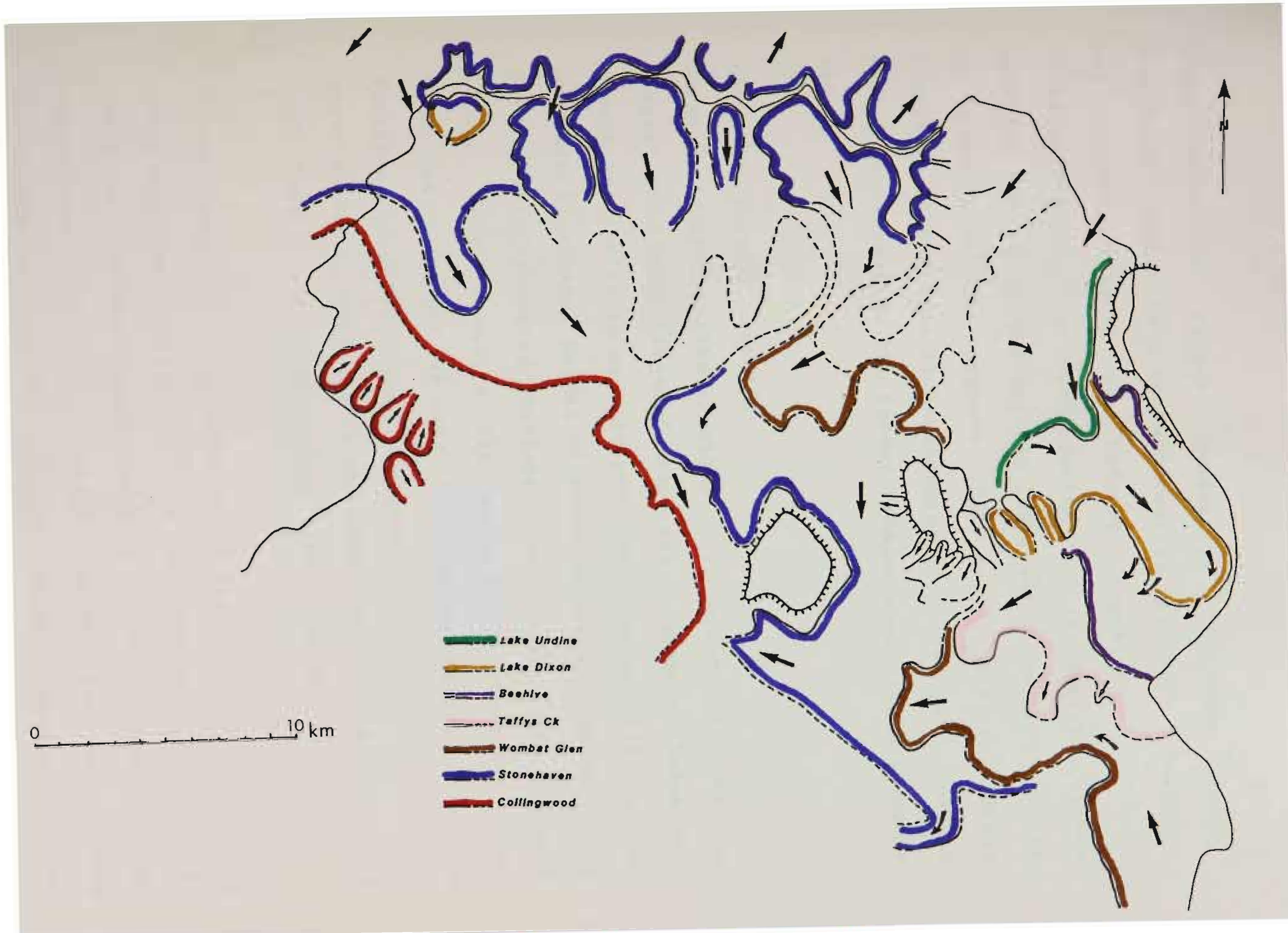
(i) Raglan erratics

Erratics of Ordovician conglomerate occur on the Raglan

Figure 12.5 Large karst depression on the western flank of Mt. Gell in the Cheyne Range. This depression formed close to the upper margin of the Alma Glacier.

Figure 12.6 Thick slope deposits overlie dolerite erratics (beneath figure) at Redan Hill in the Collingwood Valley.

overleaf
Figure 12.7 Principal ice limits in the upper Franklin River Valley.


N
Lake Undine
Lake Dixon
Beehive
Taffys Ck
Wombat Glen
Stonehaven
Collingwood
0
10 km

Range above the cirque headwalls. Peterson (1969) attributed these to a westward moving ice sheet. However, the apparent absence of bedrock exposures of Owen Conglomerate to the east coupled with the fact that these erratics lie far beyond any ice limit which can be supported by other evidence suggests that until more widespread knowledge of the area is forthcoming they are best regarded as lag boulders from the former Permian cover rocks (chapter one).

(ii) Collingwood erratics

Large Jurassic dolerite erratics that occur high on the slopes of the Collingwood Range and on adjacent Redan Hill cannot be Permian in age. At the base of Redan Hill they are overlain by several metres of slope deposits which have been derived from weathered schists (MacIntyre, 1964) (Figure 12.6). These erratics lie up to 150 m above the valley floor and 8.5 km up the Collingwood Valley from the Franklin junction. In the deeper Franklin Valley erratics occur at 600 m altitude on the summit of Artists Hill, and imply 240 m of ice in the valley 3 km from the junction.

Several important conclusions arise from the distribution of these erratics. Firstly, if the ice reached to near the crest of the Collingwood Range at Redan Hill then previous suggestions that the Franklin Glacier did not extend downstream of the Collingwood - Franklin confluence (Spry and Zimmerman 1959) are almost certainly in error. Moreover, some of the Collingwood erratics lie so close to the crest of the range that the possibility of ice having spilled directly into the Lucan Valley and thence into the Franklin must be kept open.

If the ice in the Collingwood Valley was a distributary lobe of the Franklin Glacier which flowed up into the Lower Collingwood as has previously been argued (Spry and Zimmerman 1959) then the failure of the main Franklin Glacier to penetrate further downstream than the confluence becomes even more unlikely. If the ice did reach here from the Franklin then it would undoubtedly have also extended further to the south, possibly far enough to have been confluent with the glaciers that descended from Frenchmans Cap (cf. Peterson, 1969).

Spry and Zimmerman (1959) argued that the lower Collingwood ice came from the Franklin because they had failed to find erratics further upstream in the Collingwood than the base of Redan Hill. However erratics are now known to be present 150 m above the valley floor at this point. Hence, the ice would have had to extend considerably further up this gentle valley if the source had lain to the east. The absence of erratics further upstream may be due to their remobilisation and burial in subsequent slope deposits and outwash. Two moraines that extend across the lower Collingwood Plain and are concave to the northwest indicate that a major glacier flowed down the Collingwood Valley.

Finally, it seems curious that erratics do not occur to similar elevations on top of the Mt.Mullens ridge if the Collingwood ice came from the east. This raises the possibility that the Collingwood supported a major glacier in its own right and that it may have been larger than the Franklin Glacier in this locality at this time (Figure 12.7).

(iii) Stonehaven Moraines.

A sequence of moraines, till and glaciofluvial sediments is present in the Franklin Valley near Stonehaven Creek and also occurs in the lower Collingwood and lower Alma valleys.

Till is well exposed in cuttings along the Lyell Highway from Stonehaven Creek to near the crest of Doherty's Hill (Figure 12.8). Glaciofluvial sediments are also revealed in these sections. These deposits were probably laid down by ice that flowed from the Alma Valley through the broad gap between Mt. Alma and the foothills of the Cheyne Range (figures 12.1 and 12.2). This gap carries little water today and may represent a former route by means of which the Alma River once flowed into the Franklin.

These deposits are far more deeply weathered than even the Clarence Till in the Nive Valley (chapter ten). At Artists Hill they are overlain by 4m of slope deposits that also contain deeply weathered dolerite clasts. Till mantles the northern slopes of Mt. Mullens to a height of 420 m or some 100 m above the floor of the Franklin Valley. This is at least 100 m lower than the ice limit neccessary to account for the Collingwood erratics. The downvalley gradient of the ice surface in this vicinity was about 30 m /km. A section cut by the Franklin River just upstream of the

Figure 12.8 Deeply weathered till of the Stonehaven phase at Double Barrel Creek, near the confluence of the Franklin and Collingwood rivers.

Figure 12.9 The till that forms the moraines of the Beehive phase is much less weathered that that of the Stonehaven phase.

flying fox on the Frenchmans Cap track (Tasmap grid ref. 8013-155 261) reveals till, glaciofluvial gravels and rhythmites that contain dropstones.

Further upstream the till extends southwards into the open valley of Carbonate Creek. Large boulders of Owen Conglomerate 3 km south of the Franklin River (Tasmap grid ref. 8013-193 227) are of uncertain origin. They were regarded as erratics by Spry and Zimmerman (1959) and as bedrock by Gulline (1965), but could be lag boulders from earlier Permian cover rocks. Till and outwash occur 1 km to the south adjacent to the meltwater channel that leads into the Loddon Valley. The apex of one of the gravel aprons that floor the Loddon Valley lies at the outlet to this channel.

At least 30 m of glacigenic sediments floor the lower Collingwood Valley. Two end moraines extend across it. Both are concave to the north-west. Outwash terraces occur up to 60 m above river level. Ice-contact glaciofluvial sediments occur near Scarlett Creek. A large latero-terminal moraine extends across the mouth of the Alma Valley. The full extent of the till in the Alma is not yet known but it appears to be considerable.

An impressive sequence of outwash terraces is present along the middle reaches of the Collingwood downstream of the Balaclava River junction. An end moraine crosses the Balaclava Valley at about 480 m This seems likely on morpho-stratigraphic grounds to be broadly equivalent to the

Stonehaven moraines. It is concave upstream and was constructed by ice that moved down the Balaclava. Boulders of Jurassic dolerite, Owen Conglomerate and Permo-Triassic rocks occur in the till and as erratics up to 3 m on the plain upstream. Neither Owen Conglomerate nor dolerite appear to outcrop in the Balaclava Valley.

The absence of any convincing evidence for Owen Conglomerate bedrock to the east suggests that these clasts must be either lag boulders from the former Permian cover rocks or were derived from the West Coast Range by late Cainozoic glaciers. The West Coast Range lies beyond the broad expanse of the King River Valley and carriage of conglomerate erratics on the eastern margin of the King Glacier is inconsistent with its established erratic distribution pattern (Kiernan, 1980). The conglomerate erratics probably have been reworked from the Permian rocks that are likely to have contained many conglomerate erratics so close to their source area.

The Jurassic dolerite erratics may have been derived as lagstones from the dolerite that originally capped the range at the head of the Balaclava Valley, or alternatively they may have been carried into the Balaclava by ice that extended southwards from the Eldon Range. Four lines of evidence favour the latter interpretation. Firstly, no mention of residual dolerite is made in the geological report on the range by Read (1964). Read attributed the presence of dolerite boulders on the Eldon - Balaclava divide to glacial deposition and photographs of the boulders presented in his report certainly give an impression of glacial rounding. Secondly, the rounded appearance of the Last Hill - Rocky Hill range is entirely consistent with its having been over-ridden by ice that flowed from the north or northeast. Thirdly, ice is known to have moved southwards out of the South Eldon Valley into the King Valley a few kilometres to the west. This ice left erratics on the Little Eldons, which form an extension of the Last Hill-Rocky Hill ridge (Kiernan, 1980). Finally, the cirques at the head of the Balaclava are not well situated with respect to either altitude or snow-fences. The development of a major valley glacier in the Balaclava Valley would seem unlikely unless the glacial climate was so intense that the entry of ice into the valley from elsewhere was almost inevitable. All these considerations point towards ice from the King Glacier System having spilled into the Balaclava Valley.

As a footnote, supplementation of the Collingwood ice from the King Glacier must have been greater at the time the Collingwood erratics were deposited than it was during the Stonehaven phase. As diffluence diminished the Balaclava Glacier would have declined markedly. It probably descended steeply with an ice surface gradient of at least 100 m /km during Stonehaven time. Ice from the King system probably failed to spill into the Balaclava subsequently, hence only small valley and cirque moraines were later constucted there.

(iv) Wombat Glen Moraines.

Major rock spurs extend southwards from Mt. Gell and northwards from the Loddon Range. Large lateral moraines have been constructed by the Franklin Glacier on both these spurs, enclosing the small plain known as Wombat Glen. An extensive apron of glaciofluvial gravel and sand extends downvalley of the Wombat Glen Moraine. These deposits are inset between the lateral moraines that were constructed during the Stonehaven phase. These glaciofluvial sediments are well exposed in sections cut by the Franklin River.

A latero-terminal moraine occurs in the Alma Valley 5 km north-east of the Alma - Collingwood confluence. This appears to be plastered against the flank of a bedrock ridge that was over-ridden by the Stonehaven ice. The Vale of the Mists is floored by an outwash plain that extends downstream from this moraine. Fragments of a hard, dense grey basal till occur as clasts in the moraine. The basal till is very similar to the Nive Till (chapter ten) but in the Alma Valley it includes includes numerous quartz erratics (Table 11.1). Although apparently ancient this basal till is so compact and water-tight that dolerite within it is unweathered. Some freshly fractured samples of this sediment react vigorously to 10% HCl. The carbonate has probably been derived from the dolomite on the western slopes of the Cheyne Range 2 km. to the north-east, or from the poorly mapped Precambrian basement rocks elsewhere in the valley.

During the Wombat Glen phase the ice appears to have been unable to extend far beyond the margin of the St. Clair Surface. Much of the ice that flowed down the Franklin and Alma valleys probably entered through the cols at their heads and via King William Saddle (chapter 7). The flow of ice from the Franklin Valley onto the St. Clair Surface was impeded south-east of Dolly Hill by the ice from Lake St. Clair (chapter seven) which deflected the Franklin Glacier westwards.

The Franklin Glacier is likely to have always been bigger than the Surprise Glacier but it only reached the confluence of the Surprise and Franklin rivers at this time. This suggests that the Surprise Glacier could not have eroded its impressive glacial trough during the Wombat Glen phase. Therefore, the Surprise trough must have been carved during an earlier and more intense phase of glaciation.

The gradient of the ice surface in the terminal zone was probably about 40 m./km. when the Wombat Glen moraines were constructed. Once the ice had retreated to King William Saddle additional meltwater would have been deflected around the northern side of Dolly Hill, deepening the meltwater channels at Burns Ck. and in the Upper Franklin Gorge.

(v) Taffys Creek Drift

Weathered till forms a lateral moraine near the northern margin of Wombat Glen but the end moraine appears to have been washed out by meltwater discharging from the upper Franklin gorge. Wombat Glen is floored by an outwash plain that formed downstream of this moraine. Comparably

Vicinity No.	20	26	31	36
%				
dolerite	50.0	6.0	20.0	36.0
quartz	22.0	79.0	40.0	27.2
Triassic sandstone	-	6.0	10.0	15.1
argillites	-	3.0	12.0	3.2
schist	15.0	3.0	8.0	12.4
other/unident.	7.0	3.0	10.0	6.1

Table 12.1 Lithology of glacial deposits in the Franklin River Valley. Site locations are indicated on Figure 12.1.

weathered till occurs along the northern side of the Surprise Valley west of Taffys Ck. Erratics are exposed 160-180 m above the floor of the Surprise Valley at several points along the southern slopes of Mt. Arrowsmith, at up to 890 m on its eastern flank and in scattered localities to the north-west. Some erratics have been incorporated into slope deposits. Weathered till also occurs just on the western side of King William Saddle at 820 m.

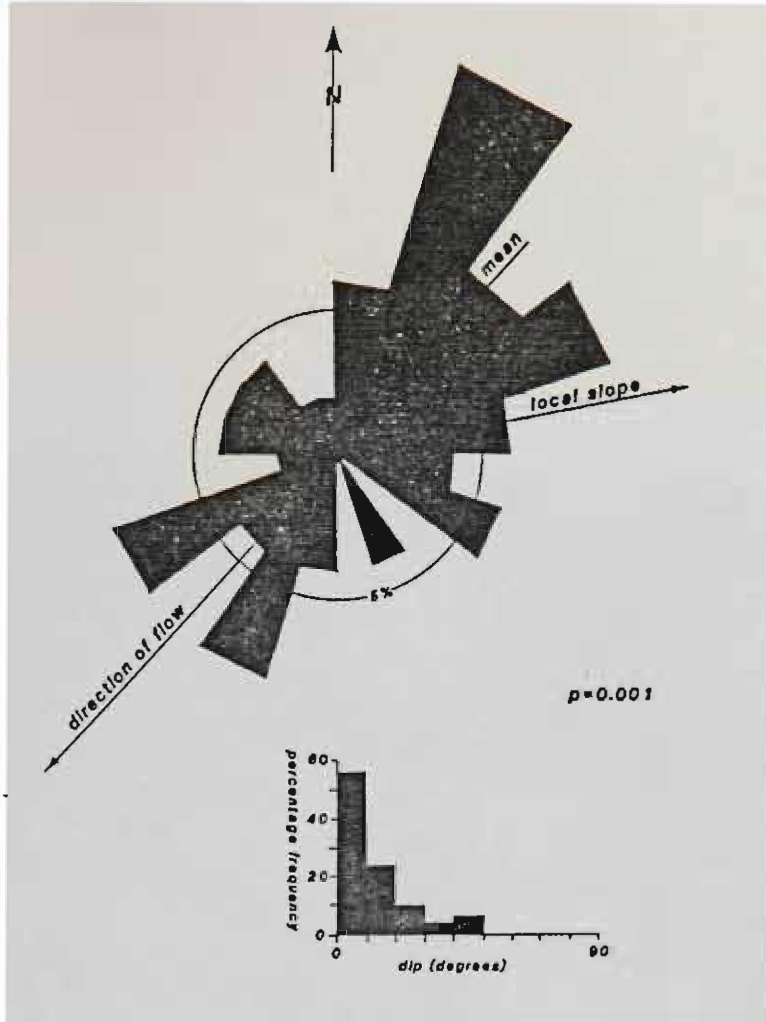
Little moraine morphology has survived as most of the sediments were deposited on steep slopes. However, it appears that a narrow tongue of ice flowed down the upper Franklin Gorge and expanded onto the broad valley floor at the northern end of Wombat Glen. Ice also passed through the col between Mt. Arrowsmith and Calders Lookout, and spilled over King William Saddle.

(vi) Beehive Moraines:

The upper Franklin Valley turns sharply westward and commences to descend steeply from the St. Clair Surface a short distance south of Lake Dixon. A large proportion of the Franklin Glacier continued out onto the St. Clair Surface east and west of Dolly Hill, which forms a minor obstacle to continued southward flow. During Beehive time the Franklin Glacier spilled onto the plains east of Dolly Hill. Well preserved end moraines occur between Beehive Ck and the Franklin - Navarre divide (Figure 12.9). Some of the ice bulged up onto Dolly Hill. The Beehive ice must also have spilt some distance down the Upper Franklin Gorge (Figure 12.10) but the very steep slopes here were not

Figure 12.10 Fabric diagram from basal till on the southeastern slopes of Dolly Hill indicates that the ice was flowing towards King William Saddle onthe edge of the St. Clair Surface. The angles at which the pebbles dip (Andrews and Smith, 1970) suggest that the velocity of the ice was only moderate.

Figure 12.11 Lake Dixon and Dolly Hill lie close to the edge of the St. Clair Surface, and are viewed here from Mt. Rufus. The entry to the upper Franklin gorge trends sharply to the right just beyond Lake Dixon. Mt. Arrowsmith and King William Saddle lie to the left.


N
mean
local slope
direction of flow
5%
p=0.001
60
40
20
0
percentage frequency
0
90
dip (degrees)

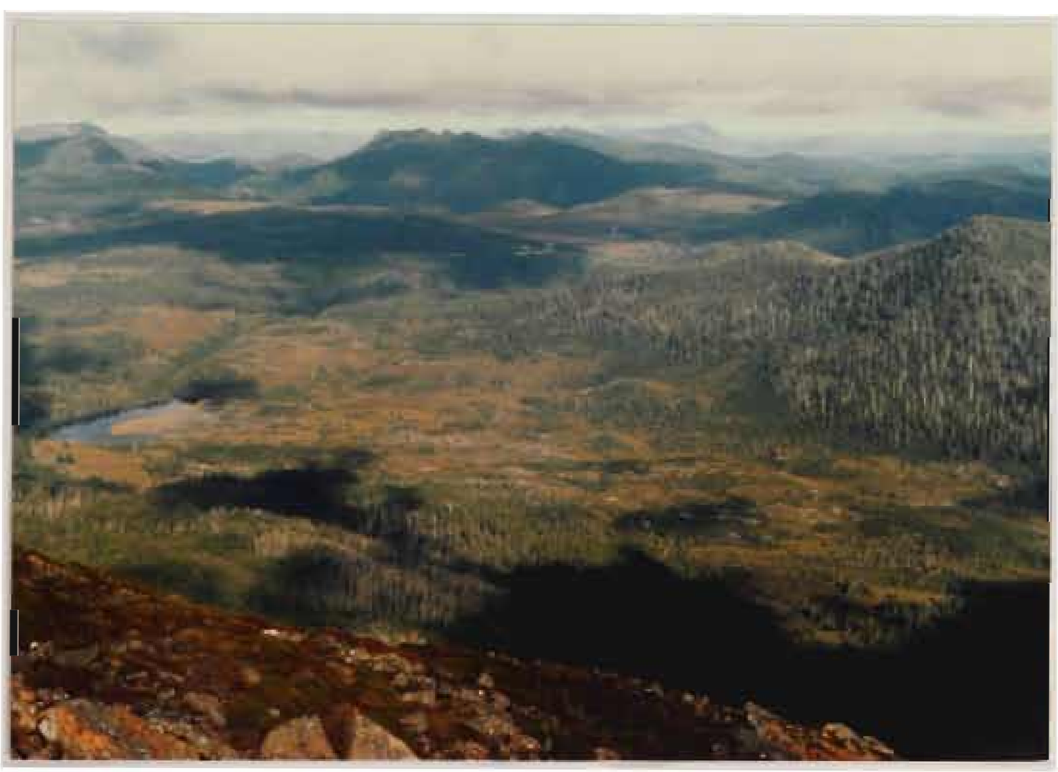


Figure 12.12 Solifluction terraces on the western slopes of Mt. Rufus.

conducive to the preservation of moraines.

During the Beehive phase the ice cover appears to have barely attained the threshold dimensions that were neccessary for it to spill through King William Saddle. Evidence for this is provided by the patchy occurrence of Beehive till near the saddle and by the failure of the Beehive ice to remove totally older glacial deposits in the lee of bedrock protruberances near Griffiths Ck. Glaciofluvial and glaciolacustrine sediments were deposited between the terminal moraines and the crest of King William Saddle as the ice retreated.

(vii) Dixon moraines

Lake Dixon is impounded behind an extensive end moraine complex near the edge of the St. Clair Surface (Figure 12.11). This complex rises over 50 metres above the valley floor and is over 1 km broad. Its crest lies at 720 - 740 m. It records a phase during which the Franklin Glacier did not descend below 680 m. The till of which this moraine complex is composed is little weathered. During Dixon time outwash gravels were discharged down the Franklin gorge from the western margin of the glacier snout. Burns Ck. delivered some meltwater from the eastern margin of the ice around the glacier snout and into the gorge.

A lateral moraine occurs at 820 m on the western side of the valley 2 km upstream from Lake Dixon. Small latero-terminal moraines cross the valley about 1 km upstream from the lake. The survival of Lake Dixon without its being filled by proglacial sediment when the ice lay further upstream is probably due to sediment traps having been formed by lesser end moraines that were plastered against roches mountonées.

The dramatic retreat of the Franklin Glacier after the Dixon phase was probably brought about by the cessation of diffluent ice flow into the valley from the Cuvier area. Fluted drift in the col at the head of the Franklin probably dates from this time.

(viii) Undine moraines

The downstream basin of Lake Undine is impounded behind a series of roches moutonées and low moraine ridges with crests at 700 m. Small outwash plains occur downstream from these moraines. The moraines associated with the small lakes on the slopes of the Cheyne Range are generally small, and this coupled with the fact that Lake Undine has not been filled by proglacial sediment suggests that the retreat from the Lake Undine limit was rapid.

Non-glacial sediments

(i) Slope deposits

Slope deposits are widespread. Many metres of angular rubbles overlie the erratics at the base of Redan Hill (Figure 12.6). Up to 4 m of quartzitic slope deposits overlie till in the lower Alma area and overlie both till and ice-smoothed bedrock at Artists Hill. No more than 30 cm of slope deposits have accumulated within the Dixon ice limits.

Two generations of slope deposits are exposed in a cutting on the southern side of Calders Lookout. The lower unit comprises angular rubble with some rounded and deeply weathered dolerite erratics in a silty clay matrix that comprises about 50 % of the sediment. This is overlain by a plug of openwork scree that fills a local topographic depression in the surface of the older scree. A thin podsolic palaeosol near Shirleys Pool at 580 m on the southern slopes of Mt. Arrowsmith is developed on till in which dolerite clasts are comparably weathered to those in the lower unit at Calders Lookout. The palaeosol is overlain by up to 1.5 m of small calibre slope deposits.

Non-sorted stone steps (Washburn, 1979) occur at 1320 m on the western slopes of Mt. Rufus and have been formed from dolerite rubbles derived from the residual summit blocks (Figure 12.12). In order to ascertain whether these terraces were actively moving forward, three rows of dolerite pebbles were marked with red paint in July 1981. One row lay along the foot of the riser, another along the edge of the tread, and the third lay 10 cm back from the edge of the tread. When the site was casually examined in April 1983 no displacement of the pebbles was evident, but upon inspection in November 1985 only three of the pebbles could be located, each of which formed part of the riser. This indicates active forward movement of the terrace.

Kirkpatrick (1984) has reported that non-sorted stone steps and stripes are well developed on the Permo-Triassic rocks that form the summits of Rocky Hill and Pyramid Mountain, and at Camp Hill, High Dome and Goulds Sugarloaf. Athrotaxis logs on Rocky Hill have been buried beneath solifluction terraces. The logs were interpreted by Kirkpatrick as indicative of a recent period of severe landscape instability after fire, but the wood has not been radiocarbon assayed and the date of any such event is unknown.

The surface morphology of a series of blocky deposits that extend down the western slopes of the Cheyne Range near the large sinkhole suggests that they are fossil rock glaciers (Derbyshire, 1973). Some of these rock glaciers descend to as low as 1150 m. Unlike their counterparts on Mt. Olympus, a number of them are associated with cirques and may have been ice cored rather than ice cemented forms. Talus accumulations below this level which Derbyshire considered may have been older rock glaciers seem more likely to be collapsed talus that accumulated on the margin of the Alma Glacier during the Stonehaven phase.

Slab toppling is once again the principal mechanism of slope retreat in the dolerite mountains. Extensive accumulations of dolerite talus underlie most of the free faces outside the limits of the Dixon phase ice, but little talus occurs within those limits.

(ii) Alluvial deposits

Up to 2 m of silt and clay has accumulated in depressions on the St. Clair Surface and silt has also accumulated as an overbank deposit adjacent to the main rivers. Less than 1 m. of silt has accumulated in Shirleys Pool on the southern side of Mt. Arrowsmith.

DATING AND DISCUSSION

The youngest deposits are alluvial silts. These silts overlie solifluction deposits that extend into Shirleys Pool from the slopes of Mt. Arrowsmith. The solifluction deposits overlie a palaeosol that is developed on weathered till. Charcoal from the palaeosol has been radiocarbon assayed at 13,000 +- 640 BP (SUA 1959).

The Undine end moraine and outwash plain lies well inside the ice limits of the Dixon phase. During Dixon time the Franklin Glacier extended only 12 km from the col at its head and attained a maximum thickness of less than 150 m. The large size of the end moraine complex downstream of Lake Dixon suggests that this moraine represents an important ice limit. Thick solifluction deposits do not occur within the ice limit of the Dixon phase but are widespread beyond it. This indicates that the main phase of solifluction predates retreat from this ice limit.

Moraines of the Beehive phase can be traced into the Navarre basin where they represent parts of the Sawmill Ridge and Coates Creek-King William Creek complexes (chapter seven). During the Taffys Creek phase ice probably reached the northern end of Wombat Glen. An outwash plain was constructed inside the ice limit that is defined by the end moraine of the Wombat Glen phase.

During the Wombat Glen phase the Franklin and Alma glaciers were respectively 25 km. and 16 km. in length. The outwash plains that extend downstream from Wombat Glen are enclosed by the older lateral moraines of the Stonehaven phase.

The Stonehaven phase is represented by moraines in the Franklin, Alma and probably Balaclava valleys. The Franklin Glacier terminated about 27 km downvalley from the diffluence col at its head, while the Alma Glacier extended for at least 16 km. At least 120 m of ice flowed through King William Saddle. Ice also passed between Calders Lookout and the Beehive and possibly over-topped both. The Surprise Glacier was at least 200 m thick south of Mt. Arrowsmith.

Deposition of the highest Collingwood erratics required a considerably greater extent of ice than that documented by the subsequent Stonehaven Moraines. Several lines of evidence suggest that the glacier in the Balaclava Valley to the west played a major role during the Collingwood phase and that the Balaclava Glacier was largely nourished by diffluent ice from the King Glacier.

The Collingwood Glacier was thickened to at least 150 m by the Alma ice. At least 100 m of ice must have reached the rock bars at the mouth of the Middle Franklin gorges north of Frenchmans Cap, and ice probably extended well into the gorge. One lobe of the Alma Glacier flowed around the northern side of Mt. Alma to merge with the Franklin

Glacier. A second lobe continued downvalley to merge with the Collingwood Glacier which was also confluent with the Franklin.

If the relative proportions of the Collingwood and Stonehaven phases were the same in the Balaclava Valley as in the Alma-lower Collingwood, then the Balaclava Glacier was probably confluent with the Collingwood Glacier during Collingwood time. If this was the case then the glacier in the Collingwood is likely to have been larger than that in the Franklin. However, it would have been susceptible to very rapid decline once the ice flow from the Eldon and King valleys was cut off. This is an important point since confluence of the King and Franklin glaciers at this time may provide a benchmark which offers the possibility of better correlation between the glacial deposits that occur towards the west coast and those that occur in the Central Highlands.

The degree of postdepositional modification of the landforms and sediments is presented in Table 11.2.

Shallow A Cox Cu type soil profiles are developed on the Undine and Dixon moraines. All the earlier moraines are characterised by thick B horizons. The B horizons in the Beehive moraines are up to 3 m thick. They are at least 4 m thick in the moraines of the Wombat Glen and Stonehaven phases.

Weathering rinds that have formed on dolerite clasts in the

moraine[a]	C	C	S	S	S	S	S	S	S	S	WC	WC
vicinity no.	1	2	3	4	5	6	7	8	9	10	11	12
material[b]	B	E	O	O	O	O	T	E	T	T	O	T
A. MORPHOSTRATIGRAPHY												
relative position	outside 8	outside 8	outside 12	outside 3	outside 8	outside 8	outside 11	inside 5	inside 7	inside 9	outside 8	inside 11
crest/site elevation(m)	360	430	440	320	390	400	400	430	400	400	400	390
max.depth of burial (bv) (m)[c]	-	-	-	S3	S7	S3	S1	sq4Sacb)	Sq 2	S4	-	-
B. SURFACE CLASTS												
dolerite rinds-max.	-	-	24.2	9.2	53.8	22.8	89.0	63.7	67.4	10.0	10.9	30.5
-min.	-	-	9.1	1.3	2.2	1.5	37.3	30.2	14.8	2.0	1.3	1.7
-mean	-	-	14.0	4.1	16.9	7.9	49.1	40.4	34.1	5.1	5.4	11.9
- SD	-	-	4.6	2.1	13.7	4.3	12.8	10.3	10.8	1.7	2.3	7.1
-hardness (1-5)	-	-	2	2	2-3	2-3	3	3-4	3	3	1-3	2-3
coherence of argillites (1-5)	-	-	4	4-5	4-5	4	4	3	4	3	3	3-4
clast surface colours	-	5YR	7.5-5YR	7.5-5YR	5-7.5YR	7.5-5YR	5YR	5YR	5YR	7.5-5YR	7.5-5YR	5YR
C. SUBSURFACE MATRIX												
profile type	-	-	A Bt	A Bt	-	-	A Bt	A Bt	A BT	A Bt	A Bt	A Bt
colour of B horizon	-	-	5-7.5YR	7.5YR	5-7.5YR	5YR	5YR	5YR	5-7.5YR	7.5-5YR	7.5YR	7.5YR
depth of B horizon(cm)	-	-	> 200	>150	-	-	> 400	> 300	> 200	> 250	> 250	> 300
depth to Cox (cm)	-	> 500	> 200	> 150	-	-	> 400	> 300	> 200	> 250	> 150	> 300
pH at ~1m	-	-	4.4	-	-	-	-	4.6	-	-	5.6	5.6
clay films (1-5)	-	-	2	2	-	4	4	4	3	-	1	-
basal till lithification (1-5)	-	-	-	-	-	-	-	3	-	-	-	3
fine matrix:	-	-	(a) (b)									
% sand	-	-	50.1 71.4	-	-	-	-	68.4	-	-	70.8	-
% silt	-	-	26.4 15.7	-	-	-	-	18.7	-	-	17.9	-
% clay	-	-	23.3 8.9	-	-	-	-	12.9	-	-	11.3	-
clast socket staining(1-5)	-	-	4	4	5	3-4	4	5	3-4	3	-	4
D. SURFACE MATERIALS												
E. DEPOSITIONAL LANDFORMS												
moraine crest width (m)	-	-	-	-	-	-	-	30	-	-	20	10
fluvial dissection(1-5)	10	-	3	3	4	4	4	3	-	4	2	1
F. EROSIONAL LANDFORMS												
width of joint opening (cm)	-	-		-	-	-	-	-	-		-	-
depth of joint opening (cm)	20	-	-	-	-	-	-	-	-		-	-
depth of scree(m)	-	-		-	-	-	-	-	-		-	-
height of tors (m)	-	-		-	-	-	-	-	-		-	-

moraine[a]	WC	WC	TC	TC	TC	TC	TC	TC	TC	B	B	B
vicinity no.	13	14	15	16	17	18	19	20	21	22	23	24
material[b]	T	O	O	T	T	T	ST	T	B	O	O	T
A. MORPHOSTRATIGRAPHY												
relative position	inside 14	inside 8	inside 12	inside 15	inside 15	inside 15	inside 15	inside 14	inside 12	outside 7	inside 13	inside 19
crest/site elevation(m)	800	390	390	790	590	400	825	470	918	50	380	870
max.depth of burial (bv) (m)[c]	S1	S1.5	-	S2	S1	-	S2	-	S1.5	AF4.5	-	S1.5
B. SURFACE CLASTS												
dolerite rinds-max.	23.8	26.2	13.1	11.6	16.0	13.1	8.0	13.6	-	2.0	3.8	6.6
-min.	3.1	4.2	2.7	2.1	1.9	1.4	3.5	3.8	-	0.8	0.1	1.3
-mean	12.0	8.9	6.1	7.0	7.2	6.4	5.4	7.1	-	1.4	1.12	3.56
- SD	4.7	4.6	2.7	6.3	2.7	3.1	1.2	5.2	-	0.4	0.53	1.01
-hardness (1-5)	2-4	2	3	2-4	2-3	2-3	2	3	-	1	1	2
coherence of argillites (1-5)	4	3	3	3-4	4	3-4	3	4	-	1	2	3
clast surface colours	5YR	7.5YR	7.5-5YR	5YR	5YR	5YR	7.5-5YR	5YR	-	-	-	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
colour of B horizon	7.5-5YR	7.5YR	7.5YR	7.5-5YR	7.5-5YR	7.5-5YR	-	7.5YR	-	10YR	10YR	7.5YR
depth of B horizon(cm)	> 500	> 200	> 200	> 300	> 100	> 400	-	> 300	-	> 180	?	> 100
depth to Cox (cm)	> 300	> 200	> 200	> 300	> 100	> 400	-	> 300	-	> 180	-	> 100
pH at ~1m	-	-	5.4	-	-	5.2	-	4.6	-	4		4.4
clay films (1-5)	4	-	-	4	4	-	-	3	-	-	1	2
basal till lithification (1-5)	2	-	-	1	-	-	-	-	-	-	-	-
fine matrix:								(a) (b)				
% sand	-	-	-	-	-	-	-	46.7 63.1	-	-	68.0	-
% silt	-	-	-	-	-	-	-	31.2 21.1	-	-	17.8	-
% clay	-	-	-	-	-	-	-	22.7 15.8	-	-	14.2	-
clast socket staining(1-5)	4	-	-	5	3	3	-	3	-	-	-	3
D. SURFACE MATERIALS												
E. DEPOSITIONAL LANDFORMS												
moraine crest width (m)	-	-	-	-	-	20	-	20	-	-	-	-
fluvial dissection(1-5)	4	1	2	4	4	1	-	1	-	1	2	3
F. EROSIONAL LANDFORMS												
width of joint opening (cm)	-	-	-	-	-	-	-	-	25	-	-	-
depth of joint opening (cm)	-	-	-	-	-	-	-	-	?(q)	-	-	-
depth of scree(m)	-	-	-	-	-	-	-	-	1.5	-	-	-
height of tors (m)	-	-	-	-	-	-	-	-	-	-	-	-

moraine[a]	B	B	B	B	B	B	LD	LD	LD	LD	LD	LD
vicinity no.	25	26	27	28	29	30	31	32	33	34	35	36
material[b]	T	T	T	T	T	T	T	T	B	F	S	O
A. MORPHOSTRATIGRAPHY												
relative position	inside 19	inside 25	inside 19	inside 19	inside 17	inside 19	inside 30	inside 31	inside 31	outside 9	inside 33	outside 31
crest/site elevation(m)	880	800	780	890	780	800	670	660	800	45	820	650
max.depth of burial (bv) (m)[c]	-	S1	-	-	_	S1	S0.1	-	S0.1	-	-	-
B. SURFACE CLASTS												
dolerite rinds-max.	3.7	9.2	5.9	6.5	7.2	6.2	2.3	3.8	-	0.9	2.4	1.0
-min.	1.7	4.1	1.4	2.9	2.2	2.2	0.1	0.4	-	0.1	0.8	0.1
-mean	2.9	6.7	3.8	4.1	3.9	3.6	1.6	1.5	-	0.51	1.56	0.32
- SD	0.57	1.8	1.2	1.3	1.4	1.2	0.3	0.7	-	0.24	0.38	0.19
-hardness (1-5)	2	2	2	2	2	2-3	1	1	-	1	1	1
coherence of argillites (1-5)	3	3	3-4	3	3	3-4	2	1	-	1	1	1
clast surface colours	7.5-5YR	5YR	7.5YR	7.5YR	7.5YR	7.5YR	10-7.5YR	7.5YR	-	-	10-7.5YR	10YR
C. SUBSURFACE MATRIX												
profile type	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 Cox cu
colour of B horizon	7.5YR	7.5YR	7.5-5YR	7.5YR	7.5YR	7.5YR	10YR	10YR	-	10YR	10YR	10YR
depth of B horizon(cm)	> 100	> 300	> 200	> 100	>100	>150	-	-	10	< 30	20	< 40
depth to Cox (cm)	> 200	> 300	> 200	> 100	>100	>150	-	-	10	> 30	>30	< 30
pH at ~1m	4.6	5.4	5.3	5.3	4.8	-	4.9	-	-	-	-	-
clay films (1-5)	2	-	-	4	3	3	1	1	-	-	-	-
basal till lithification (1-5)	-	2	-	-	2-3	-	-	1	-	-	-	-
fine matrix:												
% sand	-	-	-	-	65.3	-	-	-	-	-	-	-
% silt	-	-	-	-	20.2	-	-	-	-	-	-	-
% clay	-	-	-	-	14.5	-	-	-	-	-	-	-
clast socket staining(1-5)	2-3	3	3-5	3-4	2-3	2-3	1	2	-	-	-	-
D. SURFACE MATERIALS												
E. DEPOSITIONAL LANDFORMS												
moraine crest width (m)	-	-	10	-	30?	20	5	15	-	-	-	-
fluvial dissection(1-5)	4	3	3	3	3	3	1	1	-	3	-	-
F. EROSIONAL LANDFORMS												
width of joint opening (cm)	-	-	-	-	-	-	-	-	-	-	-	-
depth of joint opening (cm)	-	-	-	-	-	-	-	-	-	-	-	-
depth of scree(m)	-	-	-	-	-	-	-	-	-	-	-	-
height of tors (m)	-	-	-	-	-	-	-	-	-	-	-	-

(a) C= Collingwood; S= Stonehaven; WC= Wombat Glen; TC= Taffys Creek; B= Beehive; LD= Lake Dixon.

(b) B= bedrock; E= erratics; T= till; ST= solifluctad till; O= outwash gravels; F= fluvial gravels; S= solifluction deposit.

(c) Refers to burial of the same depositional unit in the same vicinity but not at the exact site where subsurface data were collected. The latter have only been recorded from unburied sites except for site 10. S= slope deposit (q= quartzite; schs schist); AF= fluvial gravels and sands.

Table 12.2 Post-depositional modification of deposits in the Franklin River Valley. Site locations are indicated on Figure 12.1.

Dixon moraines reach a maximum thickness of 3.8 mm. This is half the thickness of the maximum rinds in the Beehive moraines. The thickest rinds in the deposits of the Taffys Creek phase reach 16.0 mm , while those in the Wombat Glen moraines reach 30.5 mm. Rinds up to 89.0 mm thick occur in the Stonehaven moraines. These tills are more deeply weathered than any others that have been encountered in the study area.

Systematic variation also exists in the degree to which the glacial landforms have been eroded. The Undine and Dixon moraines are well preserved, but those of the Beehive phase generally have a subdued topography. Slope deposits that contain weathered clasts occur within the Wombat Glen ice limits and probably date from the Beehive phase. The Stonehaven moraines are overlain by slope deposits that are up to 4 m thick where they were derived from quartzitic bedrock and up to several metres thick where they were derived from schists. The Collingwood phase is represented solely by erratics and no moraine morphology remains. The Collingwood River has cut 10 m into resistant bedrock below the base of the till since the ice withdrew. Several metres of slope deposits derived from weathered schists overlie the erratics at the base of Redan Hill.

These post-depositional characteristics suggest that the earliest glacial deposits in the Franklin Valley are very old.

PART C

CONCLUSIONS

EVOLUTION OF THE MOUNTAIN LANDSCAPE

chapter thirteen
DATING AND REGIONAL COMPARISONS

".. whether the evening light or the peculiar configuration of the Frankland Range emphasised the fact I know not - but I was struck with the absolute clearness, on the panorama there unfolded, of the evidence for two distinct and superimposed glaciations."

-A.N.Lewis (1926)

This chapter explores the relative dating of the late Cainozoic deposits and the probable significance of the post-depositional differences between morphostratigraphic units in terms of time. Relative ages are suggested for the deposits. The evidence is then reviewed in the context of previous studies of Tasmanian glaciation and comparisons are drawn with the glacial histories of other mountain landscapes in temperate southern latitudes.

1 Relative dating of the glacial deposits

The field relationships between moraines and outwash terraces provide the basic means for establishing the relative sequence of the glacial events. No radiometric assays or organic deposits suitable for pollen analysis have been obtained which aid in the dating of these events. The differentiation and correlation of the glacial deposits has been facilitated by the morphostratigraphy and in particular the continuity of some drift sheets between valleys. This has been reinforced by comparing the extent of post-depositional modification of the landforms and deposits. Five separate drift sheets have been differentiated on the basis of these criteria. From youngest to oldest they are: the Cynthia Bay Drift; the Beehive Drift; the Powers Creek Drift; the Clarence Drift; and the Stonehaven Drift (Table 13.1).

The degree of post-depositional modification also forms the basis for numerical estimates of the age difference between the deposits. The parameters scrutinised have been outlined in chapter three. Non-systematic variability has been evident in all of these parameters. Subsurface parameters have revealed the most consistently systematic changes. Relative dating has therefore been based upon moraine and terrace relationships; altimetric and morphological relationships of reconstructed ice margins; the weathering of subsurface argillitic clasts; the thickness of dolerite weathering rinds; the depth and degree of oxidation of the drift; pedogenic horizonation and clay enrichment of the drift; and the degree of lithification of basal till.

A Morphology

The spatial relationships between erosional landforms and glacial deposits provides some evidence for multiple glaciation. All the glacial deposits within the steep mountainous terrain in the glacier source areas lie well inside the maximum ice limits suggested by the erosional morphology. Examples of this include lateral moraines in the Cuvier Valley that lie inside the Olympus Col, and other lateral moraines that are constructed upon or within the

Valley →	St.Clair	Cuv.Hug	Traveller	Navarre	Guelph	DB to Wyat.	Nive	Gordon	Franklin
Event ↓									
CYNTHIA BAY									
	Stoney Ck.				Sally J			Cirque M.	
	Rangers Hut	Shadow Lake							
	Ida		Midlake	Slacker					L.Undine
	Peninsula	MidCuvier	Park						
		Hugel Track							
	Cynthia Bay	Watersmeet	Cynthia Bay	Cynthia Bay	Rufus R.		Cynthia Bay	L.Richmond	L.Dixon
BEEHIVE									
	Bedlam W.	East Rufus	Bedlam W.	Bedlam W.	Top.Middle	Bedlam W.	Plateau	Camp I	Beehive
			North Charles	Little Nav.		Mt.Charles III			
			Broken Leg		Long Bay	Mt.Charles II			
			Clarence Col	Coates/KW	Gorge	Mt.Charles I		Divide?	
POWERS CK									
						Sawmill II			Taffys Ck
				Sawmill		Sawmill I			
				Guelph	Guelph	Guelph			
						Butlers G.			
						Cascade	Laughing J.		
					Hobhouse?	Mossy M	Powers Ck.	Hobhouse?	
CLARENCE									
							Clarence		Wombat G.
STONEHAVEN									
		Nive Till	Nive Till			Wayatinah?	Nive Till		Stonehaven Collingwood

Table 13.1 Approximate correlation of the glacial phases identified in the study area.

arms of cirques that have previously been ice smoothed or over-ridden.

The tills of Cynthia Bay age generally occur in reasonably close proximity to the highest cirques or within troughs that were eroded by earlier glaciers. The incision of immature cirque-in-cirque forms such as Sunrise Hollow was associated with the deposition of tills during the Cynthia Bay phase. The Beehive, Powers Creek, Clarence and Stonehaven tills occur beyond the Cynthia Bay ice limits. The Clarence and Stonehaven deposits often occur as degraded moraines or isolated erratics beyond the limits of landforms produced by glacial erosion.

The topography of the Cynthia Bay moraines is sharp and fresh in all cases. Enclosed depressions in the moraines, such as the small ponds in the Rangers Hut moraine complex north of Lake St. Clair, have not been filled by sediment. In the Derwent Valley the steep flanks and sharp crests of the Cynthia Bay moraines contrast with the more massive and rounded moraines of the Bedlam Wall and earlier stages. Periglacial slope processes have rounded the topography and smoothed the lower slopes of hills outside the Cynthia Bay ice limits. Solifluction has obscured most of the glacial deposits downstream of the Cascade Moraine in the Derwent Valley and beyond the ice limit of the Powers Creek phase in the Nive Valley. In the Franklin Valley the slope deposits that overlie the Redan Hill erratics are comparable in thickness to those that occur within the Stonehaven ice limits. This suggests that the Collingwood and Stonehaven phases are approximately of the same age.

B Postdepositional modification of the sediments

(i) Weathering of argillitic clasts

Variation in the lithology, mineralogy and grain size of Permian and Triassic argillitic clasts militates against the usefulness of measuring the weathering rinds that have developed on them. However, a systematic trend in clast weathering is evident, whereby the argillites are unweathered in the Cynthia Bay deposits but totally decomposed in the Clarence Till. This progression is evident only in the tills. Argillites are seldom heavily weathered in the glaciofluvial deposits probably because the paucity of fine matrix inhibits the retention of moisture.

(ii) Dolerite weathering rinds

The thickness of dolerite weathering rinds varies from the mean by up to 50% at one standard deviation, and in extreme cases by up to 80%. Much of this variance is probably due to differences in mineralogy brought about by differentiation processes. Variations at one standard deviation very rarely overlap between the Cynthia Bay and Beehive tills, and never overlap between the Clarence and Stonehaven tills. Mean rind thickness at one standard deviation is not mutually exclusive in the case of the Beehive, Clarence and Powers Creek tills (Table 13.2).

Because some variation in the mineralogy probably confounds the mean values, a case exists for basing differentiation upon extreme values that reflect either the most readily

Valley →	St.Clair	Cuv.-Hugel	Traveller	Navarre	Guelph	DB to Wyat.	Nive	Gordon	Franklin
Event ↓									
CYNTHIA BAY									
max.	4.6	5.5	2.3	2.8	4.3		2.1	1.4	3.8
min.	0.1	0.2	o.6	0.1	0.3		0.2	0.6	0.4
mean	1.5	1.9	1.5	1.6	1.5		1.3	1.2	1.6
SD.	0.5	0.8	0.5	0.8	0.7		0.4	0.9	0.5
BEEHIVE									
max.	6.9	5.1	6.4	11.7	8.8	5.2	6.8	6.1	9.2
min.	0.5	0.8	1.8	0.5	1.0	0.9	2.8	1.6	4.1
mean	3.0	4.5	3.0	4.8	5.1	2.9	3.8	3.1	4.1
SD.	1.5	0.8	1.1	1.7	1.7	1.5	1.5	1.4	1.2
POWERS CK									
max.				16.9		15.8	17.5		15.1
min.				3.4		1.0	0.1		1.9
mean				8.1		5.1	6.4		7.4
SD.				3.9		2.2	2.5		3.8
CLARENCE									
max.							37.0		30.5
min.							2.8		1.7
mean							13.1		12.0
SD.							3.1		5.9
STONEHAVEN									
max.						70.0			89.0
min.						8.2			14.8
mean						23.4			41.2
SD.						10.2			11.3

Table 13.2 Thickness of dolerite weathering rinds in the glacial deposits, in millimetres. Total number of rinds recorded in each valley is provided in Table 13.4.

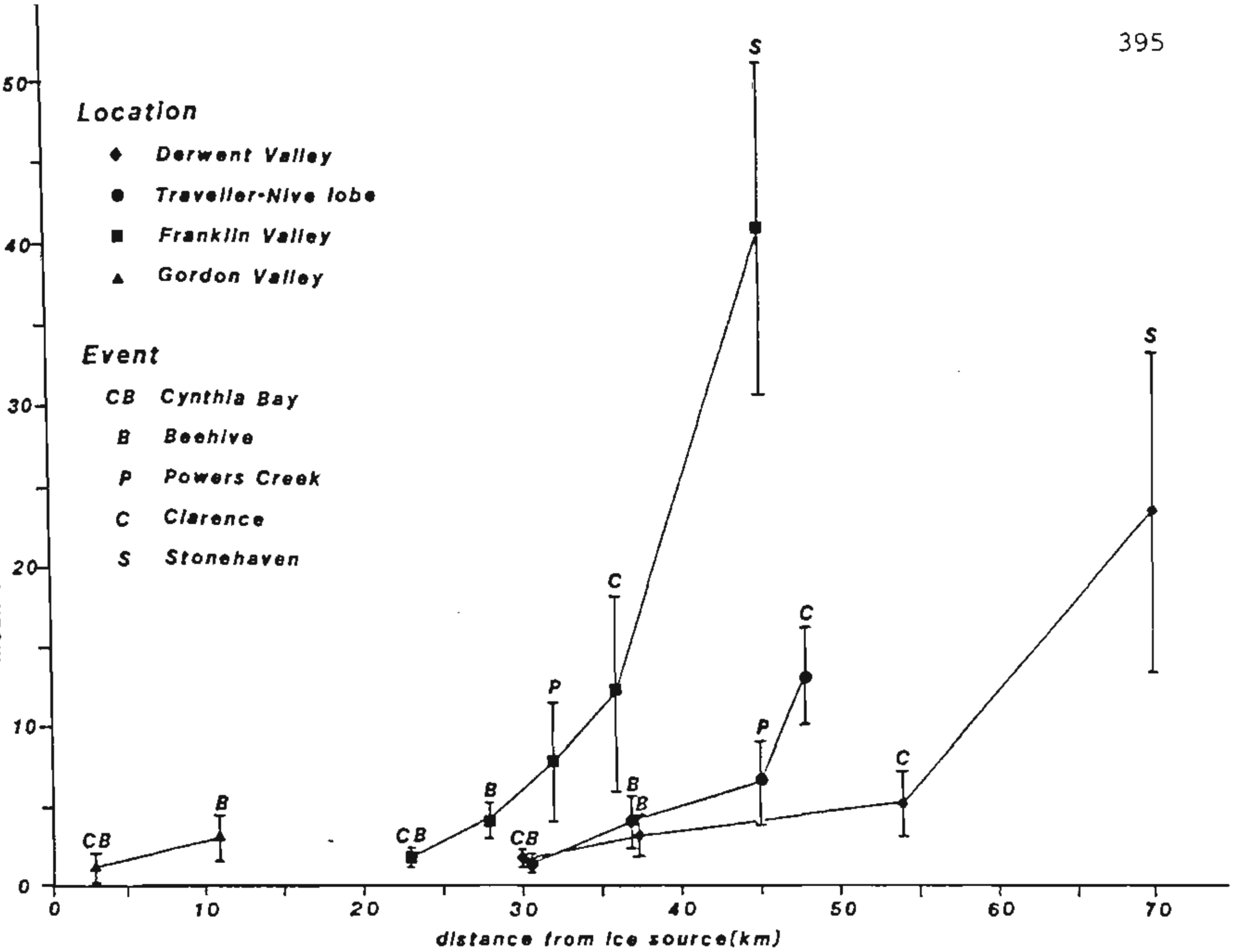
weathered or least readily weathered clasts (Figure 13.1).

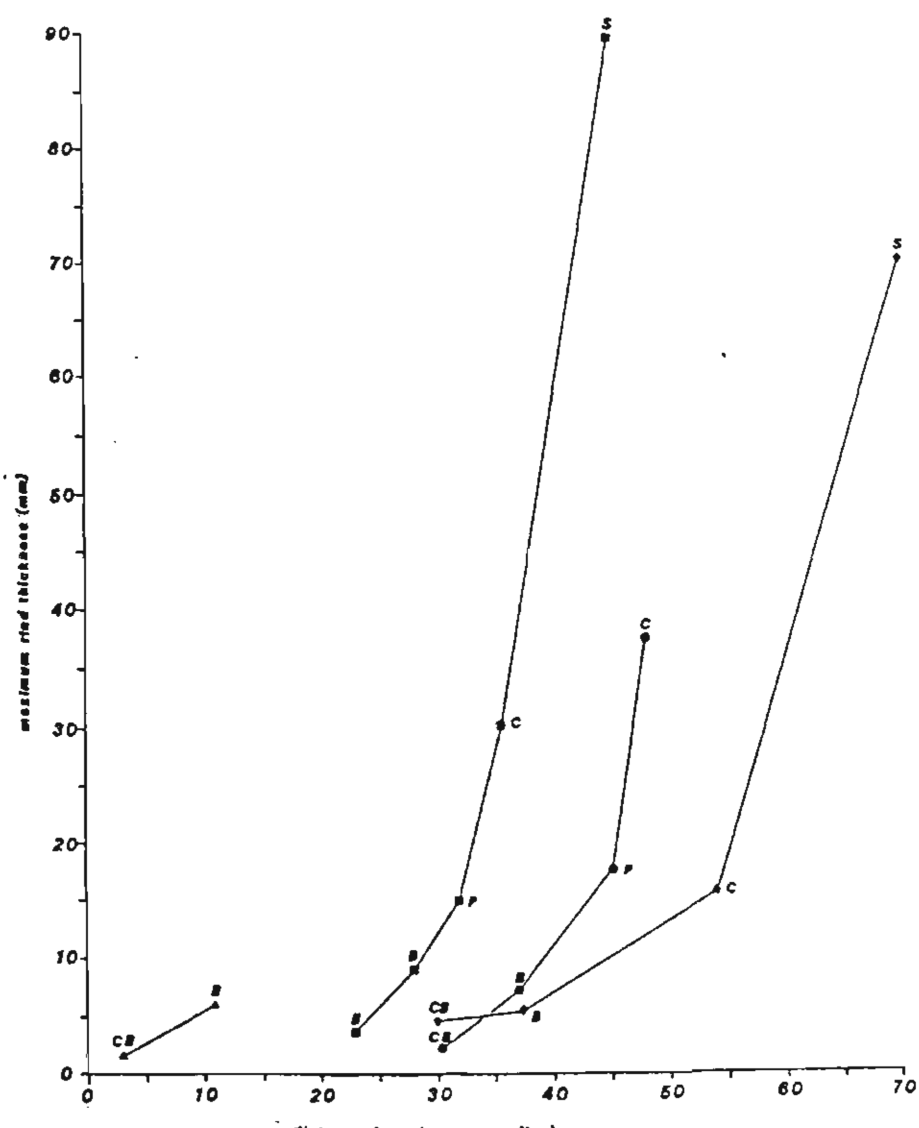
A clearer systematic trend is evident in maximum rind thickness, although it is not wholly consistent. The drift sheets most distant from the ice source areas exhibit the thickest rinds although the range of maximum values within each of the defined drifts is highly variable. Birkeland _et al._, (1979) argue that post-depositional data generally provides only a minimum age for the till and that those criteria which display the greatest differences might better reflect the true age. The thickest rinds in the Powers Creek Till are, on average, at least twice as thick as those in the Beehive Till. The thickest rinds in the Clarence Till are, on average, at least twice as thick as those in the Powers Creek Till (Figure 13.2).

(iii) Weathering rinds on calc-silicate hornfels clasts

Weathering rinds on hornfels clasts show a generally consistent trend with the thickest rinds most distant from the ice source areas. The difference in mean rind thickness between the Cynthia Bay and Powers Creek tills is not significantly different at one standard deviation. The thickest rinds in the Beehive till are about 80% thicker than in the Cynthia Bay Till, and the difference between the Beehive and Powers Creek Till is of a similar order. As less than seventy hornfels rinds have been recorded there is a double need for caution in the interpretation of these results.

Figure 13.1 Systematic variation in dolerite weathering rind thickness with distance from ice source areas


Location
Derwent Valley
Traveller-Nive lobe
Franklin Valley
Gordon Valley
Event
CB Cynthia Bay
B Beehive
P Powers Creek
C Clarence
S Stonehaven
rind thickness (mm) mean & standard deviation
distance from ice source(km)


maximum rind thickness (mm)
distance from ice source (km)

(iv) Oxidation of the drift

Use of this parameter as an aid to dating is hampered by the shallow exposure of most of the glacial deposits. Oxidation has penetrated deeply into glaciofluvial deposits. Thin to moderately developed iron pans are present at depths of as much as 2 m in Cynthia Bay outwash, and strongly developed pans are present at over 4 m in Powers Creek and Stonehaven outwash deposits.

Thin iron pans occasionally extend to 1 m depth in the Cynthia Bay Till, to over 2 m in the Beehive and Powers Creek tills and at least 4 m in the Stonehaven Till.

There is an overall tendency for the older tills to be redder in colour than the younger deposits. The colour of the Cox horizon in the Cynthia Bay Till is generally in the 10 YR colour range. The colour of this horizon in the Beehive, Powers Creek and Clarence till is generally in the 7.5 YR - 5 YR range, with the stronger red occurring in tills that are rich in dolerite. The colour of the Stonehaven drift is almost invariably in the 5 YR range unless dolerite is sparse.

The surface colour of dolerite clasts in the Cynthia Bay Till is generally in the 7.5 YR range while those in the older tills are almost invariably in the 5 YR range.

(v) Pedogenic horizonation and clay enrichment

The soils developed on Cynthia Bay Till are characterised by highly immature A Cox Cu type profiles (Birkeland, 1984) in which dolerite clasts show no sign of patina development below 10 - 30 cm depth. All the older tills exhibit thick textural B horizons. The depth of these B horizons has not been ascertained because sections are not sufficiently deep to reveal the base of this horizon, which commonly lies deeper than it is feasible to excavate in the moraines. Weathered dolerite clasts in a Bt soil horizon occur to depths of over 3 m in the Beehive Till on the northern slopes of Dolly Hill and in the Powers Creek Till near Butlers Gorge. Clay-rich B horizons with weathered dolerite clasts are found to depths of 6 m in the Clarence Till at Wombat Glen and to even greater depths in the Stonehaven Till.

The clay-sized fraction in the older profiles is derived in part from the decomposition of small dolerite clasts. Many dolerite clasts that are less than about 5 cm in diameter occur in the Cynthia Bay Till. They are unweathered. In the Beehive Till they are generally only moderately weathered although a few very coarse grained clasts may be totally rotted. In the Powers Creek Till many of the smaller dolerite clasts have been totally decomposed. Field sieving near the Powers Creek and Clarence limits in the Clarence Valley revealed no small dolerite clasts but innumerable small clay balls that are probably decomposed dolerite.

(vi) Basal till lithification

Basal till of Cynthia Bay age remains slightly pliable to the extent that in some cases a fingernail or pencil can penetrate the matrix. Basal till of Beehive, Powers Creek and Clarence age is harder and more dense. It can sometimes be broken manually due to its blocky subangular pedal structure, although occasionally hammer blows are required. The Nive Till, which is interpreted as the basal facies of the Stonehaven Drift, is very tough and dense and can only be fragmented by repeated hammer blows. It could fairly be described as a tillite.

C Significance of the post-depositional differences

The post-depositional modification of the glacial landforms and deposits indicates that the tills of this area vary considerably in age and points towards a generally three-fold sequence. The Cynthia Bay till is only lightly weathered and the moraines are well preserved. It represents the most recent phase of glaciation. The degree of weathering is consistent with the tills of the Dante Glaciation in the central West Coast Range that were deposited shortly after 18, 800 +- 500 BP (ANU 2533)(Kiernan, 1980, 1983a). The Cynthia Bay Glaciation is therefore regarded as the maximum glaciation of the late Last Glacial Stage (isotope stage 2 of Shackleton and Opdyke, 1973)

The Beehive, Powers Creek and Clarence moraines are degraded and the tills are weathered. Thick textural B horizons characterise the soils that are developed on the pre-Cynthia Bay tills and suggest that they all predate the Last Interglacial Stage. The Stonehaven Till is very deeply weathered and in many cases the depositional landforms are

SITE	COMMENTS	MAX.	MIN.	MEAN ± SD
Roaring Beach, Tasman Peninsula	gravels, overlain by aeolian sands, and slope deposits	2.0	0.7	1.21 ± 0.42
Great Bay, Bruny Island	gravels beneath aeolian sands	4.6	3.7	4.13 ± 0.37
Forth, Northern Tasmania	gravels in terrace	3.1	1.4	2.75 ± 1.04
Forth, Northern Tasmania	gravels in fossil beach ridge	2.8	1.4	2.07 ± 0.97

Table 13.3 Thickness (in millimetres) of dolerite weathering rinds in raised beach deposits of presumed Last Interglacial age from sites in northern and southeastern Tasmania.

severely degraded. It is clearly ancient. The presence of erratics beyond the limits of continuous drift at Redan Hill and elsewhere suggests that the Stonehaven moraines represent a retreat phase of a more extensive glaciation.

None of the measured parameters resolves the status of the Beehive, Powers Creek and Clarence tills in a way that is entirely satisfactory. In this thesis they are referred to collectively as the Butlers Gorge complex. The most useful comparative data is provided by the dolerite weathering rinds. In an effort to ascertain whether the Beehive Till dates from the early part of the Last Glacial Stage (isotope stage 4 of Shackleton and Opdyke, 1973) or antedates the Last Glacial Stage the thickness of weathering rinds in the Beehive Till was compared with that of the dolerite rinds in raised beach deposits of presumed Last Interglacial age that occur on several parts of the Tasmanian coastline (Chick, 1971; Van de Geer et al., 1978). This comparison revealed that rinds in the Beehive Till are up to twice as thick as those in the presumed interglacial marine deposits at Roaring Beach on Tasman Peninsula, Great Bay on Bruny Island and near the mouth of the Forth River in northwestern Tasmania (Table 13.3). This suggests that the Beehive Till probably predates the Last Interglacial Stage (isotope stage 5). However, the deposits are not strictly comparable in terms of local climate, the Beehive Till is much richer in clays than the marine deposits, and the dating of the beach deposits is uncertain.

Because the mean rind thickness in the Beehive, Powers Creek and Clarence tills is not mutually exclusive at one standard deviation it could be argued that all three reflect retreat phases of the same glaciation. However, the maximum rind thicknesses change by a factor of two between each of these drifts. A change of this magnitude in numerical values of at least some relative dating criteria is regarded by Birkeland et al. (1979) as the minimum basis for attributing deposits to different glaciations.

An approximation of the relative age of all the tills may be made on the basis of evidence from the Franklin Valley where the sequence is most complete (Table 13.4). For simplicity, the weathering rinds of Cynthia Bay age are assumed to have developed in the 10,000 years since deglaciation, or the last 14,000 years since the glaciers probably were at their maximum extent. If the rate of rind development is assumed to be a linear function of time, all but the Stonehaven Till could date from the Last Glacial Stage. However, the thick textural B horizons on the Beehive moraines suggest that the youngest of the pre Cynthia Bay deposits must antedate the Last Interglacial Stage. If a rind thickness of 4.1 mm (mean) and 9.2 mm (max.) has taken 130 ka to develop it seems highly unlikely that rind thicknesses nearly twice that (Powers Creek) or thrice that (Clarence) could have developed in deposits that date from earlier phases of the same glaciation.

The post-depositional characteristics of the Stonehaven Till bear comparison with the Linda Till of the central West Coast Range and elsewhere, and with the Bulgobac Till in the lower Pieman Valley both of which are palaeomagnetically reversed (Kiernan, 1983a; Colhoun and Augustinus, 1984; Colhoun, 1985). If a minimum age of 730 ka is assumed for the Stonehaven Till and a linear rate is assumed between then and the known weathering at 10 ka the age of the Beehive Till would still appear underestimated (Table 13.4).

These simplifying assumptions both seem to represent a probable source of underestimation of the age differences between the tills. The Cynthia Bay till may predate the 10 ka estimate for cirque deglaciation which is based on the subsequent accumulation of organic materials in the cirques. The cessation of frost shattering in the lower Franklin Valley about 15 ka BP (Kiernan _et al._, 1983) suggests that the coldest glacial conditions may have ended well prior to 10 ka BP. This would imply a slower weathering rate and hence a greater age for the glacial deposits. If rind development is enhanced under warm and moist interglacial conditions then the onset of cold glacial climates may periodically slow the weathering rate.

The assumption of a linear weathering rate is probably a more significant source of error. Numerous field and laboratory studies have shown that the rate of chemical weathering is not linear but slows over time as the build-up of residues impedes the evacuation of solutes (Colman, 1981). Insufficient radiocarbon dated dolerite till deposits are available in Tasmania to permit construction of a sound rind development curve. It has been argued that the

ASSUMPTIONS			ESTIMATED AGES (Ka. BP)								
rate	known values (ka)	rind thickness	St.Clair	Cuv.Bugel	Traveller	Navarre	Guelph	DB-Wayat.	Nive	Gordon	Franklin
CYNTHIA BAY											
linear	10 or 14	max	10-14	10-14	10-14	10-14	10-14	10-14	10-14	10-14	10-14
linear	10 or 14	mean	10-24	10-14	10-14	10-14	10-14	10-14	10-14	10-14	10-14
decelerating	0,14,750	max	-	-	-	-	-	-	-	-	14
decelerating	0,14,750	mean	-	-	-	-	-	-	-	-	14
$A=K.t^{2.0}$	K = 5.8	mean	13	21	13	15	13	(13)	10	8	15
No. of clasts			40	300	20	60	20	0	40	20	40
No. of sites			2	15	2	3	2	0	2	1	2
BEEHIVE											
linear	10 or 14	max	15-21	10-13	28-39	42-58	20-29	11-16	32-45	44-61	24-34
linear	10 or 14	mean	20-28	24-33	20-28	30-42	34-48	19-27	29-42	22-36	26-36
decelerating	0,14,750	max	-	-	-	-	-	-	-	-	60
decelerating	0,14,750	mean	-	-	-	-	-	-	-	-	60
$A=K.t^{2.0}$	K = 5.8	mean	52	117	52	134	151	49	84	56	97
No. of clasts			80	40	160	100	60	60	20	80	140
No. of sites			4	2	8	5	3	3	1	4	7
POWERS CREEK											
linear	10 or 14	max	-	-	-	60-84	-	34-48	83-113	-	40-56
linear	10 or 14	mean	-	-	-	50-76	-	34-48	31-69	-	46-65
decelerating	0,14,750	max	-	-	-	50-76	-	34-48	31-69	-	46-65
decelerating	0,14,750	mean	-	-	-	-	-	-	-	-	110
$A=K.t^{2.0}$	K = 5.8	mean	-	-	-	381	-	131	238	-	318
No. of clasts			0	0	0	140	0	300	160	0	80
No. of sites			0	0	0	7	0	15	8	0	4
CLARENCE											
linear	10 or 14	max	-	-	-	-	-	-	176-247	-	80-112
linear	10 or 14	mean	-	-	-	-	-	-	101-141	-	75-105
decelerating	0,14,750	max	-	-	-	-	-	-	-	-	232
decelerating	0,14,750	mean	-	-	-	-	-	-	-	-	200
$A=K.t^{2.0}$	K = 5.8	mean	-	-	-	-	-	-	995	-	835
No. of clasts			0	0	0	0	0	0	140		60
No. of sites			0	0	0	0	0	0	7		2
STONEHAVEN											
linear	10 or 14	max	-	-	-	-	-	152-212	-	-	234-329
linear	10 or 14	mean	-	-	-	-	-	156-212	-	-	257-374
decelerating	0,14,750	max	-	-	-	-	-	750	-	-	750
decelerating	0,13,750	mean	-	-	-	-	-	750	-	-	750
$A=K.t^{2.0}$	K = 5.8	mean	-	-	-	-	-	3,175	-	-	9,845
No. of clasts			0	0	0	0	0	20	0	0	80
No. of sites			0	0	0	0	0	1	0	0	4

Table 13.4 Estimated ages of the glacial deposits, based on different possible weathering rates. See text for explanation. Because glacial deposits of the Cynthia Bay phase do not occur between Derwent Bridge and Wayatinah the age estimates in that valley are based on the rind development rate calculated for the Lake St. Clair area.

rate of weathering rind development is a square root function of time (Andrews and Miller, 1980; Colman and Pierce, 1981).

Caine (1983) has used the rind thickness on subsurface clasts at Ben Lomond to estimate the age of the tills thus:

$$A = k.t^{2.0}$$

where A is age in ka.

k is a proportionality constant

and t is rind thickness in mm.

He estimates the proportionality constant as 5.8, based on the rinds at a dated Holocene site at Mt. Field in southern Tasmania (5.7) and rinds in the cirque moraines at Ben Lomond which he estimates at 10 ka (5.9). Applying this equation to the mean rinds in the Franklin Valley suggests ages for the most recent events which are plausible but unlikely ages for the earlier events (Table 13.4).

In summary, nonsystematic variation in rind thickness and imprecise age control on Tasmanian glacial deposits impedes the use of weathering rinds for numerical dating. The freshness of the Cynthia Bay deposits indicates that they date from the Late Last Glacial Stage. The weathering of the Bedlam Walls Drift is probably attributable to the Last Interglacial Stage. The calculations of relative age based on weathering rates probably give a reasonable impression of the order of magnitiude of the age differences. They suggest that the Beehive, Powers Creek and Clarence tills may represent separate glaciations. However, it is unwise to rely on a single relative dating criterion (Blackwelder, 1931). A more cautious approach suggests that they should be viewed as no more than stadia within a single glaciation until such time as more adequate evidence is available.

The Stonehaven drift represents a glacial climatic stage that must considerably predate the Clarence advance. While the contrast between the preservation of the Stonehaven deposits and those of the Collingwood phase suggests a major age difference the evidence is insufficiently strong to invoke two glaciations to account for this contrast. Uncertainty also surrounds the Nive Till, which has been assumed to be the basal facies of the Stonehaven Drift although unequivocal stratigraphic evidence is lacking. It could be older.

2. Age of the extraglacial landforms and deposits

A Slab topples

Slab topples are well developed outside the ice limits of the Cynthia Bay Glaciation but within those limits only incipient rock crevasses have been recorded. The partly detached topples at Big Gun Pass in the Du Cane Range and on the margin of the Mt. King William I Plateau overlooking Lake George occur on the upper margins of the Cyntia Bay ice in both cases. This suggests undercutting by the glaciers of the Cynthia Bay phase and subsequent collapse following the retreat of the supporting ice margin. The main phases of toppling therefore predate the Cynthia Bay phase.

B Other talus deposits

The block aprons that line the foot of the Mt. Olympus cliffs immediately north of Lake Helen consist largely of rockfall talus. This talus has been reworked into the northern lateral moraine of the Helen Glacier and is rare within the ice limit that the moraine defines. The moraine dates from the Cynthia Bay Glaciation. This pattern is consistent over much of the area and suggests that the major phase of rockfall predates final deglaciation.

Large joint-bounded blocks and smaller more angular blocks extend down steep slopes inside the Cynthia Bay ice limits in the upper Cephissus Trough, in the Cuvier Valley and elsewhere. These blocks have been interpreted as talus that was deposited in contact with the ice margin and collapsed following retreat of the ice. Therefore, their presence is not inconsistent with a general cessation of rockfall during the Holocene. Protalus forms a substantial component of the uppermost moraines at Sunrise Hollow that are of Cynthia Bay age. No evidence has been found of any later phase of protalus accretion.

There are minor exceptions to the general pattern of postglacial stability. A large rockfall has occurred on the margin of the Cynthia Bay ice on the slopes of Falling Mountain. This appears very fresh and almost certainly occurred during the late Holocene, as did another rockfall at the head of Lake Rufus in the King William Range. Under Holocene conditions major slope failures commonly occur in response to high magnitude storm events (Wolman and Miller, 1960; Renwick, 1977; O'Loughlin _et al._,1982) and it is probable that these isolated events are of such an origin. Minor seismic events have also been recorded in western Tasmania (Underwood, 1979) and may have contributed to rockfalls.

The small talus chutes that descend steeply from many alpine clifflines commonly originate in narrow gullies and re-entrants within the cliffs. They are the result of small scale rockfall and some avalanche activity during the Holocene.

C Rock Glaciers

The rock glaciers on the western flank of Mt. Olympus extend below the maximum ice limit in the Cuvier Valley and so must postdate that phase of glaciation. Their fresh form and their apparent contempraneity with the small cirque glaciers on the opposite side of Mt. Olympus suggests that they date from the late Last Glacial maximum.

D Solifluction mantles and other slope deposits

Solifluction mantles are abundant outside the ice limits of the Cynthia Bay Glaciation, poorly developed within the maximum limits of the Cynthia Bay ice and absent in the valley heads. This indicates that the main phase of solifluction predated final deglaciation but was still in progress while the glaciers were present. The weathering rinds in most of the slope mantles are comparable in thickness to those in the Cynthia Bay tills, but at some sites they are thicker. As entrainment in solifluction lobes and sheets is less likely to remove totally all pre-existing rinds than is glacial transport, comparisons between mantles may be better based upon mean rather than maximum rind values (Table 13.5).

No clear stratigraphic relationship between older, more weathered solifluction deposits and younger, less weathered mantles has been demonstrated, perhaps because sufficiently deep exposures have not been located. However, in the Franklin Valley the widespread quartzitic slope deposits are overlain in one section by an openwork scree. Dolerite erratics occur within the lower slope deposits. These have weathering rinds that are comparable in thickness to those in till that is exposed near Shirleys Pool on the slopes of Mt. Arrowsmith. Fresh slope deposits that overlie the till postdate 13,000+-640 BP (SUA 1959).

More recent slope instabilty is reflected by landslide scars and rubble deposits on some steep slopes, notably on the walls of the Cephissus Trough and on the eastern slopes of Mt. Olympus. In both cases the landslides occur within the limits of the Cynthia Bay ice. It is argued in chapter fifteen that these slope failures probably reflect an adjustment of the steep till-mantled slopes to the withdrawal of a supporting ice margin, coupled with storm events during the Holocene.

E Alluvial deposits

Radiocarbon assays indicate that the deposition of fine sediment succeeded the construction of outwash terraces well

Site	Broken Leg Ck.	Bedlam Wall	Mossy Marsh Ck	Clarence R.	Wayatinah
context	overlies Beehive Till	overlies Beehive Till	overlies Laughing Jack Till	overlies Clarence Till	overlies Till
thickness (mm)	1.45 ± 0.8	2.5 ± 0.5	3.1 ± 0.1	5.9 ± 0.7	4.1 ± 2.2
a) Linear estimate (ka)	9.6	16.7	20.6	45.0	
b) exponential estimate (ka)	12.2	36.3	55.7	202.0	97.0

Table 13.5 Weathering rinds (mean and standard deviation) in some solifluction deposits, together with age estimates: (a) based on a linear rate of rind development over 10ka for the rinds in the till of the Cynthia Bay phase; (b) assuming rind development to be a square root function of time (k=5.8). See text for explanation.

prior to 7,650 +- 250 BP (SUA 2079) and that alpine shrubs were well established on the Gould Plateau by 7920 +- 250 BP (SUA 2080). The deposition of minerogenic silt was in progress at Laughing Jack Lagoon before 1,540 +- 60 BP (SUA 1958). Organic silts that overly the minerogenic sediments were deposited after 810 +- 60 ka BP (SUA 1957). These dates, though scattered through the Holocene, merely serve to indicate the types and limited nature of Holocene sedimentation when compared to the more rigorous conditions of the Pleistocene.

F Small scale periglacial activity

Small scale periglacial landforms in the study area include solifluction terraces, patterned ground, nivation hollows and subnival boulder pavements.

The small solifluction terraces that occur on Mt. Rufus and on Mt. King William I are presently active. Small sorted stone nets have developed at 1000 m altitude on Mt. King William I on slopes that were devegetated no earlier than 1978. Similar forms occur on the tread of solifluction terraces above Lake George, on Mt. Rufus and in the Traveller Range. None of this patterned ground lies below 1000 m alhough the formation of pipkrake has been recorded to elevations as low as 680 m.

Most of the small nivation hollows that occur on many of the higher mountains are vegetated which indicates that nivation is responsible for little erosion at the present time. Unvegetated nivation hollows are present in a few cases as on Mt. King William I where one is cut into a sheet of solifluction deposits just outside the ice limit of the Cynthia Bay phase. At this site short grooves occur upslope of 10 cm. clasts on the floor of the hollow. These indicate that basal shear stress beneath the seasonal snowpack is sufficient to move small clasts down a slope of 10 degrees under present day conditions.

3 Climatic change through the late Cainozoic

The glacial history of this area can be considered most profitably in the context of information that is available from other parts of Tasmania. The post-depositional characteristics of the glacial deposits in this area suggest that they fall into essentially three groups, namely an ancient deeply weathered drift, a less weathered complex that is intermediate in age, and a young drift that is only lightly weathered. This is consistent with the pattern that has now been established in several Tasmanian valleys (Kiernan, 1983a).

Because of the complexity of the intermediate complex and the difficulties that exist in correlation it may be premature to formally assign it a formation name. Hence, while it is accepted as representing at least one glaciation, in this thesis it is assigned the status of a weathering zone (Boyer, 1972; Pheasant and Andrews, 1972). Weathering zones are "surficial rock units delimited from one another on the criteria of the degree of surface physical and chemical weathering exhibited on bedrock, on till, and other surficial material. Weathering zones also include certain mophological elements related to weathering, such as the presence of tors " (Boyer and Pheasant, 1974). This approach is valid under the conventions of the International Subcommission on Stratigraphic Classification which defines "zone" as "...a general term which may be used in any kind of stratigraphic classification....The boundary between two zones is considered to be isochronous " (Hedberg, 1972). The informal assignment of all the Tasmanian tills to weathering zones is useful for comparative purposes.

The Stonehaven Drift is ancient and represents part of zone I. Estimates of the possible age of this zone range from 230 ka to several million years depending upon the weathering rate assumed. Taken together, the weathering evidence from around Tasmania indicates that the tills of this zone must predate 600 ka (Kiernan, 1983a). Palaeomagnetic results from the zone I tills in the Pieman Valley and Linda Valley suggests that they predate the Brunhes-Matuyama reversal (730 ka) (Colhoun and Augustinus 1984; Colhoun, 1985a). If the Nive Till is a correlate of the Stonehaven Drift then it too is of Linda age.

The Butlers Gorge Complex represents zone II. The weathering evidence favours a pre Last Interglacial age for the Beehive, Powers Creek and Clarence tills. The zone II tills in the Pieman Valley predate 43,800 ka BP (SUA 1047) (Colhoun, 1985b). Organic deposits of interglacial type that have been radiocarbon assayed at between 41,900 +1000 -900BP (SUA 2277) and 43,000 +1200 -1100 (SUA 2278) occur between the ice limits defined by the zone II and Zone III tills in the Langdon Creek Valley (E.A.Colhoun, pers. comm.).

The weathering evidence suggests that the Beehive phase may have been contemporaneous with the Plateau glaciation on Ben Lomond (Caine, 1983). However, the exponential rind development equation that forms the main basis for dating the Ben Lomond event appears to underestimate the age of the Beehive Drift and very seriously overestimate the age of the Stonehaven Drift.

The possibility that the climate ameliorated during deposition of the zone II moraine complexes has previously been raised by postdepositional differences and organic evidence from the central West Coast Range (Kiernan, 1980), and by postdepositional differences in the Pieman Valley (Augustinus, 1982) and at Dale Creek on the Western Tiers (Kiernan, 1984) (figure 13.6). However, this study of the glacial deposits in the Central Highlands has failed to reveal any critical evidence regarding this question.

The Cynthia Bay tills form part of zone III which includes the deposits of the Dante (Margaret) Glaciation in the central West Coast Range. Driftwood in proglacial silts beneath the outwash gravels of Dante age in the King Valley has been radiocarbon assayed at 18,800 +- 500 BP (ANU 2533) (Kiernan, 1980). Two further assays have now been obtained from the same section (figures 2.1a & 2.1b). Part of an intact specimen of the cushion plant Donatia novae zelandiae

Criterion	CENTRAL WEST COAST RANGE (Kiernan, 1980)	PIEMAN VALLEY (Augustinus, 1982)	WESTERN TIERS (Kiernan, 1984)
	(1) Comstock (6 sites)	(1) Boco II	(1) Dale Ck. Member (3 sites)
dolerite weathering rinds (max)	10.0mm	19.0mm	16.2mm
chert weathering rinds	-	2.3mm	-
percentage absorption of clasts	-	3.5-6.5mm	-
	- radiocarbon age of wood in palaeosol between two tills of Comstock age in Queen Valley is 30,050 ± 200BP (ANU2535). Wood is severely discoloured by younger humic acid contaminants and assay can be interpreted as a minimum age. Associated pollen taxa have a high alpine/sub-alpine to rainforest ratio - interstadial?		
	(2) Lower Linda (2 sites)	(2) Boco I	(2) Dale Ck. Member (2 sites)
dolerite weathering rinds (max)	21.0mm	40.0mm	38.0mm
chert weathering rinds	-	7.3mm	-
percentage absorption of clasts	-	6.0-12.0mm	-

Table 13.6 Evidence from three other Tasmanian sites for possible multiple glacial events during the deposition of the zone II moraine complexes.

that occurs on a weakly developed palaeosol beneath the proglacial silts has been assayed at 21,180 +- 370 BP (SUA 2154). Branchlets from silts that probably underlie the palaeosol in another part of the section have been assayed at 20,100 +- 470 BP (SUA 2155). These assays indicate that the maximum of the Dante Glaciation, and by analogy the Cynthia Bay event, probably postdates 19 ka BP.

No evidence of the age of deglaciation has been obtained. The earliest deglaciation date that has been reported from Tasmania comes from the Ooze Lake Cirque at 880 m in the Southern Ranges where charcoal at the base of lake deposits has been assayed at 17,700 +- 400 BP (SUA 1359). However, the inwashing of older charcoal from surrounding slopes may have resulted in an excessively early date being obtained (Macphail and Colhoun, 1985). Radiocarbon assay of charcoal in presumed cryoclastic roof-fall rubble in Kutikina Cave on the lower Franklin River indicates that it accumulated during cold climatic conditions between 19,700 +- 850 BP (ANU 2785) and 14,840 +- 930 BP (ANU 2781) (Kiernan et al., 1983). Notwithstanding the possibility that aborigines may have somehow accentuated the roof-fall activity these dates probably broadly define the period of most intense cold.

The radiocarbon assay of 13,000 +- 640 BP (SUA 1959) obtained from charcoal beneath the upper slope mantle on Mt. Arrowsmith indicates that slopes remained unstable, possibly in response to sparse vegetation as aridity increased (Macphail and Colhoun, 1985) or to the impact of aboriginal burning (Kiernan et al., 1983). The dating of this mantle bears comparison with the assay of 11,200 +- 700 BP (Gak 186) on charcoal from beneath a similar mantle at 500 m in the Florentine Valley, 40 km to the south (Davies, 1974).

A number of radiocarbon dates from rock basins in western and south central Tasmania indicate that deglaciation may not have been complete at low elevations until about 11 ka BP and that the highest cirques may not have been ice free until 10 ka BP (Macphail and Peterson, 1975; Colhoun, 1979; Macphail and Colhoun, 1982). All of these results are based on the earliest organic deposits in the rock basins and may therefore give an excessively young estimate of the age of final deglaciation. However, if final deglaciation did occur well after the most intense phase of glacial cold had come to an end, it raises the possibility that climate amelioration may have occurred in two stages (Berger, 1985).

With the cessation of cold glacial conditions slopes were stabilised by vegetation, runoff was slowed, predominantly mechanical weathering gave way to predominantly chemical weathering and the calibre of material being transported by running water declined dramatically (Colhoun, 1982). Macphail (1979) proposed that a climatic optimum occurred from 8 - 5 ka BP after which the pollen record suggests a change to cooler conditions. This interpretation is complicated by the difficulty in separating temperature from precipitation affects. A stalactite from Lynds Cave in the Mersey Valley has provided an oxygen isotope record for the period 12.6 - 2.8 ka BP which should respond only to changes in temperature. This suggests a temperature variation of 4° C during the Holocene. It suggests that temperatures were up to 1° C higher than present from 12 - 9.3 ka BP and have been as much as 3° C lower than present since that time (Goede and Hitchman, 1983).

Derbyshire (1967) proposed that the incision of immature cirques such as Sunrise Hollow into the backwall of older cirques, and the development of small moraines rich in protalus may have been attributable to neoglacial cold. He also argued that a phase of slope instability occurred after 2,900 +- 80 BP (Gak 1020) at Monpeelyata Canal near Lake Echo (890 m). He attributed the presence of leaves of <u>Astelia alpina</u> in the section to colder temperatures than at present (Derbyshire, 1972). Macphail and Hope (1985) have claimed that an apparent resurgence in mire development after about 3.5 ka BP at Browns Marsh near Lake Echo resulted from a regional decrease in mean temperature.

However, Sunrise Hollow and the protalus accumulations now appear to date from the Cynthia Bay Glaciation. Colder temperatures cannot be interpreted from the presence of <u>Astelia alpina</u> in the section at Lake Echo (890 m). The distribution of this plant is dependant upon moisture as well as temperature. It has been found growing at 710 m during the course of this study, and has been recorded as low as 460 m (J.B.Kirkpatrick, pers. comm.). The dating of this section and the origin of the upper slope deposit is also in question. A further assay of 30,400 +- 2,300 BP (Gak 1163) has been obtained from charcoal fragments in a palaeosol at the same level in the section as the Astelia alpina leaves (Derbyshire, 1968b; Colhoun, 1985b). A number of other workers have proposed that slopes in various parts of Tasmania were unstable during the Holocene (Goede, 1965; Caine, 1968b, 1978; Wasson, 1977). However, none of the evidence which has been put forward neccessarily demands the onset of colder climatic conditions. All could be interpretted as a response to storm events, wildfires or aboriginal burning.

At least some of the solifluction terraces on Mt. Rufus are active to a minor extent under present day conditions. Hence, these terraces, and similar phenomena on other Tasmanian mountains are not neccessarily relict forms. Disturbance of the landscape by periglacial processes could have resulted from disruption of the vegetation cover rather than from the onset of colder conditions. Evidence from the mountains of west central Tasmania indicates that humankind is the most significant geomorphic agent of the Holocene interglacial.

4 Regional comparisons

The climates of southeastern Australia, New Zealand and Patagonia are likely to have responded to broadly similar hemispheric influences during the late Cainozoic (figure 1.1). Therefore, some consideration may be given to the possibility of comparing major events from the diverse regions.

Intense and sudden cooling occurred during the latest

Miocene (Kemp, 1978). This is reflected in Tasmanian marine deposits (Gill, 1961). Glaciers developed in the Transantarctic and Gamburtsev mountains in Antarctica during the Oligocene and an ice sheet had developed in Antarctica by the start of the Miocene (Andrews, 1979). The earliest glacial deposits outside Antarctica occur in the Mesata del Lago Buenos Aires in Argentine Patagonia where tills are sandwiched between basalts that have been K/Ar assayed at 7 ma. and 4.6 ma BP. The tills are therefore of late Miocene or early Pliocene age (Mercer and Sutter, 1982). The ice cover in the Antarctic region expanded about 3.7 - 3.35 ma BP after which glaciation became widespread in mountains around the world (Hambrey and Harland, 1981) There appear to have been further glaciations in Patagonia at 3.5 ma and 1 - 1.2 ma BP. (Mercer _et al._, 1975; Mercer, 1976). In New Zealand the earliest glaciation, the Ross Glaciation, may date from the late Tertiary (Gage, 1961; Bowen,1967; Fleming,1975).

The age of the earliest Tasmanian glaciation is uncertain. If Colhoun (1975)is correct in his interpretation of the Lemonthyne sequence and the diamicton is a tillite then glaciers were present in Tasmania during the late Tertiary and some of the now extinct Tertiary flora withstood at least one stage of intense glacial cold.

Pollen collected by the writer from non-glacial sediments which are overlain by the moraines in the Linda Valley (zone I) indicates a rainforest vegetation that included some species that are believed to have been extinct since the

Tertiary (M.K.Macphail, pers. comm.). The Bulgobac moraines in the Pieman Valley and the Linda moraines in the central West Coast Range antedate the Brunhes - Matuyama reversal (730 ka BP) (Colhoun & Augustinus, 1984; Colhoun, 1985a). The Stonehaven Glaciation may be broadly equivalent in age to the Porikan Glaciation in New Zealand which probably occurred about 750 ka BP (Mildenhall and Suggate, 1981). On the other hand it may be very much older. The Nive Till has only tentatively been interpreted as the basal facies of the Stonehaven Drift. It is as lithified as the Lemonthyne diamicton and can be described as a tillite. The writer is unaware of any proven Quaternary tills that have been lithified to tillite.

It has long been suggested that the intermediate (zone II) drifts in Tasmania represent more than a single phase of glaciation (Lewis, 1945; Kiernan 1983a)(see figure 2.5 and table 13.6). The deep sea record indicates that a succession of glaciations have occurred at intervals which average about 100 ka during the last 800 ka (Shackleton and Opdyke, 1973; Emiliani and Shackleton, 1974). Thus the intermediate drifts could represent one glaciation with several major stages or separate glaciations of early and middle Quaternary age.

Comparison of the post-depositional characteristics of the Beehive and Cynthia Bay drifts suggests that the time period between their deposition was considerably longer than the Holocene. This implies a pre-Last Interglacial age for the Beehive phase. An early Last Glacial advance that has been proposed in New Zealand has yet to be substantiated by the acquisition of definitive organic evidence between the mapped early Kumara and late Kumara glacial limits (Gage and Suggate, 1958; Suggate, 1965; Suggate and Moar, 1970; Burrows, 1978). The early Llanquihe Glaciation in Patagonia has also been claimed to be of early Last Glacial age, but this interpretation rests primarily on radiocarbon assays that are so close to the maximum limit of radiocarbon dating as to be convertible from an infinite result to a finite result by the addition of as little as 1 per mil of modern carbon (Porter, 1981).

The Powers Creek and Clarence drifts are very much more weathered than the Beehive Drift and may be roughy equivalent to the Waimean and possibly even Waimaungan deposits of New Zealand (Gage, 1979) or to the early Llanquihe and pre Llanquihe deposits of Andean Patagonia (Mercer, 1976; Heusser and Flint, 1977; Porter, 1981). But precise correlation of either age or sequence is not yet possible.

The Cynthia Bay Drift (zone III) was deposited during the Margaret Glaciation. If the period of most intense cold during the late Last Glacial Stage in Tasmania was from about 19 - 15 ka. BP the record from the southeastern Australian highlands and the Kosciusko range in particular ought to reflect this. A radiocarbon assay of 35,200 +1600 - 2150 BP (ANU 76) was obtained by Caine and Jennings (1968) from a stump beneath a block stream in the Toolong Range. It suggests climatic cooling after this time. A radiocarbon result of 20,200 +- 165 BP (NZ 435) that was obtained on organic deposits from a cirque at 1980 m. on Mt. Twynam has been assumed to indicate that deglaciation was complete by that time (Costin, 1972). However, this result is now known to be incorrect due to a laboratory error (J.I. Raine, pers. comm.). Two assays of 12,920 +- 470 BP (ANU 2679) and 11,180 +- 440 BP (ANU 2680) have been obtained from close to the original sampling position in the same profile. A further assay of 15,540 +- 420 BP (SUA 272) has also been obtained from this site (Martin and Polach, 1983; A. Martin, pers. comm.). In addition, a wood fragment overlying a diamicton at 1830 m in the headwaters of the Snowy River has yielded a result of 15,000 +- 350 BP (NZ 399)(Costin, 1972). These results suggest that deglaciation of the Australian mainland was not complete before 15 ka. BP.

The Tasmanian evidence corresponds moderately well with the evidence from New Zealand where the Kumara 2(2) advance during the Otira Glaciation occurred between about 22.3 and 18.5 ka. BP. (Suggate and Moar, 1970). It corresponds even better with the record from Patagonia where the late Llanquihe maximum is dated to 19.4 - 17.3 ka BP. (Mercer, 1976). The evidence for the maintenance of vigorous mechanical weathering and slope instability in Tasmania until 14 - 13 ka BP bears comparison with the record from New Zealand where the last main Pleistocene glacial event terminated about 14 ka. BP, although there were significant events during the Holocene (Burrows and Gellatly, 1982).

Active glaciers remain extant in New Zealand and Patagonia. The southeastern highlands of the Australian mainland were deglaciated no later than 12 ka BP and Tasmania was deglaciated no later than 10 ka BP. (Macphail and Peterson, 1975). A series of minor glacial advances occurred in New Zealand throughout the Holocene. Several of these occurred after 4.2 ka BP (Burrows, 1979; Burrows and Gellatly, 1982; Gellatly, 1984). Advances during the late Holocene have also been reported from Patagonia (Mercer, 1970; Heusser, 1974). Williams (1978) has argued that slopes in the Southern Tablelands of New South Wales were unstable from 4 - 1.5 ka BP. Costin (1972) has proposed that slope instability in the Snowy Range indicates that temperatures were 3^{o} C colder than present from 3 - 1.5 ka BP. The speleothem studies of Goede and Hitchman (1983) suport Costin's proposition.

However, the geomorphic evidence from Tasmania is equivocal and unconvincing. If climatic cooling did occur during the Holocene, it does not appear to have been sufficient to produce geomorphic effects in this part of the Tasmanian Central Highlands. Small scale periglacial activity on these mountains is permissive of a deterioration of climate during the late Holocene, but does not demand it.

chapter fourteen

PALAEOGLACIOLOGY AND PALAEOCLIMATOLOGY

The spatial relationships between the alpine landforms and deposits indicate that a very extensive ice cover has at times previously existed in the mountains and valleys of Tasmania's Central Highlands. This chapter presents an integrated map of the maximum ice cover during the main glacial events (Table 13.1 and 14.1, figure 14.1). It discusses the implications of the maximum ice cover in the Central Highlands; defines the patterns of ice movement; and addresses the environmental nature of the glaciations.

EXTENT OF THE ICE COVER AND THE PATTERNS OF GLACIER FLOW

A The maximum ice cover of the Stonehaven Glaciation.

While deposits of the Collingwood and Stonehaven phases define the maximum ice cover in the west, younger drifts define the easternmost margin of the ice. This probably does not imply that the earlier ice did not extend as far eastwards as the later ice. Rather, it suggests that less abundant deposits were laid down by the slightly continental glaciers of the east and that these deposits may have succumbed more rapidly to redistribution by periglacial and other processes. The glacier limits that have been recognised in the east are therefore likely to be conservative.

The ice-abraded summit rocks of Walled Mountain (figure 4.10) indicate that the ice was at least 600 m thick in the adjacent tributaries of the Murchison River. The rubble summit block of Mt.Gould indicates a maximum ice thickness of 420 m in the Cephissus Trough. This implies an ice surface that declined SSE at 33 m/km between Mt. Gould and Walled Mountain. This is confirmed by the erosional morphology of the Parthenon (1200 m) which lies on the Derwent- Murchison divide and has been over-ridden by ice from the north that left erratics near its summit.

The ice surface also declined towards the east. The Narcissus and Cephissus glaciers were confluent through the col between Mt. Geryon and the Acropolis. The eastern ridge of the Acropolis has not been over-ridden which suggests that the ice in the col was no more than 100 m thick and that its surface sloped downwards into the Narcissus Trough. The ice in the head of the trough immediately adjacent to the ridge cannot have been more than 400 m thick.

While some parts of the Narcissus headwall may have been over-ridden by ice at the maximum phase the northern Du Cane Range generally formed a major divide between ice that flowed northwards into the Mersey and ice that flowed southwards into the Derwent. Jennings and Ahmad (1957) have shown that the ice divide on the Central Plateau lay approximately in the position of the present fluvial divide. Some of the ice that flowed northwards into the Mersey spilled through Du Cane Gap into the Narcissus valley head, while plateau ice spilled directly into the Lake St. Clair trough along almost the entire margin of the Traveller Range.

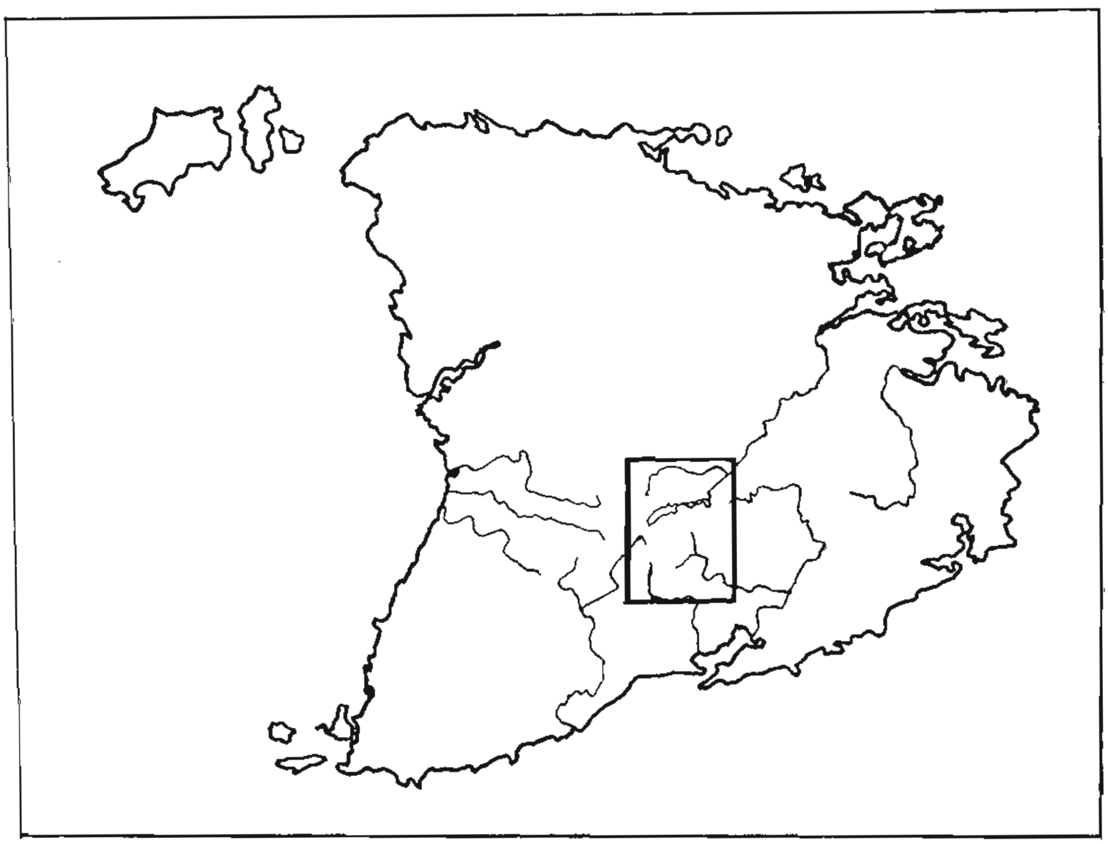
VALLEY	EXTENT OF ICE COVER (km^2)				
phase:	Cynthia Bay	Beehive	Powers Creek	Clarence	Stonehaven
St.Clair	92	100	100	100	100
Cuvier-Hugel	41	48	50	50	50
Traveller	45	72	72	72	72
Navarre	6	23	23	23	23
Guelph	17	61	91	95	100
Derwent Bridge	-	6	124	124	180
Eastern Nive	114	125	216	> 238	> 262
Upper Gordon	15	27	> 37	> 40	> 40
Franklin	72	101	115	187	326
Total extent in Derwent Basin	201	310	460	464	520
TOTAL EXTENT	402	563	> 828	> 929	> 1153

Table 14.1 Extent of the ice cover during the principal phases of glaciation. The figure for the Stonehaven phase is based on the ice limit implied by the Collingwood erratics. Further ice existed in the Nive, Gordon and Franklin valleys outside the area studied in this thesis.

The Cephissus Glacier was deflected southwards around the western side of the Mt. Olympus nunatak by the Narcissus Glacier. It spilled into the Cuvier Valley via the Byron and Cuvier gaps. Ice from the Cuvier spilled in turn into the heads of the Alma and Franklin valleys. Derbyshire (1972) has argued that the ice at the head of the Cuvier was 425 m. thick but Mt. Byron does not appear to have been abraded more than 320 m. above the valley floor. The passage of 100 m. of ice out of the Cuvier Valley via the Olympus Col implies that the Cuvier Glacier was 320 m. thick at that point. Clasts of Nive Till that occur in a later moraine at 1340 m in the East Rufus Cirque indicate that a glacier at least 300 m thick existed there.

The ice probably extended eastwards across the Central Plateau to Pine Tier Lagoon. On the St. Clair Surface it probably reached close to the Clarence - Nive confluence. Independant glaciers would have arisen on the Wentworth Hills, King William Range, Mt. Rufus, Cheyne Range, Loddon Range and other ranges to the west. The Derwent Glacier probably extended nearly 70 km from its headwall in the Du Cane Range. It may have been a narrow ice tongue in the Derwent Gorge over its final 10 kms. Strong east-west asymmetry of the ice surface would have been maintained by the entry of the Guelph Glacier and possibly also some ice from the Hobhouse Range.

An ice thickness in excess of 300 m was attained in the glacial troughs of the King William Range. The Derwent



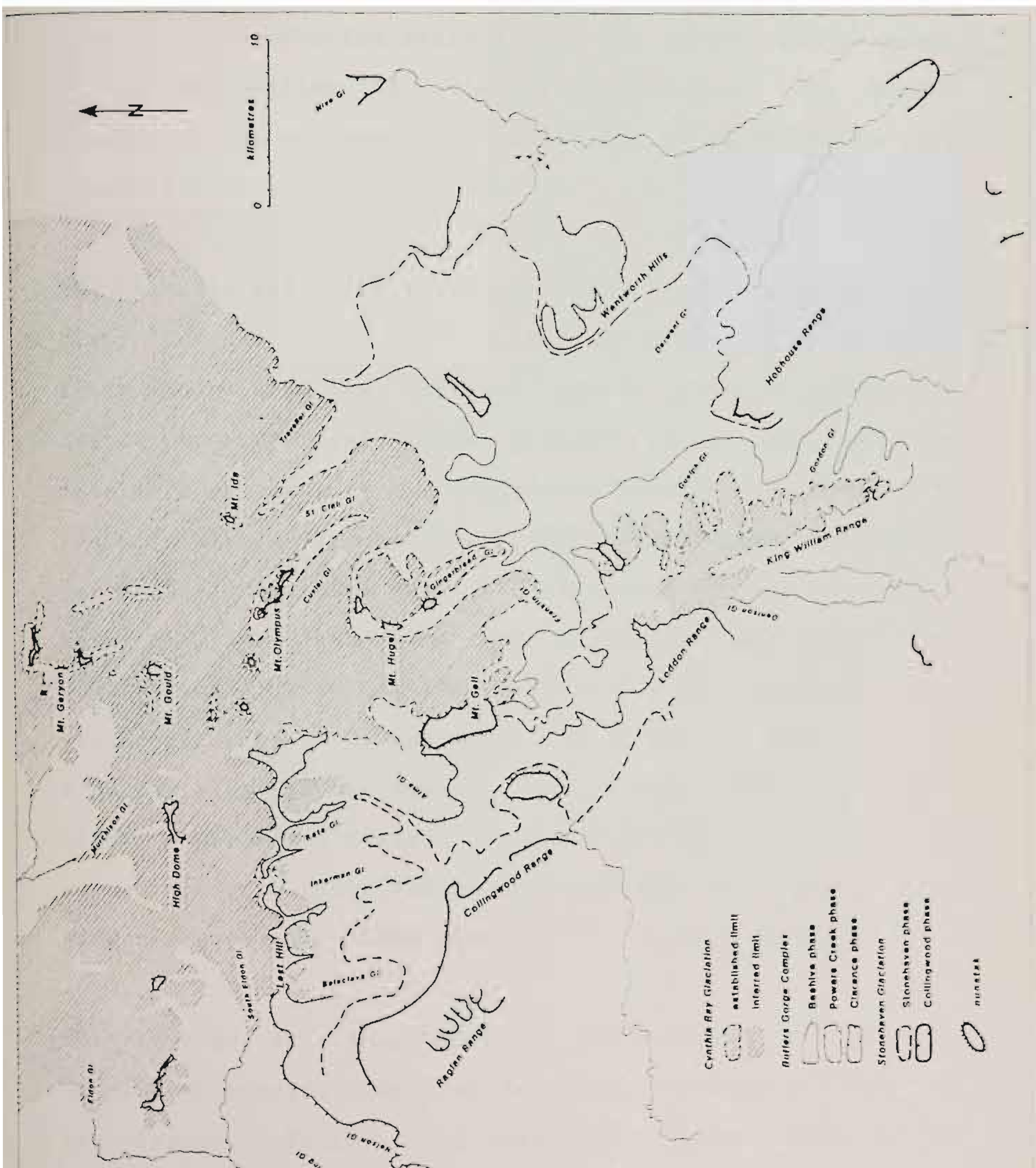


Figure 14.1 Ice limits during the principal glacial events. Ice limits in far west after Kiernan (1980).

Glacier deflected much of the Guelph ice southwards into the head of the Gordon Valley where it merged with the glaciers that formed in the southern part of the King William Range. Some of the Gordon ice spilled into the Denison Valley south of Mt. King William III. The maximum limits of the Gordon Glacier have not been ascertained but it is likely to have extended far to the south.

The Franklin and Derwent glaciers were confluent on the St. Clair Surface south of Mt. Rufus. A glacier up to 500 m thick flowed down the Surprise Valley from a major ice divide at its head. Other glaciers flowed east and south from this divide into the Guelph and Denison valleys. The Franklin Glacier reached the Franklin-Collingwood confluence and probably extended further. It merged at this confluence with ice from the Alma and Collingwood valleys. The Collingwood Glacier was 150 m. thick 8.5 km upstream from the confluence. It was nourished in part by diffluent ice from the King Glacier system that flowed over the South Eldon - Balaclava divide. An ice thickness of at least 440 m. in the South Eldon Valley north of this point would have been neccessary for this divide to be swamped by ice.

Ice covered at least 520 km^2 in the Derwent Valley. The Stonehaven moraines imply an ice cover of about 275 km^2 in the Franklin Valley, but the ice limits implied by the Collingwood erratics suggest that ice probably covered nearly 330 km^2 during the maxinum phase. Much more fieldwork is required to establish the ice limits in the Nive Valley, but if a glacier reached Pine Tier Lagoon then several hundred square kilometres of the upper Nive catchment must have been covered by ice.

B The Butlers Gorge Complex

(i) The Clarence phase (Figure 14.1)

Ice streams from the transection glacier in the highlands flowed down the Franklin Valley to Wombat Glen and in the east extended to within 5 km of the confluence of the Clarence and Nive rivers. This glaciation was more extensive than any which occurred after it.

The thickness of the glaciers at this time is difficult to determine but from the gradient of the Clarence Col lobe it is unlikely to have been less than 300 m thick at the southern end of Lake St. Clair. The ice cover in the Derwent Valley totalled 460 km^2. A comparable area of the adjacent Nive, Upper Gordon and Franklin catchments was also covered by ice.

The broad patterns of ice flow were similar to those of the Stonehaven Glaciation but with two important differences. Firstly, the ice in the eastern part of the Nive catchment was dominated by diffluent lobes of the Derwent Glacier, whereas the lithology of the Nive Till which underlies the Clarence drift suggests that during the Stonehaven phase most of the ice here came from the Central Plateau. The second major difference was that diffluent ice from the King Valley no longer spilled into the Balaclava Valley.

The maximum length of the Derwent Glacier was at least 54 km. The Clarence lobe was at least 160 m thick south of Clarence Lagoon and its surface declined eastwards at about 20 m/km. The Franklin and Alma glaciers were respectively 19 km and 16 km in length.

(ii) Powers Creek phase (Figure 14.1)

During this event the Derwent Glacier probably extended as far as the headwaters of Mossy Marsh Ck. Distributary lobes once again extended through the Clarence and Laughing Jack cols. The ice surface gradient of the final 8 km of the Derwent Glacier was about 15 m/km. Ice from the Guelph would once again have been deflected into the upper Gordon however it probably extended no further than 7 km and had a very steep gradient to the south of about 70 m/km. The ice cover in the Derwent Valley totalled 460 km^2. The total ice cover in the study area was at least 830 km^2.

Ice spilled over King William Saddle and extended some distance down the upper Franklin gorge. Difluent ice from the Derwent Glacier System is also likely to have extended well down the Alma Valley.

(iii) Beehive phase (Figure 14.1)

The ice cover during the Beehive phase was far more restricted than that during the earlier events although the patterns of ice flow were broadly similar. The Derwent and Franklin glaciers formed discrete piedmont lobes on the St. Clair Surface. The Derwent Glacier was divided into two lobes by the Bedlam Wall ridge. The eastern lobe was contained by the Clarence Col for all but the earliest part of this event. The western lobe spread out onto the Navarre Plains.

The maximum length of the Derwent Glacier was 37 km. The Franklin Glacier was only 15 km long. The glaciers of this time were characterised by ice surface gradients that declined steeply eastwards. This is evident from the spacing of the retreat moraines on either side of Bedlam Wall (Derbyshire, 1971a). The ice cover in the Derwent Valley totalled 310 km^2 and about 250 km^2 in the adjacent valleys was also covered by ice.

C Cynthia Bay Glaciation (Figure 14.1)

The trend towards shorter and steeper glaciers was continued during the Cynthia Bay Glaciation. The Derwent Glacier terminated 1.2 km southeast of the present shoreline of Lake St. Clair and therefore had a maximum length of 30 km. The longitudinal surface gradient over its final 16 km was about 25 m/km. The Franklin Glacier at this time was no longer than 11 km. Glaciers extended over 200 km^2 of the Upper Derwent Valley and the ice in the adjacent Nive, Upper Gordon and Franklin Valleys covered a further 200km^2.

The intervention of diffluence cols resulted in the glaciers in many valleys retreating out of phase with one another. A major retreat of more than 6km occurred in the Cuvier Valley after the flow of ice through the Byron and Cuvier gaps was severed. This led in turn to retreat of the Franklin Glacier from the Lake Dixon limit as diffluence from the Cuvier into the Franklin also came to a halt. The emergence of Du Cane Gap through the downwasting ice surface probably played a significant role in retreat of the Narcissus Glacier. In the Guelph area the Top End Glacier diminished from being the largest of the main glaciers to the smallest after the Top End Gap emerged through the downwasting and strongly asymmetric Surprise Glacier.

Despite the loss of the ice from the Central Plateau the Narcissus Glacier persisted for some time in a vigorous condition. This suggests that the Central Plateau was secondary to the valley head as a source of ice. The general lack of hummocky moraine (Derbyshire, 1967) and the presence of a large tract of fluted drift indicates that the ice remained highly active even during final retreat. Persistance of ice in the valleys and the incision of immature cirques such as Sunrise Hollow into larger pre-existing headwalls supports the contention of Derbyshire that the glaciers became increasingly dependant upon shading and wind-drift accumulation.

2. EMERGING PATTERNS OF TASMANIAN GLACIATION

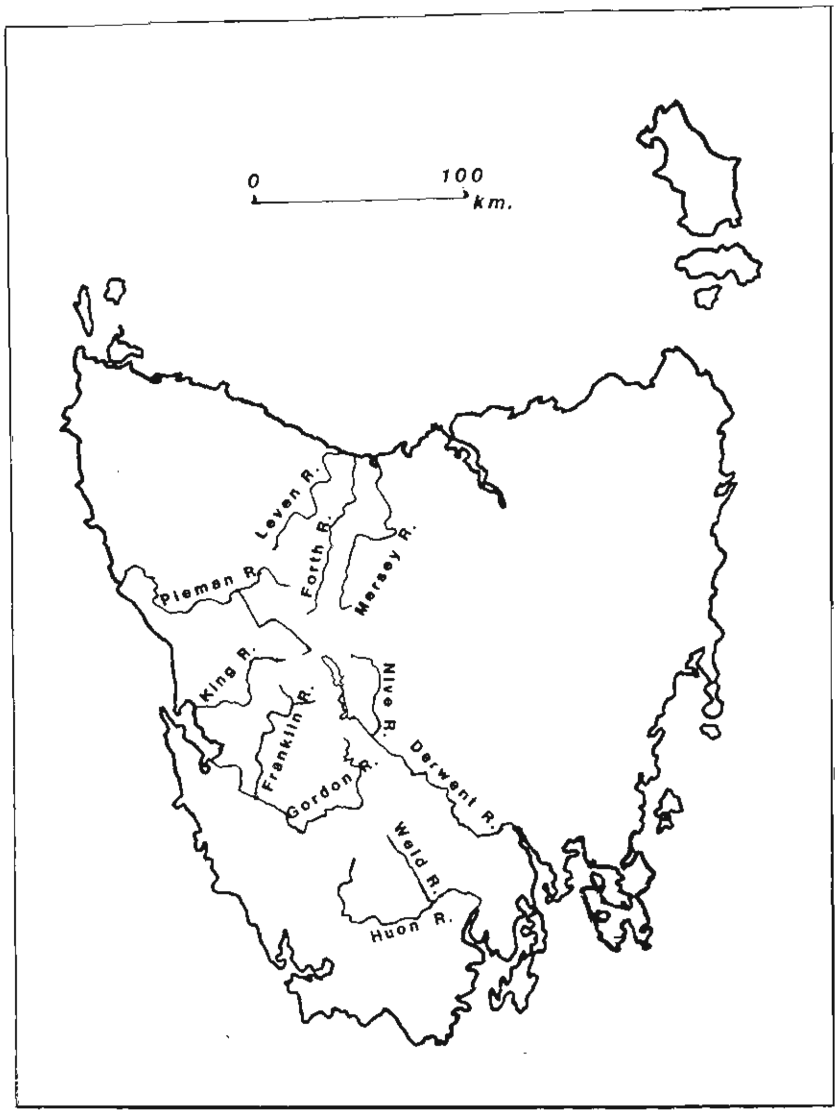
Evidence presented in the preceeding section indicates that the thickest ice cover lay near Walled Mountain just west of the present Pieman - Derwent divide and that the ice surface gradient declined to the south and east. This evidence supports the suggestion (Davies, 1969 figure 84; Derbyshire, 1972) that the upper Murchison area may have been completely inundated by ice. It adds to the growing evidence (Sansom, 1978; Colhoun, 1979, 1985a; Augustinus, 1982) that the ice sheet in the Central Highlands and that in the West Coast

Range were probably confluent (Figure 14.2; cf. Figure 2.1). It suggests that the central Tasmanian ice cap was much thicker west of the Du Cane Range than anywhere to the east.

This suggests that an ice cap of over 5000 km^2 extended continuously between the mountains of central and western Tasmania. This ice cap was drained by a series of distributary glaciers. The known limits of the Pieman Glacier lie 55 km WNW (Augustinus, 1982; Colhoun and Augustinus, 1984). The ice in the upper Murchison was sufficiently thick to over-ride the Murchison - South Eldon divide and form part of the King Glacier. This glacier also drained the West Coast Range ice cap and during the Comstock Glaciations (= Butlers Gorge Complex?) merged with ice from Mt. Jukes close to the King - Andrew, divide 45 km southwest of Walled Mountain . This divide may have been over-ridden during the Linda Glaciation (Kiernan, 1980)

Evidence presented in this thesis indicates that diffluent ice from the King Glacier merged with the Franklin Glacier near the present Collingwood - Franklin confluence. A substantial proportion of the ice in this area extended into the Franklin Valley from the Derwent Glacier via the Alma and Franklin cols. In excess of 100 m of ice reached the mouth of the Franklin Gorge upstream of Frenchmans Cap and the possibility of confluence with the glaciers that descended northwards from that massif remains open.

Equivocal erosional evidence suggests that parts of the Du



0
100
km.
Leven R.
Forth R.
Mersey R.
Pieman R.
King R.
Franklin R.
Nive R.
Derwent R.
Gordon R.
Weld R.
Huon R.

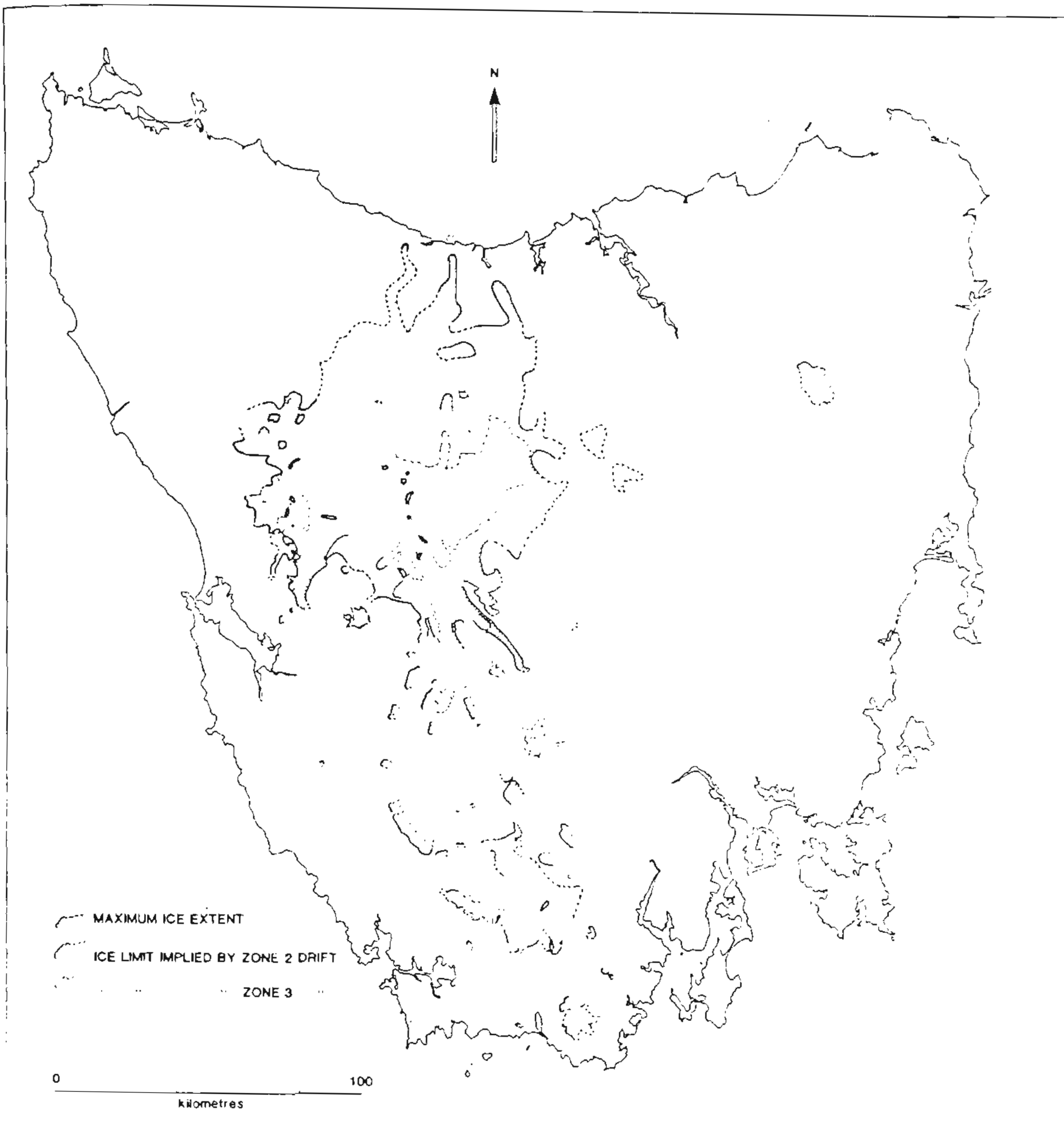


Figure 14.2 .Extent of late Cainozoic glaciers. Ice limits in the far south are based on Colhoun and Goede (1979), Kiernan (1982b, 1983a) and unpublished data.

Cane Range may have been over-ridden from the north, but it is probable that for most of the Pleistocene the northern part of the range formed a major divide between ice that extended 81 km northwards via the Forth and Mersey Valleys (Colhoun 1976b; Kiernan, 1982, 1983a, 1984; Hannon, in prep.) and the Derwent Glacier that extended at least 70 km. to the south. The Mersey Glacier appears to have been 10-15% longer than the Derwent Glacier during the main phases of glaciation. Part of the reason for this may lie in the existence of the col between Mt. Massif and Mt. Ossa through which the western ice-cap was able to discharge into the Mersey Valley. This col lies 170 m below the lowest col through which the western ice could discharge into the Derwent Valley. The Derwent Glacier was supplemented by the smaller ice-cap that developed on the Central Plateau. The Derwent Glacier was also joined by ice from the King William Range that was confluent with glaciers in the Surprise, upper Denison and upper Gordon valleys.

Despite much easier access than in the west and widespread excavation during hydro-electric construction activities the maximum ice limits to the east are more difficult to discern. Glacial deposits occur 42 km southeast from the Narcissus headwall in the lower Clarence Valley where a lobe of the Derwent Glacier was confluent with local ice that arose in the Wentworth Hills. Fairbridge (1949) has claimed on the basis of the erosional morphology around Victoria Valley that ice extended a further 28 km in this direction but no glacial deposits that support this are known. Nor have glacial deposits been found at the deeply incised

Monpeelyata Canal 50 km east of the Du Cane Range (Derbyshire, 1967).

The evidence of glacial activity is equally patchy higher on the Central Plateau. Till close to the western shoreline of Great Lake (Derbyshire, 1968c; Kiernan, 1983a) indicates that the Central Plateau ice reached at least 50 km west of the Du Cane Range but other till north-east of the lake appears to have been depositied by local glaciers (Banks, 1973). Dense grey basal till very similar to the Nive Till is present on the plateau near Lake Mackenzie 36 km north of the Du Cane ice divide and it is possible that much of the ice that reached further down the Mersey may have originated on the plateau near here.

The abundant glacial deposits that occur in the west contrast strikingly with the paucity of deposits in the east, and suggest a sluggish glacier regimen inland (Derbyshire, 1972; Davies, 1974). Deeply weathered tills define the maximum ice limits in the west but not in the east. This could be due to the younger ice limits having over-stepped the older ice limits in the east. However, this seems unlikely as the earlier ice limits are much more extensive along all the other margins of the ice cap. The fact that the deposits which define the limits to the north and south are also well weathered means that the limited weatherng in the east is unlikely to be the result of climatic differences between these areas since deposition of the drift.. Alternatively, if few glacial sediments were deposited in the east they could easily have been reworked or buried during later periods of periglacial activity. This suggestion is supported by the presence of solifluctcd and weathered till that is overlain by more recent solifluction deposits close to the western shoreline of Great Lake (Derbyshire, 1968c; Kiernan, 1983a). These facts suggest that earlier ice may have extended further eastwards than the remaining deposits indicate.

Derbyshire (1967) noted a broad similarity between those areas of Tasmania that presently receive >250 mm/pa (water equivalent) of snow and those that were glaciated (Figure 2.3). Because the Central Plateau did not fit this pattern he concluded that a mere lowering of the snowline would have been insufficient to account for the distribution of the ice. He therefore proposed that a greater proportion of the solid precipitation budget had been derived from north of west. Because the Plateau no longer appears to be the heart of the ice cap it is no longer neccessary to infer such a major shift in the direction of snow bearing winds to explain the pattern of glaciation.

CHARACTERISTICS OF THE GLACIERS

The importance to glacier formation of snowfences that are aligned across the prevailing westerly airstreams has received considerable emphasis (Derbyshire, 1968a). The concept of a large ice sheet west of the Du Cane Range is inconsistent with this pattern (Derbyshire, 1972). In recent years increasing evidence has become available that suggests glacier formation on the windward flanks of many ranges (Colhoun and Goede, 1979; Kiernan, 1980; Corbett, 1980). Cirques may be better developed on leeward slopes only because ice can persist there under more marginal conditions and hence there has been a much greater duration there of glacial erosion than on the windward slopes.

A transition from maritime glaciers in the western mountains to subtropical-continental type glaciers has been recognised (chapters two and thirteen)(Derbyshire, 1967; Peterson, 1968, 1969). Derbyshire (1972) contended that precipitation and temperature gradients across the area were at least as marked as those that prevail today. Even if this were not the case a general diminution of precipitation eastwards would be inevitable.

Peterson (1968) has observed that the western Tasmanian glaciers were characteristic of the warm infiltration ice formation zone of Shumsky (1964). This zone is characteristic of maritime environments that experience heavy precipitation, small amplitude temperature fluctuations, a relatively warm winter and slight freezing of the ice. Given the lesser solid imput to be anticipated on the more inland mountains compared to the mountains nearer the coast quite different glaciological regimes are implied. Because glaciers in inland areas recieve less precipitation than those in coastal situations continental glaciers are more dependant upon colder temperatures for their survival.

A Glaciological regimes

(i) Cryogenesis

Ice formation by infiltration demands the presence of an unmelted residue of solid precipitation, a heat imput to provide sufficient meltwater to fill the pores in the residue, and a cold reserve or cold influx capable of freezing that meltwater. In the warm ice infiltration zone the heat imput is sufficient to bring the entire active layer to melting point. The abundant precipitation allows considerable firn to accumulate and facilitates very active glaciation. Regelation is the basic process in cryogenesis, and where the solid imput is sufficient compaction plays a major role. Meltwater is important in mechanical rounding, lubrication and packing of snow and firn grains. The mean annual temperature can exceed the melting point by as much as 1.5° C (Shumsky, 1964)

If precipitation diminishes inland warm infiltration may be replaced by infiltration congelation even though air temperature remains the same. Ice formation by congelation requires a sufficient heat influx to cause meltwater runoff or melting of the entire solid imput and a cold reserve or cold influx sufficient to congeal that meltwater. Melting, therefore, does not lead to a loss of ice in all cases. Cryogenesis is much less dependant upon compaction and is slower to occur (Lliboutry, 1956). According to Shumsky (1964) infiltration congelation processes are most prominent in continental areas where the mean annual temperature is a few degrees below freezing and the mean annual temperature of the warmest month is close to 0°C.

Shallow open cirques such as those that occur around Lake St. Clair tend to be formed under conditions of strongly localised wind drift accumulation of snow, high sun angles and low air temperatures (Garcia Sainz, 1949; Derbyshire, 1972). Mechanical damage to wind blown snow dendrites is likely to aid initial snowpack settling and aid in reducing grain surface area (Perla and Martinelli, 1976). Wind-packed snow has a density of 350-400 kg/m^3 whereas the density of undrifted fresh snow is only 50-70 kg/m^3 (Seligman, 1936). However, this is still only half the density that can result from packing by meltwater. Sublimation is likely to be of greater significance in continental areas and to greatly aid cryogenesis since this is the main process involved in initial grain bonding by sintering (Hobbs and Mason, 1964; Perla and Martinelli, 1976). Sublimation will ultimately be inhibited by diminished porosity.

(ii) Ablation processes

In theory the ablation of snow by airflows should be accentuated if the air is derived from sea level and has a high rather than low relative humidity (Leighly, 1949). Under cloudy maritime conditions ablation results mainly from the transfer of heat from the atmosphere (Ahlmann, 1919, 1948). Melting will occur from the surface where ablation results from heating by atmospheric air and condensing water vapour (Shumsky, 1964).

Conditions in more continental situations are likely to be generally colder, cloud cover is likely to be diminished and absolute humidities to be lower. While the infiltration zone is warmed significantly by the infiltration and regelation processes (Maohuan et al., 1982) only a small amount of meltwater may need to be produced before refreezing produces impermeable ice layers that prevent further infiltration (Shumsky, 1964). Radiation is likely to be the predominant factor in ablation in more continental areas (Platt, 1966). Radiant energy will induce melting in the upper 10-20 cm of the snowpack (Shumsky, 1964) but under dry conditions radiant energy is likely to be expended upon sublimation rather than melting. Troll (1942) has argued that under dry conditions in the Chilean and Argentine Andes melting did not occur despite air temperatures as high as 15°C.

Rain is not a major source of ablation. Each gram of water contains only 4.19 joules for each degree above freezing whereas the melting of snow requires 335.0 joules (Perla and Martinelli, 1976). Increased free water would aid in the packing of grains, increase the weight of the snowpack and therefore aid cryogenesis to some extent. Nonetheless, the freezing of 1 g. of water releases enough heat to warm 157.6 g of ice by 1°C. This would accelerate the pace of free water production by solar radiation due to the greater absorbtive capacity of wet snow. Therefore, the effect of rain would be to warm and deepen the infiltration zone. But nocturnal refreezing may produce impermeable horizons that inhibit further infiltration that would warm the snowpack (Shumsky, 1964; Paterson, 1981). However, precipitation during the summer is likely to be in the form of rain rather than snow. While some of this rain may be converted to superimposed ice only a very small proprtion of it is likely to add to the glacier mass (Ostrem et al., 1967)

The similar width to depth ratios of the glaciers and their host cirques in the inland mountains during the Cynthia Bay phase suggests that the form of the glaciers was controlled by the cirques (Graf, 1976). The glaciers typically did not extend far out of the cirques and troughs in which they arose. This general pattern of reliance upon the topography is illustrated by the continued vigour of the Derwent Glacier after retreat of the Central Plateau ice cap had severed the flow of ice into the Derwent via the Ida valleys and Du Cane Gap. However, the ice cover was very much more extensive prior to the Cynthia Bay phase. It is to the landforms eroded by more active ice during these earlier phases of glaciation that the build-up and persistance of much of the later ice must be attributed.

(iii) Inland humidity

Palynological evidence points towards cold and dry conditions during the Last Glacial Maximum (Macphail 1975, 1976, 1979; Colhoun 1985a). The continentality of Tasmania may have been increased as a result of glacial sea level lowering, but the effects of this upon precipitation would have been mitigated by the narrowness of the continental shelf off the west coast. Bowden (1983) has calculated that windspeeds in northeastern Tasmania during the Last Glacial Maximum may have been 30% greater than those that prevail today. Such increased windspeeds may have counteracted somewhat the tendency towards reduced evaporation and precipitation in maritime areas brought about by the lower sea surface temperatures.

The abundant protalus in some of the high cirque moraines suggests that the ice surface was hard (Derbyshire, 1972) and is consistent with limited cloud cover and aridity shortly prior to deglaciation. Caine (1983) has argued that the cessation of frost shattering at Ben Lomond in northeastern Tasmania may have been a response to increasing aridity rather than rising temperatures. However, apparently frost shattered debris ceased to be produced after about 15 ka BP at Kutikina Cave on the lower Franklin River where moisture is never likely to have been a limiting factor. However, because this halt in debris production coincided with the abandonment of the cave by aborigines the possibility exists that the human occupants of the cave contributed somehow to production of the debris (Kiernan _et al._, 1983). Recently it has been suggested from the pollen assemblage at the Ooze Lake cirque in southern Tasmania that the replacement of _Astelia_ bog by _Plantago_ herbfield was good evidence for a period of effectively drier climate between approximately 16.5 - 13.5 ka BP. On the basis of this evidence it has been suggested that the early deglaciation of that site may have been due to increasing aridity (Macphail and Colhoun, 1985). However, _Plantago_ often occurs in very wet situations along watercourses or where snowlie is protracted. Hence, the advent of _Plantago_ herbfield at Ooze Lake does not neccessarily imply a transition to drier conditions.

Derbyshire (1973) has argued that the development of rock glaciers on the windward slopes of Mt. Olympus (1463 m) and Mt. Gell (1432 m) attest locally cold and dry conditions. Rock glaciers that move with the aid of interstitial ice are commonly interpreted to indicate of the presence of mountain permafrost (White, 1976; Barsch, 1977, 1978; Gorbunov, 1978; Washburn, 1979; Caine, 1983). Rock glaciers develop in continental climates where there is high incoming radiation, sublimation, evaporation and little snowfall (Karte and Liedtke, 1981). A snow depth of as little as 25 cm, equivalent to 10 cm of rain that is received in winter, is sufficient to limit permafrost (Krinsley, 1963; Williams, 1975). Karte and Liedtke (1981) consider that rock glaciers will not develop where the mean annual precipitation exceeds 1200 mm. These facts suggest that the windward slopes of Mt. Olympus and Mt. Gell experienced a continental climate with little snowfall or rapid redistribution of snow by strong winds during the Cynthia Bay phase.

Derbyshire (1973) attributed the failure of rock glaciers to develop on Mt. Hugel (1386 m), which lies midway between these two mountains, to the critical minimum altitude for rock glacier formation having lain at 1372 - 1432 m. However this thesis has shown that the rock glaciers on Mt. Olympus postdate an earlier glaciation that over-rode the flank of the mountain. The apparent absence of older rock glacier deposits dating from earlier glaciations at lower elevations elsewhere in the study area hints that conditions may not have been as dry during the earlier episodes, and that the greater ice extent may have been the result of greater precipitation rather than colder temperatures.

High humidity will have a disproportionately great effect upon continental glaciers compared to those in more maritime situations because the increased cloudiness will promote ablation by heat conducted from the atmosphere (Shumsky, 1964). It follows from this that if the more extensive ice cover prior to the Cynthia Bay Glaciation was due solely to greater precipitation the total precipitation budget would have had to have been considerably greater. Because moisture availability is likely to diminish in response to any major decline in sea surface temperature climatic cooling alone seems unlikely to produce large glaciers in the east. From this it would seem that greater precipitation may have reached further inland during the earlier glaciations, acting in concert, perhaps, with slightly cooler temperatures. However, the Cynthia Bay Glaciation was only of very short duration. The weathering evidence suggests that the advances that preceeded it occurred over a much longer time scale. Differences in the duration of the glaciations may therefore account for at least part of the difference in ice cover.

C The thermal regime of the glaciers

The glaciers of western Tasmania were temperate rather than polar in nature (Peterson, 1968). Glacial deposits in the central West Coast Range confirm that the ice bodies there were characterised by rapid mass throughput with high rates of accumulation and ablation, and were analagous to the present temperate maritime glaciers of south island New Zealand and western Patagonia (Kiernan, 1980) (figures 14.3 & 14.4). The evidence from the Lake St. Clair area suggests limited precipitation, colder temperatures and cryogenesis by congelation. This indicates that the thermal regime of these glaciers was different to that further west, but the question arises as to the extent of this difference.

The rounded clasts that occur in the Nive Till in the Clarence valley and on Mt. Rufus indicate that water was abundant when this basal sediment was deposited. This implies that the thermal regime at the ice base was temperate rather than polar. The later glacial deposits are predominantly either terminal or supraglacial in origin and hence do not provide evidence of meltwater at the glacier sole (Dreimanis, 1976). Meltwater channels that are probably subglacial in origin occur near some cirques but need not have been produced during phases when glaciers were most extensive.

The development of moraines of Thule-Baffin type downstream of Lake St. Clair, together with the evidence that localised permafrost existed on Mt. Olympus, hints at conditions having been mildly continental in the upper Derwent Valley during the Cynthia Bay Glaciation. However, while the inland glaciers were more continental in character than those in the western mountains, they were still not of truly continental character. Nonetheless, the survival of

Figure 14.3 The glaciers that developed in western Tasmania were characterised by high rates of accumulation and ablation with rapid mass throughput. Steep ice-falls would have existed where glaciers descended the scarps between preglacial erosion surfaces. (Upper Fox Glacier ice-fall, Westland, South Island New Zealand.)

Figure 14.4 Like the glaciers of the Patagonian Andes and South Island, New Zealand, the glaciers of west-central Tasmania were of temperate maritime type. However, mildly continental conditions probably existed in parts of the Central Highlands where localised permafrost is indicated by the presence of fossil rock glaciers. The largest glaciers were those that flowed from the ice cap that developed west of the Du Cane Range. (Grey Glacier and the South Patagonian ice cap, far southern Chile.)

saprolite on the floor of the Derwent Valley near the Cascade Moraine raises the possibility that the glacier which overrode it may have been cold based (chapter fifteen). This would be consistent with the observation that some ice caps have a firn warmed central region and cold ice in the peripheral zone (Weertman, 1961; Shumsky, 1964). It is conceivable that the Derwent Glacier could have maintained its flow and pressure melting at its sole in response to its gradient further upstream but become colder further downstream as it spread out and became stranded on the St. Clair Surface. However, the survival of the saprolite may be explicable in other ways and all the other evidence hints that temperate thermal conditions are to be anticipated. If cold based ice did occur it was probably only a local phenomenon.

PALAEOTEMPERATURES

The calculation of palaeotemperatures from geomorphic evidence is a longstanding tradition in glacial geomorphology. The real worth of the figures derived is open to serious doubt beause of the difficulty in separating out temperature and precipitation effects (Wright, 1961; Soons, 1979). The most common approach is to estimate the temperature based upon approximations of the equilibrium line altitude (ELA) which are in turn based upon the height of cirque floors or the distribution of depositional landforms.

The observed gradient of cirque floor elevation rises across Tasmania (Davies, 1967; Peterson and Robinson, 1969) broadly parrallel to the present day precipitation gradient. Within the Lake St. Clair area cirque floor elevations have been locally conditioned by preglacial topography and bedrock geometry. As a result cirque floors do not provide a useful estimate of the ELA. Only conservative values will be obtained where glaciers extend beyond their cirques. Valley geometry may also condition glacial limits (Burbank and Fort, 1985). Estimates of the altitude of the firn line of valley glaciers are complicated by the inability to take tributary glaciers into account. Meierding (1982) tested six methods for approximating late Pleistocene ELA's in the Colorado Front Range and concluded that none provided a reliable estimate.

Extrapolation of temperature figures from ELA data entails further difficulties. Wind drifting and topographic interference with the meso-scale climate will be emphasised by snowline lowering, increased albedo and frost (Soons, 1979). Lapse rates are not usually constant with increasing elevation (Cole, 1975). There is no guarantee that present day lapse rates were applicable during the glaciations. Few workers take into account the lowering of the atmospheric envelope in response to sea level decline which of itself would reduce mean temperatures by 0.7-0.9oC.

If a palaeotemperature calculation is based upon the apparent ELA deduced from the altitudinal range of the Derwent Glacier then a snowline at 1020 m and depression of mean annual temperature by 6.2°C is indicated for the maximum of the Cynthia Bay Glaciation. A figure of 6.5°C is indicated by analysis of the Franklin Glacier. These estimates compare closely with the figure of 6.5°C which Derbyshire (1973) advanced on the basis of rock glacier limits on Mt.Olympus. These figures are consistent with the observation by Brown (1967a) that occurrences of discontinuous permafrost in Canada are bounded by the -1°C isotherm. They are less consistent with the fact that sporadic permafrost at 40°N latitude in Colorado is related to a mean annual temperature of -3.9°C and a mean temperature of -13.2°C for the coldest month (Ives, 1973). Local conditions exert important controls upon the distribution of permafrost. For instance, permafrost persists at an altitude of 4140 m in the summit crater of Mauna Kea in Hawaii despite a mean annual air temperature of 3.6°C at an elevation of 4,200 m elsewhere on the same mountain (Woodcock, 1974).

The snowline data from the Franklin Valley suggests that mean annual temperature was depressed by 8.7°C during the Wombat Glen (Clarence) phase and 8.9°C during the Stonehaven Glaciation. Part of this effect may be due to the greater duration of these earlier glaciations. The figures for the Cynthia Bay Glaciation are comparable to that proposed by Kiernan (1980) for the Dante Glaciation in the central West Coast Range, but during the earlier glaciations the temperature appears to have been as much as 2oC colder further inland than it was in the West Coast Range. This would be consistent in general terms with the usual temperature relationship between the two ice formation zones (Shumsky, 1964).

However, long term glacier behaviour is a function not only of the heat and mass budgets but also of bedrock configuration and ice dynamics. A glacial environment similar to that envisaged at Lake St. Clair is found today on the slopes of Mt. Kenya in east Africa where Platt (1966) claims that 90% of the ablation is due to radiation. A sensitivity analysis of the Lewis Glacier on Mt. Kenya revealed that recession during historical time could have been produced by a change in solid precipitation of 100cm/pa; in cloudiness of 10%; albedo of 3%; air temperature of 1°C ; or relative humidity of 10% (Hastenrath 1984).

This highlights the need to treat the proposed temperature depression figures with extreme caution. Nonetheless the general impression within a degree or two should be substantially correct. The temperature depression figures from the western mountains assume the snowline to coincide with a mean annual temperature of 0°C , but in such a strongly maritime environment the 1.5°C isotherm would probably be more appropriate (Shumsky, 1964). This would suggest that during the late Last Glaciation mean annual temperature in the west was depressed by little more than 5 C., and this emphasises the suggested temperature gradient.

chapter fifteen
MOUNTAIN GEOMORPHOLOGY

The form of any mountain landscape is the result of an interplay between geological characteristics, time and geomorphic processes (Derbyshire _et al._, 1979; Caine, 1983). The mountain landscape of west central Tasmania is geologically diverse. The legacy of geomorphic processes that were associated with cold climates that prevailed during the Quaternary is superimposed upon an ancient landscape. This chapter briefly explores the geomorphology of this part of the Tasmanian Central Highlands and in particular reviews the relative importance of glaciation in shaping the landscape and the ways in which glacial landscape modification has been conditioned by environmental factors. It also examines the impact of glaciers in conditioning postglacial change in the landscape. Man emerges as the most significant agent of postglacial change.

THE PREGLACIAL LANDSCAPE

A Geological determinants

Geological structure has strongly influenced the development of the preglacial landscape. The landscape of the fault province is dominated by the orientation of joints and faults in the dolerite. The detailed pattern of lineaments in the dolerite on the Central Plateau has been mapped by Jennings and Ahmad (1957) (Figure 2.2) who found no evidence to suggest that the broad drainage pattern and drainage divides on the Central Plateau had been altered by glaciation. The generally southeastward drainage of the plateau (Figure 2.2) is the result of tilting in that direction and was well established prior to 26 ma BP (Banks 1973).

The course of the Derwent River east of Mt. Hobhouse is controlled by joints and faults in the dolerite in which the valley has been eroded. The upper Gordon Valley, which is cut in Permian rocks to the west of Mt. Hobhouse, follows a course that is parallel to the Derwent. This suggests that the Gordon Valley has been superimposed from the original dolerite cover rocks (Davies, 1959, 1965). The upper Franklin, Alma, upper Surprise and upper Denison rivers also parrallel this trend but have been superimposed still further onto pre-Carboniferous rocks that are overlooked to east and west by dolerite mountains.

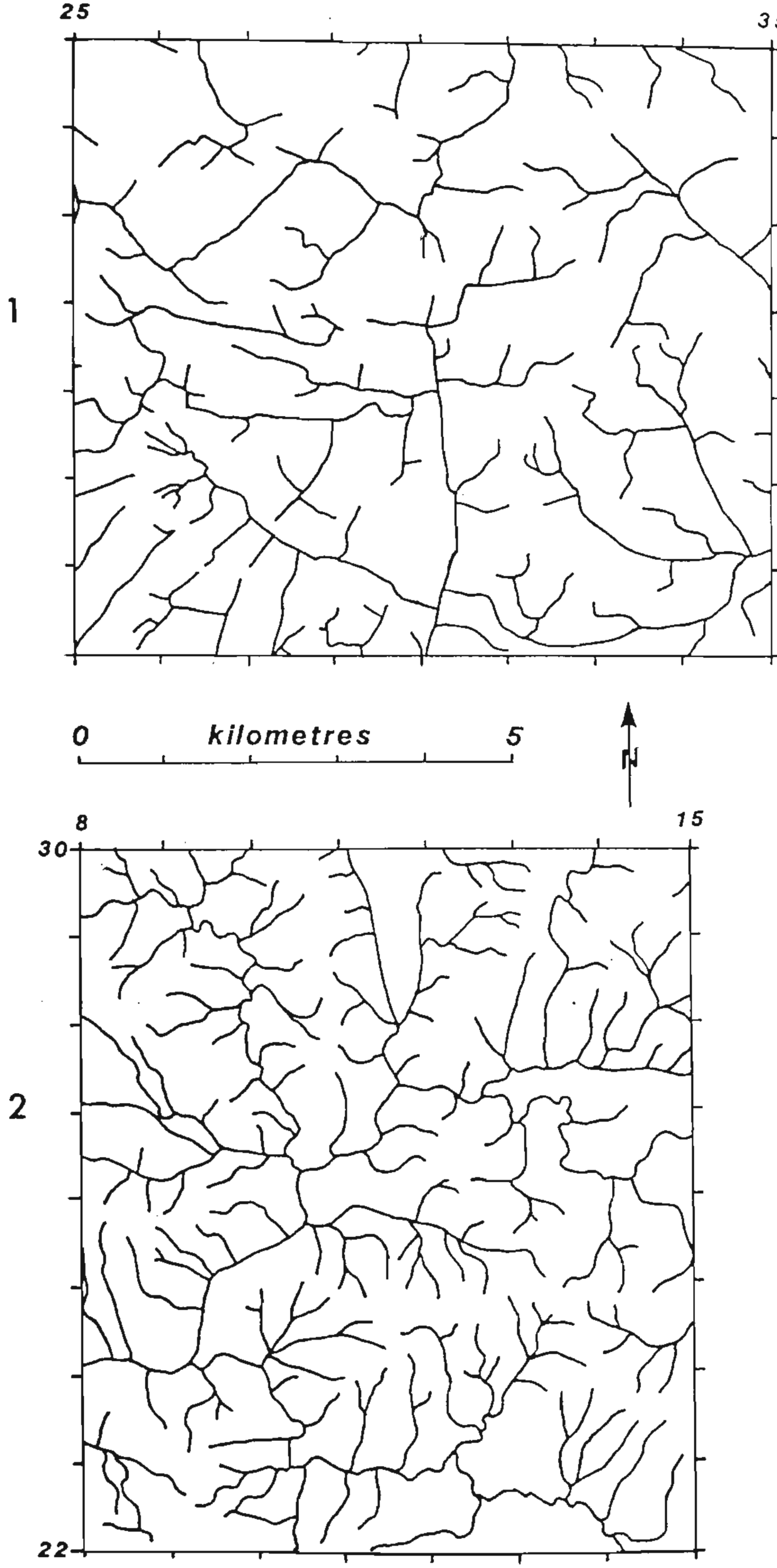
Other major lineaments in the dolerite are oriented approximately east to west. Within the study area the most striking of these is the Top End Lineament which cuts through the King William Range. The Laughing Jack and Clarence cols, and the King William Saddle also reflect this structural trend. The cols at the head of the Cuvier and Franklin valleys have developed along faults. The preglacial Derwent River between Mt. Olympus and the Traveller Range exploited a fault, and faultline scarps form the valley walls. Such scarps are an important feature in the topography of the fault province.

The influence of the former dolerite cover rocks extends well into the fold structure province. The Top End lineament can be traced across the edge of the fold province and down the Loddon Valley. The lower Surprise Valley follows a similar trend, as does much of the Franklin Valley between Lake Dixon and the Collingwood junction. This drainage trend cuts across the dominant trend of compositional and transpositional layering in the metamorphic rocks. This is consistent with the proposition (Davies, 1965), that the trans-structural orientation of the principal drainage trends in the fold province is the result of superimposition. Nonetheless, the detailed form of the landscape and the dense trellis pattern of the drainage in the fold province is dominated by the underlying structure (MacIntyre, 1964) to the extent that virtually no other influence can be discerned (Figure 15.1). The only exception to this lies in minor diversions of streams by glacial ice such as occurred in the lower Collingwood Valley.

B Historical development

The local exhumation of the pre-Carboniferous surface in the upper Franklin Valley (Banks, in Derbyshire, 1966) has revealed the oldest elements in this mountain landscape. The pre-Carboniferous surface occurs at 912 m on Last Hill, 851 m on Pyramid Hill and 730 m in the Franklin Valley (Banks, 1962b). This surface appears to have been extensively revealed only where it coincides with later erosion surfaces, particularly the St. Clair Surface (730-825 m) (Davies 1959). The legacy of Permo-Carboniferous glaciation remains remarkably strong. In addition to ice

Figure 15.1 Drainage patterns (1) on dolerite at the northern end of the Traveller Range; (2) on Precambrian schists in the Franklin Collingwood area.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abraded bedrock in the upper Franklin Valley there are widespread exposures of glacial and glaciomarine sediments. Resistant erratics that were carried eastwards from Antarctica by the Gondwana ice sheets remain in the landscape as reworked deposits and isolated lag boulders (Figure 15.2). Some have been carried a short distance westwards again by late Cainozoic glaciers (chapter seven). The bedouin boulders on King William Saddle stand as testimony to the resistance of some rock materials and to the polycyclic nature of landscape evolution.

The erosion surface morphology is particularly well preserved in the eastern part of the study area where the attitude of the rocks is close to horizontal. It is less well preserved on the deformed rocks of the fold province. To the west of King William Saddle the broad surfaces give way to apparently accordant summits of discrete mountain ranges.

The proposition that the erosion surfaces predate the late Cainozoic glaciations is compelling. However, the suggestion that all are between Miocene and late Pleistocene in age (Davies, 1959) may be refined. The uplift resposible for the stepped relationship between the surfaces probably dates from about 65 ma BP (Griffiths, 1971). The presence on the Upper Coastal Surface (365-460 m) of glacial deposits that are early Pleistocene if not Pliocene in age (Kiernan, 1983a; Colhoun and Augustinus, 1984) implies that the second youngest of the surfaces considerably predates the late Pleistocene. Deposits apparently equivalent in age have

Figure 15.2 Small rock basin lakes and roches moutonées are well developed on the Labyrinth Plateau. Mt. Gould is in the background.

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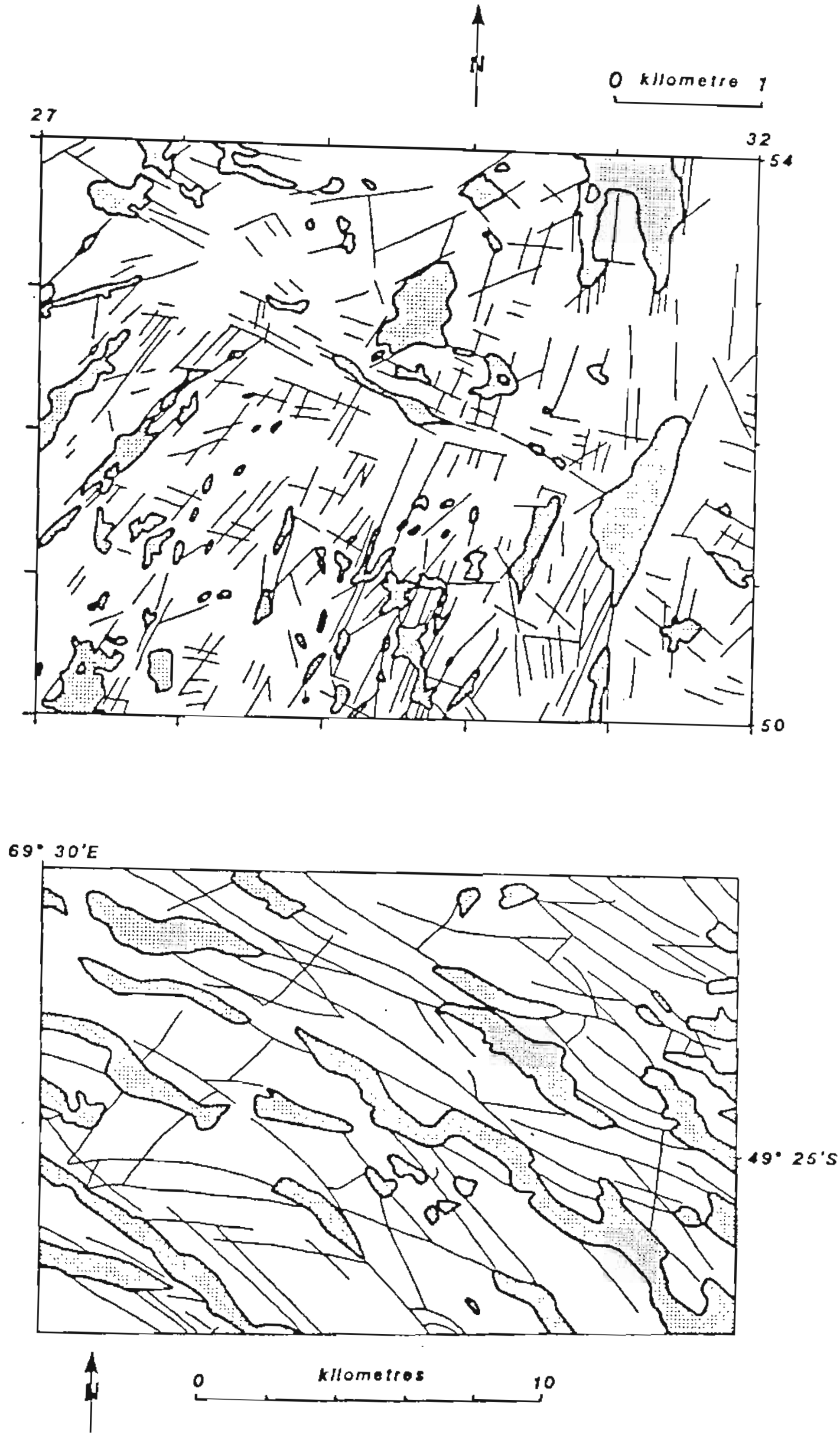

Figure 15.3 Strong structural control of (1) rock basins and roche moutonees on the ice abraded plain of the Traveller Range bears comparison with (2) the central plateau of Kerguelen (redrawn from Nougier, 1972).

been found at the same level during this study (chapter thirteen). The Upper Coastal Surface truncates basalts that overlie late Oligocene to middle Miocene limestones at Granville Harbour and elsewhere in western Tasmania (Quilty, 1972; Banks, et al., 1977). It is therefore probably between late Pliocene and middle Miocene in age.

Volcanic outpourings occurred on the St. Clair Surface and Lower Plateau Surface in the Nive Valley during late Oligocene - Miocene time and provide a limiting date for those surfaces. Similar basalts at Great Lake have been dated to 21.8-23.6 ma BP (Sutherland et al., 1972). The effect of these lava flows was to locally disrupt drainage patterns, displacing the Nive River eastwards (Prider, 1948) and producing a small twin stream system southeast of Dyes Marsh. Small conical hills composed of strongly unsaturated basalt that occur near Laughing Jack Lagoon probably represent denuded volcanoes that existed prior to the mid Tertiary flow sequences (Wyatt, 1971).

Revised topographic maps indicate that some levels which have previously been regarded as part of the same erosion surface are not accordant. While the origin and significance of these surfaces is problematic they remain very important features in the landscape. Hence, the status of some of these surfaces warrants review. It is the summit of Mt. Alma that best accords with the St. Clair Surface (730-825 m), not its shoulders (about 640 m) as Davies suggested. Scott (1960) argued that the generally accordant summits of the Collingwood Range and a number of other ridges formed part of the St. Clair Surface but an accordance at around 600 m is more prominent in each case. This surface extends westward into the Franklin Hills. It is therefore suggested that an additional erosion surface, the Collingwood Surface (580 - 640 m), forms an important part of the landscape in this part of the fold province.

C Geomorphic processes and climate

Fossil evidence of rainforest vegetation that included species now restricted to tropical areas is preserved beneath the basalt near Tarraleah (Prider, 1948). This is consistent with conditions having been warm and moist during the middle Tertiary. Under an established forest cover weathering is likely to have been predominantly chemical in nature. Erosion is likely to have proceeded by fluvial processes. Warm and moist conditions are also suggested by the presence of saprolitic dolerite beneath the glacial deposits downstream from Butlers Gorge in the Derwent Valley (Edwards, 1955). At Wayatinah deeply weathered dolerite is overlain by angular dolerite slope deposits typical of those that formed in the extraglacial areas during the glacial phases. This indicates a transition from predominantly chemical to predominantly physical weathering processes as climate deteriorated during the late Tertiary.

The basalt that occurs around Bronte in the Nive Valley is unlikely to be less than 21.8 ma in age (Sutherland _et al._, 1973). The rate at which the scarp has retreated can be estimated if the same age is assumed for the basalt that crops out close to the foot of the scarp between the St.

Clair Surface and the Lower Plateau Surface near the Clarence River. On this basis the foot of an interfluve on the scarp has been worn back at a mean rate of less than 18.5 mm/ka over the past 21.8 ma. A broad re-entrant in the scarp immediately to the east has been worn back at a mean rate of not more than 46.9 mm/ka. No ice descended the scarp at this point and slab toppling is not evident. These figures indicate that the warm and moist conditions of the Tertiary were also times of relatively little erosion and mass movement, even in areas of steep gradient. Stability over a lengthy period is also implied by the depth to which the bedrock is weathered in the Derwent Valley. The angular slope deposits that overlie the saprolite at Wayatinah reflect increasing instability as the forest biomass diminished.

Suggestions that the deeply weathered mantles in southeastern Australia date from at least the early Tertiary (Hills and Carey, 1949; Pillans, 1977) are compatible with a prolonged period of landscape stability in the mountains of west central Tasmania. It is probable that this ancient landscape remained essentially unchanged until the advent of Pleistocene cold.

GLACIAL MODIFICATION OF THE LANDSCAPE

It is particularly difficult to single out the impact of geological structure, preglacial form, glacier dynamics and stage in the evolution of the landscape. All are complexly inter-related. In broad terms there exists a gradient of glacial modification that emphasises erosional landforms at higher altitudes and depositional landforms at lower levels. However, the efficacy of glacial erosion is not entirely consistent with this gradient due to variations in lithology, structure and preglacial morphology. Structure has strongly influenced the preglacial form of the landscape. This has in turn conditioned glacier dynamics and the rate and manner in which glaciers have remodelled the landscape. This section reviews the likely impact of these factors on glacial action.

A Geological and morphological influences

(i) Geological influences

The fundamental geological control upon glacial modification of this landscape was the transition from the subhorizontal post Carboniferous rocks in the east to the strongly deformed preCarboniferous rocks in the west. Snow was able to accumulate on benches formed of Permian and Triassic rocks that constitute important landforms in the fault province (Figure 4.7). The glacial stairway in the Cheyne Range illustrates the importance of horizontal jointing in dolerite that also gives rise to benches. Structural benches are less well developed in the fold province.

The most widespread rock of the mountain summits is Jurassic dolerite, a hard, well jointed rock that is highly prone to glacial plucking (Davies, 1969). One of the most striking areas of glacially eroded landscape is the plain that has been formed on the dolerite sill of the Central Plateau (Jennings and Ahmad, 1957). The plateau is characterised by rock basins amid an otherwise almost unbroken landscape of roches moutonées that vary in size from a few metres to over a kilometre in length. A similar landscape has developed on the much smaller Labyrinth plateau (Figure 15.2). Erosion in both these areas has been focussed along lineaments in the dolerite (Linton, 1963) and the glacial landforms are strongly controlled by geological structure. This relationship is particularly strong where the ice flow was parallel to the major joint trends and the probable preglacial topography (Jennings and Ahmad, 1957). The rock basins on the Central Plateau occur in the deepest parts of elongate depressions that reach kilometres in length. Observations from the Traveller Range suggest that closely spaced minor joints may be a major determinant of basin location within these larger depressions. The resulting pattern is very similar to that reported from the central plateau of Kerguelen by Nougier (1972) (figure 15.3).

Elsewhere the vertical jointing of the dolerite enables it to form steep cirque headwalls as at Mt. Gell. The headwalls are often extremely steep where the cirque floor lies beneath the base of the dolerite sill and cirque erosion has cut back into the Permian and Triassic rocks undermining the dolerite columns (figure 15.4).

The glacial troughs north and south of Mt. Ida and re-entrants along the plateau margin where ice spilled into the Lake St. Clair trough are all developed along structural lineaments that are exposed on the plateau at their head. The western slopes of the Traveller Range south of Mt. Ida demonstrate the capacity of the dolerite to form steep

Figure 15.4 Longitudinal profiles of (1) a cirque formed wholly in dolerite on the eastern flank of Mt. Gell; (2) the Lake Richmond cirque and (3) the Lake Rufus cirque, both of which are cut in Permian rocks at the base of the dolerite sill. The steeper cirque headwalls at Lake Richmond and Lake Rufus are largely the result of sapping at the base of the columnar dolerite. Compare also with the Lake George cirque (Figure 15.7) which is cut wholly in dolerite.

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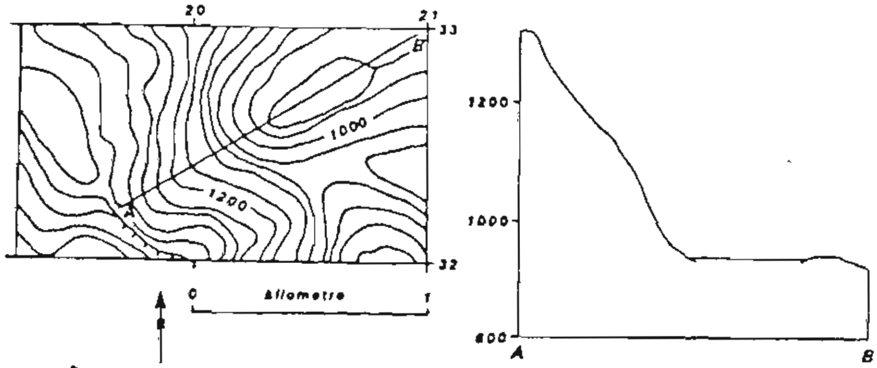
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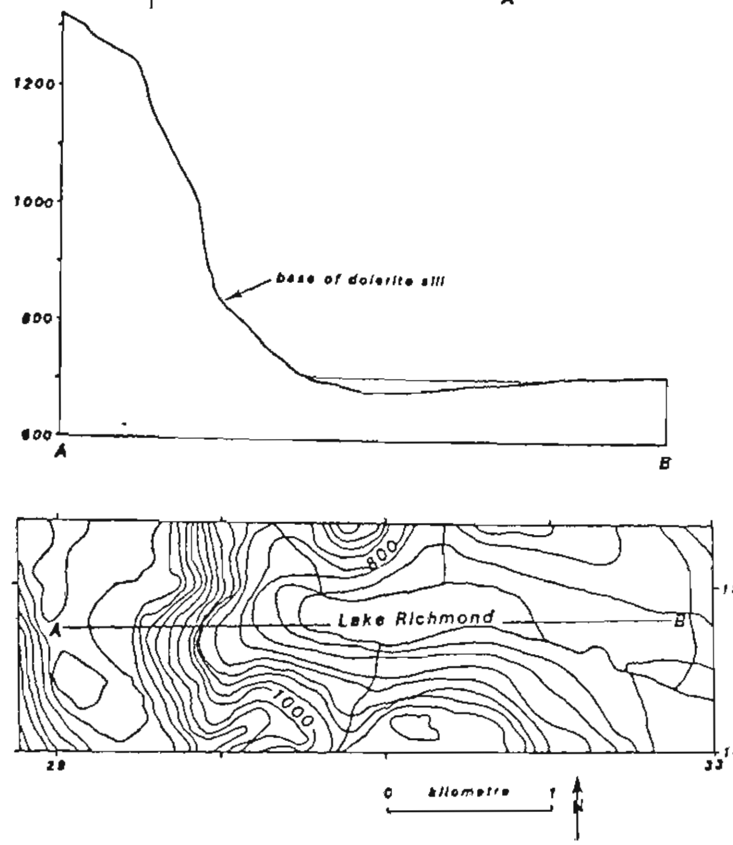
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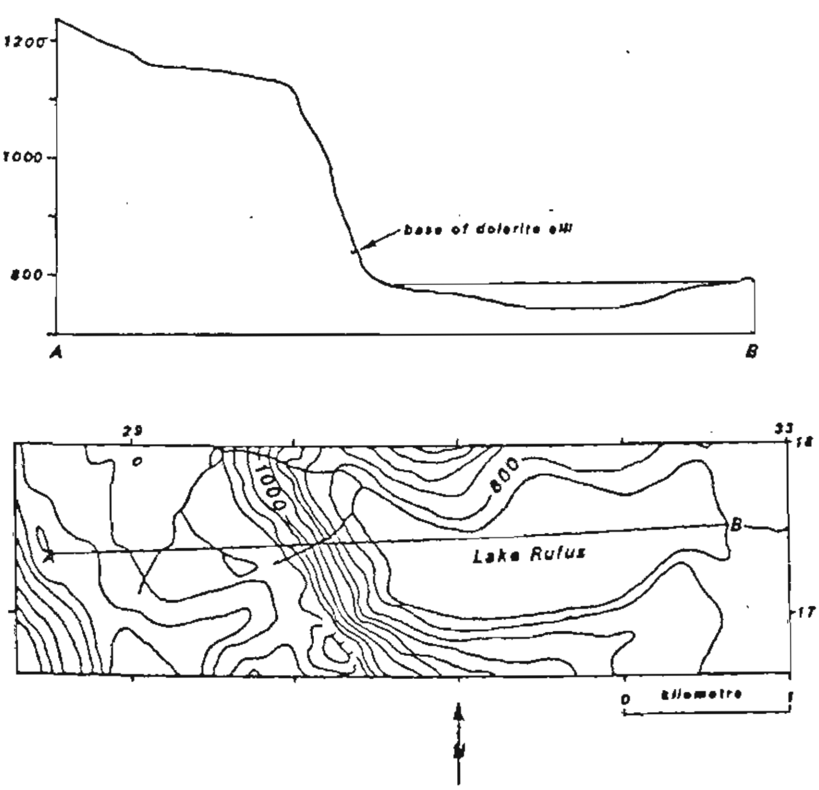
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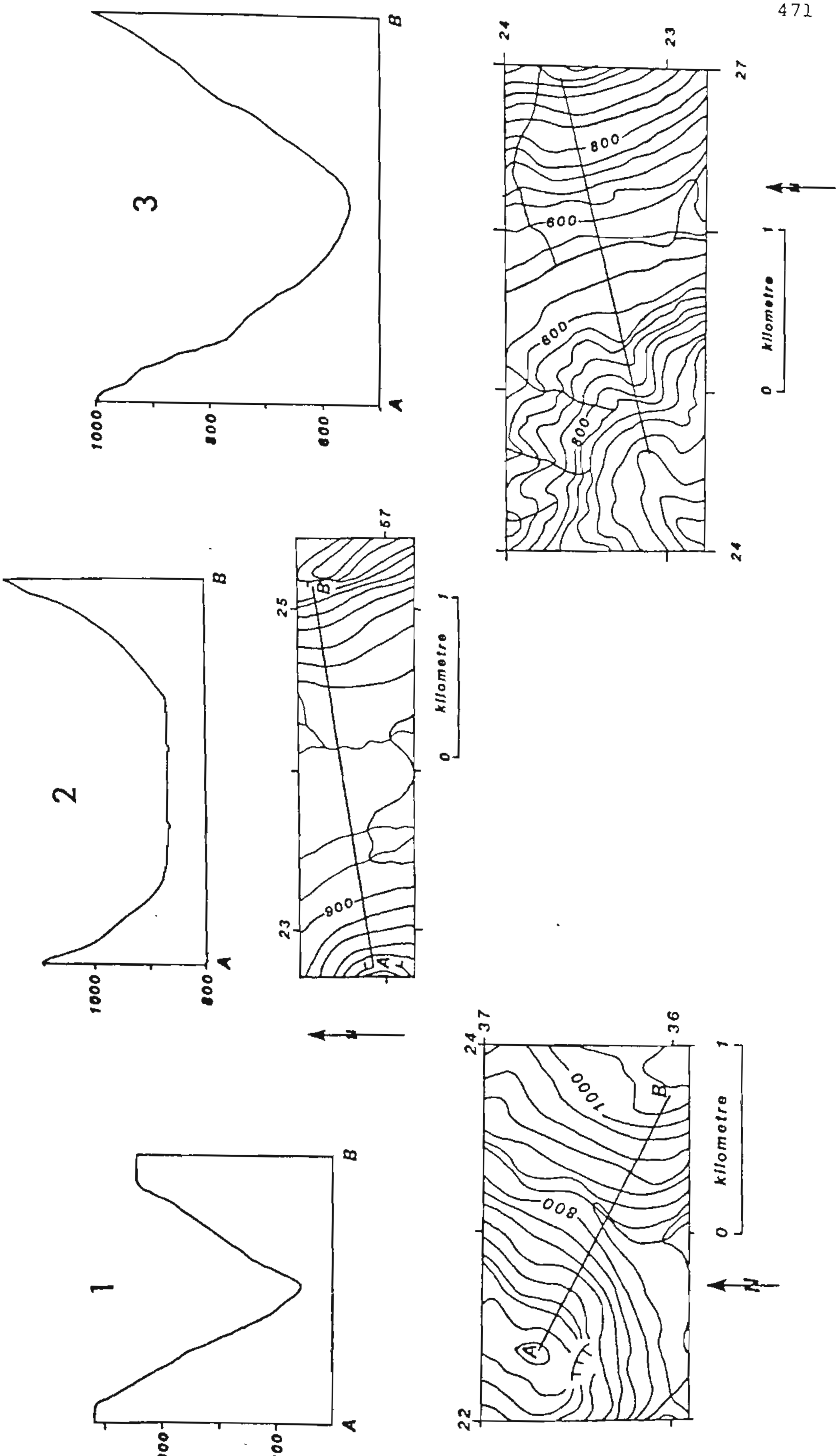
trough walls. However, the dolerite is prone to fragmentation by frost processes and as a consequence the profile tends to consist of two facets, namely a steep scarp formed in bedrock and a talus apron at its foot. At least some of the talus was probably deposited in contact with the ice in the Lake St. Clair trough and lowered during glacier retreat.

Because the most resistant rocks coincide with the steepest valley gradients it is not possible to evaluate independantly the impact of these two factors upon trough morphology. Comparison of the cross profile of the glacial trough upstream from Lake Undine in the Franklin Valley with that of the Surprise Valley trough near King William Saddle seems to support the contention of Matthes (1930) and King (1959) that troughs become narrower and deeper where the bedrock is resistant to erosion (Figure 15.5). However, the evidence from Mt. Ida suggests that troughs also become narrower and deeper where well developed vertical joint networks are present. Vertical jointing is markedly less well developed in the Permian and Triassic rocks than in the dolerite. Troughs cut in these rocks commonly approach the classical U-shaped form, as in the upper Cephissus Valley. Trough asymmetry is frequently the result of contrasting bedrock on the opposing walls. The upper Alma trough has a steep southeastern wall cut in dolerite and a more gentle northwestern wall cut in Permian rocks (Figure 15.6).

(ii) Morphological influences

Because the frequency of glaciers above the regional

Figure 15.5 Transverse profiles of (1) the glacial trough formed in dolerite north of Lake Undine in the Franklin Valley; (2) the upper Narcissus trough, which is formed in Permo-Triassic rocks; (3) the Surprise Valley trough, which is formed in Precambrian dolomite and schist.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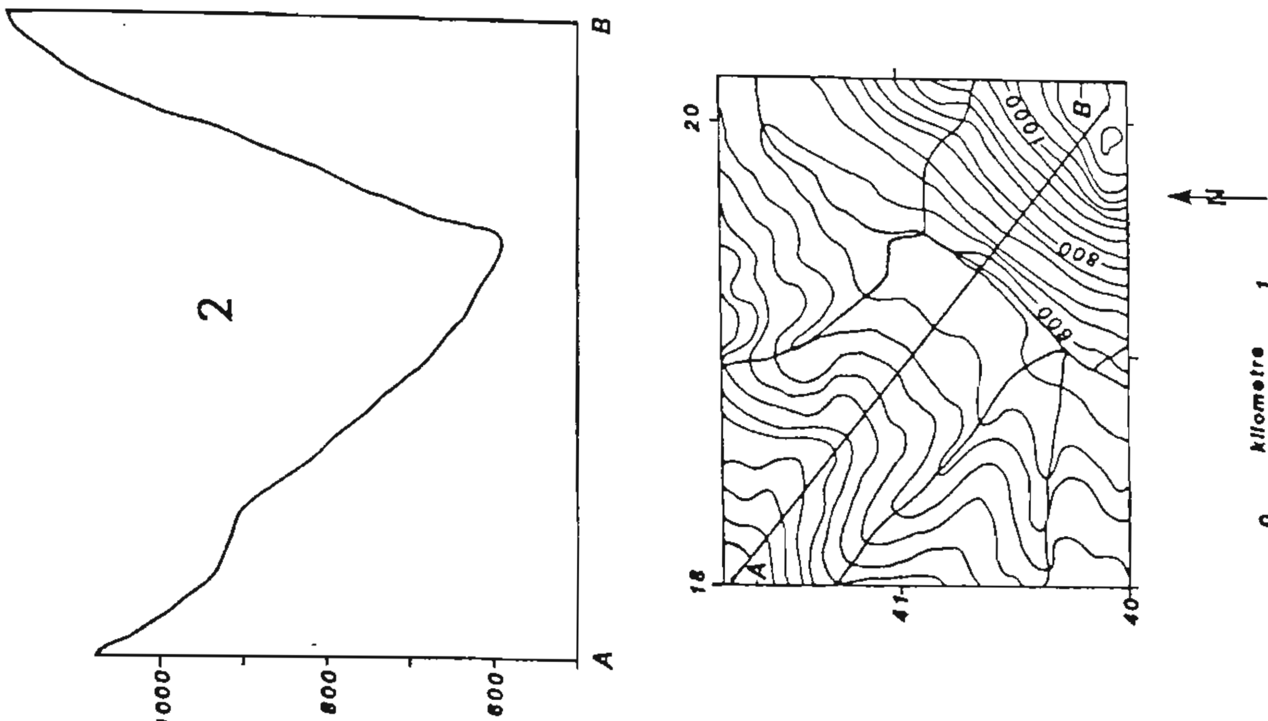
equilibrium line altitude is determined by the availability of suitable topography rather than increasing altitude (Graf, 1976) the preglacial erosion surface morphology has strongly influenced the glacial geomorphology. In the Du Cane Range the High Monadnocks (Davies, 1959) have formed important snow fences. On the Central Plateau remnants of the Higher Plateau Surface formed the divide between the ice which flowed northwards into the Mersey and that which flowed southwards into the Derwent, Traveller and Nive basins. The Central Plateau ice cap east of Lake St. Clair lay on the Lower Plateau Surface. Some of the largest valley-head cirques formed at the level of the St. Clair Surface. Piedmont glaciers spread out on this surface as they emerged from their alpine troughs. The Franklin Glacier cut a deep gorge as it spilled from the edge of the St. Clair Surface to the Upper Coastal Surface.

For all but the Wayatinah phase the great breadth of the St. Clair Surface prevented the Derwent Glacier from descending to a lower altitude where conditions would have been warmer. The Derwent Glacier was stranded on this high and cold surface. As the surface gradient of the glacier was low (chapter fourteen) the stresses that were produced in the ice by the force of gravity were probably also low. As a consequence, the Derwent Glacier may have been unable to maintain a high rate of flow and continuous pressure melting at its base (see below). Hence, the preglacial form of the landscape may have also influenced the thermal regime at the ice base.

Figure 15.6 Transverse profiles of (1) the Cephissus trough, which is cut through Permo-Triassic rocks, and (2) the upper Alma trough, the steep southeastern wall of which is cut in dolerite, the less steep northeastern wall in Permo-Triassic rocks.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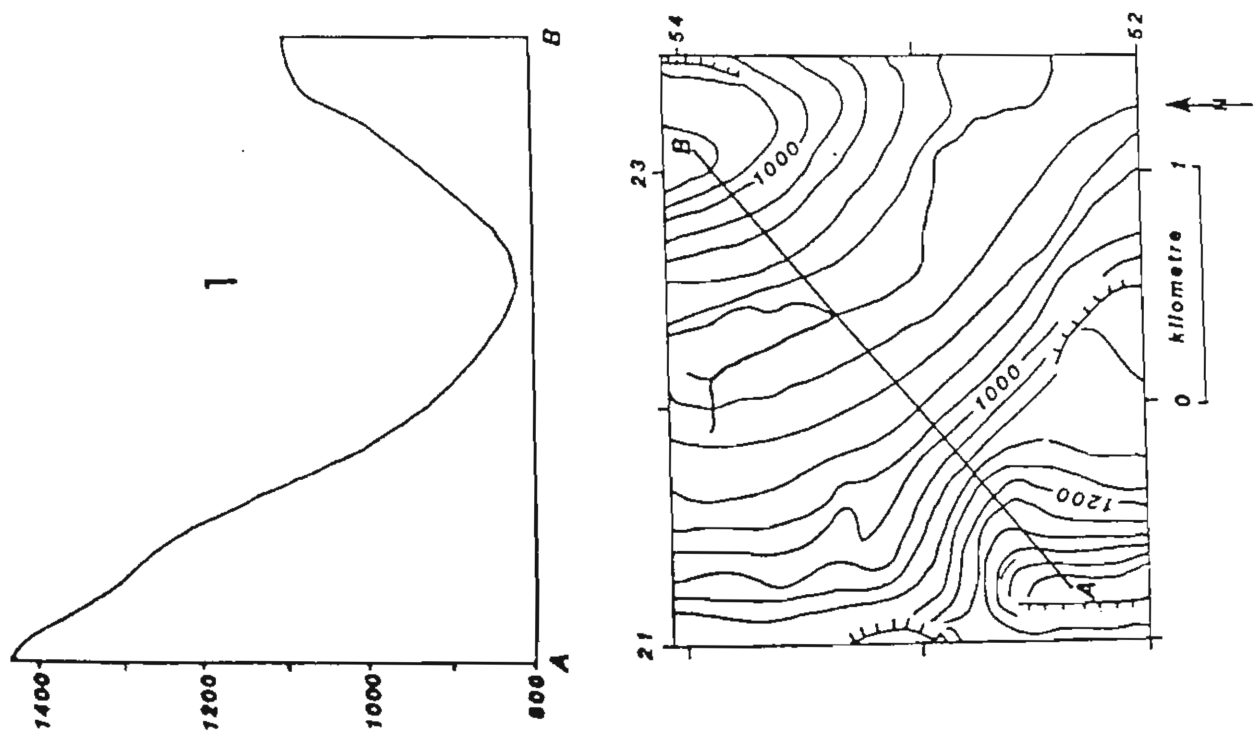
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Because topography may directly control glacier power through its influence upon the ice surface gradient (Andrews, 1975) the preglacial erosion surfaces played a major role in conditioning the intensity of glacial erosion. Intense linear erosion was focussed on the edge of the St. Clair Surface in the Franklin Valley but erosion was less intense on the Central Plateau, where it seems probable that little more than the regolith may have been stripped in some areas (Caine, 1967; Davies, 1969). Some of the larger roches moutonées on the Central Plateau are streamlined remnants of the Higher Plateau Surface. Hence, the evidence from the Central Plateau is inconsistent with the suggestion (Von Engeln, 1937; Lewis, 1947) that roches moutonées are equilibrium forms that are the characteristic product of glacial erosion . The large roches moutonées on the Central Plateau are clearly the result of smoothing of the preglacial landscape (Rudberg, 1954; Zumberge, 1955; Jennings and Ahmad, 1957).

Few of the lakes on the Central Plateau appear to be particularly deep and it is likely that preglacial valleys have been widened to a far greater extent than they have been deepened. Rudberg (1954) has noted a similar trend in Vasterboten, and the suggestion seems consistent with the observation that lakes in the Traveller Range are focussed where jointing is most frequent. Plucking along the sides of roches moutonées (Jennings and Ahmad, 1957) has probably played a major role in this widening. The western part of the plateau offers some support for the contention that lakes in areally abraded landscapes are more frequent where the topography is flatter (Tanner, 1938).

Small dolerite plateaux such as the Labyrinth, Cheyne Plateau and King William I Plateau have a similar form to the Central Plateau. The absence of any large elevated plain cut in pre-Carboniferous rocks prevents direct comparison between the effects of areal erosion on dolerite and on the rocks of the fold province. Comparable terrain occurs in the fold province only on a preglacial erosion surface cut across Owen Conglomerate on the Dante Plateau in the central West Coast Range (Kiernan, 1980). Remnants of a dolerite cap occur on the adjacent slopes of Mt. Sedgwick hence some of the landforms on the adjacent plateau may have been superimposed from the original dolerite cover-rocks. However, the presence of a Permian roche moutonée beneath the dolerite (Edwards, 1941; Kiernan, 1980) indicates that at least part of the surface was ice-abraded well before the intrusion of the dolerite. This suggests that the preglacial topography has been the primary determinant of the erosional morphology of the Central Plateau, although the structure of the dolerite has also been important.

The narrow divides characteristic of a "fretted upland" (Hobbs, 1911) that are evident on some of the quartzite mountains to the west are probably the result of steep preglacial fluvial divides rather than progressive glacial erosion (Davies, 1969). Rock benches have been important to snow accumulation in the east but the linear form of the faultline scarps has offerred little shade and has probably hindered the development of cirques (Peterson, 1968, 1969).

Two-storyed cirques in the southern part of the King William Range are difficult to accept as the result of progressively rising nivation levels (Taylor, 1922; Lewis, 1945b). If this had been the case the rock basin that sometimes occurs in the lower cirque would probably have been filled by proglacial sediment. It seems more likely that this form is the result of preglacial topography which has been emphasised by glacial erosion (Klmaszewski, 1964).

B The significance of stage

As the chronology of glaciation in this area was reviewed in chapter thirteen, the present discussion merely focusses upon selected examples of progressive landscape modification.

A geomorphic feedback system enables glaciers to persist or reform more readily once cirques have been developed (Graf, 1976). The transition from a cirque scalloped morphology to a fretted morphology has been postulated to be sequential (Hobbs, 1911) and there is evidence for this being the case in the Tasmanian Central Highlands (Davies, 1969). The relationship between the snowfences and cirques in the King William Range is scallopped, while parts of the Du Cane Range are fretted (Figure 15.7).

The col between Mt. Geryon and the Acropolis has developed by headward retreat of the Narcissus cirque towards the Cephissus trough. The Du Cane Gap also originated as an inosculation col. Small cirques east of Du Cane Gap on the southern slopes of Falling Mountain and on the northern

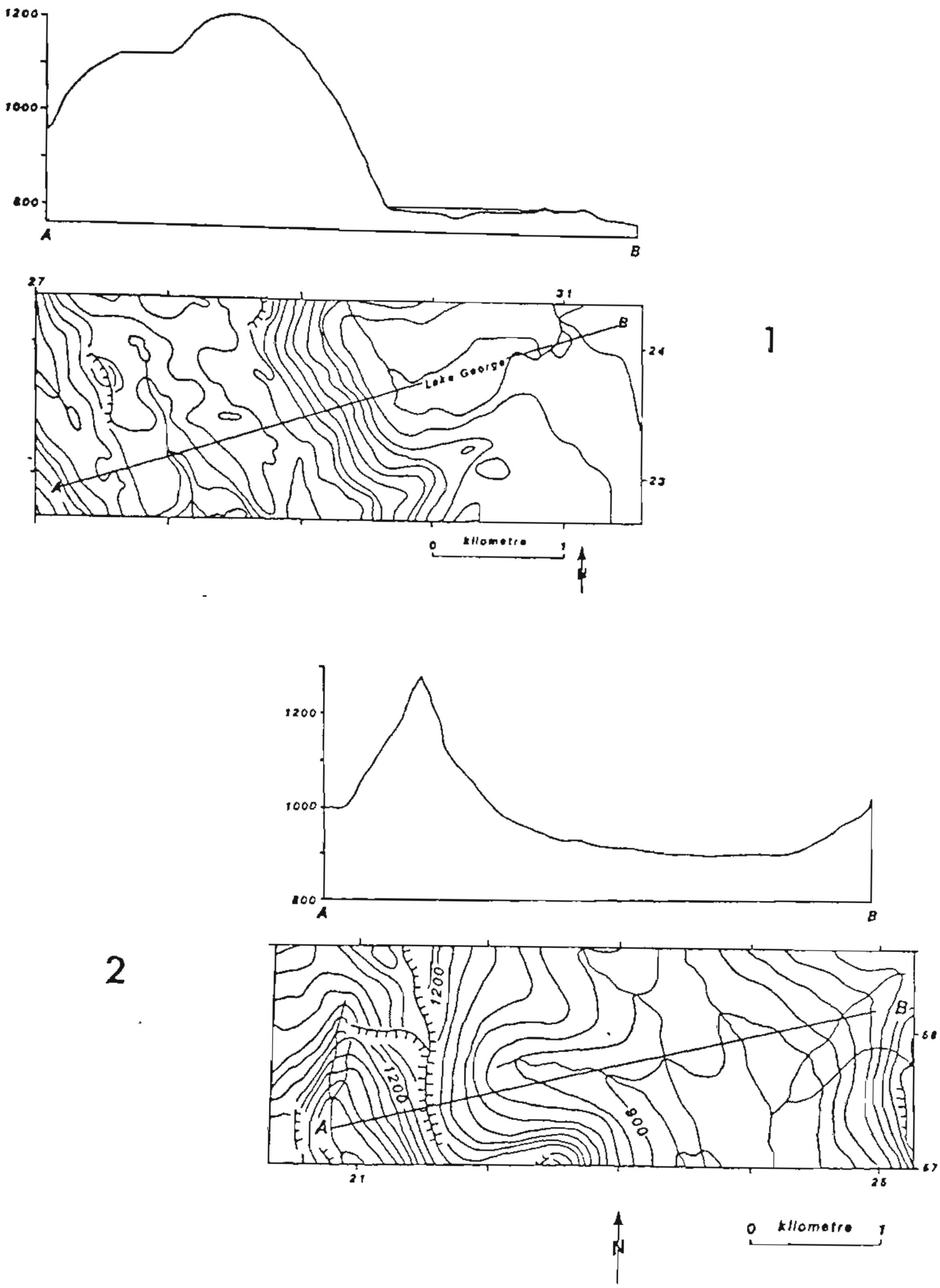


Figure 15.7 Contrasts in cirque and snowfence relationships: (1) scalloped morphology at Lake George in the King William Range (bathymetry after Derbyshire, 1967); (2) fretted morphology at Mt. Geryon in the Du Cane Range.

shoulder of the Traveller Range, together with the favourable snowfence offerred by the ridge that formerly extended between the two, suggest that Du Cane Gap was previously the site of a cirque that fed ice into the Mersey Trough. This cirque lay on the outside of a right-angle bend in the Mersey trough, and with the build-up of the ice surface some of the Mersey ice flowed into this cul de sac. The ice from the Mersey then breached the divide at the head of the cirque leading to the development of a transfluence col.

A second example of progressive development is offerred by the Top End Gap which conducted the preglacial drainage from part of the Loddon Range eastward through the King William Range. Ice that subsequently accumulated at the head of the Top End Valley could not be accomadated by the narrow defile through the range and spilled northwards over a low divide of Precambrian rocks. Incision into this divide and headward erosion by the cirques of the Loddon Range led to the capture of most of the Top End Glacier by the Surprise Glacier (Figure 15.8a). The floor of the Surprise Valley at this point now lies 50 m below the Top End Gap. Because sone ice flow was maintained through the Top End Gap the Top End River discharges in the opposite direction to that which is usual for a hanging valley. This is the first example of drainage diversion due to glacial erosion that has been recognised in Tasmania.

A short distance to the south the drainage from Lake Montgomery into the Guelph basin has been diverted southwards into the Gordon Valley by the construction of an end moraine across the eastern end of the lake (figure 15.8b). Deflection of the drainage was probably initiated when a distributary lobe of the Guelph Glacier extended through the Gordon Gap. Several other cases of drainage diversion by glaciers and glacial deposits have previously been recognised in Tasmania (Colhoun, 1980; Kiernan, 1982, 1983b; Colhoun and Augustinus, 1984).

C The impact of process

The topography of the Central Plateau is the result of glacial abrasion and intense glacial plucking of the well jointed dolerite. An ice thickness of at least 200 m is required to permit erosion of the high roches moutonées east of Lake Payanna. However, during the latter phases of glaciation ice spilled from the western margin of the Traveller Range in only a few localities. This suggests that the ice cover was generally much less than 200 m. Shear stresses at the base of the ice cap on the plateau would have derived primarily from the ice surface gradient which, given the thin ice, would generally have been very low. Local thickening of the ice in hollows would have favoured glacial abrasion of well jointed dolerite but plucking would have been impeded by high ice pressures (Sugden and John, 1976). Lower ice pressures in the lee of eminences would have facilitated regelation and plucking of the roches moutonées under conditions of extending flow.

The importance of basal sapping of the dolerite in conditioning the angle of cirque headwalls has previously

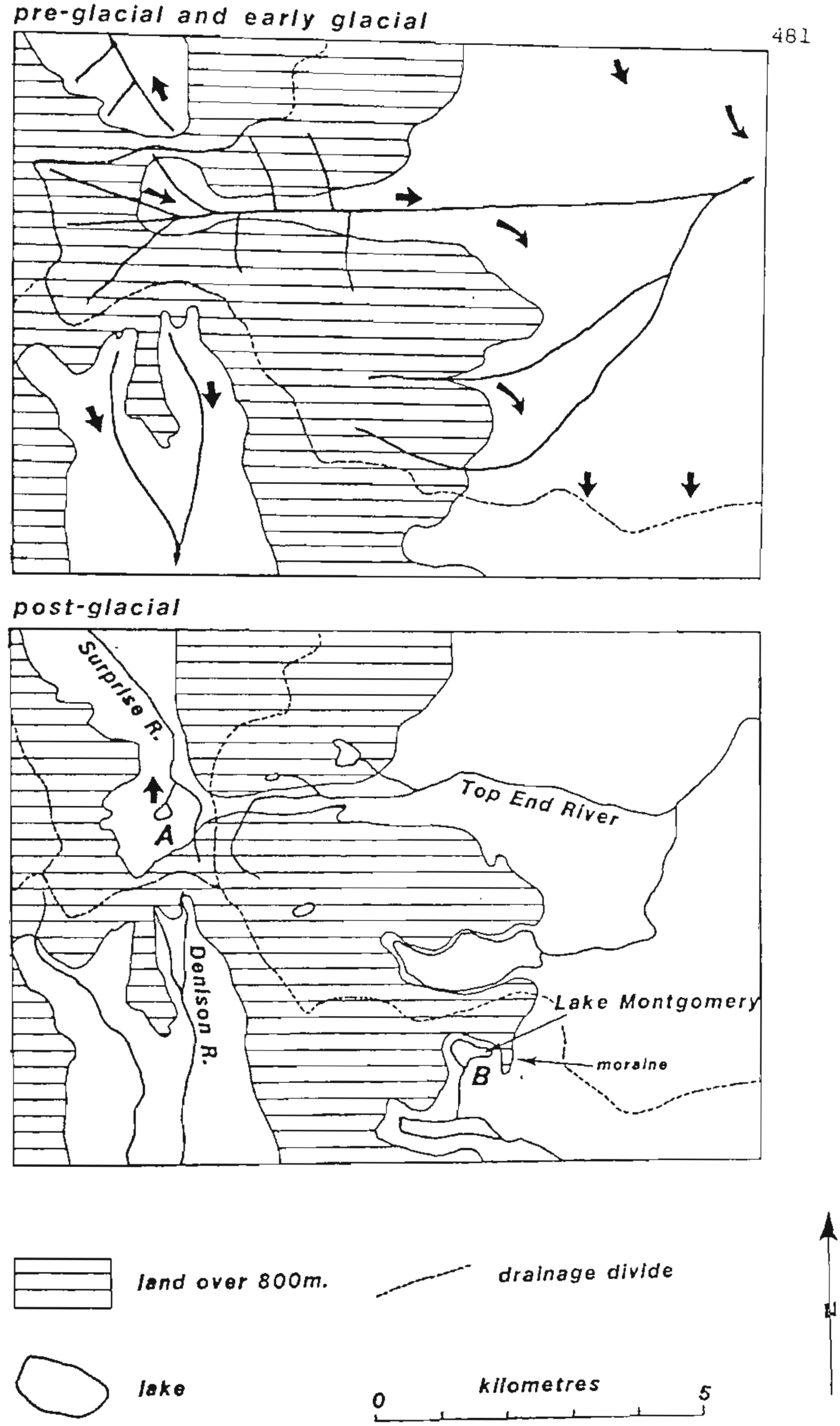


Figure 15.8 Diversions of drainage by glacial erosion in the upper Surprise-Top End Gap area (site A) and due to moraine construction at Lake Montgomery (site B). The principal directions of ice flow are indicated.

been cited. The steep headwalls in the King William Range are all likely to have resulted from this process. Plucking was clearly important in the formation of cirques in the dolerite mountains but the headwalls of those cirques that are formed entirely in dolerite are less steep than those of cirques that are floored by Permian or Triassic rocks (figures 15.4 & 15.9).

In some cases cirque development has been impeded by the slowing of ice flow from tributary valleys by the heavy discharge from trunk glaciers. The Hamilton cirque provides a good example of this, diffluent ice from the Cephissus trough having passed across the Hamilton Valley en route to the cols at the head of the Cuvier Valley.

It has also been argued that in some cases cirque development has been inhibited by later inundation beneath a major ice sheet (Peterson, 1969). Most of the evidence for this has been limited to the erosional morphology of the cirques which most workers would regard as being equivocal evidence that should be interpreted with caution. Derbyshire (1972) theorised that a major ice sheet may have existed west of the Du Cane Range and speculated that the absence of cirques from this area may have been due to inundation by ice in this manner. Erosional and depositional evidence presented in chapter fourteen of this thesis indicates that the ice was much thicker west of the Du Cane Range than it was further east. The depositional evidence from the Franklin Valley (chapter twelve) confirms that elevated divides were over-ridden by an ice sheet.

Figure 15.9 The steep eastern face of Mt. Geryon is the result of intense glacial erosion in Permo-Triassic rocks at the base of a sill of columnar dolerite.

Coupled with considerable evidence that is now available from surrounding areas (Sansom, 1978; Colhoun, 1979; Kiernan, 1980, 1983a; Augustinus, 1982) this is consistent with the thesis of cirque inhibition in such a manner.

Intense linear erosion by glaciers has occurred where flowing ice was focussed into channels and the stress imposed on the ice mass by gravity was high. The majority of the trough-ends are alpine in character but the Cephissus trough has a more Icelandic form (Linton, 1963). Its head consists of steep mammallated rock that has been over-ridden by ice from the Labyrinth Plateau and the ice sheet to the west. The Cephissus trough broadens downstream of a low pass to windward. This pass would have permitted further ice to enter the trough from the ice sheet and would have facilitated wind-drifting of snow at times when the glaciers were less extensive (Graf, 1976).

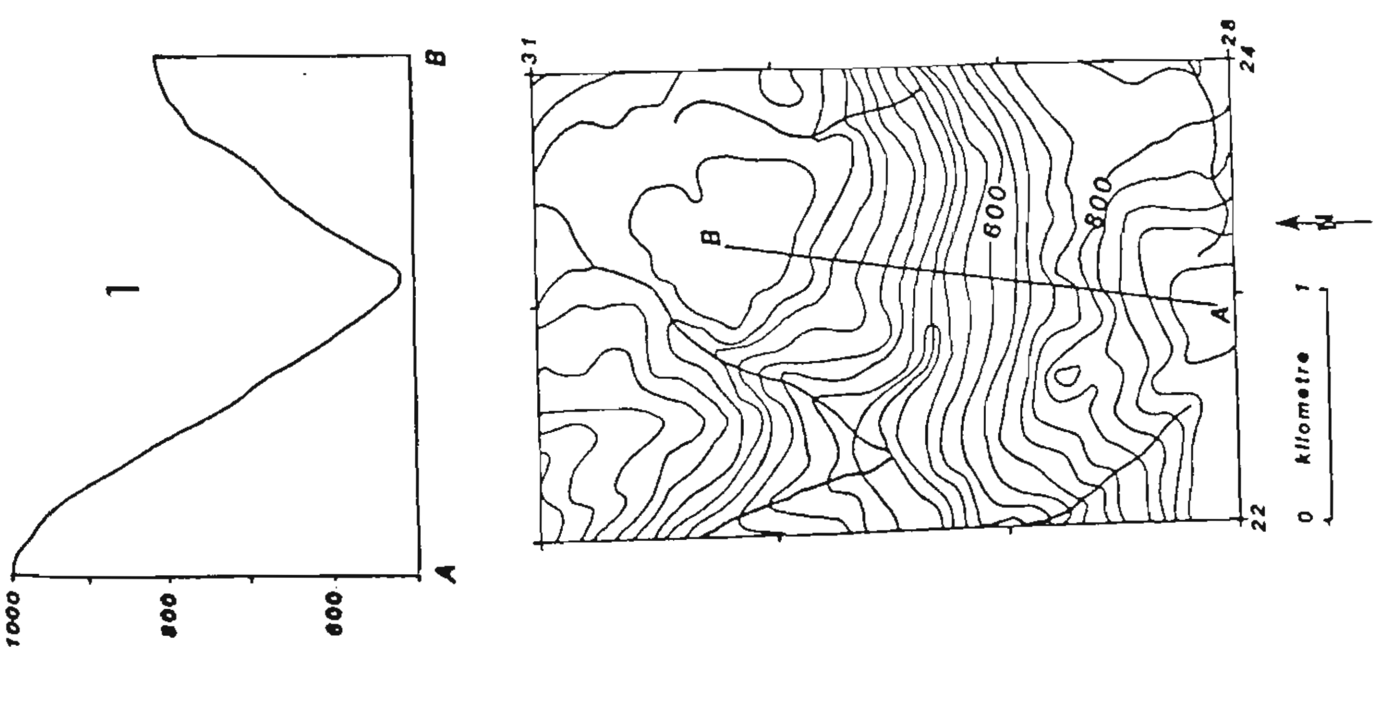
The longitudinal gradient of the troughs and, hence, the intensity of glacial processes (Graf, 1970), appears to have been an important determinant of trough form. The deep and narrow trough between Lake Dixon and Wombat Glen in the Franklin Valley is consistent with the suggestion that troughs assume such a form where the gradient is steep and a progressively increasing ice thickness emphasises erosion of the thalweg (Lewis, 1947; Veyret, 1955; Graf, 1970). Where gradients are more gentle, U-shaped cross profiles have developed, as in the Surprise Valley. Glacial troughs have developed least well where the gradient is low as at Travellers Rest Lake on the Central Plateau and in the

Franklin Valley downstream of Wombat Glen (Figure 15.10).

Few of the troughs are totally symmetrical. In some cases this is due to differences in lithology and structure on opposing valley walls (Figure 15.6), but where no major differences of this sort exist asymmetric troughs may be due to variations in glacier dynamics. The western flank of the Cuvier trough has been smoothed by transfluent ice from the St. Clair area that flowed across it to the Alma and Franklin cols. Asymmetry along parts of the Alma trough is due to erosion by short valley glaciers that descended its northwestern flank. The deepest part of the Lake St. Clair trough lies close to the western shoreline. This has probably resulted from glacial erosion having been focussed in this position due to interference with the flow of the Derwent Glacier by tributary ice from the Traveller Range. However, the preglacial Derwent River may have exploited a fault in this position. Hence, the form of the preglacial landscape and the geological structure that gave rise to it are likely to have been predisposing influences in the development of this asymmetry (Figure 15.11).

The strongly rectilinear form of the landscape on the northern part of the Central Plateau gives way in the south to a zone of glacial deposition (Jennings and Ahrad, 1957) where some lake shorelines are curvilinear. These curvilinear shorelines have resulted from the construction of ice pushed ramparts composed of dolerite boulders (Griggs, 1909) as at Lake Ina, or of terminal moraines as at Travellers Rest Lagoon. Glacial deposition has generally

Figure 15.10 Transverse profiles of the Franklin Valley (1) near the Beehive, and (2) near the Franklin-Collingwood junction. In both cases the glacial troughs are cut in Precambrian schists. The differences between in form between the two sites have resulted mainly from the gradient of the thalweg.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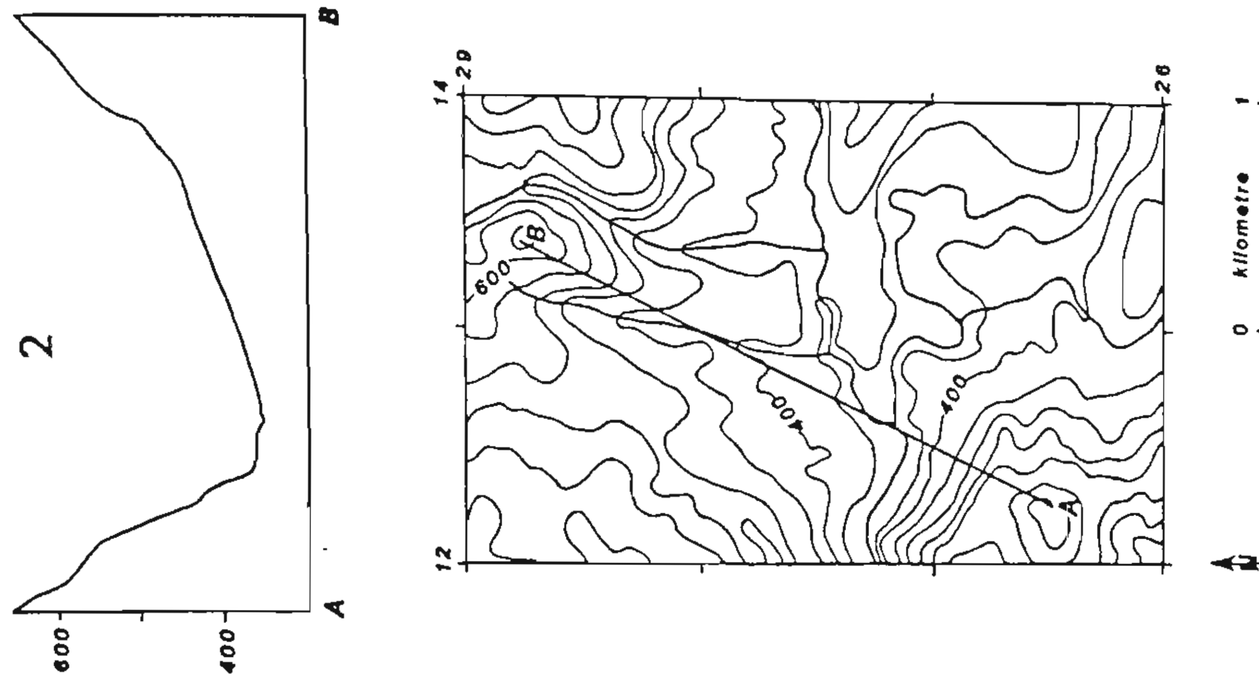
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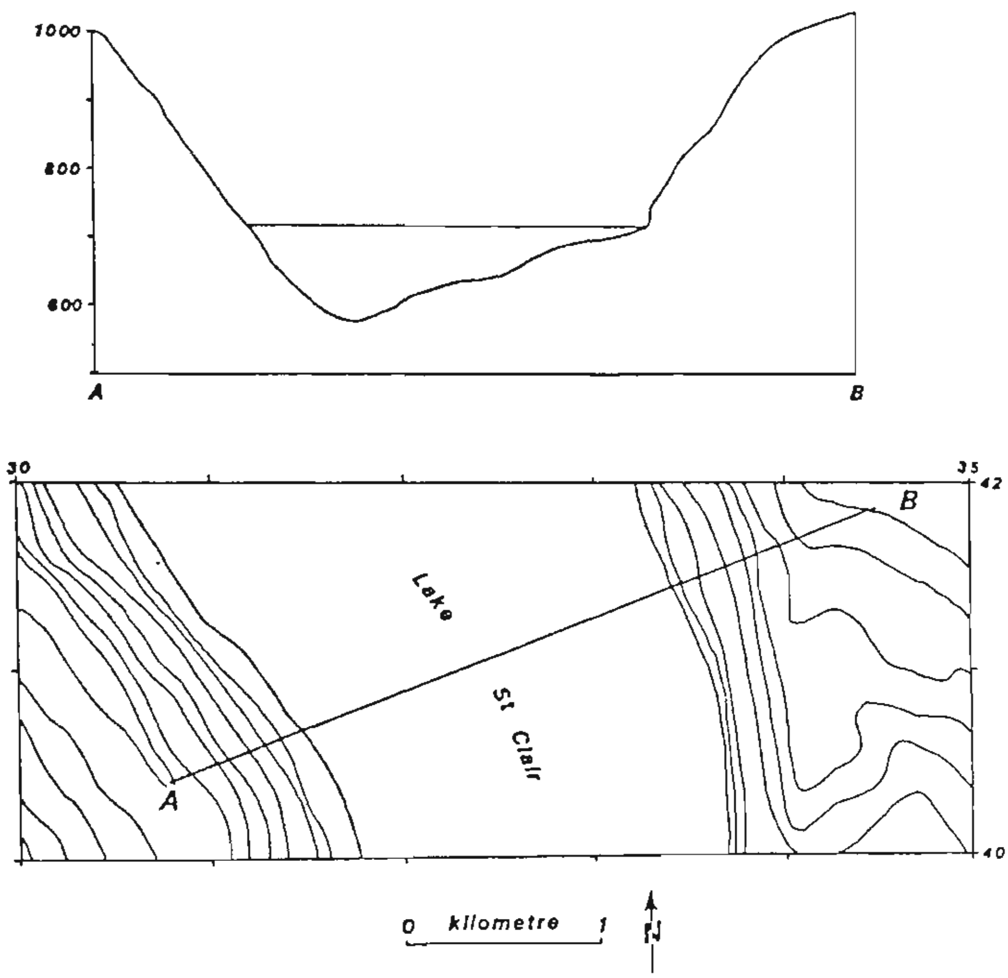

Figure 15.11 Transverse profile of the Lake St. Clair trough (bathymetry after Derbyshire, 1971a).

produced less spectacular modification of the landscape than has glacial erosion. Nevertheless, substantial quantities of material have been deposited and the piedmont moraines constructed by the Derwent and Franklin glaciers on the St. Clair surface are striking in plan. Numerous lakes such as Travellers Rest Lagoon are solely the result of impoundment behind end moraines, while others such as Lake St. Clair are rock basins that have been increased in depth by the construction of a moraine barrage.

D The impact of climate

The high ranges that trend north-south across the westerly airstreams have been deeply eroded on the leeward side by glaciers. The importance of snowfence orientation to the pattern of glacial erosion has been well emphasised by Derbyshire (1968b). The intensity of glacial erosion is related to the trend of the regional snowline which is generally lowest in the southwest of Tasmania and rises to the northeast (Davies, 1969). A similar pattern is evident on a smaller scale on the Central Plateau (Jennings and Ahmad, 1957). Local variation from this pattern occurs where low saddles permit moist maritime air masses to penetrate furher inland (Porter, 1977).

In broad terms the contrast between deep cirques in western Tasmania and more shallow open cirques in the east has previously been interpreted as being due largely to the type of glaciers that developed. This view asserts that the glaciers in the west developed under more maritime conditions and were more active than the glaciers in the east (Derbyshire, 1967; Peterson, 1968). The character and distribution of the glacial and periglacial deposits has also been seen as a response to differences in glacial climate (Derbyshire, 1972, 1973). Field evidence gathered during the preparation of this thesis generally supports this view. A minor qualification is that the clay-rich dolerite slope deposits and tills in the east were probably more prone to soliflcution than the quartzite and schist mantles of the west and hence the two are not directly comparable.

Most of the evidence points towards temperate glaciological regimes. However, the saprolitic dolerite that underlies the Cascade Moraine in the Derwent Valley suggests that some local anomalies may exist. The depth of weathering and the presence of deeply weathered dolerite deposits beneath the glacial deposits further downstream indicates that the weathering occurred prior to the development of glaciers in the late Cainozoic. The saprolite has not been eroded away by the glacier that extended far downstream of this point. Its survival cannot be explained as the result of its having been protected from erosion by the surrounding topography. If the ice was actively eroding its substrate then the saprolite may represent a remnant of a previously much deeper weathered profile. Alternatively, the presence of cold based ice has sometimes been invoked to explain the preservation of deeply weathered rock or delicate subaerial forms (Gauthier, 1978; Dyke, 1983). Hale (1958) raised the possibility that the material beneath the Cascade Moraine may have been frozen, but noted that the shape of the depression in its surface indicated that it was soft at the time the overlying till was deposited. However, this depression, and the laminated clays that occur at the site, need only reflect the thermal conditions at the glacier snout at the time the Cascade Moraine was constructed.

The possibility of cold-based ice has only once before been suggested anywhere in Tasmania (Caine, 1966) but this suggestion has since been discarded (Caine, 1983). Because cold based ice is claimed to have still had the capacity to deform competent sandstone in at least one case (Grant, 1981) it does not neccessarily provide a perfect explanation for the preservation of the saprolite in the Derwent Valley. Nonetheless, predominantly warm based glaciers may have localised cold based areas (Weertman, 1961; Sugden and John, 1976), and this could account for its survival.

GLACIAL CONDITIONING OF SUBSEQUENT NONGLACIAL PROCESSES.

The glaciers of west central Tasmania generated meltwater and glaciofluvial sediment that facilitated erosion on steep proglacial slopes. The sediments have insulated the substrate against weathering and erosion where the slope was more gentle. The impact of the glaciers themselves upon the patterns and processes of postglacial change in the landscape appears to have been even more profound. Once again, these patterns of change varied according to the rock materials, the stage of development, and the processes and climate.

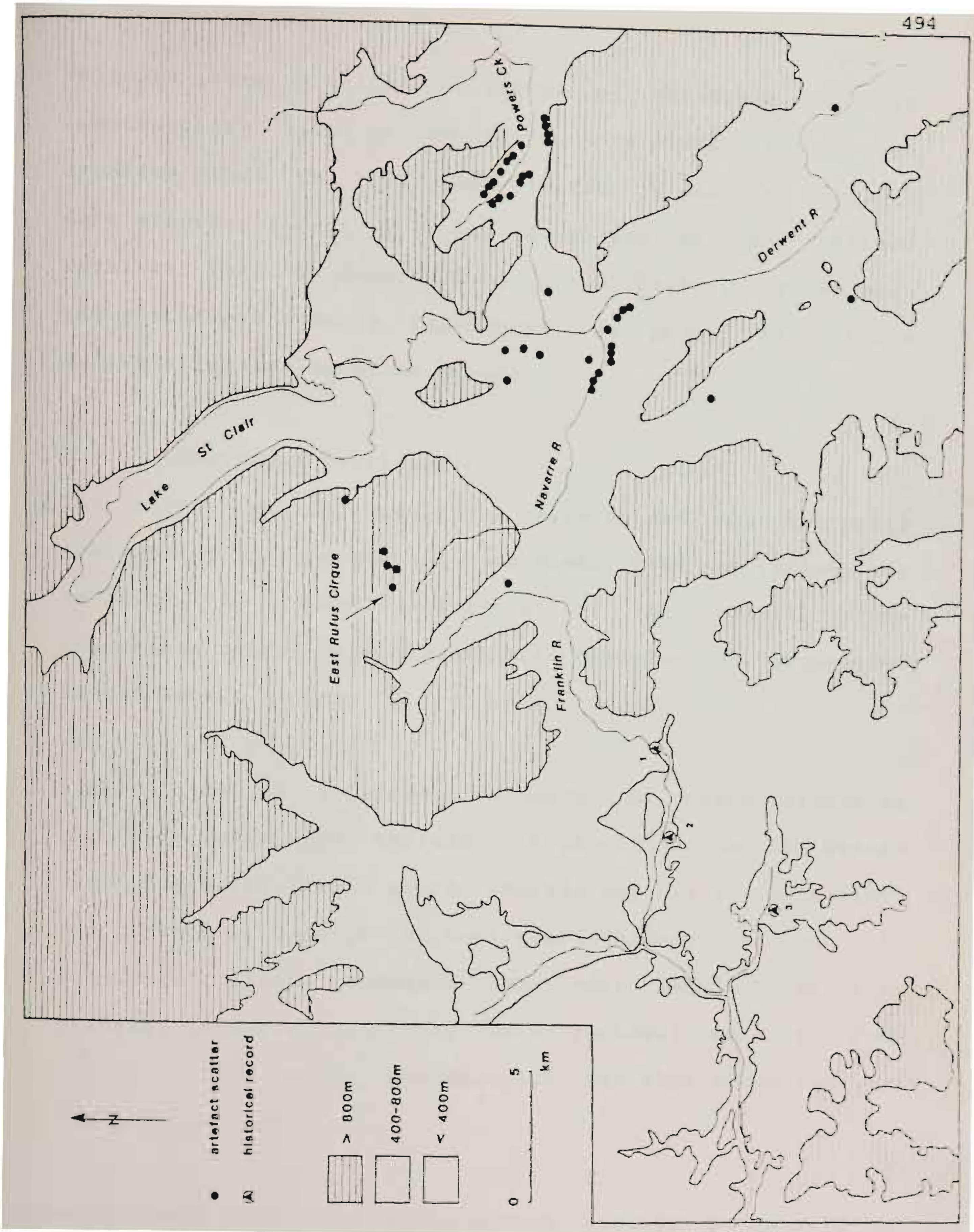

Figure 15.12 Distribution of archaeological sites recorded during this study, and ethno-historical evidence of aborigines in the area. (1) huts (Calder, 1849); (2) huts (Burn, 1842); (3) burnt plain attributed by Surveyor Sharland to aboriginal firing (Binks, 1981).

deposits have provided a source of materials for road construction. Soils are developing more slowly on outwash deposits than on tills that are rich in silt and clay. A more vigorous forest cover has developed on well drained moraines than has developed on outwash plains. Hence, past glaciation has greatly influenced the growth of forests suitable for exploitation by man.

B Historical development

The chronology of nonglacial events and the sediments produced by the geomorphic processes that prevailed was reviewed in chapter thirteen of this thesis. This discussion merely outlines selected examples of sequential development.

Rockfall is an important mechanism of slope retreat in glacially steepened terrain. Rockfall talus is scarce within the ice limits of the Cynthia Bay phase, but litters the surface of some of the innermost moraines, as at Lake Richmond. This suggests that most rockfall activity predates deglaciation, that some rockfall occurred from trough walls during deglaciation, and that there has been limited rockfall since.

Large scale slab toppling is almost totally restricted to sites that have been subject to glacial steepening (figure 15.13) (Clemes, 1925; Derbyshire, 1973). Its absence within the ice limits of the Cynthia Bay phase is consistent with the evidence from elsewhere in Tasmania. Topples in the Mole Creek area have been mapped as part of the Central

Plateau Formation which is regarded as pre Last Glacial in age (Kiernan, 1984), while those at Ben Lomond may not have stabilised until 18 ka BP (Caine, 1983).

Where glacial steepening is less advanced processes of postglacial weathering and erosion have proceeded more slowly. Toppling is absent from the tops of plateaux, although dilation trenches occur near some plateaux margins, often with rock crevasses behind them as on Mt. Hugel. The concentric dilation of mammalated dolerite surfaces has not been recorded within the ice limits of the Cynthia Bay phase, which suggests that it occurs slowly. Therefore, while some slope failures in steep terrain follow rapidly upon the loss of supporting ice, processes associated with dilation on the margin of the steep terrain occur over a very much longer time scale. It is these processes, principally slab toppling, that have most significantly modified the morphology of the cliffs.

C Process and climate

Where free faces are present the retention of moisture is inhibited. Hence, mechanical weathering is likely to be of major importance in the recession of free faces. This is likely to be most rapid during episodes of cold climate. Caine (1983) has proposed that the dolerite at Ben Lomond can support vertical faces of up to 120 m, but that this height can be considerably extended when ice supports the columns. The importance of dilation and collapse in the re-establishment of a stable profile in areas that have been glacially steepened has already been stressed. The stability of the topples during the Holocene interglacial suggests that their collapse is associated largely with cold glacial climates. Wedging by snow, ice and rock that accumulated in dilation features is probably of major importance. Similarly, almost all the rockfall talus has accumulated during periods of cold climate.

Fluvial erosion, including landslide activity, is the most important mechanism of retreat where slopes consist of unconsolidated glacial sediments. Landslides on Mt. Olympus and in the Cephissus trough probably occurred in response to pore water pressures on the edge of rock benches and along stream banks having exceeded critical limits during storm events. The widespread failure of steep moraine slopes on Mt. Olympus since a wildfire in the mid 1970's emphasises the importance of the vegetation in maintaining slope stability. The stability of glacial deposits on slopes would have been low while the climate was cold and the vegetation cover was sparse, as the evidence from Mt. Arrowsmith shows. Open conditions would also have facilitated the movement of aborigines through the landscape, their fires probably having further destabilised the material on the slopes by removal of the vegetation.

The deposition of coarse gravels by vigorous meltwater streams that derived their bedload from areas where vigorous mechanical weathering was operative, gave way during the early Holocene to the deposition of finer calibre materials that were the product of chemical weathering (Colhoun, 1982). In large measure this transition would have been a response to the development of a more continuous vegetation cover as the climate ameliorated. Primary production of coarse sediment was much diminished, although considerable bank erosion by streams has led to reworking of earlier deposits and the solute load is likely to be much higher under the warmer Holocene climate (Slaymaker and MacPherson, 1977).

Stripping of the regolith from more gentle slopes by glacial erosion has exposed the underlying rock to weathering processes from which it would otherwise have been insulated. Hence, the result of glacial erosion has been to expedite non glacial processes. As exposure to subaerial weathering follows retreat of the ice, this weathering is likely to occur under warm conditions and be predominantly of a chemical nature. Evidence from Lake Helios indicates that glacial striae form readily on dolerite, so their absence from most of the area is probably due to chemical weathering (Jennings and Ahmad, 1957). Their removal implies lowering of the landscape by at least a few millimetres since deglaciation and this is confirmed by the presence of sandstone pedestals beneath dolerite erratics on the Gould Plateau (Figure 4.11).

The Central Highlands are subject to only very limited periglacial activity under the present climate. Solifluction lobes and terraces occur on some summits and the evidence from Mt. Rufus indicates that one terrace is actively advancing. Lobes and terraces in alpine environments are commonly assumed to be periglacial landforms that develop when downslope movement of the regolith is maintained by frost heave and gelifluction (Washburn, 1969). While diurnal freezing on Mt. Rufus is sufficient to form pipkrake and disturb the regolith, and the soil may freeze to depths of a few centimetres under extreme conditions, there is no evidence for deeper seasonal freezing that would form an impermeable substrate facilitative of gelifluction in the manner described by Benedict (1976). The mean annual temperature on Mt. Rufus is unlikely to be less than 3.5°C which is 5.5° higher than the temperature considered neccessary by Karte and Liedtke (1981) for gelifluction lobes and terraces to develop. While insufficient data is available to permit an accurate estimate of the speed at which the Mt Rufus terraces are advancing, a mean rate of up to 3-4 cm /yr seems possible. This is much more rapid than is commonly the case for gelifluction terraces (Benedict, 1976), possibly because the clay-rich mantles on this dolerite summit are facilitative of solifluction. Solifluction does not demand the presence of ice (Andersson, 1906) and can occur even in tropical environments (Ruxton, 1969), while stone banked lobes and terraces can form in desert environments, probably due to the shrinking and swelling of clays (Figure 15.14). These considerations suggest that fossil terraces similar to those at Mt.Rufus which occur on other dolerite mountains in Tasmania are not reliable palaeoclimatic indicators.

The most significant changes in the landscape during the Holocene have been wrought by man. Frost heaving of the

Figure 15.13 Part of a dilation trench in dolerite on the glacially-steepened margin of the Mt. King William I plateau.

Figure 15.14 Stone banked terraces near Mt. Hopeless on the Strzelecki Track to Coopers Creek, northern South Australia. These terraces have formed in a desert environment close to present sea level, probably in response to the shrinking and swelling of clays.

Figure 15.15 Soil erosion after logging operations and the burning of slashed debris on State Forest near Dyes Marsh.

soil has been facilitated by fires and construction activity that have exposed the regolith. Construction activities have physically altered the landscape and some forms of land-use, such as timber harvesting, have promoted considerable soil erosion (figure 13.15). Recreational use of walking tracks in the national parks has also led to some localised soil erosion (Calais, 1981)

SYNTHESIS

The processes of change in this landscape have not occurred evenly over either space or time. It is difficult to assess in more than qualitative terms the importance of glacial erosion in remodelling this landscape and in conditioning postglacial change. Table 15.1 estimates the rates of change in more qualitative terms and attempts to indicate at least their order of magnitude. These figures give an overall impression of stability prior to the onset of glaciation and during the Holocene but greater instability during the glacial stages. Caine (1982) has calculated that slab toppling during the last 100 ka. has led to recession of the free faces around Ben Lomond at the very rapid rate of 0.25 mm/yr. Given the similar form of some of the dolerite mountains in west central Tasmania to those in the northeast of Tasmania it is reasonable to assume that recession by toppling occurred here at a broadly similar rate. This emphasises the fact that long term estimates of denudation based on short term process measurements invariably obscure the episodic nature of geomorphic change (Gage, 1970). Because of these facts Table 15.1 should be interpreted with some caution

EROSION RATES

regime	site	substrate	description	Extent of development	assumed occurrence (Ka BP)	rate (B)	areal scale (m^2)
interglacial	Gould Plateau	sandstone	development of pedestals	5cm	14-0	3.6	10^{-1}
	Mt. Rufus	sandstone	development of solution pans	30cm	14-0	21.4	10^{-1}
	Dyes Spur	solifluctate	soil erosion after logging	10cm(est.mean)	0.01-0	10000.0	10^2
late glacial	Mt. Gould	sandstone	widening of dilation trench	10m	14-10	2000.0	10^2
primarily glacial	Dervent Gorge	dolerite	glaciofluvial erosion	15m	1000-10	15.0	10^2
	Mt. Geryon	mudstone & dolerite	backwearing of cirque	100m	1000-10	101.0	10^3
	DuCane Gap	dolerite	lowering of diffluence col	150m	1000-10	151.5	10^3
	Lake St. Clair	Permian sediments	deepening of rock basin	160m	1000-10	161.6	10^3
	Surprise Valley (at Top End Gap)	quartz schist	deepening of trough	50m	1000-10	50.5	10^3
	Lake George	dolerite & mudstone	deepening of trough	180m	1000-10	181.9	10^3
	Lake Rufus	mudstone	deepening of trough	400m	1000-10	404.0	10^3
	Lake Richmond	mudstone	deepening of trough	360m	1000-10	363.6	10^3
	Mt.King William I	dolerite	widening of dilation trench	50m	130-10ka.	416.7	10^2
	Lake St. Clair	dolerite	retreat of faultline scarp	1600m	1000-10	1616.2	10^3 (a)
	Wentworth Hills	sandstone & dolerite	retreat of faultline scarp	1300m	1000-10	1313.1	10^3 (b)
glacial & pre-glacial	Lake St. Clair	dolerite	retreat of faultline scarp	1600m	65000-0	24.6	10^3 (a)
	Wentworth Hills	sandstone & dolerite	-"-	1300m	65000-0	20.0	10^3 (b)
primarily pre-glacial	Dyes interfluve	dolerite	-"-	400m	21600-0	18.5	10^3
	Dyes re-entrant	dolerite	-"-	1013m	21600-0	46.9	10^3

DEPOSITION RATES

regime	site	sediment	thickness	assumed occurrence (Ka BP)	rate (B)	areal scale (m)
interglacial	Mt. Gould tarn	organic silt	0.85m	7.9-0	107	10^1
	Marcissus R.	alluvium	1.1m	7.6-0	144	10^2
	Laughing Jack Lagoon	organic silt	0.5m	0.8-0	625	10^2
	Dareys Bluff tarn	"	1.25m	10-0	125	10^2
	Upper Nive River	alluvium	1.0m	10-0	100	10^2
glacial	Bedlam Wall	solifluctate	2.0m	20-12	250	10^3
	"	frost shattered debris	0.8m	19-14	160	10^2
	Cynthia Bay	till	40.0m	19-14	8,000	10^3

Table 15.1 Estimated rates of erosion and deposition at selected sites. Rates are in Bubnoff units (m.3/km-2./yr-1. or 1 mm./1 ka. denudation) (Fischer, 1969)

Nonetheless, it is clear that erosion and deposition proceeded far more rapidly under the cold conditions of the late Cainozoic than during preglacial and postglacial time. That these rapid processes of change were unable to fundamentally alter the landscape emphasises the great antiquity of the theatre in which these events unfolded.

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A Geological and morphological influences

The principal consequence of glacial erosion has been a steepening of mountain slopes by cirque formation. This has increased the stress placed upon rock materials by gravity. Longitudinal valley profiles have been flattened or overdeepened, which has reduced the stress along the thalweg but increased the stress upon valley walls. The most probable result of increases in stress is enhancement of sediment production and transport, while reductions in stress are likely to have the opposite effect. Glacial deposition has generally decreased the stress but glacial deposits that were unstable following glacier retreat have been reworked. The glacial geomorphology has also influenced the modification and use of the landscape by man.

The effect of glacial steepening in the mountains of west central Tasmania has varied according to rock type. Where steep slopes have been cut in Permian and Triassic rocks there is often evidence of only limited postglacial rockfall although dilation following the release of confining ice pressure has produced small 'rock crevasses' and dilation trenches on Mt. Gould. Steep slopes cut in massive Jane Dolomite in the Surprise Valley also appear to have remained stable although minor dilation is again evident in a few places. In contrast, the vertical jointing in the dolerite has facilitated failure of steepened slopes on a large scale.

The erosional morphology has been important to human activity. The deepening of Butlers gorge and the formation of the Powers Creek gorge by meltwater erosion provided damsites that could be used for hydro-electric energy. The resultant storages have changed the face of the landscape by inundating about $50km^2$, providing sediment sinks and locally decreasing stream energies. Former meltwater routes have been re-established as part of these developments, as at the Rufus Canal, while larger scale diversions such as the Butlers Gorge canal have been made possible.

The nature and distribution of glacial deposits has influenced their response to postglacial processes. Where rock materials were deposited in steep terrain on the margins of valley glaciers their reworking by solifluction or fluvial processes has been widespread. Even where slopes were less steep clay-rich dolerite tills have been reworked by solifluction. The thick slope deposits on the Hedlam Wall ridge must consist mostly of reworked till although this origin is no longer discernible from their general character. The identity of glacial deposits on the plateau to the east has been lost in a similar manner.

The distribution and morphology of glacial deposits has also influenced the impact made by man. Glacial transport has presorted rock materials and deposited the most resistant fine grained rock types in accessible locations, thereby attracting prehistoric man and his fires (figure 15.12). The piedmont moraines have provided well drained routes across the St. Clair Surface which in modern times have been utilised in road location. Glacial and glaciofluvial

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Addenda

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- W.E. Bowman (1956)
<u>The Ascent of Rumdoodle</u>
Parrish, Sheffield.

Kiernan, Kevin

Late Cainozoic glaciation and mountain geomorphology in the central highlands of Tasmania