LATE WISCONSIN GLACIATION OF TASMANIA

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(with two tables, four text-figures and one plate)

COLHOUN, E.A., HANNAN, D. & KIERNAN, K., 1996 (xi): Late Wisconsin glaciation of Tasmania. *In* Banks, M.R. & Brown, M.F. (Eds): *CLIMATIC SUCCESSION AND GLACIAL HISTORY OF THE SOUTHERN HEMISPHERE OVER THE LAST FIVE MILLION YEARS. Pap. Proc. R. Soc. Tasm.* 130(2): 33–45. ISSN 0080-4703. Department of Geography, University of Newcastle, Callaghan, NSW, Australia 2308 (EAC); Department of Physical Sciences, University of Tasmania at Launceston, Tasmania, Australia 7250 (DH); Forest Practices Board of Tasmania, 30 Patrick Street, Hobart, Tasmania, Australia 7000 (KK).

During the Late Wisconsin, icecap and outlet glacier systems developed on the West Coast Range and on the Central Plateau of Tasmania. Local cirque and valley glaciers occurred in many other mountain areas of southwestern Tasmania. Criteria are outlined that enable Late Wisconsin and older glacial landforms and deposits to be distinguished. Radiocarbon dates show Late Wisconsin ice developed after 26–25 ka BP, attained its maximum extent c. 19 ka BP, and disappeared from the highest cirques before 10 ka BP. Important Late Wisconsin age glacial landforms and deposits of the West Coast Range, north-central and south-central Tasmania are described. Late Wisconsin ice was less extensive than ice formed during middle and earlier Pleistocene glaciations. Late Wisconsin snowline altitudes, glaciological conditions and palaeoclimatic conditions are outlined.
Key Words: **glaciation, Tasmania, Late Wisconsin, snowline altitude, palaeoclimate.**

INTRODUCTION

An answer is still sought to the question of the extent of glacial ice during the maximum of the last glaciation (Derbyshire 1972: 86).

Evidence for development of Pleistocene ice in Tasmania was first noted in 1859–60 by Charles Gould in the Cuvier Valley near Lake St Clair (Banks *et al.* 1987). Subsequently, numerous geologists recorded evidence for ice in the Central Plateau, the West Coast Range and in mountains throughout southwestern Tasmania. Most data related to landforms of glacial erosion with only limited recognition of glacial deposits; e.g. the map and description of erosional forms on the Central Plateau by Jennings & Ahmad (1957) and the description of glacial deposits near Queenstown by Gregory (1904). Only Lewis (1945) integrated his results in a sequence-model of glaciations that had affected western Tasmania. Lewis suggested three stages: the first characterised by an extensive ice sheet over much of western Tasmania; the second by cirque and valley glaciers; the third by cirque glaciers only. None of the deposits relating to the three stages of glaciation were dated. Gill (1956) obtained a radiocarbon date of 26 480 ± 800 yr BP (W-323) on wood from sands and gravels at Linda. The deposits were interpreted as glacial in origin, and a Late Wisconsin age was suggested for the glaciation. Jennings & Banks (1958) considered there was not stratigraphic evidence for three glacial stages in Tasmania and, when Banks & Ahmad (1959) showed that the deposits of Lewis's oldest glaciation at Malanna were fluvial and lacustrine, the hypothesis of multiple glaciation became seriously questioned.

The evidence for glaciation in Tasmania remained fragmentary until Derbyshire *et al.* (1965) compiled *A Glacial Map of Tasmania* from glacial landforms observable on the first aerial photograph coverage. The map indicated the extent to which Tasmania had been glaciated, but it did not differentiate Late Wisconsin from older glacial landforms.

During the late 1970s, 1980s and early 1990s, field mapping and stratigraphic investigations demonstrated that Tasmania had a complex history of multiple glaciation, most of which pre-dated the Late Wisconsin (Kiernan 1982, 1983a, 1989, 1991, 1995, Colhoun 1985, Augustinus & Colhoun 1986, Hannan & Colhoun 1987, Barbetti & Colhoun 1988, Kiernan & Hannan 1991, Fitzsimons & Colhoun 1991, Fitzsimons *et al.* 1992). A number of criteria were used to separate Late Wisconsin from older glacial landforms and deposits. The limits of Late Wisconsin ice were first recorded for the West Coast Range system (Colhoun 1985) and then extended to the northwestern Central Plateau and Mersey Valley (Hannan & Colhoun 1987), and the southern Central Highlands and Lake St

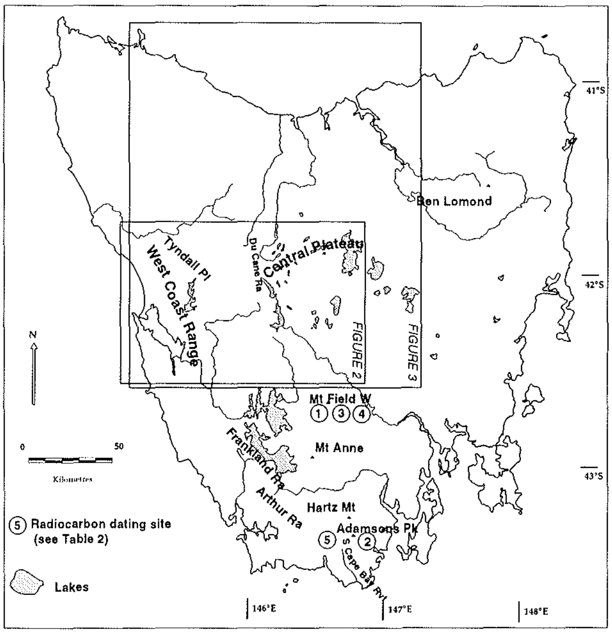

FIG. 1 — Tasmania, showing locations of figures 2 and 3, and radiocarbon dates not shown on the larger figures.

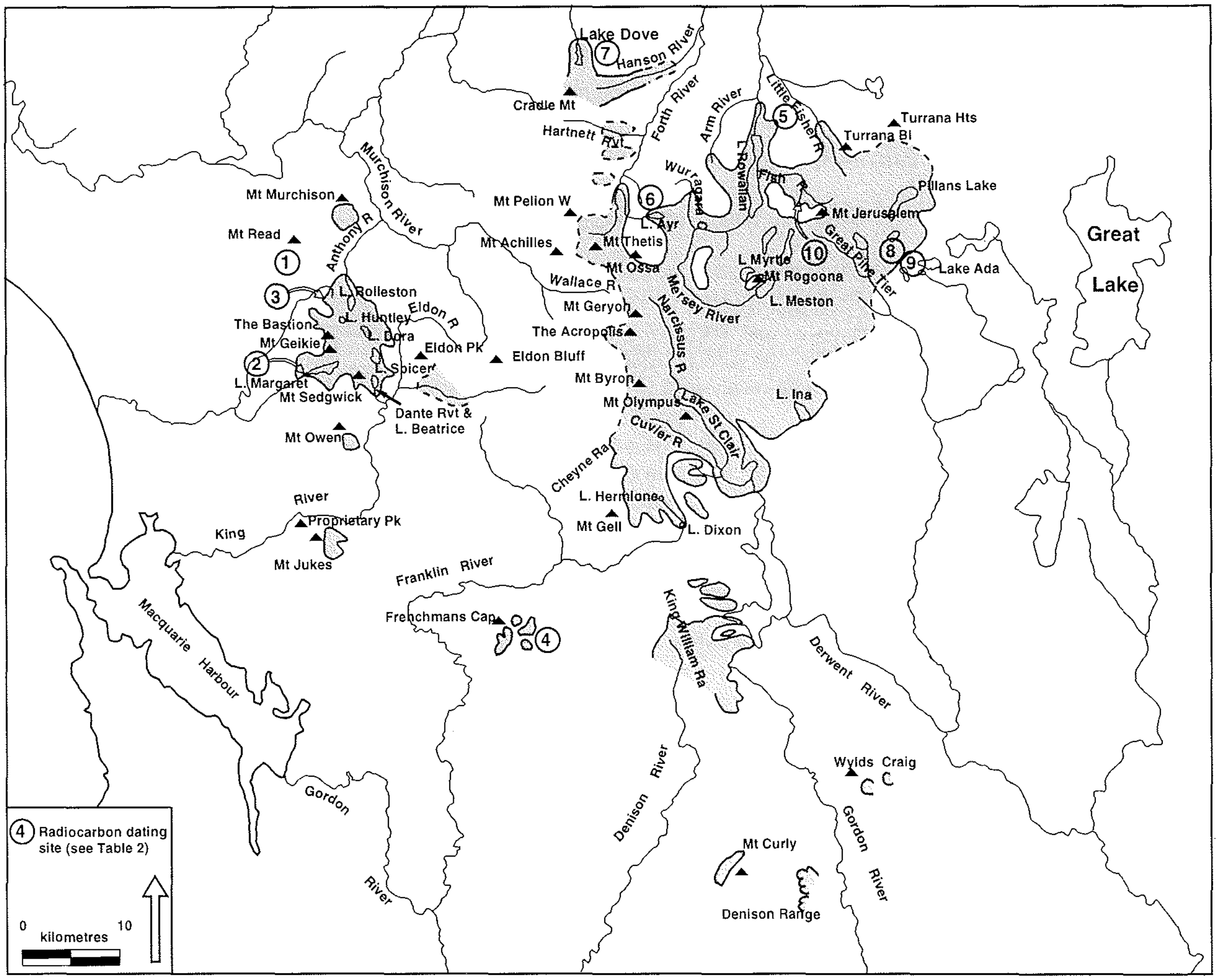

FIG. 2 — The major areas of glacier ice during the Late Wisconsin.

Clair area (Kiernan 1990). The Late Wisconsin ice limits are not yet known in all parts of Tasmania, e.g. the ice limit shown in figure 2 for the southeastern Central Plateau is approximate, and only some of the mountain areas of southwestern Tasmania have been mapped recently. Nevertheless, many new data have been obtained for the extent of Late Wisconsin ice, which allow revision of the small-scale map of Davies (1967).

This paper illustrates, with examples, criteria used to separate Late Wisconsin from older glacial landforms and deposits. It describes evidence for the Late Wisconsin ice limits in the West Coast Range, north–central and south–central parts of highland Tasmania (figs 1, 2), and discusses the Late Wisconsin glacial environment. The greater extent of middle and early Pleistocene ice is shown in figure 3.

SEPARATION OF LATE WISCONSIN FROM OLDER GLACIAL LANDFORMS AND DEPOSITS

The criteria used to separate Late Wisconsin and older glacial landforms and deposits are indicated in table 1.

Throughout the mountains and plateaux of Tasmania landscapes glaciated during the Late Wisconsin can be recognised by ice-scouring in the source areas, and deposition of end moraines and outwash terraces at the ice limits. Ice developed on the high western parts of the Central Plateau, on the small plateau-summit of the Tyndall Range (pl. 1A), and in cirques and valley heads, where snow borne by westerly and southwesterly winds accumulated leeward of watershed-ridges.

The source areas are distinguished by glacially abraded rock surfaces, scraped almost bare of soil and debris. In addition, ice-plucking has emphasised rock structures, as shown by the NW–SE and NE–SW joint-controlled development of the rock-basin lakes of the Central Plateau (Jennings & Ahmad 1957). Quartzites and other siliceous rocks in the western mountains exhibit numerous striations, which are rare on the dolerites of the Central Plateau due to postglacial chemical weathering.

The glaciers that flowed from the Tyndall Range and Central Plateau through deep valleys terminated in conspicuous end moraines with steep ice proximal/contact slopes, as did cirque and valley glaciers in the mountains of southwestern Tasmania. Most end moraines are 10–100 m in height, and some form impressive scenic features. The highest is the Hamilton End Moraine that has 300 m of relief, though it was deposited at the edge of a bedrock shelf which accounts for about half the relief (pl. 1B). Outside the Late Wisconsin ice limits, moraines formed

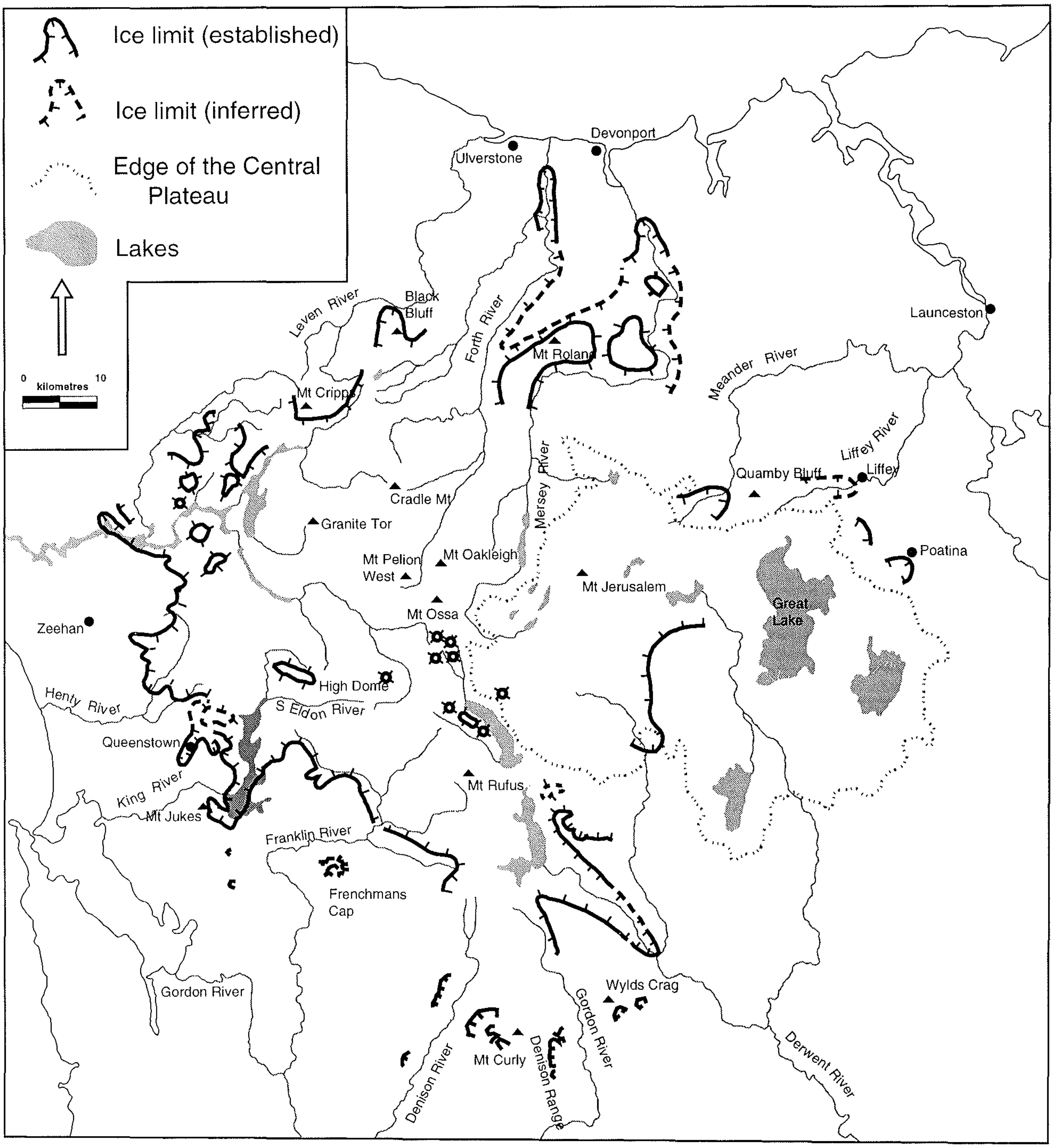

FIG. 3 — The maximum extent of Pleistocene ice in western and central Tasmania.

during middle and early Pleistocene glaciations do not exhibit steep ice-contact slopes, but have more rounded forms. Their slopes are dissected by minor, water-cut valleys. Outwash terraces or fans, undissected except for incision of the main postglacial river courses, extend down-valley from many of the Late Wisconsin end moraines. The terraces consist of bedded gravel and sand, and may extend several kilometres before ending, usually at a constriction in valley cross-profile.

Outside the end moraines and above the lateral moraines of both Late Wisconsin and earlier glaciations, the plateaux are covered with rock debris. The debris was formed from weathered regolith of Jurassic dolerite that contained many corestones, and from bedrock by periglacial processes. Though largely undated, the rock debris accumulated over periods much longer than single glacial events. Thick periglacial debris is never found within Late Wisconsin ice limits. The development of extensive blockfields and blockstreams on Ben Lomond, in northeastern Tasmania, was used by Caine (1983) to assign glaciation of the plateau to a pre-Wisconsin stage, but the extent of such glaciation is not known.

Between the zone dominated by glacial erosion and the end moraines, Late Wisconsin till and associated deposits mantle the valley floors and lower slopes, often discontinuously. The till is variable in thickness and consists of boulders and cobbles supported in an unweathered, compact matrix, derived mainly from local rocks (pl. 1C).

TABLE 1

Late Wisconsin forms and deposits	Pre-Late Wisconsin forms and deposits
Landforms	
Cirque headwalls, valley sides and plateau surfaces ice-scoured with rock basin lakes. Virtually no soil and little rock detritus	
Large,sharp-crested, undissected end and lateral moraines	Moraines have rounded crests and are dissected by minor channels
Extensive, undissected outwash terraces	Extensive outwash terraces, little to moderately dissected by minor streams
	Extensive boulder fields, blockstreams and tor-forms on high plateaux, with screes and solifluction deposits on slopes
Deposits	
Geographically the innermost drift sheet	Occur geographically outside the Late Wisconsin ice limits
Igneous clasts in till and outwash not weathered below shallow soil prpfile of <1–1.5 m depth	Igneous clasts in till and outwash moderately to strongly weathered; soil profiles may be >1–1.5 m in depth
Only moderate iron and manganese staining of deposits	Moderate to very strong iron/manganese staining of deposits. Reduced specific gravity of volcanic clasts
Only rare, poorly developed, iron-pans formed below soil	Thick and multiple iron-pans may be formed below soil
Dating	
Mean weathering rind thickness of subsurface Jurassic dolerite clasts usually <1 mm at 300 mm depth (max. 5 mm)	Weathering rinds on subsurface Jurassic dolerite clasts up to 200 mm thick
Younger and contemporary organic materials datable by radiocarbon	Associated organic materials beyond radiocarbon range. Wood in interglacial deposits assayed by amino-acid method
	Oldest glacial deposits have reversed magnetisation

Few easily recognised erratics occur in northern and south-central Tasmania, but in the glacial deposits of western Tasmania igneous erratics are common. Only rarely are stratigraphic contacts found that show Late Wisconsin till overlying till of an older glaciation. The best example occurs in the upper Arm Valley of northern Tasmania, where Late Wisconsin Rowallan Till overlies Arm Till of the preceding glaciation (probably of Isotope Stage 6 (pl. 1D). The Hamilton End Moraine (pl. 1B) also shows complexity of stratigraphic development. Till containing strongly weathered clasts of the Mt Read Volcanic Group occurs on the surface of Boulder Hill in the centre of the moraine arc, but the major part of the moraine formed of Owen conglomerate boulders is of Late Wisconsin age (E.A. Colhoun & P. Augustinus, pers. obs.).

The Late Wisconsin till, ice-contact and outwash gravel and sand deposits can be distinguished from older glacial deposits by the relative lack of weathering of igneous clasts (see dating, table 1). In addition, Late Wisconsin age deposits do not have thick iron-pan horizons, and iron and manganese oxide staining does not extend below the soil profile. At freely drained sites, postglacial soil development is minimal, and the C_{ox} horizon does not exceed 1–1.5 m depth (Hammond 1985). Older glacial deposits are characterised by strong weathering of igneous clasts, iron-pan formation, iron and manganese oxide precipitation, and deep soil formation on freely drained and uneroded sites.

Dating and age differentiation of Late Wisconsin from older glacial deposits have employed several methods. Kiernan (1983a) used mean thickness of weathering rinds on erratic dolerite clasts in tills of the central West Coast Range to assess the age of deposition. He showed that, in Late Wisconsin tills, weathering rinds on dolerite clasts at 300 mm depth were < 1 mm in mean thickness, and that two older tills gave mean weathering-rind thicknesses of 4.9 ± 2.8 mm and 55.8 ± 21.6 mm (fig. 4). Augustinus & Colhoun (1986) also showed the effects of increased weathering of igneous clasts with age by demonstrating that volcanic clasts of the Mt Read Group have reduced specific gravities in stratigraphically older tills. These studies, combined with the work of Bowden (1974) and Colhoun (1985), convincingly demonstrated pre-Late Wisconsin age multiple glaciation of western Tasmania, as hypothesised by Lewis (1945). Kiernan used a linear relationship, between weathering rind thickness and time, to suggest that the oldest glacial deposits in western Tasmania were more than 600 000 years old. Kiernan's method has also been applied to the glacial deposits of north-central and south-central Tasmania. In both regions, the results show that Late Wisconsin glacial deposits are little weathered and older deposits are highly weathered (Hannan & Colhoun 1987, Kiernan 1992).

Radiocarbon age assay of the Late Wisconsin glacial stade in Tasmania has been obtained by dating organic-rich deposits in three situations. Firstly, dates have been obtained from organic deposits in lakes and bogs outside the inferred Late Wisconsin ice limits. Several of these sites have pollen/vegetation histories that allow estimation of the date of commencement of the last cold stade. Secondly, an important section occurs at Dante Rivulet which has permitted close dating of the maximum advance of Late

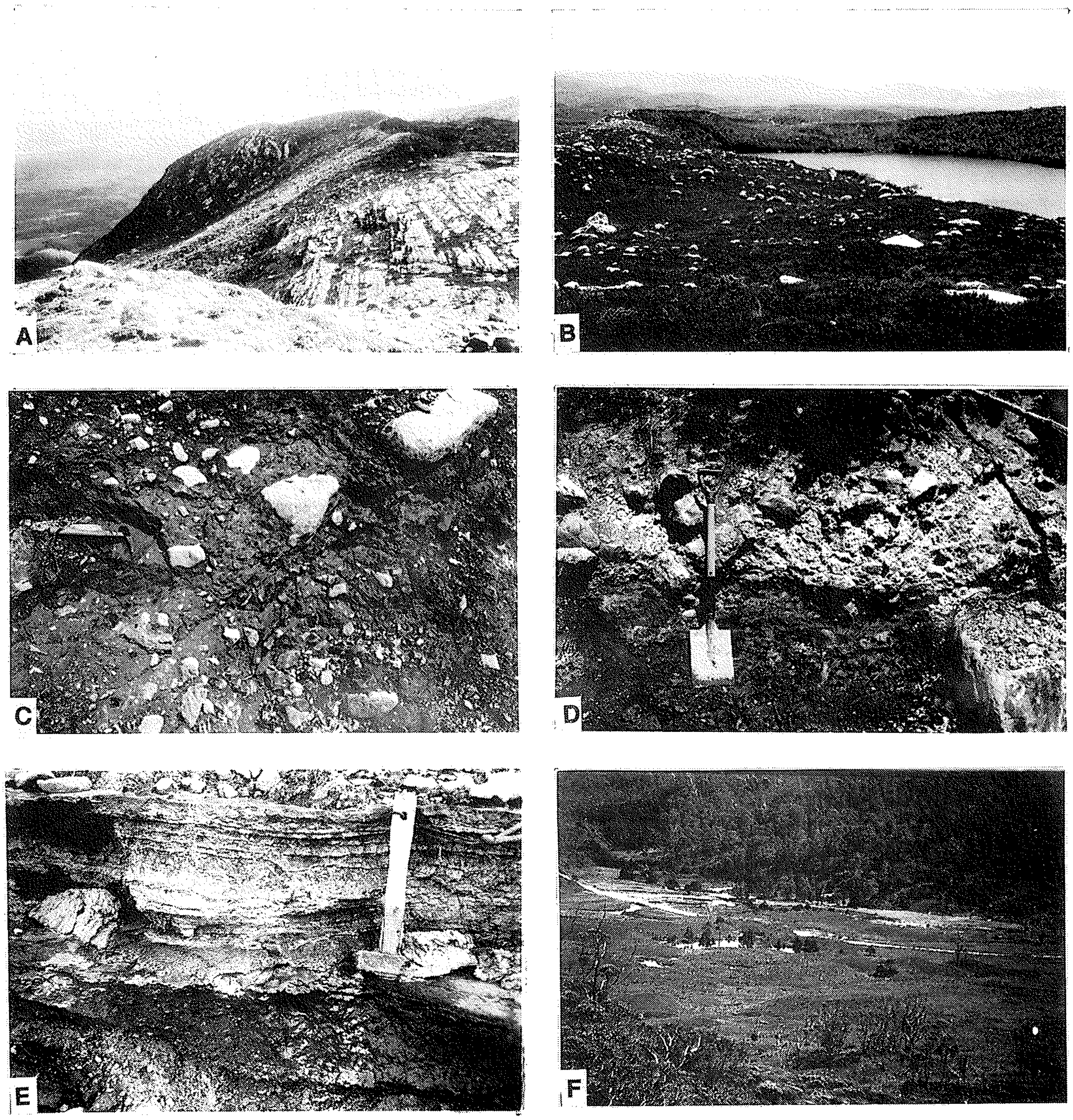

PLATE 1

(A) Ice smoothed conglomerate and sandstone of the Tyndall Plateau with end moraine on the plateau edge, looking north from the Bastion on Mt Geikie (Photo A.Bowden, 1974). (B) The Hamilton End Moraine surrounding Basin Lake showing the steep eastern ice-contact slope (right) and the boulder-mantled distal slope (left).(C) Late Wisconsin age basal till deposits in the Sedgwick Valley, showing the chemically unweathered surfaces of boulders and compact matrix. (D) Boulder-rich till of the Late Wisconsin Rowallan Glaciation disconformably overlying strongly weathered older till of the Arm Glaciation in the upper Arm Valley.(E) Fossil alpine humus soil with dated bolster of Donatia novae-zelandiae *underlying the Dante outwash fan and interstratified King River alluvial silts in the King Valley.(F) Hummocky end moraines with kettle holes deposited by the Dove and Crater lakes glaciers on the floor of Cradle Valley.*

Wisconsin ice from the West Coast Range (Gibson *et al.* 1987, Colhoun & Fitzsimons 1996). Thirdly, numerous dates have been obtained for the oldest organic sediments deposited in postglacial lakes and bogs within the inferred Late Wisconsin ice limits (Macphail & Peterson 1975; table 2). These dates provide minimum ages for deglaciation.

The first radiocarbon assay of 26 480 ± 800 yr BP (W-323) obtained by Gill (1956) has been found to have been obtained from derived Pliocene wood contaminated by young humic acids (Macphail *et al.* 1995). The date of commencement of Late Wisconsin glacial conditions is inferred from pollen-rich sediment sequences located outside

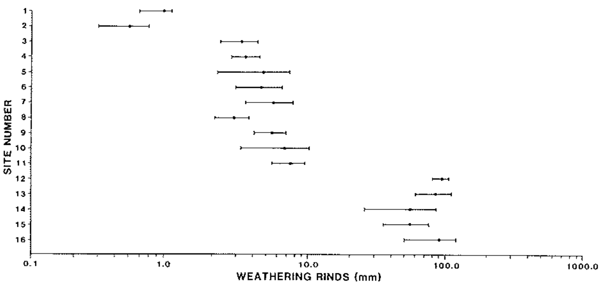

FIG. 4 — Mean and standard deviation of thicknesses of dolerite weathering rinds from the central West Coast Range plotted on four-cycle semi-log base. Where possible 20 sample rinds were recorded at each site. Sites 1–2, Last Glacial Maximum; 3–11, Middle Pleistocene; 12–16, Early Pleistocene or Late Tertiary. After Kiernan (1983b).

but close to the ice limits. Severe cold is indicated by maximum development of vegetation dominated by Gramineae, Asteraceae and Chenopodiaceae, and very few pollen grains of rainforest taxa. An age of 26 000–25 000 years is estimated from two sites in the West Coast Range: Lake Selina, which occurs 2.5 km north of the Rolleston end moraine (see "West Coast Range"), and Tullabardine Dam, which occurs in the Mackintosh Valley and 20 km north of the ice limit (Colhoun & van de Geer 1986, Colhoun *et al.* in press). These sites provide the best estimate for the onset of severe cold during the Late Wisconsin in Australia, as pollen analysis of sediment sequences at sites distant from ice sources are unlikely to signal as sharp a climatic change.

At Dante Rivulet in the King Valley, outwash sands and gravels deposited while ice of the West Coast Range glacial system was at its maximum extent, are interstratified with organic-rich fluvial silts deposited by the King River. These deposits overlie a humus-rich palaeosol, with fossil bolsters of the alpine cushion plant *Donatia novae-zelandiae* (pl. 1E). A radiocarbon assay of 19 100 ± 170 yr BP (SUA-2856) on wood fragments from the river silts within the outwash, and assays of 20 100 ± 470 yr BP and 21 180 ± 370 yr BP (SUA-2155, SUA-2154) on the underlying *Donatia* bolster, indicate that the peak of the Late Wisconsin glacial stade in Tasmania was around 19 ka. This date correlates closely with dates for the maximum extent of ice during the Late Wisconsin in western New Zealand and in Chile (Suggate 1965, Suggate & Moar 1970, Porter 1981). The date confirms the almost synchronous occurrence of the Late Wisconsin glacial maximum in southern hemisphere middle latitudes.

The oldest postglacial date was obtained from the small cirque Ooze Lake at 880 m altitude in southeastern Tasmania. Ice did not extend outside the cirque during the Late Wisconsin and had melted before 17 700 ± 400 yr BP (SUA-1359) (Macphail & Colhoun 1985), indicating that the mountains of southeastern Tasmania were only marginally glaciated. The youngest assay of 9080 ± 200 yr BP (I-7571) occurs at 960 m at Adamsons Peak, which lies 18 km to the northeast (Macphail 1979). The difference seems to have little bearing on the age of initial deglaciation in southeastern Tasmania, unless the Ooze lake date is erroneous.

In the West Coast Range region, the oldest postglacial date from within the ice limits is 11 530 ± 240 yr BP (I-7683) from 560 m altitude at Lake Vera, and the youngest 9050 ± 120 yr BP (SUA-1358), from 1000 m altitude at Tarn Shelf on the Tyndall Plateau (Macphail 1979, 1986). However, deglaciation may have commenced earlier, judging from a date of 13 010 ± 130 yr BP (SUA-2793) for the commencement of postglacial organic deposition at 180 m altitude, at Governor Bog in the King Valley (Colhoun *et al.* 1991a), and the cessation of strong frost weathering of the limestone roof of Kutikina Cave (formerly Fraser Cave) in the Franklin Valley, c. 14 840 ± 930 yr BP (ANU-2781) (Kiernan *et al.* 1983). Older glacial and interglacial deposits in western Tasmania have been separated from Late Wisconsin age deposits by amino-acid dating of wood and by palaeomagnetic investigations (Colhoun & Fitzsimons 1990, Pollington *et al.* 1993).

Dates of similar minimum age for deglaciation occur in north–central Tasmania and on the Central Plateau. The date of 13 400 ± 600 yr BP (SUA-2188), obtained from an infilled lake inside an end moraine at Dublin Bog, at 575 m altitude in the Mersey Valley, indicates retreat from the ice limits before this time (Colhoun *et al.* 1991b). The date of 12 420 ± 150 yr BP (SUA-3046), from Wild Dog Creek, occurs at 1170 m altitude just inside the upper margin of ice that failed to override the Mt Jerusalem area (Hannan & Colhoun 1991). The date indicates retreat of the upper ice margin on the western Central Plateau before 12.4 ka, while the later dates near Lake Lanka and Talinah Lagoon in the east suggest retreat of the eastern ice margin before 11 ka (table 2, fig 2).

Several dates, taken from short sediment cores in cirque lakes in the Mt Field National Park, show that deglaciation at the highest altitudes had taken place by c. 10 ka, e.g. ages of 12 960 ± 950 yr BP (I-7684) at Eagle Tarn, 11 420 ± 205 yr BP (SUA-325) at Beatties Tarn and 9725 ± 190 yr BP (I-8008) from an unnamed tarn on the Tarn Shelf (Macphail 1979). At present, no evidence has been found for a distinct Younger Dryas-type re-advance during the Late Glacial, though a block moraine with a high protalus component at 1100 m altitude east of Lake Newdegate could have been formed after the main deglaciation.

LATE WISCONSIN ICE EXTENT AND GLACIAL LANDFORMS

The description of glacial landforms and ice limits that follows is an updated and integrated version of detailed records for the West Coast Range Bowden (1974), Kiernan (1980), Colhoun (1985), Hammond (1985), Fitzsimons & Colhoun 1991, and Fitzsimons *et al.* (1992), the northwestern Central Plateau and Mersey Valley (Hannan & Colhoun 1987; 1991), the Forth Valley (Kiernan & Hannan 1991), Cradle Mountain (Colhoun 1980), the Central Highlands and Lake St Clair area (Kiernan 1990a, 1991) and the Franklin Valley (Kiernan 1989). It is not possible to show all names on the small-scale maps (figs 2, 3) of the main areas covered by ice during the Late Wisconsin, and maximum extent of ice in pre Wisconsin time. Most localities referred to are shown on the 1:100 000 scale maps of the Topographic Survey of Tasmania.

TABLE 2
Selected post-glaciation radiocarbon dates

Site	Altitude (m)	Date ^{14}C yr BP	Lab no.	Source
Western Tasmania				
1 Lake Johnson	875	9380 ± 110	SUA-2987	Anker (1991)
2 Poets Hill	600	11 420 ± 770	GAK-6297	Colhoun (1992)
3 Tarn Shelf, Tyndall Mtns	1000	9050 ± 120	SUA-1358	Macphail (1986)
4 Lake Vera	560	11 530 ± 240	I-7683	Macphail (1979)
Northern Tasmania				
5 Dublin Bog	575	13 400 ± 600	SUA-2188	Colhoun *et al.* (1991b)
6 Lake Ayr	870	11 020 ± 280	SUA-3060	Wilson & Hannan (1995)
7 Lake Dove	934	10 620 ± 60	Beta-82244	Dyson *et al.*, pers. comm.
8 South of Lake Lanka	1190	10 960 ± 230	Beta-68716	D. Hannan, pers. obs.
9 North of Talinah Lagoon	1160	10 380 ± 110	SUA-3056	D. Hannan, pers. obs.
10 Wild Dog Creek	1170	12 420 ± 150	SUA-3046	D. Hannan, pers. obs.
Southern Tasmania				
11 Adamsons Peak	960	9080 ± 200	I-7571	Macphail (1979)
12 Beatties Tarn	990	11 420 ± 205	SUA-325	Macphail (1979)
13 Eagle Tarn	1033	11 400 ± 235	I-7684	Macphail (1979)
14 Tarn Shelf, Mt Field	1158	9725 ± 190	I-8008	Macphail (1979)
15 Ooze Lake	880	17 700 ± 400	SUA-1359	Macphail & Colhoun (1985)

West Coast Range

During the Late Wisconsin, ice accumulated on the 1000 m high plateau area in the central part of the Tyndall Range (fig. 2), in the cirques at the head of Tyndall Valley and occupied by Lake Huntley to the north and northeast, and in cirques east of the Bastion and Mount Geikie to the south. In addition, ice accumulated in cirques leeward of Mt Murchison, in the northern part of the West Coast Range, and leeward of the summits of Mt Owen and Mt Jukes, in the southern part. The extent of ice did not exceed 105 km^2 (Colhoun 1985: 41). The icecap was thin, with probably no more than 50–100 m of ice cover on the upland plateau areas of the Tyndall Range and around Lake Dora. The ice attained a maximum thickness of about 250–300 m in the adjacent cirques and deep valleys.

The icecap that accumulated on the Tyndall Range abraded the rock surface, formed small ice-eroded rock basins and moved both westwards and southeastwards. On the western side, the ice limit is marked by distinct end moraines. These extend from the plateau margin a short distance downslope into the minor valleys cut into the western scarp of the Tyndall Range (pl. 1A). To the east, ice flowed across the Lake Dora plateau towards the Eldon Valley, and southeastwards via Lake Spicer to Lake Beatrice. Several areas of fluted drift occur on the plateau, as near Farquhar Lookout. The ice margin east of Lake Dora is marked by a line of large Owen Conglomerate (Ordovician) erratics flanking a small nunatak, and the margin east of Lake Spicer by small moraines. South of Lake Beatrice, the ice margin formed an end moraine and the associated Dante outwash fan. A lateral moraine-crest extends downslope, from 800 m altitude on the southeastern shoulder of Mt Sedgwick to 400 m at the Lake Beatrice end moraine. South of Mt Sedgwick, a small glacier extended down the Sedgwick Valley to 600 m altitude. The limit is marked by an end moraine that shows dark grey-coloured till (pl. 1C) that has not been weathered despite containing dolerite clasts derived from Mt Sedgwick.

The ice that accumulated east of the Bastion and Mt Geikie flowed through Lake Margaret to the Hamilton End Moraine (pl. 1B). The ice flow scoured the basin and eroded the large rock sill at the western end of the lake, before expanding into a bipartite piedmont glacier that extended down to 550 m altitude on the northern side and 300 m on the southern. South of Lake Margaret, lateral moraines descend from 950 m altitude westwards to the very large end moraine, that has a crest at 640 m and base at 300 m, adjacent to Lake Margaret Village. This is the highest end moraine in Australia.

North of the West Coast Range, the Tyndall Glacier flowed 2.5 km northwards to form a 70–100 m high end moraine. In the adjacent valley to the east, the Rolleston Glacier flowed 5 km northeastwards, from Lake Huntley cirque via the deep glacial valley occupied by Lake Rolleston, and terminated in a large complex of end moraines, south of the flat-floored Anthony outwash plain.

Further north, and leeward of Mt Murchison, two composite, deeply incised cirques were sources of south-easterly flowing glaciers. The cirques are now occupied by several tarns, the two largest being Lake Sandra and Lake Gaye. The glaciers descended steeply downslope and formed end moraines, with crests at 440 m altitude in the southern complex and 520 m in the northern.

Several other cirque and short valley glaciers occur leeward of high summits in the West Coast Range. A small cirque glacier formed at 900–1050 m altitude south of Mt Read and eroded the basin occupied by Lake Johnston. East of the southern watershed of the West Coast Range, a small steep-sided moraine occurs within the cirque southeast of

Mt Owen. The cirque form clearly pre-dates the Late Wisconsin stage and has been occupied by ice during more than one glacial event. Similarly, two distinct moraines composed of unweathered deposits occur at the northern and southern ends of the complex of cirques that extends from Proprietary Peak to South Jukes Peak. These cirques also have a multiglacial origin (Fitzsimons *et al.* 1992).

The cirque, valley glacier and icecaps that formed in western Tasmania during the Late Wisconsin glacial stade were not coterminous with ice masses formed in north-central Tasmania, at Cradle Mountain and in the upper Forth and Mersey Valley regions or in south-central Tasmania, in the Lake St Clair, upper Derwent and Franklin Valleys.

North-Central Tasmania

The maximum extent of Late Wisconsin ice in north-central Tasmania has been traced from Great Pine Tier northwest into the Mersey and Forth Valleys and the Cradle Mountain area.

Late Wisconsin ice covered 550 km^2 of the Central Plateau, of which 175 km^2 was north of the Great Pine Tier. The Central Plateau consists of Jurassic dolerite and is inclined southeastwards. The high western part attains approximately 1200 m altitude, and isolated peaks of 1400 m occur in the northwest. The Great Pine Tier is a 150–200 m high escarpment that extends 25 km southeast from the Walls of Jerusalem National Park and divides the Central Plateau into northern and southern parts. The northernmost and eastern parts of the Central Plateau were not glaciated, while in the northwest the iceflow surrounded the Walls of Jerusalem National Park (Hannan & Colhoun 1991). A thin icecap formed near the western edge of the plateau and flowed eastwards and southeastwards. The ice eroded the strongly jointed dolerite surface to produce a multitude of roches moutonnées, whaleback ridges and rock basins now occupied by lakes (Jennings & Ahmad 1957). To the northwest, the Mersey and Forth Valleys are deep, steep-sided and often flat-floored, and their glaciers were fed by ice from the higher surrounding plateaux (Hannan & Colhoun 1987).

End moraines commonly mark the Late Wisconsin ice limits, but are not continuous around the former icecap. Instead, they are concentrated in certain areas, such as west of Lake Ada and east of Pillans Lake, where they form hummocky mounds. Surface boulders are abundant and often exceed 4 m in diameter. Moraines seldom exceed 6 m in height and consist of light grey, poorly-sorted cobbles and boulders in a fine matrix. Weathering rinds on subsurface clasts are less than 1 mm thick. There is no basal till on the eroded rocky surface of most of the glaciated area, and only small amounts have been deposited on the floors of some minor valleys near the periphery.

The Late Wisconsin ice limit has been mapped using moraines and surface boulders. These moraines are higher and have steeper slopes than older moraines, and they contain a high proportion of fine sediment. The older moraines can be devoid of fine material on the surface and may consist only of rounded dolerite boulders. Boulders inside the Late Wisconsin ice limits are rounded, entire and generally show only slight surface weathering. Those outside the limit show greater effects of exposure in the form of surface shattering, exfoliation and an accumulation of weathered material around their bases, but the limits attained by ice during earlier glaciations have not been determined in the eastern Central Plateau.

A Late Wisconsin maximum ice thickness of 80 m can be deduced for the northern part of the Central Plateau from the interrelationships of moraine trimlines, and the lower limits of blockfields and screes in the Turrana Bluff–Turrana Heights region. When the differences in altitude of the former ice margin and plateau surface are considered, they suggest the divide between west and east-flowing ice was close to the western edge of the Central Plateau, and that most ice flowed towards the east and southeast. As a result, very little ice spilled off the plateau to the northwest. This is demonstrated by the short distance that ice travelled down the Little Fisher River and by the absence of Late Wisconsin ice from the central part of the Walls of Jerusalem, west of Mt Jerusalem (Hannan & Colhoun 1991). The thickest ice on the Central Plateau built up in the lakes Adelaide–Meston–Louisa–Myrtle area. Mt Rogoona formed a large nunatak rising above the Central Plateau icecap between lakes Myrtle and Meston. The contrast between the blocky dolerite screes on its upper slopes and the ice-moulded landscape below shows that the ice was about 200 m thick at Lake Myrtle. The southwestward flow of ice from this area of deeply ice-eroded valleys with ribbon lakes formed the main supply for the Mersey Glacier. Stratigraphic evidence in the Fish Valley shows that the 300–350 m thick Mersey Glacier reached the lower Fish Valley before the thinner ice flowing northwestwards past the Walls of Jerusalem, entered the valley head.

The Mersey Glacier comprised a reticulate system of ice fed from the Central Plateau via the Junction Lake area, Moses Creek, Lake Louisa and the Fish Valley. Ice in the Little Fisher Valley did not reach the main valley during the Late Wisconsin (Hannan & Colhoun 1987). The main Mersey Glacier terminated in lobes at Wallace River, Lake Ayr Valley, Wurragara Creek, the upper Arm Valley, Dublin Bog, and in the Mersey Valley, near its confluence with the Arm Valley. Moraines mark the ice limits and vary from hummocky moraines at Lake Ayr (Wilson & Hannan 1995) and lateral moraines at Wurragara Creek, to distinct end moraines at Dublin Bog and in the Mersey at the Arm Valley confluence. Around Rowallan Dam, the glacier eroded and striated the bedrock extensively. In the upper Arm Valley, dolerite till deposits of the Late Wisconsin glaciation stratigraphically overlie till of the preceding glaciation (pl. 1D), and at Rowallan Spillway, a 70 m section shows a Late Wisconsin sequence of unweathered sand, gravel and till deposits overlying an older sequence of more weathered deposits, that include rhythmites, sand, gravel and till. A Uranium-series assay of a thick ironpan between the two sequences of deposits gave a date of $95^{+80}/_{-51}$ ka (LH-1566) (P. Augustinus, pers. comm. 1995), indicating that the lower sequence is of pre-Late Wisconsin age.

During the Late Wisconsin glaciation of the Forth Valley, ice accumulated in small cirques at Mt Achilles, Mt Pelion West and Mt Thetis, and formed a 13 km long glacier that terminated near the junction of Oakleigh Creek, where a number of small end moraines occur (Kiernan & Hannan 1991). At this stage there was no connection through the Lake Ayr Valley between the Mersey and Forth glacial systems (Kiernan & Hannan 1991, Wilson & Hannan 1995). Further north, ice that accumulated in Waterfall Valley, south of Cradle Mountain, flowed 9 km down the

Hartnett Rivulet Valley. Similarly, ice that accumulated east of Cradle Mountain and ice that breached Hansons col from Lake Dove, flowed 10.5 km down Hansons Valley towards the Forth Valley.

North of Cradle Mountain ice accumulated in the cirques occupied by Lake Wilks and by Crater Lake. The ice from Lake Wilks flowed into Lake Dove and was over 200 m thick. East of Lake Dove, the summit of Mt Campbell is covered with quartzite blockfield debris and was not overridden by Late Wisconsin ice. Mt Campbell deflected the ice northwards, and the eastern trimline of the former Dove Glacier declines steeply along the western face of the mountain toward the hummocky moraines on the floor of Cradle Valley, that mark the ice limit (pl. 1F) (Colhoun 1980). West of Cradle Mountain, a small moraine bounds Suttons Tarn on the western edge of the Cradle Plateau and, north of Cradle Mountain, snow from Cradle Plateau formed a relatively large isolated cirque glacier that occupied Crater Lake. The glacier flowed northwards into Cradle Valley, where it merged with the terminal part of the Dove Glacier to the east.

South-Central Tasmania

In south-central Tasmania, the main directions of iceflow were southeast across the Central Plateau south of Great Pine Tier, southwards from sources in the Du Cane Range through the Lake St Clair and upper Derwent Valley, and southwestwards towards the upper Franklin Valley.

Southeast of the Walls of Jerusalem, ice flowed along Powena Creek towards the gap in the Great Pine Tier, south of Lake Fanny, but did not cover the southeastern part of Great Pine Tier. Ice that formed close to the western rim of the Central Plateau between the northwestern part of Great Pine Tier and the Mountains of Jupiter flowed 10–14 km southeast towards ice limits which are defined by low end moraines, as at Lake Ina (Kiernan 1990b). The general southeasterly iceflow is demonstrated by the NW–SE orientation of the many ice-eroded lake basins on the plateau surface. Only in Travellers Rest Valley did the ice reach the margin of the Central Plateau, and here at least 14 moraines record retreat of the ice margin northwards (Kiernan 1985, 1991). Some of the ice formed southwest of the Mountains of Jupiter flowed southwestwards to join the Derwent Glacier that occupied the deep trough of Lake St Clair.

The main source of the Derwent Glacier lay in the Du Cane Range, where high ridges, such as Mt Geryon and The Acropolis, acted as natural snowfences to the prevailing westerly weather, leeward of which considerable volumes of ice accumulated. The largest glacier to reach Lake St Clair flowed down Narcissus Valley. Tributary glaciers flowed southeastwards along Pine, Marion and Hamilton Valleys, and became confluent with Narcissus ice. Some diffluent ice from the Mersey Glacier also passed southward through Du Cane Gap into the head of Narcissus Valley. During the maximum of the Late Wisconsin Glaciation, ice in the upper Narcissus and lower Pine Valleys was over 300 m thick, and the down-valley surface gradient east of Mt Olympus was 25–30 m/km. The Derwent Glacier extended to 1.2 km south of Lake St Clair, where it was 30 km distant from the Du Cane Range. At the head of Lake St Clair, some ice spilled through the gap between Mt Olympus and Mt Byron into the Cuvier Valley (Kiernan 1985, 1992).

The glacial landforms in the upper Derwent Valley area result from multiple Pleistocene ice advances. Only in a few cases is it possible to ascribe erosional landforms solely to the Late Wisconsin. Cirques occur at the heads of the principal valleys and their tributaries. These are cut in Jurassic dolerite, the columnar jointing of which has produced very steep headwalls. The steepest headwalls occur where erosion has cut into the underlying Permian mudstone and sandstone, which has allowed strong basal-sapping of the columns and valley-widening. The morphology of the glacial troughs varies according to rock type, valley cross-profile and thalweg gradient. Parabolic cross-profiles are most nearly attained where the troughs are cut in subhorizontally-bedded Permian sediments. The troughs tend to be narrower where the opposing walls are dolerite, and where the thalweg is steep. Lake St Clair has been deeply eroded by ice and dammed by moraines, to give a maximum depth of 167 m (Derbyshire 1971). The Labyrinth plateau area, west of upper Pine Valley, was ice-scoured by westward-flowing ice during the Late Wisconsin, which eroded many rock basins now occupied by shallow lakes such as Long Lake. During earlier glaciations, the Labyrinth was overridden by ice from the Murchison Valley that became confluent with the Derwent Glacier (Kiernan 1990b).

North of Lake St Clair, fluted drift occurs on the floor of Narcissus Valley, and moraine ridges have been deposited around the head of the lake in the area of glacial confluence. South of Lake St Clair, an impressive series of steep and narrow end moraines occurs. These moraines are generally less than 10 m high and are spaced 50–100 m apart; some can be traced for over 3 km (Derbyshire 1963, 1965, 1967). The steep distal slopes suggest they were formed by numerous minor re-advances rather than by gradual retreat of the glacier terminus, and the large number indicates that the glacier was very sensitive to short-term changes in mass balance. In contrast to the West Coast Range, the Derwent Glacier deposited abundant outwash sediments between the moraine ridges. The greater quantity of outwash deposits is probably due to the larger mass of ice in the southern Central Highlands than in the West Coast Range, and to greater storage of the outwash on the broad valley-plains than in the deep valleys of the west (Kiernan 1985, 1992).

During the Late Wisconsin, ice from the Derwent Glacier overflowed into the Cuvier Valley, and from there into the head of the Franklin Valley, where it merged with local ice that accumulated along the eastern slopes of the Cheyne Range from Lake Hermione to Mt Gell. The Franklin Glacier extended for 12 km to Lake Dixon, and constructed one large end moraine complex that contrasts with the many individual moraines formed south of Lake St Clair. Ice formed in the Mt Gell cirques flowed into the upper Franklin Valley floor at Lake Undine, where it was c. 150 m thick. The upper Franklin Valley is a well-developed glacial trough. It has a steep thalweg and is considerably narrower than the troughs around Lake St Clair, probably due to being cut in dolerite. Further west, individual small glaciers existed in the upper parts of valleys tributary to the Franklin River (Kiernan 1985, 1989).

Highlands, gave values of 1050 m for the Mersey and 1020 m for the Derwent. These values are consistent with ELAs of major glaciers being lower than the regional snowline on adjacent icecaps. The 30 m increase in ELA between the Derwent and Mersey glaciers parallels the northward increase in regional snowline on the Central Plateau. The northward or northeastward rise in the regional snowline suggests that precipitation inputs from the southwest contributed to the much greater development of ice in the upper Derwent and Franklin Valleys than north of the Du Cane Range, in the Forth Valley–Cradle Mountain area. Comparison of the ELA levels with modern freezing levels indicates reduction of Late Wisconsin snowlines by c. 950 m and temperatures by around 6.2°C, relative to present, in both the Mersey and Derwent Valleys.

The absence of Late Wisconsin ice from the surface of the Ben Lomond Plateau in northeastern Tasmania indicates that the regional snowline was above 1400 m (Caine 1983). At Legges Tor in the northeast, the snowline must have exceeded 1500 m, otherwise Legges Tor would have had a small icecap. The small cirque glaciers at Denison Crag in the south, and at Hamilton Crag and Broken Bluff on the eastern scarp give ELAs of 1250–1260 m. These values, however, reflect orographically induced snowline levels, which could be as much as 300 m below the regional snowline.

CONCLUSION

The Late Wisconsin glaciation of Tasmania was smaller than middle and early Pleistocene glaciations of the island. Icecaps and associated outlet valley glaciers developed in two centres; the Central Plateau and adjacent mountains of the Central Highlands, and in the West Coast Range. The onset of glaciation commenced after 26–25 ka BP and peaked at 19 ka BP and all known ice had decayed before 10 ka BP. The regional snowline increased in altitude across the island from below 800 m on the west coast to over 1500 m in the northeast. The large glaciers of the West Coast Range and Central Highlands and cirque glaciers of the western mountains were more active than the thin ice cap on the Central Plateau, and the isolated cirque glaciers formed leeward of mountain ranges and plateaux scarps in the east. The Late Wisconsin snowline was lowered by about 1000 m, and mean temperature was depressed by about 6–6.5°C from present. The system reflects development under westerly and southerly onshore, oceanic, climatic conditions, that were modified by increasing effects of continentality inland and precipitation-shadowing by blocking mountain ranges. The blocking effects produced the unglaciated enclave of the central part of the Walls of Jerusalem National Park and the thin, relatively undynamic ice sheet of the eastern part of the Central Plateau.

ACKNOWLEDGEMENTS

This paper is Contribution No. 8 of the Geomorphology and Quaternary Science Research Unit of the Department of Geography, University of Newcastle. The authors thank Dr M.R. Banks and the referees for helpful comments on this paper.

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(accepted 14 May 1996)