MOUNTAIN GEOMORPHOLOGY AND THE LAST GLACIATION AT LAKE ST CLAIR

by Kevin Kiernan

(with four text-figures)

KIERNAN, K.R., 1992 (31:x): Mountain geomorphology and the Last Glaciation at Lake St Clair. *Pap. Proc. R. Soc. Tasm.* 126: 47–57. ISSN 4703–0080. Department of Geography and Environmental Studies, University of Tasmania, GPO Box 252C, Hobart, Tasmania, Australia 7000.

During the Last Glaciation, the glaciers that developed in the Lake St Clair area were largely controlled by the topography and played only a minor role in shaping the landscape, the most striking glacial features of which were produced during earlier glaciations. The glaciers were subject to continental influences, which resulted in some distinctive depositional and erosional landforms, but were still of essentially temperate type. In apparent contrast to those in the West Coast Range, the glaciers in the Lake St Clair area retreated systematically in clear stages.
Keywords: glaciation, palaeoclimate, Lake St Clair, Tasmania, geomorphology, landforms, till, cirque, glacial trough, mountains.

INTRODUCTION

The scenery of the Lake St Clair area in the Cradle Mountain–Lake St Clair National Park owes much to past glaciation. Lake St Clair was formed by the Derwent Glacier downstream of the confluence of the Narcissus, Cephissus (Pine), Marion and Hamilton tributary glaciers (all of which arose in the Du Cane Range), where that glacier was joined by ice that moved westwards from the ice-abraded Central Plateau (Jennings & Ahmad 1957). This paper explores the glacial geomorphology of the upper catchment area of the Derwent Glacier (fig. 1). It also examines the uppermost reaches of the adjacent Murchison, Mersey and Nive valleys in an effort to identify the glacial divides during the Last Glaciation, known locally as the Cynthia Bay Glaciation (Kiernan 1985, 1991) and regarded as equivalent to the Margaret Glaciation of western Tasmania.

The area lies amidst some of Tasmania's highest mountains, the study area ranging in altitude from c.1470 m in the Du Cane Range to c. 700 m south of Lake St Clair. These mountains consist of Jurassic dolerite that overlies Permian and Triassic sedimentary rocks of the Parmeener Supergroup (Banks 1973). Former erosion surfaces are conspicuous features of the landscape. Davies (1959) recognised five principal accordances, together with higher monadnocks. The most conspicuous surfaces are the St Clair Surface (730–825 m), which forms the plains south of Lake St Clair and the floors of some of the larger valleys, and the Lower Plateau Surface (915–1065 m) and Higher Plateau Surface (1200–1350 m) that are represented on the Central Plateau and Mt Olympus. The monadnocks form some of the higher summits, including Mt Geryon and the Acropolis in the Du Cane Range. The present daily median temperature at the southern end of Lake St Clair (735 m) is 7.95°C and annual precipitation totals about 1200 mm.

Moraines were first recognised in the area by Gould (1860) and the glacial origin of Lake St Clair was recognised by Officer *et al.* (1895). Clemes (1925) suggested that four glaciations had taken place, but Lewis (1939, 1945) argued that there had been only three glaciations. Subsequent studies on the adjacent Central Plateau prompted the conclusion that only one glaciation had occurred (Jennings & Ahmad 1957), while Derbyshire (1963) also interpreted glacial deposits that occur downstream of Butlers Gorge as representing a single glaciation. However, Derbyshire (1967, 1968) later suggested that two glaciations had occurred. In other papers, Derbyshire described additional aspects of the glacial and periglacial geomorphology of the Lake St Clair area (Derbyshire 1965, 1968a, 1968b, 1971, 1972, 1973; Derbyshire *et al.* 1965). The most recent studies have increased the evidence for multiple glaciation of the area (Kiernan 1983, 1985, 1989, 1990a, 1991). Some previous researchers into past glacial environments and the effectiveness of glaciation as an agent of landscape change have interpreted all the landforms and sediments as the product of a single glaciation, the Last Glaciation (e.g. Peterson & Robinson 1969, Davies 1969). More recent studies (Kiernan 1985, 1991) have permitted the differentiation of features attributable to different glaciations (fig. 2), and this paper provides an assessment of the extent, impact and nature of the Last Glaciation.

EROSIONAL LANDFORMS

A variety of erosional landforms occurs in the mountains of the Lake St Clair area (fig. 2). A composite valley-head cirque 3 km wide lies at the head of the Narcissus Valley. This cirque is floored by Permian sediments, and its headwall is cut in Triassic sandstone that is overlain by columnar dolerite. The pronounced scalloping of the north–south aligned ridge of Mt Geryon and the Acropolis, which forms the western margin of this cirque, indicates that this was a particularly important part of the Du Cane Range snowfence. Rock type has exerted a major control on the form of the cirques in this area. Whereas headward sapping of the Triassic sandstone beneath the dolerite columns has permitted the development of a steep 500 m-high headwall in the Narcissus Valley, the head of Cephissus Valley cut only in massive dolerite is steep and mammillated rather than vertical. A steep arête occurs between the two valleys. The floors of the Marion and Hamilton cirques are also cut in sedimentary rocks of the Parmeener Supergroup and, again, have vertical headwalls.

Each of these valley-head cirques has its floor at 900–950 m and a headwall 350–500 m high. Only in the Marion Cirque, where there is a small rock basin lake 13 m deep

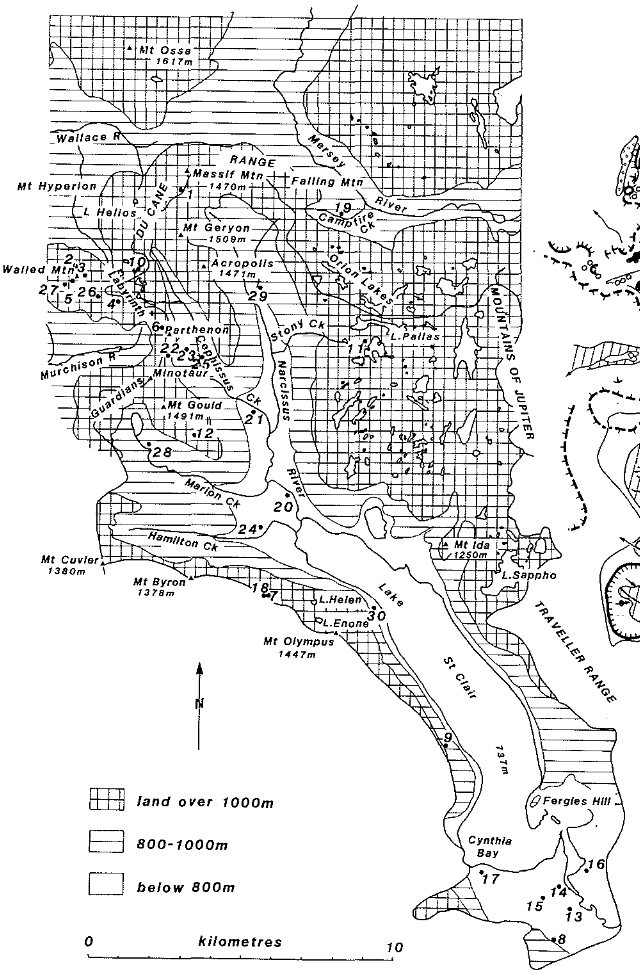

FIG. 1 — Locality map of the Lake St Clair area, showing the catchment of the Derwent River above its confluence with the Travellers Rest River, and mountain areas adjacent to its headwaters. The numbers refer to sites at which weathering rind measurements were made.

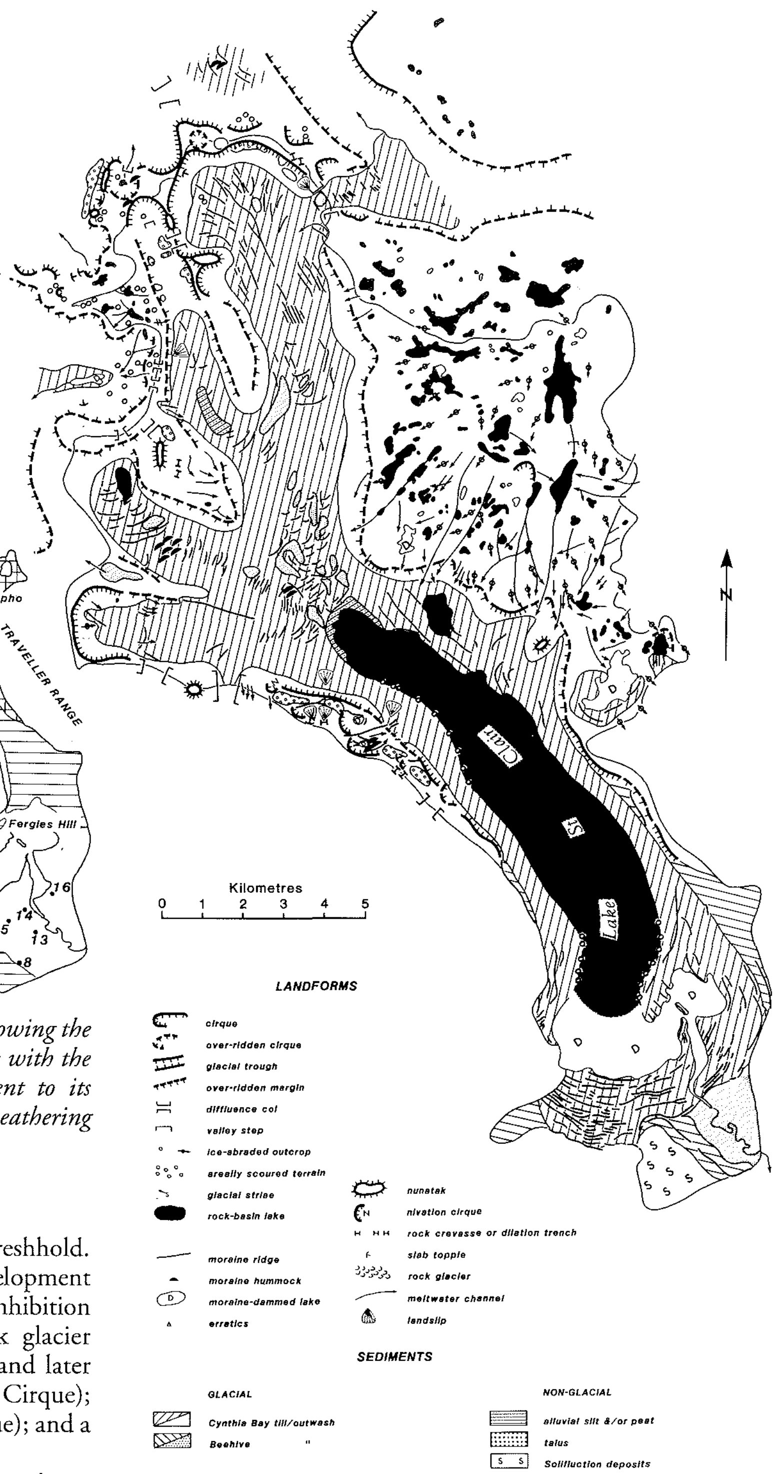

FIG. 2 — Glacial geomorphology and sediments of the Lake St Clair area.

(Derbyshire 1967), is there a clearly defined threshhold. Several factors have limited the degree of cirque development and conditioned the character of these cirques: the inhibition of tributary ice flow into the very large trunk glacier (Hamilton Cirque); over-riding of the headwall and later some downwind interception of snow (Cephissus Cirque); an unfavourably aligned snowfence (Marion Cirque); and a low and broken valley head (Hamilton Cirque).

Geological structure has played a major role in cirque location. Lake Helen (1190 m) and Lake Enone (1210 m), the latter a rock basin, occupy two prominent cirques on a sandstone bench on the eastern flank of Mt Olympus (Derbyshire 1964, 1968a). A number of smaller cirques are also present at about the same altitude further north on this mountain. Preglacial topography, in the form of benches on which snow could accumulate leeward of the snowfences, is likely to have been a major determinant of their location (Peterson 1969).

A series of cirques occurs on the slopes of the Du Cane Range at the head of the Murchison and Mersey Valleys. They indicate that the range formed a major glacial divide between ice that flowed south to Lake St Clair, west into the Murchison Valley and north to the Mersey Valley. A shallow and totally enclosed cirque, 200 m in diameter, occurs on the summit of Massif Mountain. This has a distinct bedrock lip on the northern side. The cirque drains into a pit 15 m in diameter and 8 m deep which appears to be focussed on a major joint intersection. While the development of this curious streamsink may be due to very localised joint dilation, there is little evidence to support this. An alternative is that it is the product of solution processes acting upon autometasomatic alteration products in the joints (Hale 1958, Spry & Hale 1964).

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

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

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

The bedrock geometry and preglacial topography have clearly been responsible for determining the location of the glacial troughs. South of Mt Ida, the Lake St Clair trough is bounded by dolerite to the east and subhorizontal sedimentary rocks to the west. The base of the dolerite lies over 300 m above the lake on Mt Olympus but descends below lake level at the foot of the Traveller Range which raises the possibility that the trough may exploit a fault. The orientation of structural lineaments in the dolerite of the Central Plateau is duplicated in the orientation of the tributary valleys upstream from Lake St Clair, which suggests that these valleys have been superimposed from the dolerite (Jennings and Ahmad 1957). Because the major erosion surfaces (Davies 1959) predate the glaciations, the suggestion that the principal valleys have been superimposed carries with it the corollary that the broad topography of the area is essentially the product of fluvial erosion.

Broad U-shaped troughs are well developed in sandstone and mudstone. In contrast, narrower troughs are cut in dolerite and descend to the St Clair Surface from the Central Plateau and at the head of the Cephissus Valley. The Lake

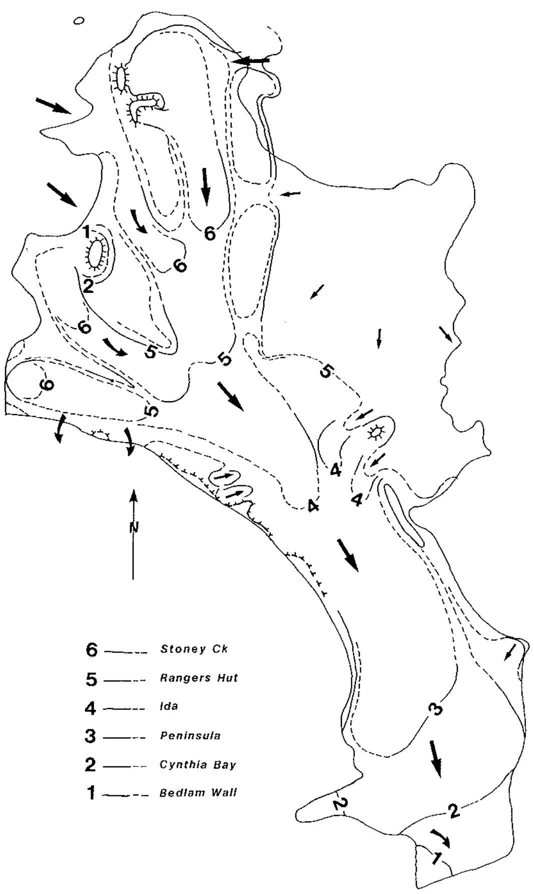

FIG. 3 — Principal ice limits during different phases of glaciation in the Lake St Clair area: scale as figure 1.

St Clair trough is strongly asymmetric and has a steep western wall. The deepest part of the lake occupies a steep, straight trench that trends NNW–SSE close to the western shoreline. This is cut in Triassic rocks. At this point, the rock bastion of Fergies Hill deflected the ice flow westward towards a re-entrant in Parmeener rocks, where the Cuvier Valley joins the St Clair trough. The rock bastion probably owes its origin largely to interference with the flow of the Derwent Glacier by ice that spilled over the margin of the Central Plateau. Therefore, the most marked over-deepening of the trough may not reflect the amount of ice that descended from the Central Plateau so much as its energy, and its interference with the main ice flow southwards (cf. Derbyshire 1971).

Du Cane Gap is a transfluence col through which ice escaped southwestwards from the Mersey Valley into the head of the Narcissus trough. A further col occurs between Mt Ossa and Mt Massif, and ice-abraded rock surfaces indicate that ice flowed eastwards through this from the upper Murchison catchment, where a very large volume of ice accumulated (Kiernan 1990a), into the Mersey Valley, which broadens appreciably downstream. However, moraine ridges that are concave to the east were constructed east of this col during the most recent phase of glaciation and indicate that diffluence through the col did not occur at that time. Cols between Mt Olympus, Mt Byron and Mt Cuvier exhibit stoss and lee features, which indicate that ice moved southwards through them from the Hamilton Valley into the Cuvier Valley. Similar evidence indicates that ice from the Cuvier moved into the St Clair trough through a col on Mt Olympus.

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

Smoothed bedrock on the summit of Walled Mountain suggests that it may have been overridden by ice in the upper Murchison catchment when the glaciers were most extensive. These summit rocks, together with other abraded dolerite surfaces at 1250 m on the eastern ridge of Walled Mountain, are exfoliating in slabs 0.05–0.2 m thick. Glacial erosion may also have played a significant role in stripping overlying rocks from the lithological benches of Triassic sandstone developed on Mt Gould and Mt Olympus.

Glacial striae are generally poorly preserved, due to the susceptibility of the rocks to chemical weathering. However, numerous striations on dolerite one metre below the normal water level of Lake Helios indicate that ice moved northwards into the valley of the Wallace River, a tributary of the Murchison River. Striations that trend towards 190°–205° on sandstone bedrock just beneath the surface of a small tarn on the Mt Gould Plateau, together with plucking of nearby outcrops, indicates that the Cephissus Glacier flowed across the plateau from the north. The generally southward flow of this ice across the Marion and Hamilton valleys towards the cols at the head of the Cuvier is supported by bedrock plucking and one lunate fracture. Lake Helios undoubtedly contained ice during the most recent glaciation, but the possibility that the Gould Plateau striae are earlier forms which have been exhumed must be kept open (cf. England *et al.* 1981).

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

The present outlet of the Derwent River from Lake St Clair was probably initiated subglacially on the eastern margin of the Derwent Glacier, where confining ice pressures were lowest (Derbyshire 1971), and probably was a proglacial channel cut through the drift before adopting its present form.

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

Impressive free faces characterise many of the scarps in the Lake St Clair area, especially in the Du Cane Range and on Mt Olympus. While many of these have resulted from direct erosion by glaciers, some have resulted from the mechanical adjustment of rock materials to unloading or basal sapping in response to previous glacial erosion. On a small scale, such processes give rise to concentric shells such as are found on the summit rocks of Walled Mountain. Slopes formed on rocks with a high degree of structural anisotropy, particularly the vertically-jointed dolerite, are prone to fail due to joint dilation, once valley sides or cirque headwalls are steepened. The cambering and ultimate disintegration of columns due to this process of slab toppling (Caine 1982) is evident on Mt Olympus (Clemes 1925, Derbyshire 1973) and in the Du Cane Range where it has effected considerable local topographic change. The development of rock crevasses and dilation trenches that parallel the steepened slope is the first obvious sign of this process. Both are present above the Narcissus Cirque and on Mt Olympus. The process is not confined to dolerite terranes as rock crevasses and small gullies formed parallel to the valley sides also occur at 1100 m on the sandstone eastern slopes of Mt Gould. The "Big Gun" in the Du Cane Range west of Massif Mountain is a large leaning topple that has slipped downslope and is close to the point of outward collapse.

Toppling is most apparent in those areas where there has been glacial erosion of sedimentary rocks that underly the dolerite. Unloading is probably the principal cause, but periglacial processes, including wedging by ice, snow and rocks, probably play a role in dislodging columns (Banks 1981, Derbyshire 1973). Removal of alteration products in joints may also be significant in some cases (Hale 1958). Some topples may be directly of glacial origin. Six metre high columns lean outwards at 30°–50° on the plateau margin just south of Mt Hyperion and have clearly been pushed by a small glacier that constructed a small blocky moraine against their foot.

Occasional fresh scars on dolerite cliffs and freshly broken rocks at their foot suggest that only limited erosion by rockfall is occurring at present. A large fresh scar on Falling Mountain delineates a recent collapse over an area of c.5 ha. Some steep eroded gullies between dolerite columns on Mt Geryon and Walled Mountain are probably the result of snow and rock avalanche. Snowpatch erosion occurs to a very limited extent, as on the northern end of Mt Olympus. It is far more effective on Triassic sandstone than on dolerite.

Solution pans have developed on the dolerite summit rocks of Walled Mountain (1411 m), and pans, runnels and sub-erratic pedestals are present on the Triassic sandstone benches of the Gould Plateau (1070 m). Present-day fluvial erosion is restricted largely to reworking of unconsolidated deposits. Some debris avalanching has occurred adjacent to a creek which descends steeply to the floor of Pine Valley southeast of the Parthenon, and landslide scars and hollows are common in areas of steeper terrain on Mt Olympus and elsewhere. Erosion runnels have developed on regolith bared by nivation in the Narcissus Cirque and on Mt Olympus. Small-scale erosion and collapse of stream-banks is widespread, as exemplified along the course of Cephissus Creek in Pine Valley.

Erosion scars and collapsed areas around parts of the Lake St Clair shoreline have resulted from the artificial raising of the lake level to facilitate generation of hydro-electricity. This has focussed erosion higher on the shore profile than was naturally the case. Moreover, this exposure has opened the moraines to the operation of other erosional processes from which they would otherwise have been insulated by the presence of a vegetation or peat cover. These processes include sapping of the bared banks by groundwater discharge when drawdown of the lake water level occurs. Erosion due to the raised lake level has also caused retreat of the dunes at Frankland Beaches, which have been further eroded by sand quarrying and the construction of a road along their crest. Serious localised erosion has also occurred along walking tracks and in areas of the Central Plateau subject to repeated fires and sheep grazing.

DEPOSITIONAL LANDFORMS AND SEDIMENTS

The distribution of depositional landforms and sediments is strongly influenced by the broad topographic framework of the area (fig. 2). The Central Plateau adjacent to Lake St Clair is dominated by erosional landforms (Jennings and Ahmad 1957). Few moraines are to be found there, which Davies (1969) suggested might be due to retreat having been gradual rather than episodic. The steep walls of the St Clair trough and the presence of Lake St Clair mean that glacial deposits are accessible only in the tributary valleys and beyond the southern margin of the lake. The principal end moraine complexes enable several phases in the Last Glaciation to be identified (fig. 3).

Moraine Complexes

(1) Bedlam Wall moraine

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

The Bedlam Wall moraine marks a phase during which the Derwent Glacier terminated c. 3 km south of Lake St Clair, close to the site of the present Derwent Bridge settlement. The moraine is well preserved but the till is weathered to a depth of at least 1.2 m. Small till knobs south of Mt Olympus, sandstone erratics at 1040 m on the western side of the Parthenon, and a talus that formed in contact with an ice margin at 1260 m on the northeastern shoulder of Mt Gould all appear to lie c. 150 m above any of the more recent ice limits and may relate to this Bedlam Wall phase. The Bedlam Wall deposits generally lie inside the maximum ice limits suggested by the erosional topography but outside the succeeding Cynthia Bay moraines. The Bedlam Wall till was deposited during the Beehive Glaciation, which occurred prior to the Last Interglacial Stage (Kiernan 1985, 1991).

(2) Cynthia Bay moraines

An impressive array of at least 25 terminal moraine ridges and latero-terminal moraines bounds the southern shoreline of Lake St Clair. The southernmost of these moraines is located 1 km from the lake shore and is believed to represent the terminus of the Derwent Glacier during the late Last Glacial Stage. These narrow and steep moraines do not exceed 10 m in height, and on average are spaced 50 m apart (fig. 4). Some can be traced almost continuously for 3.6 km. They were formed by the expanding foot of the Derwent piedmont glacier as it emerged from the St Clair trough. Given the short duration of the Last Glaciation in Tasmania (Kiernan 1983) and the likelihood that retreat of the ice occurred very rapidly, these moraines, and others of similar form that occur further upvalley, may represent annual fluctuations, although slightly longer periods could be involved.

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

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

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

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

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

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

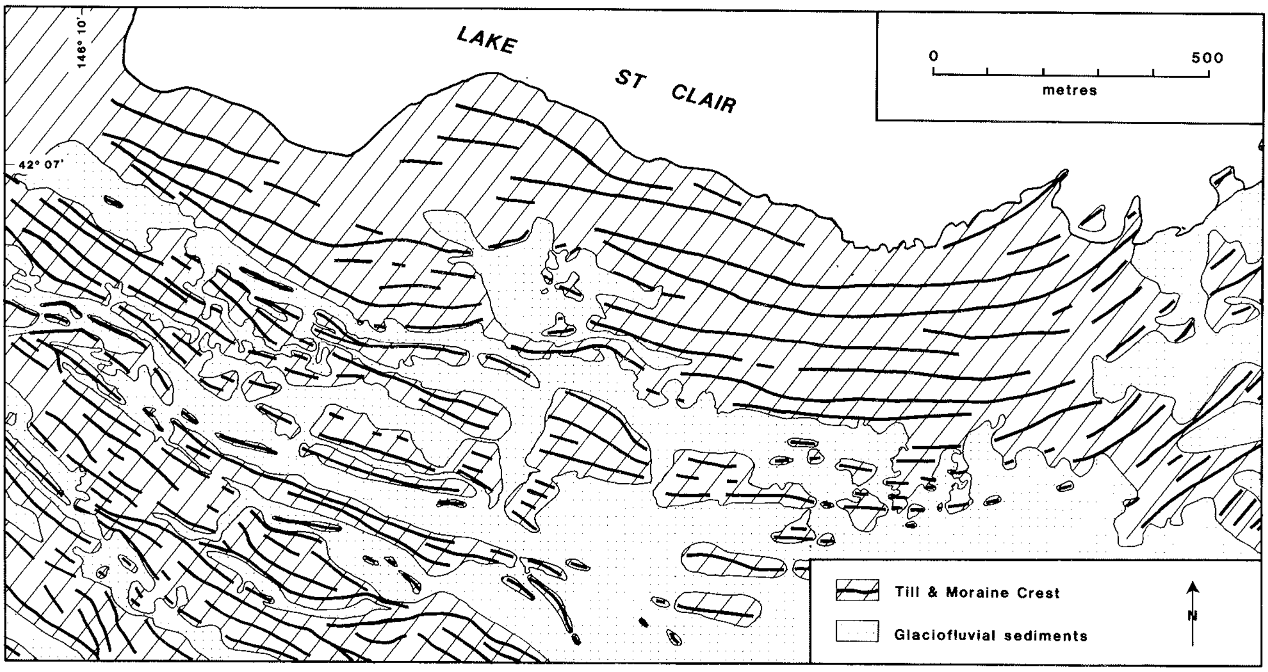

FIG. 4 — Moraine ridges in the Cynthia Bay area.

The coarser grades of the outwash gravels are dominated by dolerite, the finer grades by quartz. The dolerite cobbles are moderately well rounded and of moderate to high sphericity, but angularity increases in the finer grades, presumably due to water cushioning (Hatch *et al.* 1971, Blatt *et al.* 1972). The larger clasts occur in a loose-textured matrix which is commonly bright yellowish brown (10YR7/6). Sections cut by the Derwent River reveal foreset beds, truncated bedding and cross bedding indicative of seasonally torrential meltwater flow. Interbedded fine gravel, sand and silt lenses are common. Massive clays, either in the ice contact stratified drift or within intermorainal swales, are of glaciolacustrine origin.

The individual clasts in the moraines are little weathered and the soil profile can best be described as of A Cox Cu-type (Birkeland 1984). Nowhere do more than 0.3 m of slope deposits overlie the Cynthia Bay till. X-ray diffraction led Derbyshire to suggest that some kaolinite was present in the till matrix but the peak on his diffractograms is very close to background and has not been duplicated by analyses during the present study. All these characteristics suggest the Cynthia Bay till is considerably younger than the Bedlam Wall till. Its weathering characteristics compare closely to tills from elsewhere in Tasmania that are known to date from the late Last Glacial Stage (Kiernan 1983, 1990b).

(3) Peninsula moraines

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

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

(4) Ida moraines

Discontinuous lateral moraines descend to Lake St Clair from 800 m altitude beneath Mt Ida. These document the most recent occasion during which the Ida glaciers descended into the Lake St Clair trough. The moraines that were formed on the southern side of the North Ida Glacier are sharply recurved southwards, whereas lateral moraines of the Derwent Glacier occurring on its northern side appear to have been undisturbed by the Ida ice. This suggests that the largest of the glaciers to descend from the Central Plateau to Lake St Clair was deflected southwards by the Derwent ice.

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

(5) Hamilton Valley moraines

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

These moraines have an amplitude of less than 6 m. A fan of outwash extends into the northwestern corner of Lake St Clair.

(6) Rangers Hut moraines

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

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

(7) Marion Valley moraines

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

(8) Cephissus Valley moraines

The Cephissus Glacier constructed numerous moraines on rock ribs between the mouth of the Cephissus Valley and the Pine Valley Plains. The innermost of these moraines also exhibits kettle holes. The Cephissus ice flow then ceased, possibly due to the loss of any significant ice input from The Labyrinth. Two intermorainal outwash surfaces are overlain by 2 m of massive gleyed clays deposited by ponded meltwater (Jennings 1959). Intervening hummocky moraine indicates that deposition of the clays occurred in two stages (Derbyshire 1963).

(9) Stony Creek moraines

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

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

Unequivocal fluted drift occurs along c. 4.5 km of the valley floor, from 850 m to 1040 m, and indicates that the ice continued to maintain downvalley flow, even during the final retreat of the glacier terminus. This fluted moraine contrasts with the moraines that exhibit kettle holes and the hummocky moraines which indicate that the ice flow died in the tributary valleys. The fluting originates in the head of the Narcissus Valley and does not extend towards Du Cane Gap. This suggests that the accumulation of ice in the lee of the high Geryon–Acropolis snowfence, rather than continued diffluence through Du Cane Gap, lies behind the comparative longevity of the Narcissus Glacier.

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

Non-glacial deposits

The nature and distribution of non-glacial deposits assists in determining the extent of the former ice cover, the age relationships between the diferent phases of glaciation, and the geomorphological evolution of this glacially modified area, subsequent to deglaciation. Diamictons mantle many of the slopes outside the Cynthia Bay ice limits but are scarce within those limits. They reach a depth of 2 m on the upper slopes of the Bedlam Wall ridge. These deposits consist of fragments of local bedrock or clasts up to 2 m or more in size that have been reworked from till, and occur in a silty clay matrix. They have moved over slopes of as little as 8°. These diamictons are interpreted as having been produced by periglacial solifluction.

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

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

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

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

Alluvial, peat and dune deposits

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

Minor paludal deposits have accumulated with fibrous peats on the button-grass plains. Glacial sands have been worked into beaches around parts of the shoreline of Lake St Clair, most notably at the Frankland Beaches at the southern end of the lake. Dunes up to 4 m high have been constructed by aeolian action behind the southern beaches, in response to northwesterly winds down the length of the lake during the Holocene. This is the largest dune system available for study on a glacial lake in Australia, following inundation of the larger complex at Lake Pedder. Unfortunately, the dune system at Lake St Clair has been subject to unnatural erosion, quarrying and modification of its morphology by construction activities.

DATING AND DISCUSSION

The differentiation of glacial deposits of different ages is facilitated by their relative positions in the landscape and by the degree of postdepositional modification to which the landforms and sediments have been subject. The use of postdepositional modification, particularly contrasts in clast weathering and in degrees of soil development, as an aid to dating Tasmanian glacial deposits has been reviewed elsewhere (Kiernan 1990b). The glacial deposits of the Lake St Clair area generally occur within previously ice-eroded landforms. They do not define the limits of the ice which was responsible for that erosion. While some summits remained nunataks, including the Acropolis' eastern ridge and Mt Gould, equivocal erosional evidence suggests that, during a very intense early phase of glaciation, the summit of Walled Mountain (1410 m) and many of the summits of the Du Cane Range may have been overridden by ice. The similar width-to-depth ratios of the glaciers and their host cirques suggest that the form of the glaciers was controlled by the cirques (Graf 1976), which emphasises the influence of the pre-existing topography. The surface gradient of the Derwent Glacier near the southern end of the St Clair trough declined eastwards, which suggests that snowfence effects and shading were important influences upon its form. It also suggests that the glaciers which descended into the trough around Mt Ida were of less significance as sources of ice than were the cirques at the head of the Narcissus Valley and its tributaries. Indeed, only a comparatively small part of the Central Plateau ice cap is ever likely to have drained into the Derwent (cf. Derbyshire 1963). Retreat of the Derwent Glacier to the Rangers Hut limit was brought about largely through the decline of the Central Plateau ice cap, which resulted in a halt to the passage of ice from the Mersey Valley into the Derwent via Du Cane Gap.

Weathering rinds formed on subsurface clasts in the Bedlam Wall till vary in thickness from 0.5 to 7.4 mm thick, the mean value and standard deviation being 2.7 ± 1.51 mm. Rinds in the Cynthia Bay till range from 0.1 to 4.6 mm, the mean and standard deviation being 1.5 ± 0.46 mm. While the weathering rinds formed on subsurface dolerite clasts in the Bedlam Wall till are little thicker than those in the Cynthia Bay till, there is a major contrast between the soil profiles that have formed on these two tills. The clay-rich B horizon of the soil, developed primarily by pedogenic weathering of the Bedlam Wall till, is consistently at least three times thicker than in any of the moraines formed subsequently. This indicates that the Bedlam Wall moraines are considerably older than the Cynthia Bay till. During the Bedlam Wall phase of the Beehive Glaciation, the Derwent Glacier extended well south of Lake St Clair. The talus that accumulated in contact with the ice on the northeastern shoulder of Mt Gould at this time indicates that the Cephissus Glacier was a least 400 m thick at this point.

Moraines that abut talus 1 km south of Lake St Clair and in the Lake Helen cirque delineate the maximum extent of ice during the Cynthia Bay Glaciation. At the peak of the Cynthia Bay Glaciation, the Derwent Glacier extended 27 km from its headwall in the Du Cane Range. Fresh rock crevasses and ice-abraded surfaces on the Gould Plateau indicate that the ice was 300 m thick west of this point. Diffluent ice from the Derwent Glacier passed through two cols northwest of Mt Olympus into the head of the Cuvier Valley until the end of the Peninsula phase. This ice surplus existed partly because the Derwent Glacier was supplemented by the diffluent lobe of the Mersey Glacier, which flowed through Du Cane Gap at the head of the Narcissus Valley, and the discharge of some ice from the western margin of the Central Plateau into the Narcissus Valley.

Postdepositional modification of the landforms and sediments suggests that the glacial deposits as far downstream as the Cynthia Bay moraines do not differ significantly in age. All have very immature soil profiles of A Cox Cu-type (Birkeland 1984), in which a thin A horizon has formed over parent material that has been oxidised but undergone little chemical weathering. The average thickness of dolerite weathering rinds is less than 2 mm. The diamictons that are interpreted as solifluction deposits occur almost exclusively outside the Cynthia Bay ice limits, and the stratigraphically uppermost of them are weathered to a degree comparable to the Cynthia Bay till. This suggests that they are approximately equivalent in age.

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

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

Radiocarbon assay of charcoal fragments, obtained from 0.2 m above the base of 1.3 m of deltaic silt that overlies glaciofluvial sediments near the mouth of the Narcissus River, indicates that the climatic amelioration that allowed organic-rich silt to commence accumulating occurred prior to 7650 ± 250 BP (SUA 2079). The fragments of *Athrotaxis selaginoides* at the base of the organic silt in the Gould Plateau tarn have been radiocarbon assayed to 7920 ± 250 BP (SUA 2080). Evidence from elsewhere in the state suggests that actual deglaciation of the highest cirques was complete well prior to this time (Macphail & Peterson 1975), probably by about 9 ka BP.

The degree of postdepositional modification and the morphologic and stratigraphic context of the Cynthia Bay drift bears comparison with that of the Dante drift in the central West Coast Range, which was deposited shortly after 18 800 ± 500 BP (ANU 2533) (Kiernan 1980, 1983). Similar criteria suggest that these tills are also equivalent in age to the Dixon till in the Franklin Valley (Kiernan 1989), the Rowallan till in the Mersey Valley (Hannan & Colhoun 1988) and the Oakleigh till in the Forth Valley (Kiernan and Hannan, 1991). Hence, it is with these formations that comparison with the Cynthia Bay drift might best be made, to evaluate the nature of the glacial environment and events during the late Last Glacial Stage.

During the maximum of the Cynthia Bay Glaciation, the Equilibrium Line of the Derwent Glacier lay at c.1020 m. This implies a depression of the mean annual temperature by c. 6.2°C compared to its present value. This estimate compares closely with the figure of 6.5°C calculated by Derbyshire (1973) on the basis of the limits of rock glaciers on the western slopes of Mt Olympus. It is also comparable to the figure proposed by Kiernan (1980) for temperature depression during the Dante Glaciation.

The regular spacing and size of the Cynthia Bay moraines recalls the form of De Geer moraines, which form some distance behind an ice front that calves into water (Stromberg 1965). However, while meltwater may previously have been impounded behind the Bedlam Wall moraine, further downstream a large lake would have been neccesary for the Cynthia Bay moraines to have formed in this manner, and there is no evidence for more than localised ponding of meltwater. The development of the Cynthia Bay moraines probably more closely resembles that of Thule–Baffin moraines (Embleton & King 1975), cross-valley landforms that have been described from the wasting margins of cold ice caps in the Thule area of Greenland (Weertman 1961) and on Baffin Island (Andrews & Smithson 1966). The general absence of shear planes in sections exposed along the Cynthia Bay road and Derwent River suggests that the Cynthia Bay moraines are not subglacial forms that have developed in the general manner proposed for Rogen moraines, which involves the shearing of ice over the moraine crest (Goldthwait 1951, Derbyshire 1965, Prest 1968).

Weertman (1961) argued that Thule–Baffin moraines are formed where debris is brought to the ice surface in regelation ice. This is consistent with the limited shearing at Cynthia Bay and is essentially the interpretation favoured by Derbyshire (1965). The Cynthia Bay moraines lack the alluvial fans that commonly form on the distal face of Thule–Baffin moraines (Goldthwait 1951). The reason for this may lie in the thermal regime of the glaciers. Thule–Baffin moraines are formed on the edge of cold ice sheets, where there may be only a limited amount of basal debris entrained. Active temperate ice is likely to entrain more basal debris than will cold ice and, upon reaching the surface, this will form a thicker debris cover and will more effectively insulate the ice, thereby retarding melting and flowage.

The moraine complexes in this area attest to systematic retreat of the glaciers in clear stages, during which abundant proglacial sediments were deposited inside the earlier ice limits. This contrasts with the pattern of retreat during the Last Glaciation interpreted from Tasmania's West Coast Range by Colhoun (1985), who proposed that the large, narrow end moraines in that area were formed by glaciers that were maintained at the maximum glaciation by an approximate balance between accumulation and summer ablation that persisted for a considerable time, after which deglaciation was rapid. The slightly more continental glaciers in the Lake St Clair area may have been more susceptible to minor climatic changes, perhaps particularly in summer ablation conditions, than the very maritime glaciers in the mountains further west. The regular spacing and limited evidence for shearing suggests that the moraines upstream of the Cynthia Bay terminal moraines are all retreat features. However, insufficient sections are available to confirm this impression. Fitzsimons (1988) has suggested that the Dante Advance in the King Valley may have been preceded by an earlier advance during the Last Glacial Stage, but the evidence for this remains equivocal. The postdepositional data from the Lake St Clair area do not suggest there is any significant age difference between the tills deposited during the Cynthia Bay and Stony Creek phases and do not indicate preservation of evidence for an early Last Glacial Stage advance.

Both erosional and depositional landforms attest to the differences between the glaciers at Lake St Clair and those further west. The shallow and open cirques of the Lake St Clair area contrast with deeper cirques and rock basins in the western mountains, which are the product of higher rates of accumulation and ablation and more vigorous mass throughput. The development of moraines of possible Thule–Baffin type at Lake St Clair reflects the slightly more continental character of glaciation in this area compared to those in the more westerly mountains. These moraines are quite unlike the broader and more massive moraines that were formed further down the Derwent Valley during earlier glaciations (Kiernan 1985, 1991). No moraines of Thule–Baffin type have been recorded in the more maritime areas nearer the west coast. Abundant meltwater in the terminal zones of the glaciers around Lake St Clair is attested to by their composition of up to 75% bedded materials. The evidence from this area suggests limited precipitation, cooler temperatures than in the west, and cryogenesis by congelation (Derbyshire 1963, Peterson 1968, 1969). Nevertheless, the rounded clasts in the basal tills indicate that meltwater was abundant at the glacier sole (Driemanis 1976).

The extensive tract of fluted moraine in the upper Narcissus Valley indicates that the Narcissus Glacier remained active even during its final retreat. In contrast to this, the other tributary valleys contain small end moraines and hummocky drift, which implies that the ice there became inactive and decayed *in situ*. This contrast in late stage glacial deposition indicates that the high Geryon–Acropolis snowfence and the shading from the afternoon sun that it offered were critical factors during the Stony Creek phase.

This analysis of the glacial legacy of the Lake St Clair area therefore indicates that, during the Last Glaciation, the glaciers here were controlled by the topography and played only a comparatively minor role in shaping the landscape, the most striking glacial modification of which occurred during earlier glaciations. Other non-glacial landforms such as slab topples may date at least in part from periglacial episodes during the late Last Glacial Stage but they were facilitated by glacial steepening that occurred earlier. Erosional and depositional evidence indicates that there were slight continental influences on the glacial environment, but that the glaciers were still essentially of temperate type. In apparent contrast to the situation further towards the west coast, however, these inland glaciers retreated systematically in clear stages. The ice flow slowed significantly or stagnated in some smaller valleys, while remaining active in more favourable accumulation sites. No evidence has yet been detected that would suggest any phases of readvance interrupted this steady retreat from the maximum ice limits as the climate ameliorated.

ACKNOWLEDGEMENTS

The initial phases of this study were undertaken as part of my PhD studies in the Geography Department at the University of Tasmania under the supervision of Dr E. A. Colhoun and with support from the Australian Department of Education under the Commonwealth Postgraduate Research Awards Scheme. For support, advice and assistance during this phase of the study, I am grateful to Dr Les Wood, Dr Michael Macphail, Mick Iloski and Karen Kiernan. For assistance in various ways during both this and later phases of the study, I also wish to thank the Department of Parks, Wildlife and Heritage, particularly Dick Dwyer and Steve Hepworth, and Dr Mick Brown of the Forestry Commission.

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(accepted 14 February 1992)