A reconnaissance of the geomorphology and glacial history of the upper Gordon River Valley, Tasmania

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Abstract

A reconnaissance of the land forms and Quaternary sediments of the uppermost Gordon River Valley and adjacent Guelph River Valley indicates that at least two and probably three glaciations are represented. The most striking landforms produced by glacial erosion of the King William Range appear to date from glaciations that predate those recorded during this survey. When the ice cover was most extensive, the Gordon Glacier was nourished partly by diffluent ice from the Derwent and Guelph Valleys, but during the late Last Glacial Stage the Gordon Glacier was nourished primarily by local sources. The glacial sediments as far as 11 km from the head of the Gordon Valley are weathered to only a moderate extent and appear much younger than the heavily weathered drifts that define the maximum glacier limits in almost all other Tasmanian valleys studied previously. This suggests that the tills identified to date in the Gordon Valley are not likely to represent the maximum down-valley extent of the Gordon Glacier.

Introduction

The Gordon River is Tasmania's largest river system. Its basin of nearly 5000 km^2 contains many glaciated mountain ranges and valleys. The pioneering reconnaissance works of Peterson (1968, 1969), Derbyshire (1967, 1968), and Derbyshire *et al.* (1965) have provided most of the scant data concerning the glacial history over most of the area. The glacial sediments of the Gordon River basin have not been studied in any detail apart from those in the uppermost tributaries of the Franklin River Valley (Kiernan 1989a). This Gordon River basin constitutes a very major resource for glacial studies, with important implications for understanding Late Cainozoic climate change, the consequences for landscape evolution in the area and the nature of the environment exploited by Pleistocene human inhabitants. The Gordon River rises from the eastern slopes of Mount King William II at the southern end of the King William Range. Further north, the eastern slopes of Mount King William I are drained by the Guelph River, a tributary of the Derwent River.

Within the past two decades, important advances have been made in knowledge of the late Cainozoic glacial history of Tasmania. These initially involved detailed studies of the landforms and sediments produced by the ice masses that developed the West Coast Range (Banks *et al.* 1977; Kiernan 1980, 1983a; Augustinus 1982; Colhoun 1985; Fitzsimons 1988; Colhoun and Fitzsimons 1990). More recently, the Central Highlands have become a focus of detailed research (Kiernan 1983b, 1984, 1985, 1990a, 1991, 1992; Hannan 1989; Hannan and Colhoun 1987; Kiernan and Hannan 1991). Less attention has been paid to elucidating the patterns of glaciation and glacial chronologies in the often spectacularly glaciated landscapes of south-western Tasmania. Some broad overviews have been stimulated by land-use controversies (Kiernan 1987; Hannan 1987) but only a very few detailed studies have been undertaken (Colhoun and Goede 1979; Kiernan 1989b, 1990b,c). Collins (1990) has provided a popular account of some of the glacial landforms.

It has previously been suggested that the upper reaches of the Gordon Valley formed part of a major reticulate glacial system on the edge of the Central Highlands ice cap (Kiernan 1985, 1987; Hannan 1987). However, the extent to which outlet glaciers from the Central Highlands penetrated the Gordon River catchment remains to be ascertained.

This paper records a reconnaissance of the glacial geomorphology of the uppermost reaches of the Gordon Valley. The landforms and sediments in the headwaters of the Gordon River were investigated, together with those in the Guelph River Valley immediately to the north, which shares the King William Range snowfence (Figure 1). The results are presented in maps of the Quaternary geology and landforms of these two basins (Figure 2). An integrated map of the glacier limits during the various phases that have been recognised is presented

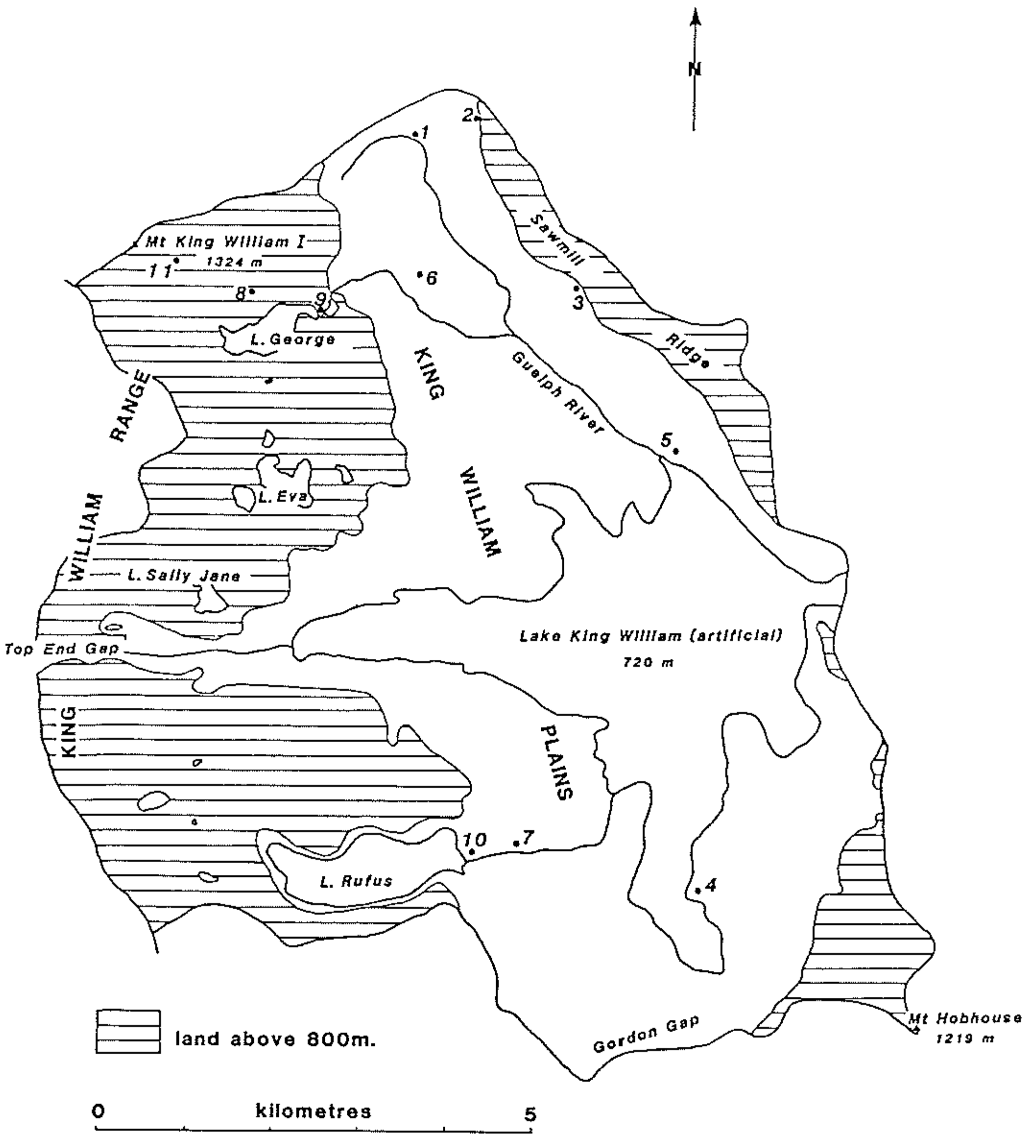

Figure 1a. Locality map of the Guelph Valley. Numbers on map refer to site locations in Tables 1 and 2.

(Figure 3). Mapping in the upper Gordon Valley has only been at a reconnaissance level and no attempt has been made to map comprehensively 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.

Erosional landforms

The preglacial topography of the Gordon and Guelph Valleys is the result of fluvial erosion that has been controlled by structural lineaments. The floor of the Guelph Valley has been eroded into Permian and Triassic sediments west of a fault that trends north–south along the edge of a resistant dolerite ridge that lies between the Guelph and Derwent Valleys (Gulline 1965). Permian and Triassic sedimentary rocks crop out to over 100 m above the northern shore of Lake George, but ice-smoothed dolerite is present on its south-western shoreline. This suggests that the valley which previously existed here may have been eroded along a fault. The upper Gordon River Valley trends south-eastwards from the Guelph–Gordon divide.

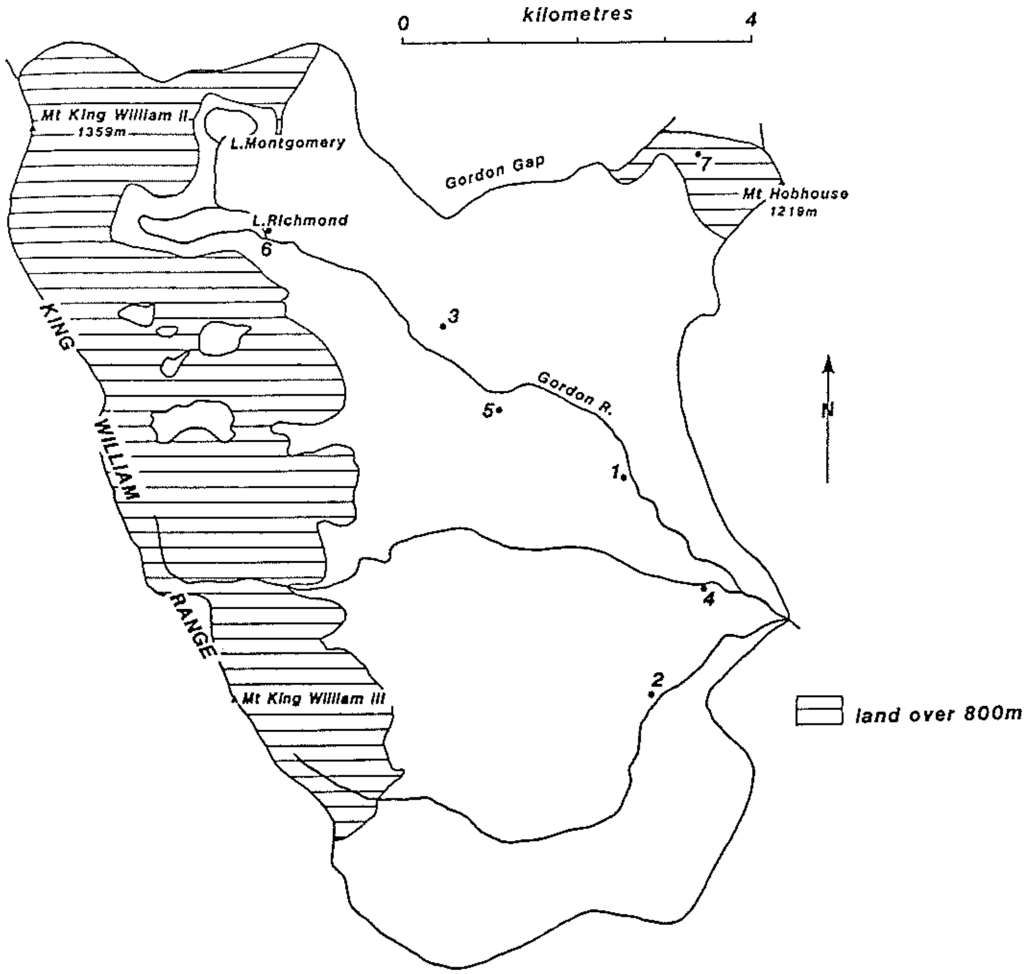

Figure 1b. Locality map of the upper Gordon Valley. Numbers on map refer to site locations in Tables 1 and 2.

Although it is eroded in Permo-Triassic rocks, the Gordon Valley parallels the Derwent River Valley to the east which is largely controlled by joints and faults in the dolerite. This suggests that the upper Gordon Valley has been superimposed from the original dolerite cover rocks that still occur on the crests of Mount Hobhouse and the King William Range.

The Guelph–Gordon divide is strongly asymmetric, with a gentle gradient to the north and a steeper slope southwards 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 Guelph–Gordon divide have been further steepened by erosion on the margin of the glaciers that have flowed out of the Lake Richmond trough.

Mount King William I is separated from the rest of the range by the deep gap of the Top End River Valley. This Top End Gap is developed along a structural lineament which extends east–west through the King William range and which can be traced eastwards to a small gorge towards the downstream end of the Guelph River. This lineament is of uncertain origin but its orientation parallels that of other troughs and ridges on the eastern flank of the King William Range which suggests that it is a major joint or fault in the dolerite. It previously conducted the main branch of the Guelph River eastward through the range until headward erosion by glaciers that flowed from the cirques of the Loddon Range west of the area under discussion in this paper, led to the capture of the Top End River headwaters by the Surprise River, a tributary of the Franklin River.

The King William Range is renowned for its impressive glacial landforms, notably cirques, glacial troughs and a variety of lakes (Peterson 1969; Davies 1969). Broad and shallow cirques have been eroded in dolerite above 1050 m altitude. One such cirque has been cut into the head of a glacial trough above Lake Richmond, and others surround the summit of Mount King William III. Large lateral moraines indicate that glaciers extended from these cirques to low altitudes in some cases (Derbyshire 1968). Ice-smoothed rock on the arms of many cirques indicates that glaciers previously extended beyond the limits of the lateral moraines constructed in the cirques (Peterson 1969).

Overall, the leeward eastern slopes of the King William Range have been markedly steepened by intense glacial erosion. The mean orientation of the cirques is 89°, which reflects primarily the north–south orientation of the range which has acted as a natural snowfence. A low threshold is common. A number of rock basin lakes are present (Photo 1), often bounded by a single end moraine as exemplified at Lake Eva. This contrasts with the general absence of end moraines in other cirques that lack any threshold, further north in the Lake St Clair area (Peterson 1969; Kiernan 1992).

The cirques and glacial troughs that have been eroded into the eastern flank of the King William Range are cut in Permian and Triassic rocks, while sapping at the foot of the overlying sill of columnar dolerite has produced steep and imposing headwalls. Valley steps up to 160 m high occur within some of the troughs. Most of these troughs are occupied by lakes. In the Gordon catchment, Lake Richmond is a narrow rock basin 1.4 km long and up to 33 m deep (Derbyshire 1968) that has been extended by the construction of a moraine barrage at its eastern end (Photo 2). The threshold of the Lake Richmond trough lies a few hundred metres west of the break of slope at the foot of the range. Further north, three glacial troughs are present in the Guelph catchment. The northernmost of these occurs at the head of the Guelph River and contains Lake George. Like Lake Richmond, Lake George is also a rock basin that has been extended by the deposition of a moraine barrage at its eastern end. Bathymetric surveying by Derbyshire (1967) reveals that the deepest

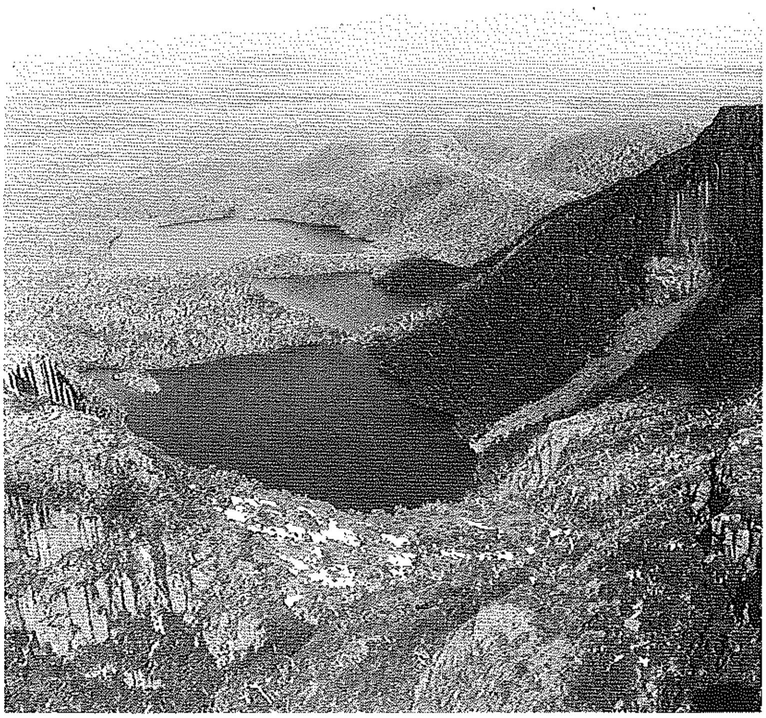

Photo 1. Cirque lakes on the eastern slopes of the King William Range, south of Lake Richmond. During the late Last Glacial Maximum, ice did not extend far beyond the confines of the cirques. Note the recent rockfall into the nearest lake.

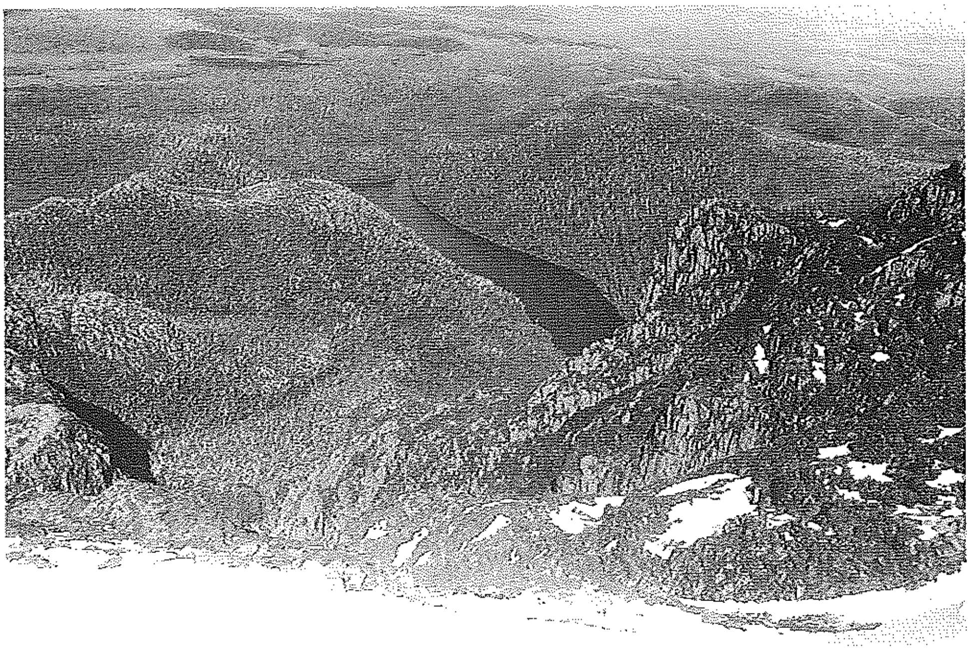

Photo 2. Lake Richmond, source of the Gordon River. Ice-abraded dolerite slopes of the King William Range can be seen in the foreground.

part of Lake George is oriented about 30 degrees more northerly than the main basin. This probably reflects some deflection of the erosion by a structural lineament.

The Top End Valley is also a glacial trough. Extending from the head of the Surprise River west of the King William Range eastwards into the Guelph Valley, this narrow 300 m defile separates the Mount King William I massif from the rest of the King William Range to the south. Shallow valleys that extend downslope of cirques on both walls of the Top End Gap have been deflected eastwards by an ice stream that flowed through the Gap from the west. It is clear from this that a considerable quantity of ice that accumulated in the lee of the Loddon

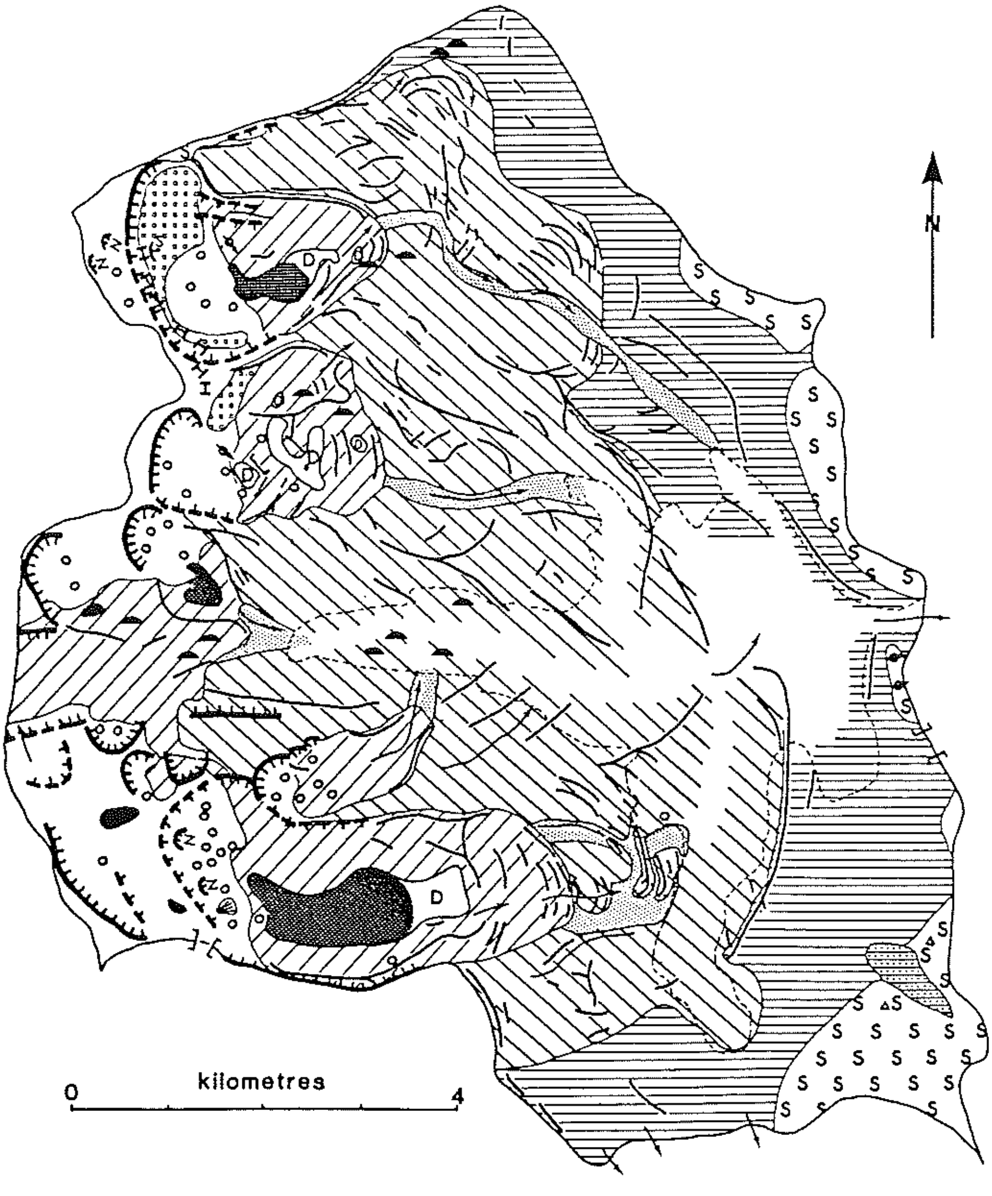

Figure 2a. Glacial geomorphology of the Guelph Valley. For key, see Figure 2c.

Range west of the King William Range at the head of the Surprise Valley passed eastwards through the Top End Gap and supplemented the glaciers that arose in the King William Range. The remainder of the ice from this part of the Loddon Range flowed WNW down the Surprise Valley. Mount King William I was isolated as a nunatak by these glaciers and those in the upper Franklin Valley and Lake St Clair areas to the north,

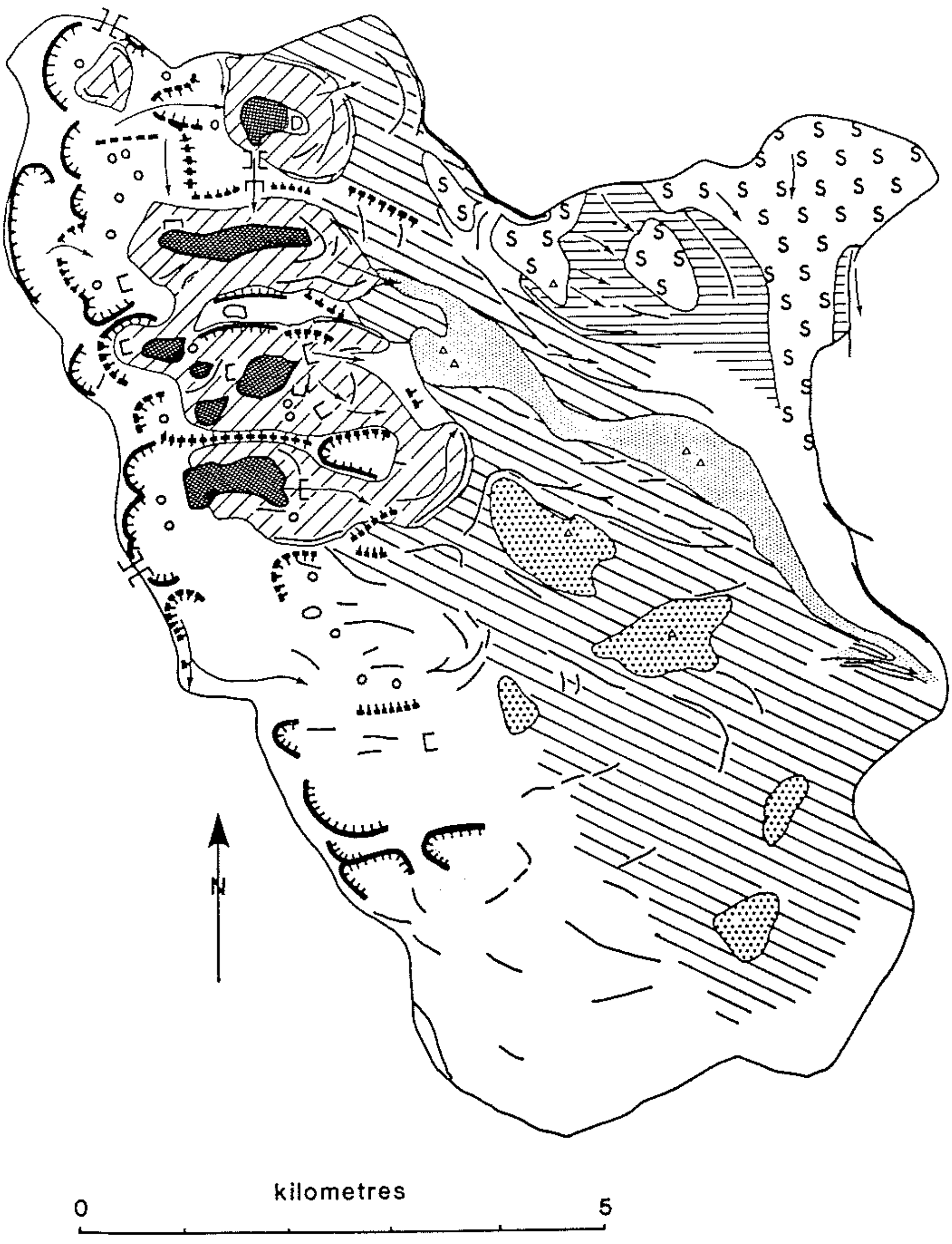

Figure 2b. Glacial geomorphology of the upper Gordon Valley. For key, see Figure 2c.

but a small ice carapace developed on its summit plateau and small cirque glaciers developed on its slopes.

The Lake Rufus trough is perhaps the most visually striking glacial landform in the study area (Photo 3). Lake Rufus lies in a well-developed glacial trough 3 km long and up to 1 km wide. The plateau margin at its head is severely ice-abraded and glacial erosion has left a small 'Half Dome' hill (Derbyshire 1967). The lake lies in a rock basin that once again has been extended by the construction of a moraine barrage. It is about 80 m deep and the trough walls rise 200 m on either side of it. Excavation of this trough has involved the removal of over 50 million cubic metres of rock (Derbyshire 1967). Ice-smoothed

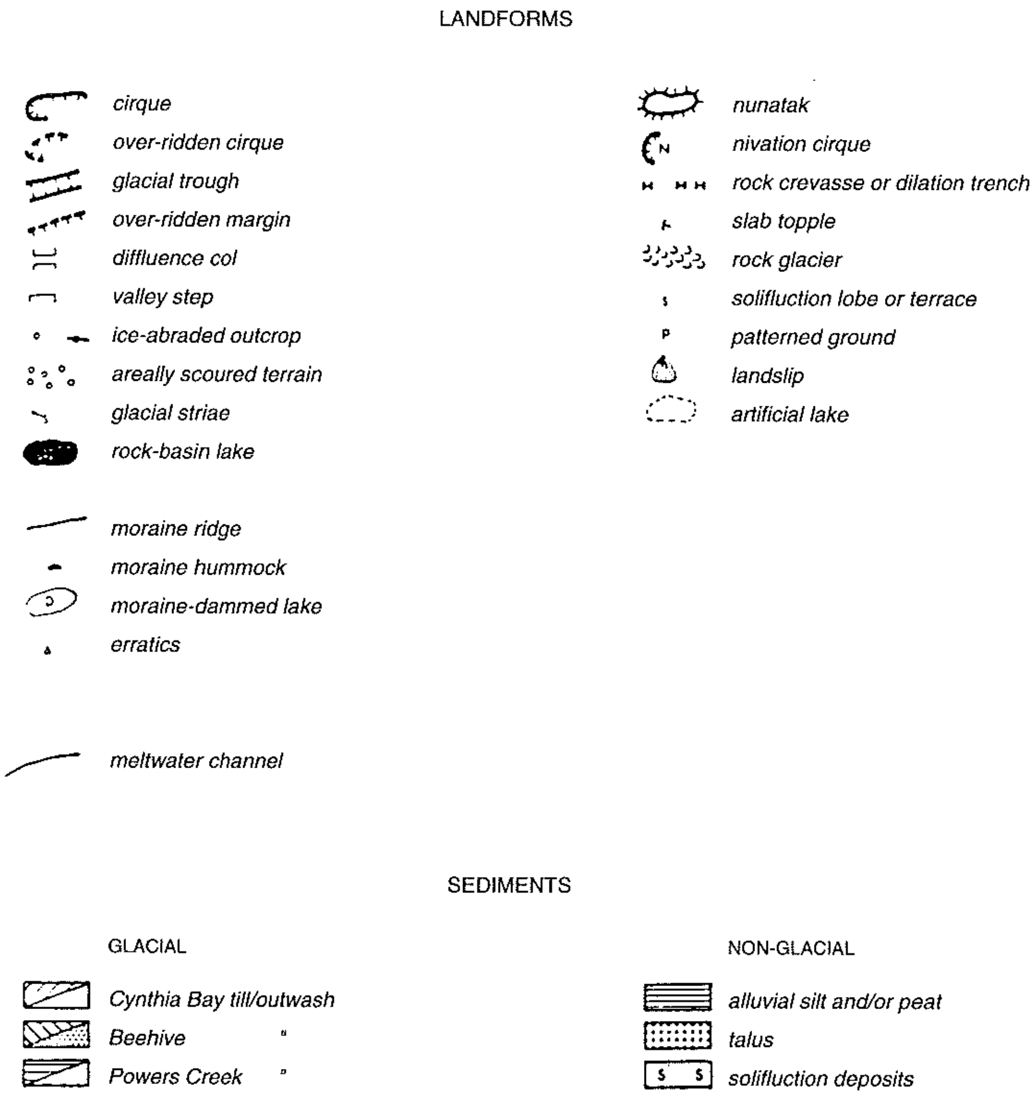

Figure 2c. Key reference for glacial geomorphology.

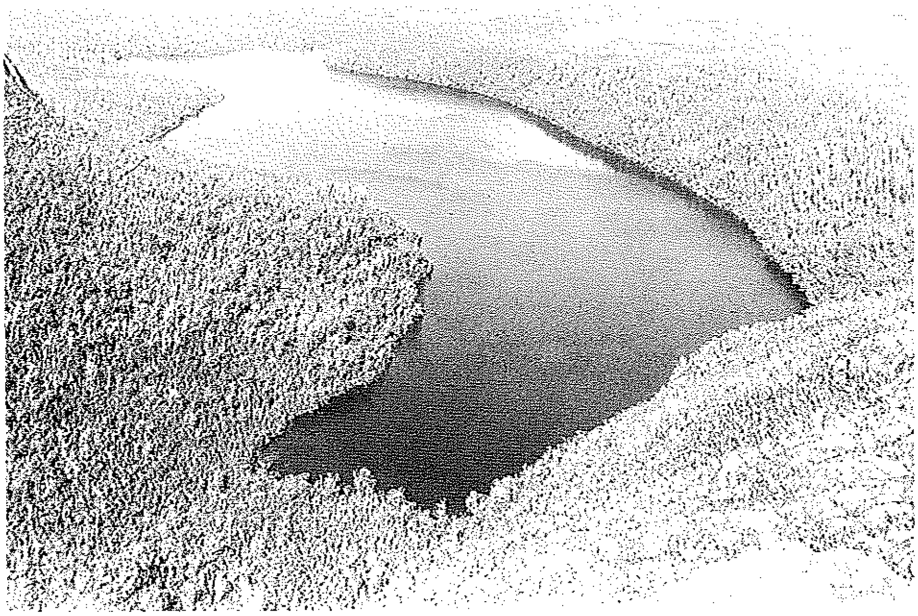

Photo 3. Lake Rufus, which occupies a small glacial trough 2 km north of Lake Richmond, is the source of the southernmost major tributary of the Guelph catchment. During the late Last Glacial Stage, ice did not extend far beyond the present lake but, during earlier glaciations, the ice stream from Lake Rufus was swept southwards into the Gordon Valley by a diffluent lobe of the Derwent Glacier.

bedrock on the arms of the trough near its head indicates that up to 300 m of ice accumulated here.

Each of these troughs terminates abruptly at the level of the St Clair Surface (Davies 1965) where piedmont ice lobes developed (Derbyshire 1967). Half a dozen smaller lakes are developed on the range, most of these probably being rock basins. Lake Montgomery, a small moraine-bounded lake north of Lake Richmond, is 25 m deep (Derbyshire 1967) and is probably also a rock basin that has been extended by the construction of a moraine barrage. Small diffluence cols occur between some of the cirques in the Gordon catchment. Cirque headwalls 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. Diffluence cols are also present in the Guelph catchment. The head of the Top End Gap is a diffluence col by means of which ice flowed eastwards from the head of the Surprise Valley outside the area under discussion in this paper, and into the Guelph basin. A smaller diffluence col occurs south-west of Lake Rufus.

Less visually conspicuous than the cirques, troughs and lakes are widespread ice-abraded surfaces. The summit of the King William I massif is an ice-abraded dolerite plateau. Elevated crests on the plateau have been little modified by glacial erosion which suggests that only a thin carapace of ice was present (Peterson 1969). Stoss and lee orientations on bedrock indicate a predominantly eastward movement of the plateau ice towards the Guelph Valley, but there are also local indications of movement westward towards the Surprise Valley. At lower altitudes, abraded bedrock also occurs near the crest of the ridge that separates the Guelph Valley from the Derwent Valley. This has been frost-shattered and has a less fresh appearance than the ice-abraded rocks that occur on the crest of the King William Range. The direction of ice movement over this ridge is

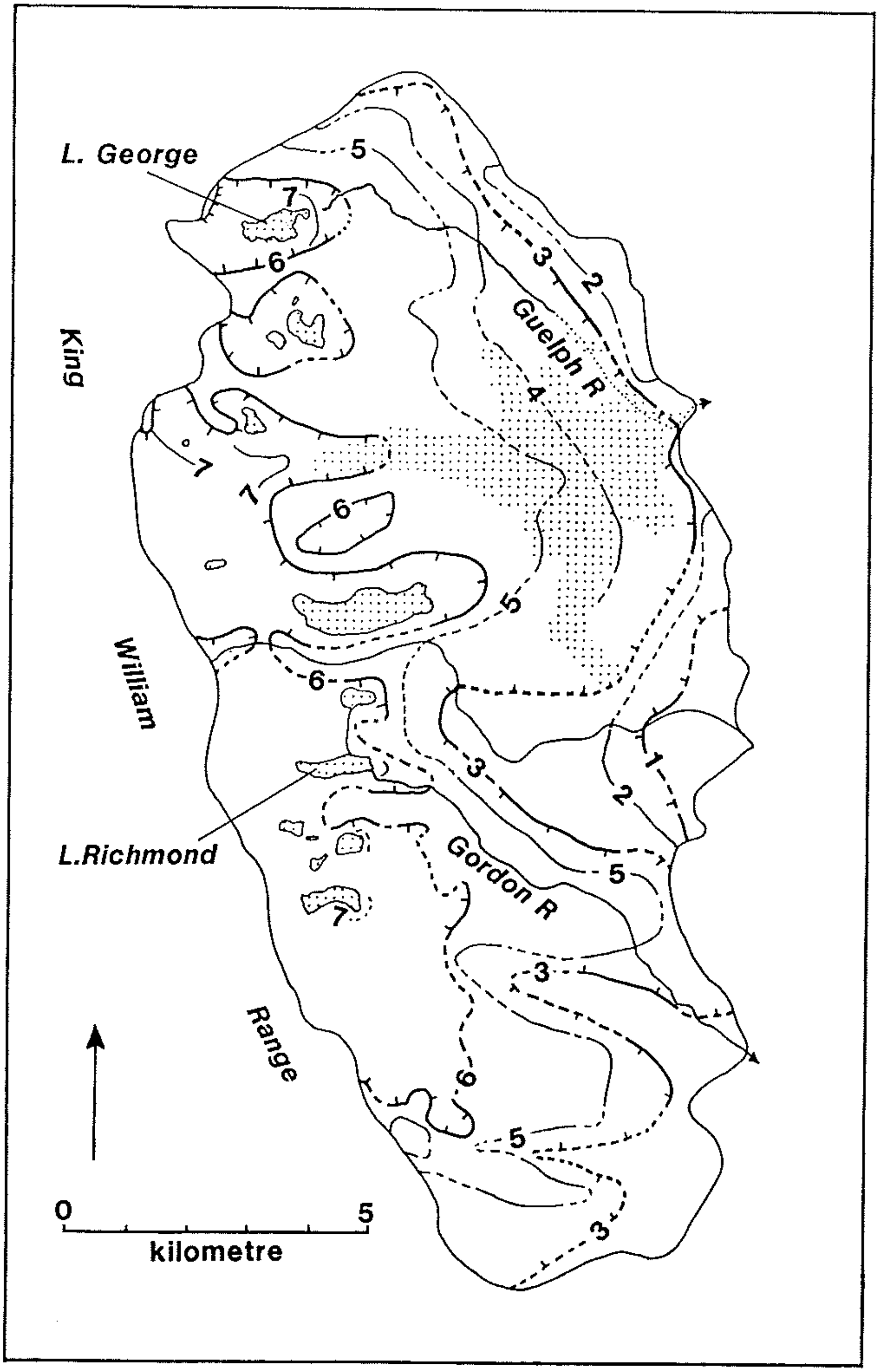

Figure 3. Integrated map of principal ice limits in the upper Gordon and Guelph Valleys: 1. Hobhouse Point; 2. Guelph Phase (pre Beehive Glaciation); 3. Divide Phase; 4. Long Bay Phase; 5. Camp I Phase (Beehive Glaciation); 6. Lake Richmond Phase; 7. Cirque Phase (Cynthia Bay Glaciation).

uncertain from the erosional evidence. Further ice-abraded surfaces have probably been drowned in construction of the artificial Lake King William formed by a dam on the Derwent River at Butlers Gorge, just downstream of the Guelph–Derwent confluence. The bathymetric chart of Lake King William (Peterson and Missen 1979) reveals a series of ridges that are aligned north–south on the lake floor upstream of the lower Guelph gorge, and a steep knoll at the gorge outlet. Abraded rock ridges that are aligned parallel to these drowned ridges occur above lake level. This suggests that many of the ridges beneath the lake that have this alignment are glacially eroded bedrock rather than moraines. In contrast to the Guelph Valley, there is little ice-abraded rock exposed on the floor of the Gordon Valley where thick drift is generally present.

Micro-erosion features are uncommon due to the susceptibility of most of the rock to post-glacial weathering. One set of striae that trends towards 135° was found at 580 m altitude (site 4 on Figure 1) adjacent to the creek that drains the small southernmost lake on the range. These striae indicate that a valley glacier moved south-eastwards across this site. Ice-abraded bedrock is present at the foot of the range in the Guelph Valley. Derbyshire (1967) reported striae near the south-western shoreline of Lake King William that indicated ice movement at 80–90° from the Lake Rufus area towards the Guelph–Derwent confluence.

Landforms produced by glaciofluvial erosion are also prominent. Meltwater channels have been cut in rock along structural lineaments in the dolerite of the King William I Plateau. Other meltwater channels plunge steeply downslope into the valley heads. Proglacial channels have been cut in rock beyond the threshold of the glacial troughs. A series of steep subglacial channels plunge downslope into the Gordon Valley from the Guelph–Gordon divide. These developed at a point where ice pressure was relieved in a zone of extending ice flow, and appear to be equivalent to the 'chutes' of Mannerfelt (1949). The lower Guelph Gorge north of Mount Hobhouse has been deepened by glaciofluvial erosion. Whereas most of these channels have been cut in bedrock, many others have been cut in drift at the foot of the King William Range. The majority of these are the result of proglacial erosion. These channels postdate ice retreat and were cut at a time when local glaciers terminated not far from their cirques. The outermost moraines on the plains at the foot of the King William Range have been heavily dissected by meltwater to an extent that often complicates the interpretation of their original morphology.

The remainder of the erosional landforms in the area are of fluvial and periglacial origin. The western slopes of the Hobhouse Range above 800 m have been fashioned primarily by fluvial erosion. Limited 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 largely inactive at present and was probably the result of frost action under colder climatic conditions than apply today. Joints have also been etched by chemical weathering at lower levels on the Guelph–Gordon divide and on the floor of the Gordon Valley.

On the King William Range a partly detached mass of dolerite at the head of the Lake George trough demonstrates the cambering of slab topples on glacially oversteepened dolerite slopes (Photo 4). The margin of the King William I Plateau is characterised by rock crevasses and dilation trenches for a distance of several hundred metres south of this point. Some of these trenches are up to 100 m wide and contain small semi-permanent ponds. Rockfall has been a slightly more active agent of scarp retreat at the head of the glacial troughs, most notably above Lake Rufus (Derbyshire 1967) and on the eastern slopes of Mount King William I. 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. Very limited rockfall has also occurred from some partly detached slab topples.

Nivation cirques and hollows are well developed on the King William I and King William II plateaux. The largest nivation cirques are seldom more than 20 m wide with a backwall height of about 8 m. Some of these cirques have developed on solifluction deposits that accumulated above the level of the plateau ice carapace. Therefore, they must postdate accumulation of the solifluction deposits. The source of Tasmania's largest river, the Gordon, lies amid the late-lying snows of a nivation cirque at 1320 m altitude that overlooks Lake Richmond.

Little erosion appears to be occurring at present where the vegetation cover remains intact. However, some severe erosion has occurred along a four-wheel drive track that has been bulldozed up the side of Mount King William I to give access to a fire tower.

Moraine sequences and other glacigenic deposits

At maximum ice conditions, glaciers that flowed from the eastern slopes of the King William Range were deflected by a lobe of the formerly very large glacier that extended down the Derwent River Valley further east. The Derwent Glacier was confluent with the Franklin Glacier north of the King William Range, and spilled into the Guelph Valley north-east of Lake George where it supplemented the ice generated from the local cirques. Further south, a low ridge oriented NNW–SSE divided the ice that flowed down the Guelph Valley from that which flowed down the main part of the Derwent Valley, now occupied by the artificial Lake King William. The Guelph ice was able to rejoin the Derwent Glacier between the southern end of this ridge and the northern slopes of Mount Hobhouse (Kiernan 1985, 1991). The main Derwent Glacier continued down the eastern side of the Hobhouse Range. However, the Derwent Glacier was at times sufficiently thick to impede the discharge of ice from the lower Guelph Valley into the Derwent Valley. This bottleneck facilitated the diversion of some of the Guelph ice into the upper Gordon River Valley through the 'Gordon Gap' between Mount Hobhouse and the King William Range.

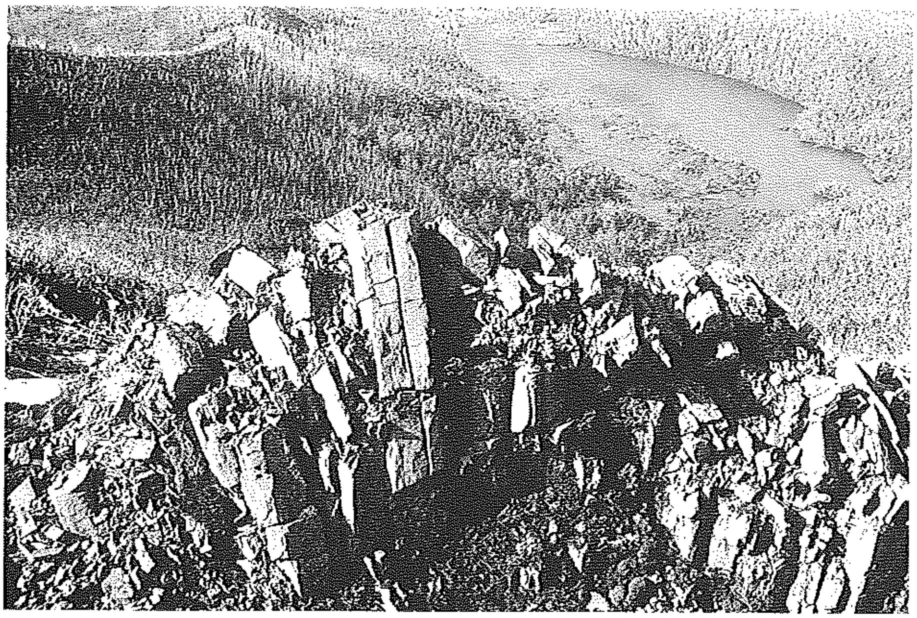

Photo 4. Lake George, source of the Guelph River. The low ridge on the left is a moraine. In the foreground is a slab topple on the edge of the Mount King William I plateau, formed as a result of post glacial failure of the cirque headwall following over-steepening by glacial erosion.

Table 1. Lithology of glacial deposits in the Guelph Valley and Gordon Valley. Site locations are indicated on Figure 1. (n = 100 for each site)

Guelph Valley							
Site no.	1	3	4	5	7	8	9
dolerite %	60.0	37.5	54.0	68.9	38.5	92.0	52.0
quartz %	32.0	47.5	29.0	26.8	46.6	4.0	38.5
quartz schist %	-	-	-	-	10.1	-	-
sandstones/ mudstones %	8.0?	15.0?	15.0?	4.3?	4.8	4.0?	9.5
other %	-	-	-	-	-	-	-

Gordon Valley						
Site 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
sandstones/ mudstones %	2.0	1.0	8.0	45.0	25.0	5.0
black schist %	15.0	-	10.0	-	-	-
other %	-	-	-	10.0	5.0	-

On the basis of evidence on Mount Hobhouse and on the floors of the Gordon and Guelph Valleys, seven phases of glaciation that appear to relate to at least three separate glaciations have been identified during this study. Diffluent ice from the Guelph Valley entered the Gordon Valley during at least three of these phases of glaciation.

1. Hobhouse Phase

In the Guelph River catchment, quartzite erratics that occur among dolerite slope deposits indicate that ice once reached 850 m altitude on the northern slopes of Mount Hobhouse. In view of the large volume of ice that was generated from the King William Range, it is reasonable to interpret these erratics as deposits of the Guelph Glacier rather than the Derwent Glacier, deposited at a time when the two glaciers were confluent and the Derwent ice would have extended far downstream of the Guelph–Derwent confluence (Kiernan 1985). Part of this Guelph ice flowed eastwards from the Lake Montgomery Cirque (55GDP 315160) which now lies in the Gordon catchment but earlier formed part of the Guelph catchment. The lithology of the clasts in the till exhibits considerable local variation (Table 1).

The uppermost lateral moraine on the eastern side of the Gordon Valley can be traced SSE from 730–630 m altitude on the south-western slopes of Mount Hobhouse. It records the passage of ice at least 100 m thick southwards through the Gordon Gap from the Guelph Glacier. The moraine crest suggests that the ice surface sloped southwards at a gradient of about 70 m/km. This moraine suggests 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 down the Gordon Valley are difficult to discern on aerial photos due to dense vegetation and the possibility that some moraine-like ridges are bedrock, perhaps only thinly mantled with till. A pronounced ridge 3.5 km south of Mount Hobhouse is probably a moraine and, if so, it suggests that at least 80 m of ice buried the valley floor at this point. This seems surprisingly little for the maximum phase of glaciers emerging from such a major snowfence as the King William Range. This is particularly so given the emerging picture of very extensive late Cainozoic ice in the Tasmanian Central Highlands (Kiernan 1985, 1990a; Hannan 1993). It is probable, therefore, that the maximum ice limits lay still higher on the Hobhouse Range but that moraines have not survived on these steep slopes.

A moraine-like ridge at 800–720 m below the cirque on the southern flank of Mount King William III lies on a rock bench and forms the present divide between the Gordon River and the Denison River, the latter of which drains the western slopes of Mount King William II and Mount King William III, outside the area under discussion in this paper. While it is probable that some ice and meltwater spilled into the upper Denison River Valley 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 previously studied in this part of Tasmania (Kiernan 1985, 1989a, 1990a, 1991). This compounds the impression that the Gordon Glacier probably once extended further than indicated there.

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 while the majority of the ice in the Gordon Valley was generated locally, a significant volume of ice also entered the Gordon Valley from the Guelph Valley.

2. Guelph Phase

In the Guelph Valley, degraded moraine remnants occur from reservoir level to 800 m altitude on the southern end of the ridge between the Derwent and Guelph Valleys. Moraines occur to a similar altitude on the foot of the slopes that extend towards that ridge from Mount Hobhouse. These moraines represent part of a complex that occurs in the Derwent Valley immediately north of the Guelph–Derwent confluence (Kiernan 1985). It records a glacial event during which the Guelph and Derwent glaciers ceased to be confluent and retreated into their respective valleys. Till deposits imply that ice probably also passed eastwards into the Derwent via two minor cols on the northern end of the Hobhouse Range.

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

3. Divide Phase

The Guelph–Gordon divide is mantled by thin till and erratics that were transported by a thin lobe of ice about 1 km wide that spilled over this divide from the north. No moraine morphology remains on the steep upper slopes south of the divide but two lateral moraines were constructed across its lower slopes on the margin of the Gordon Glacier. The uppermost of these two moraines 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 altitude. The till of which this moraine is composed contains a moderately high proportion of quartz schist clasts that may either have come from the Guelph Valley or may have been reworked from the Permian sedimentary rocks at Lake Richmond. The Divide Drift appears to represent the final occasion on which ice spilled southwards through the Gordon Gap.

Further north, in the Guelph catchment, till that is plastered against the western side of the ridge between the Guelph and Derwent Valleys appears to mark a distinct phase during which the Guelph Glacier was trapped behind the rock bar at the lower Guelph River Gorge. A small outwash plain was constructed against the inner margin of the earlier moraine at the gorge outlet (Peterson and Missen 1979). The steep flanks of the ridge have militated against the preservation of glacial sediments, but some highly degraded moraines occur on more gentle terrain to the south of Lake King William. A till mantle below about 800 m altitude indicates that the northern end of the ridge was overridden. Hummocky drift occurs in the far north-west of the Guelph area within this ice limit.

4. Long Bay Phase

The bathymetric chart of Lake King William (Peterson and Missen 1979) suggests that a probable major end moraine extends northwards across the bed of the lake from a degraded moraine on the eastern shoreline of 'Long Bay', the major southward-trending arm of Lake King William, east of Lake Rufus. The moraines north of Lake King William have been severely dissected by meltwater, but possible continuations of the Long Bay Moraines can be traced eastwards to the Guelph River and northwards along the western flank of the ridge between the Guelph and Derwent Valleys.

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

5. Camp I Phase

The crest of the inner lateral moraine south of the Gordon Gap lies at 680–720 m altitude (55GDP 8113-345 135). A probable continuation of this moraine sequence terminates 5.5 km south-east of the Gordon Gap at about 600 m altitude, but its crest has not yet been traversed continuously. Pronounced lateral moraines 3–5 km north-east of Mount King William III indicate that two separate ice lobes extended from the range at this time, while another smaller glacier also arose from the southernmost cirque on Mount King William III. 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 altitude about 4.8 km ESE of Mount King William III.

The till that forms the lateral moraine 2.5 km downstream from Lake Richmond appears to be up to 20 m thick but no sections 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 1). This is probably the result of rapid comminution of the sandstones and mudstone, but localised exceptions to this general pattern are common. Comminution till (Dreimanis 1976) is exposed at 580 m altitude on the south bank of the creek that drains the southernmost lake on the eastern flank of the King William Range. This till contains abundant angular fragments of dark siltstone amid rounded clasts of dolerite.

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

The most extensive ice body at this time probably remained the Top End Glacier which still appears to have extended 5 km eastwards of the Guelph–Surprise divide. This was the last occasion during which ice accumulated in the shallow cirque north of Lake George. Hummocky drift well inside this ice limit suggests that retreat may have been rapid and indicates that it left stranded masses of debris-rich ice to decompose *in situ* in a number of localities (Derbyshire 1968).

6. Lake Richmond Phase

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 moraines can be traced to a lateral moraine on the north-eastern side of the lake. 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 down valley (Table 1). The Lake Richmond moraines record a glacial event during which ice was generally confined above 680 m altitude and in few cases extended beyond the cirques and glacial troughs where it accumulated.

Glaciofluvial sediments that can be traced back to the Lake Richmond moraines are inset between the Camp I moraines. This end moraine and outwash apron are indicative of a major stage limit. They generally appear to mantle only thinly older ground moraine. Sections cut by the Gordon River 4 km south-east of Lake Richmond reveal up to 4 m of generally compact glaciofluvial gravels that contain moderately developed iron pans and that 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 down valley from the southern cirques and infill inter morainal swales. These sediments are invariably dominated by dolerite and quartzite clasts.

In the Guelph Valley, a complex of topographically fresh and little dissected moraines commences about 1 km down-stream of Lake Rufus and extends westwards towards the lake shoreline. Fresh lateral moraines occur 2 km east of the Top End Gap and other fresh moraines occur around Lake Eva (Photo 5) and to the north and east of Lake George. The lateral moraine on the northern side of Lake George has been constructed within the limits of a degraded outer moraine which now has little topographic expression.

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

7. Cirque Phase

Small terminal moraines form the present eastern shoreline of Lake Richmond and also bound the shores of lakes further to the south. Those moraines that stand upon well-defined thresholds are generally well preserved. Little significant outwash plain 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 up to 1 m long occur on the surface of the moraine. These slabs are highly

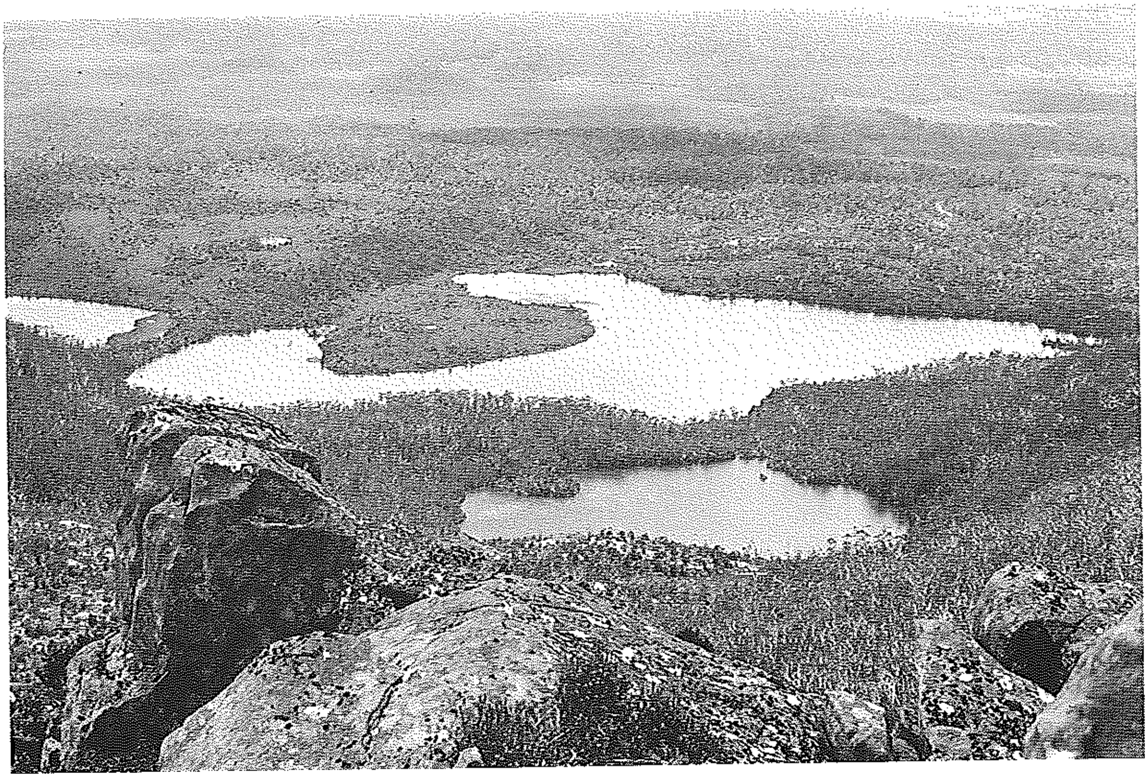

Photo 5. The Lake Eva complex, east of Mount King William I, is impounded by moraines of late Last Glacial age.

angular despite the incompetence of this rock and may therefore be of supraglacial origin. This would imply rockfall from the walls of the trough while the glacier was present. The survival of the lakes in the upper Gordon Valley without their being filled by glacial and glaciofluvial sediment suggests that final retreat of the ice occurred rapidly.

In the Guelph catchment, latero-terminal moraines just east of the Top End Gap on the slopes of the King William Range impound a number of small lakes, most notably a composite moraine at Lake Sally Jane (Photo 6). These moraines appear large given the limited extent of the ice. Older lateral moraines deposited by the Top End Glacier have probably been partly reworked and blocked flow, but it is nonetheless clear from the size and angle of the moraine that the Sally Jane Glacier was very active. Protalus has also contributed to the moraine at Lake Sally Jane, quite angular blocks being present just inside its crest.

This phase is represented elsewhere by the innermost of the cirque moraines that bound rock basins. The Top End Glacier had by this time declined from being the largest glacier in the area to virtually the smallest. The cause of this decline lay in the continued retreat of the Surprise Glacier (Kiernan 1989a) which no longer had a sufficient surplus to direct ice eastwards through the Top End Gap. Once again, little has accumulated in the moraine-bounded lakes in the Guelph catchment which seems to confirm the impression from the Gordon Valley that deglaciation was probably fairly rapid.

Non-glacial deposits

Solifluction deposits that contain abundant angular dolerite clasts cloak the western flank of Mount 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

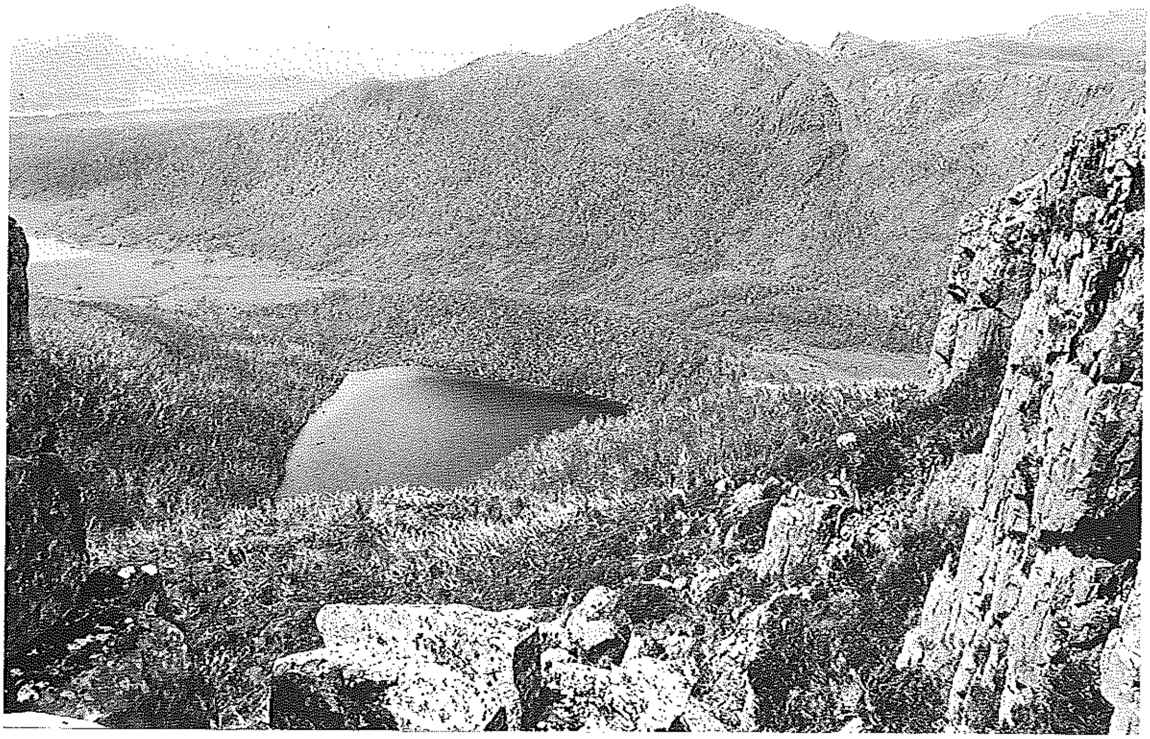

Photo 6. Lake Sally Jane, below the southern end of the Mount King William I plateau, is impounded by a prominent composite moraine most recently added to during the late Last Glacial Stage. Mount King William II is in the background with the Top End Gap separating the two.

western flank of Mount 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. Similar mantles up to 3 m thick occur on the northern flank of Mount Hobhouse where they overlie weathered till and incorporate occasional quartzite erratics. Sandstone and mudstone-rich mantles over 1 m thick cloak the upper western flank of the ridge between the Derwent and Guelph Valleys.

Few fully detached dolerite topples occur on the King William Range, where most of the plateau margin above the cirques has been over-ridden by ice during the most recent phases of glaciation. However, 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 buttresses on Mount 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. A considerable thickness of rockfall talus has accumulated beneath the summit cliffs of Mount King William I. Most of this lies above 1100 m altitude which suggests that the upper surface of the George Glacier lay below this level at the time the talus accumulated. A small protalus rampart is present in this locality.

Small solifluction terraces have developed on a partly detached topple that overlooks Lake George. Bare treads about 2 m wide are separated by risers 30–50 cm high which support small shrubs. Some primary frost sorting has occurred on the treads. Small polygonal nets up to 30 cm in diameter developed within 10 years at about 1100 cm altitude north-east of Mount King William I on a surface devegetated by bulldozing during construction of a four-wheel drive track. Small subnival pavements of closely fitted stones floor some of the nivation cirques on the King William I plateau. They have probably resulted from block movement due to the weight and creep of the snow-pack, and frost heaving (Troll 1958; White 1972).

In the Gordon Valley, alluvial sediments are confined to over-bank 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. In the Guelph Valley, swales between moraines on the King William Plains have been filled by up to 1 m of alluvial sand and silt. Organic silt has accumulated to depths of 30 cm in ponds that have formed in dilation trenches on the margin of the King William I plateau. The silt reaches 60 cm thick in the lagoon at the eastern end of Lake George.

Discussion

The relative positions in the landscape of the landforms and sediments provide the most obvious means for differentiating glacial events of different ages. Landforms and sediments that are interpreted as being of Holocene age include the organic-rich silts that have accumulated inside the most recent ice limits of the cirque phase at the heads of Lake Richmond and Lake George and in dilation trenches on the edge of the King William I plateau. Also of presumed Holocene age is the silt that has accumulated on the floors of nivation hollows on the King William Range and in swales between moraines on the plains east of the range. Small solifluction terraces have developed inside the ice limits of the Lake Richmond phase on the partly detached slab topple that overlooks Lake George.

A small ridge of protalus was constructed beneath the cliffs of Mount King William I on an earlier talus that probably accumulated in contact with the ice margin during the Lake Richmond Phase and locally slipped down-slope as the ice retreated. The upper protalus unit must postdate the Lake Richmond Phase. Protalus also forms a significant part of the cirque moraine at Lake Sally Jane. The cirque moraines appear to represent a retreat from the Lake Richmond limit and the large angular clasts reflect considerable rockfall activity, in contrast to the general stability of the free faces today. Thick 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 (or phases) of solifluction predates deglaciation. The crest of the partly detached topple above Lake George has been smoothed by glacial erosion. This erosion could not have been accomplished without collapsing the topple if it had already been partly detached. Therefore, the toppling must postdate the Lake Richmond Phase and probably occurred in response to the retreat of the supporting ice margin. Given its stability under the present interglacial conditions, this topple is interpreted as broadly equivalent in age to the moraines of the Cirque Phase.

The size of the end moraine and outwash terrace that were deposited during the Lake Richmond Phase suggests that they represent the ice limit during a major stage. During the Lake Richmond Phase, the glaciers were short and steep. The moraines in the Rufus River area occur inside the ice limits of the Camp I Phase. These glaciers formed separate piedmont lobes on the King William Plains. Outwash sand and gravel extends from the ice limits of the Lake Richmond Phase along meltwater channels that were eroded through the Camp I and earlier moraine sequences.

The Camp I moraines in the Gordon Valley enclose the outwash plain that was deposited during the Lake Richmond Phase and must predate them. The broad outwash terrace 7 km south-east of Lake Richmond records another important ice limit. During this phase, no ice from the Guelph Valley passed through the Gordon Gap. The ice that accumulated behind the King William Range within the upper Gordon Valley occurred as two main lobes and one smaller glacier.

The Long Bay moraines define a major ice limit at a time when the King William Range glaciers formed a confluent piedmont apron. The Derwent Glacier and the Guelph ice were still confluent at this time around the northern end of the ridge that separates the Derwent and Guelph Valleys. However, local sources accounted for the majority of the ice in the upper Guelph Valley.

The glacial sediments that were deposited during the Divide Phase lie inside the ice limits of the Guelph Phase. They indicate a more extensive ice cover that inundated much of ridge between the Derwent and Guelph Valleys. Diffluent ice from the Guelph Valley also spilled through the Gordon Gap into the upper Gordon Valley. The Guelph Glacier later became lodged behind the rock bar at the lower Guelph Gorge and was no longer confluent with the Derwent Glacier in that area.

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. At its maximum, the Guelph Glacier probably over-rode the ridge between the Derwent and Guelph Valleys and reached an altitude of at least 850 m on the northern slopes of Mount Hobhouse. The lateral moraine on the western slopes of Mount Hobhouse was constructed by a distributary lobe of the Guelph Glacier at least 100–150 m thick that extended southwards through the Gordon Gap. In the Gordon Valley, this ice from the Guelph was confluent with the considerable volumes of ice that accumulated in the local cirques, and a major valley glacier resulted, the maximum downstream limits of which have not yet been ascertained.

Contrasts in post depositional modification (PDM) of the landforms and sediments (Kiernan 1989b) further aids dating, notably contrasts in moraine degradation, clast weathering and the degree of soil formation (Tables 2 and 3). The morphology of the Cirque Phase and Lake Richmond Phase moraines is well preserved, but the Camp I moraines are degraded and partly buried by outwash gravels and sands. The Long Bay Phase and Divide Phase moraines are rounded in form, have been extensively dissected by stream erosion, and give an impression of greater age. Erratics high on the slopes of Mount Hobhouse are the sole remaining depositional legacy of the Hobhouse Phase.

Only A-Cox-Cu-type soil profiles or, at most, profiles with some very limited development of a Bt horizon (Birkeland 1984), have developed on the moraines of the Lake Richmond Phase. This limited pedogenic alteration of the Lake Richmond drift contrasts with deeper soil profiles on the older moraines that have B horizons more than 1 m thick. The Cox horizon reaches a depth of at least 1.5 m in the Camp I moraines. The soil profiles on the earlier moraines are characterised by very thick textural B horizons.

Weathering rinds on dolerite clasts in the Lake Richmond moraines in the Gordon Valley are no more than 1.4 mm thick. The mean thickness of weathering rinds formed on dolerite clasts in the moraines around the Rufus River is less than 2 mm and maximum rind thickness is less than 5 mm. This is half the thickness of the weathering rinds in the moraines of the Guelph Phase.

On the basis of these differences in PDM of the drifts, coupled with the published PDM data available for drifts in the adjacent valleys (Table 4), the Cirque and Lake Richmond Phases can confidently be correlated with the Dixon Glaciation in the Franklin Valley (Kiernan 1989a) and the Cynthia Bay Glaciation in the Derwent Valley (Kiernan 1985, 1990a, 1991, 1992). Both of these are regarded as dating from the late Last Glacial Stage. The Camp I and Divide Phase deposits are interpreted as being equivalent to the Beehive Glaciation deposits in these two neighbouring valleys. The Guelph and Hobhouse moraines appear to be of still greater age and are interpreted as being broadly equivalent in age to the Powers Creek Glaciation in the Derwent Valley.

The glacial sediments in the Gordon Valley indicate that at least 15 km^2 was ice covered during the Cynthia Bay Glaciation, 27 km^2 during the Beehive Glaciation and in excess of 40 km^2 during the earliest event recognised. However, it must be stressed that drift of Beehive age occurs on the valley floor to the downstream limit of the present mapping. The presence further upstream of older glacial sediments much higher on the valley walls than is attained by the Beehive drift implies that the older glaciers must almost certainly have extended much further down valley than the glaciers that developed during Beehive time. This ice would probably have been confluent with ice in the upper Denison Valley south of Mount King William III. Confluence of the Guelph–Derwent ice with the Surprise Glacier and Franklin Glacier west and north-west of the King William Range has previously been recognised (Derbyshire 1967; Kiernan 1989a) while ice would also have covered the present Surprise–Denison divide. Hence, the King William Range would have been a nunatak.

A useful outcome of the present study, besides defining with greater precision the glacial legacy in the upper Gordon Valley, is the extension of the stratigraphic model developed for the Derwent and Franklin Glaciers, the two main southern outlet glaciers from the Central Highlands ice cap, into the more outlying glaciated mountains of south-western Tasmania.

There is a striking contrast between the Derwent Glacier and the King William Range glaciers in terms of ice extent during the late Last Glacial maximum, when a relatively large glacier still formed at Lake St Clair, in contrast to the very small glaciers of the King William Range. The southward deflection by the Derwent Glacier of tributary ice streams from the Central Plateau implies that the Central Plateau was not the main ice source for the Derwent Glacier during its retreat. However, the relatively extensive ice cover in the Derwent Valley at the maximum of the Cynthia Bay Glaciation demands considerable invasion of the Lake St Clair trough by ice from the Central Plateau. In contrast, the glaciers of the King William Range lacked any external supplementation and remained very small during the Cynthia Bay Glaciation.

The lack of evidence for major post-maximum ice limits of the glaciers of the West Coast Range (Colhoun 1985) during the Margaret (= Cynthia Bay) Glaciation is contrasted strongly by the Derwent Glacier (Kiernan 1992) while two limits are also evident in the King William Range glaciers during Cynthia Bay time.

The Franklin Glacier attained a length of 12 km during the Dixon (= Cynthia Bay) Glaciation, when there was only limited invasion of the Franklin Valley by ice masses that originated elsewhere. The Franklin Glacier also lacks evidence of more than two significant limits during the late Last Glacial Stage. Although the length of the Franklin Glacier appears to contrast with the very small size of the King William Range glaciers, given the reasonably comparable height and orientation of the principal snowfences to the west, two additional factors must be taken into account. Firstly, the Franklin Glacier had the advantage of an additional broad snowfence and neve area to the north around its headwaters. Secondly, whereas the King William Range glaciers flowed eastwards onto a plain or broad valley floor and in large measure remained independent of one another, ice that flowed eastwards into the Franklin Valley was confined in a narrow valley oriented south-eastwards along the foot of the Mount Gell snowfence such that tributary ice streams augmented a single valley floor glacier. Viewed in terms of ice volumes rather than simply glacier length and discounting the additional input from the Lake Hermione area, the glacier systems in the

Table 2. Post-depositional modification of deposits in the upper Gordon River Valley. Site locations are indicated on Figure 1. (T = till; O = outwash gravels; S = solifluction deposit)

Site no. Material	1 0	2 0	3 T	4 T	5 0	6 T	7 S
A. Morphostratigraphy							
relative position	outside 2	inside 1	outside 6	inside 2	outside 6	inside 3 & 5	outside 3
crest/site altitude (m)	600	640	740	640	630	710	850
B. Surface 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 (score 1–5)	1–2	1–2	2–4	1–2	1	1	1–2
clast surface colours	7.5YR	7.5YR	5YR	5YR	5 YR	7.5YR	7.5YR
% unrecognisably weathered clasts							
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	7.5 YR	7.5 YR	7.5YR-5YR	7.5YR	7.5YR	10YR-7.5YR	7.5-5YR
depth of B horizon (cm)	-	-	>200	>100	-	-	-
depth of lowest Cox (cm)	>100	>100	>200	>100	>70	50	>70
development of clay films (score 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 (score 1–5)	-	-	-	4	-	1	-

Table 3. Post-depositional modification of deposits, Guelph area. Site locations are indicated on Figure 1. (T = till; O = outwash gravels; S = solifluction deposit)

Site no. Material	1 T	2 T	3 T	4 T	5 O	6 T	7 T	8 T	9 T	10 S	11 S
A. Morphostratigraphy											
relative position	outside 2	outside 3	inside 2	inside 3	inside 3	inside 5	inside 4	outside 9	inside 8	inside 3	inside 9
crest/site altitude (m)	770	770	760	740	740	780	760	960	810	730	1200
B. Surface clasts											
dolerite rinds (mm)											
max.	8.2	8.0	8.8	-	2.7	-	-	4.3	4.2	1.9	1.2
min	1.9	1.0	4.0	-	1.2	-	-	0.3	0.9	0.4	0.3
mean	4.7	4.5	6.1	-	1.8	-	-	1.3	1.7	0.8	0.8
SD	1.8	1.5	1.7	-	0.5	-	-	0.8	0.6	0.4	0.3
hardness (1–5)	3	3–4	2–4	1–3	1–2	1–3	1–2	1	1–2	1	1
coherence of argillites (1–5)	5	3–5	4–5	4	2	4	1	2	1	1	1
clast surface colours	5YR	7.5YR	5YR	5YR	5YR	7.5-5YR	10YR	10YR	10YR	-	-
C. Subsurface matrix											
profile type	A Bt	A Bt	A Bt	A Bt	A Cox cu	A Bt	A Cox cu	A Cox cu	A Cox cu	A Cox cu	A Cox cu
colour of B horizon	5YR	7.5YR	7.5-5YR	7.5YR	7.5YR	7.5-5YR	10YR	7.5-10YR	10YR	10YR	7.5YR
depth of B horizon (cm)	>150	>100	>150	>150	-	>150	-	-	-	-	-
depth of lowest Cox (cm)	>150	>100	>150	>150	>150	>50	<50	<80	-	-	-
development of clay films (1–5)	4	3–4	3	4	-	3–4	1	1	1	1	1
fine matrix											
% sand	-	-	29.2	-	-	-	-	-	13.7	27.8	-
% silt	-	-	37.1	-	-	-	-	-	48.8	40.0	-
% clay	-	-	33.7	-	-	-	-	-	37.5	32.2	-
degree of lithification of basal till	-	-	3	3	-	3	-	1	-	1	-
clast socket staining	4	3–4	2–4	3	-	2–3	-	1	-	1	-
D. Surface materials											
no. of surface boulders (/m²)	-	-	-	-	-	-	-	-	-	-	
and split clasts	20	-	-	23	-	16	3	-	-	-	-
E. Depositional landforms											
moraine crest width (m)	-	-	-	15	-	25	5	5	-	-	-
moraine slopes (°)											
proximal	-	-	-	<5	-	8	20	25	-	-	-
distal	-	-	-	<5	-	10	20	15	-	-	-
degree of fluvial dissection (1–5)	3	3–4	3–4	4	-	4	2	1	2	-	-

Table 4. Thickness of weathering rinds on subsurface dolerite clasts in tills of the Gordon River Valley and adjacent valleys.

	Gordon Valley	Guelph Valley	Derwent Valley (Kiernan 1985)	Franklin Valley (Kiernan 1989a)
Cynthia Bay Glaciation				
max	1.4	4.3	4.6	3.8
min	0.6	0.3	0.1	0.4
mean	1.2	1.5	1.5	1.6
SD	0.9	0.7	0.5	0.5
Beehive Glaciation				
max	6.1	8.8	6.9	9.2
min	1.6	1.0	0.5	4.1
mean	3.1	5.1	3.0	4.1
SD	1.4	1.7	1.6	1.2
Power Creek/Taffys Creek Glaciation				
max	-	-	15.8	15.1
min	-	-	1.0	1.9
mean	-	-	5.1	7.4
SD	-	-	2.2	3.8

Franklin, Guelph and Gordon Valleys appear reasonably comparable.

However, the glaciers in the Franklin Valley were significantly larger when ice invaded from outside the Franklin catchment. During earlier glaciations, additional ice from the Central Highlands Ice Cap enabled the Franklin Glacier to extend for at least 27 km. A similar situation is evident on the northern margin of the ice cap, where the Forth Glacier may once have extended for about 95 km down valley (Colhoun 1976) but extended for only about 13 km during the late Last Glaciation when nourished solely from local snowfences and cirques (Kiernan and Hannan 1991). This is consistent with the evidence from the Derwent Valley, Guelph Valley and upper Gordon Valley.

It has been argued that a major ice cap of at least 7000 km^2 developed during the most intensive phase of glaciation in the Central Highlands (Kiernan 1990b). Subsequent work on the eastern edge of the Central Plateau has suggested that even this figure may be conservative (Hannan 1993). The evidence from the Guelph and Gordon Valleys compounds previous impressions that the glaciers which formed during the late Last Glacial maximum were generally very much smaller than those which formed during some earlier glaciations (Colhoun 1976; Kiernan 1983b). The ice cap that formed on the West Coast Range during the late Last Glaciation was barely 10% the extent of the ice that accumulated there during the earlier Linda Glaciation (Colhoun 1985), while the Central Highlands ice cap was at least 50% smaller during the late Last Glaciation than it was during at least one earlier glaciation (Kiernan 1990a). The drifts which permit identification of the maximum limits of the outer glaciers in most of the valleys that have been studied to date are generally very weathered, with dolerite weathering rinds reaching more than 250 mm thick in some cases. The age of the most extensive phases of glaciation remains uncertain but the degree to which the relevant drifts are weathered suggests that they are very old (Kiernan 1983a, b). Weathering evidence suggests that the earliest deposits in the Franklin Valley could date from as early as approximately 10 ma BP (Kiernan 1985,

p. 403). Palynological evidence has recently suggested that the inception of glaciation in Tasmania may date from the earliest Oligocene (Macphail *et al.* 1993).

More work is required to determine the maximum glacier limits in the Gordon River Valley. The fairly limited weathering of the glacial deposits 11 km from the valley head contrasts with the heavy weathering characteristic of the outermost ice limit in most of the other valleys studied in the Central Highlands and West Coast Range. The King William Range is a major snowfence in its own right. It is highly probable that a significantly larger glacier than that so far demonstrated would have developed in the Gordon Valley under as prolonged or intense period of glacial cold as is implied for some older glaciations, even without the added input from the Derwent Glacier which also deflected much additional ice from the Guelph Valley southwards into the Gordon Valley.

Investigation of important outlet glaciers from the Central Highlands has been impeded by the loss of access to landforms and sediments on the floors of the Derwent, Mersey and Forth Valleys in critical locations due to the construction of hydro-electric dams, the filling of artificial reservoirs and quarrying activities. The upper Gordon River Valley offers researchers both a corridor to follow from the Central Highlands ice cap to the mountains of south-western Tasmania and perhaps a means of learning more of the main Derwent Glacier and hence the ice-cap itself.

Acknowledgements

My thanks are extended to Peter Robertson, Eric Colhoun, Steve Harris and Mick Brown for encouragement or assistance at various times. I also acknowledge support via a Commonwealth Postgraduate Research Award during the earliest phase of this study, and subsequently from Forestry Tasmania.

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