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Glaciation and cave sediment aggradation around the margins of the Mt Field Plateau, Tasmania

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Landform evolution around the Mt Field Plateau has been strongly influenced by multiple stages of cold glacial climate. Only small glaciers were present during the late Last Glacial or Global Isotope Stage 2, but degraded moraines and the distribution of erratics indicate that ice cover was more extensive earlier when ice and meltwater invaded neighbouring karst areas and meltwater streams deposited gravel in caves. Weathering evidence suggests a significant glacial advance during Global Isotope Stage 4. Uranium–thorium dating of speleothems associated with gravels in proglacial caves suggests a major phase of gravel aggradation that post-dates Global Isotope Stage 5 and pre-dates Global Isotope Stage 2.

KEY WORDS: glaciation, karst, Quaternary, Tasmania, uranium-series dating, weathering rinds.

INTRODUCTION

Mt Field is a mountain plateau of ~120 km^2 in south-central Tasmania (Figure 1). Evidence for past glaciation of Mt Field has long been recognised (Lewis 1922a, b; Taylor 1922; Derbyshire *et al.* 1965; Peterson 1969) and was interpreted by Jennings and Banks (1958) as indicative of only a single glaciation. Mackintosh (1993) attributed the deposits on the plateau to the late Last Glacial Stage of Global Isotope Stage 2 (Shackleton 1987). The pre-Global Isotope Stage 2 glacial history has not previously been investigated.

The Mt Field Plateau is capped by a sill of Jurassic dolerite that was intruded into Carboniferous–Triassic mudstone, sandstone and glacigenic sedimentary rocks of the Parmeener Supergroup, which occur around the margins of the plateau and overlie Gordon Group limestone of Ordovician age outcropping in the Tyenna and Florentine valleys (Brown *et al.* 1989). Elaborate karst has developed in the limestone, which contains some of the deepest cave systems in Australia (Kiernan 1971; Goede 1973; Eberhard 1994, 1996). The relationship between glaciers and karst evolution in Tasmania has received only limited study (Jennings 1967; Kiernan 1982, 1983a, 1990a, b). Glaciation can have important implications for karsts, and relict sediments in caves can provide insight into the dating of glacial events in cave catchments (Ford *et al.* 1983; Lauritzen 1984a, b; Ford & Williams 1989; Williams 1996).

The primary objectives of this study are to: (i) explore the evidence for glaciation of the Mt Field area prior to the Global Isotope Stage 2; (ii) determine the spatial, hydrological and genetic relationships between past glaciers and caves; and (iii) assess the usefulness of cave sediment sequences for dating glacial events. The focus has been restricted to determining past glacier limits proximal to the karst areas.

Mackintosh (1993) concluded that when the plateau glaciers were most extensive ice cover totalled only ~21–23 km^2 and temperatures were ~10°C colder than present. On the basis of dolerite weathering-rind thicknesses he interpreted the oldest tills to date from *ca* 25 ka BP and suggested smaller advances occurred at *ca* 16–15 ka BP and 10 ka BP. Some older drifts have been recorded beyond the plateau moraines (Brown *et al.* 1989; Kiernan 1989a) although systematic mapping of the Quaternary geology has not previously occurred. Many slopes are mantled by diamictons similar to those that elsewhere in Tasmania have been interpreted as periglacial solifluctate (Davies 1967). Alluvial fans also occur in the karst areas (Goede 1973) and fluvial gravels are abundant on the floors of the Florentine Valley (<400 m) and Tyenna Valley (<200 m).

The largest underground drainage system is the Junee River, which rises on the western slopes of Mt Field West (1434 m) and flows southeastwards beneath a major surface drainage divide to emerge from the Junee Resurgence (~280 m) ~16 km from the most distant tributary stream-sinks. Lawrence Rivulet, which also originates on the plateau, sinks at ~430 m altitude and rises at ~380 m, close to the Florentine River. General descriptions, hydrogeological interpretations and maps of the karst have been provided by several authors (Kiernan 1971, 1989a; Goede 1973; Hume 1991; Eberhard 1994, 1996). Goede (1973) argued that periglacial slope instability blocked stream-sinks and caused reversions to surface drainage. Alluvial deposits in Beginners Luck Cave (Figure 1) are overlain by a limestone breccia interpreted by Goede and Harmon (1983) as the product of congelifraction. Goede and Harmon (1983) suggested an age of *ca* 20 ka BP for this breccia, based on ^{14}C dating of charcoal and aspartic acid dating of bone from within it. They concluded from weathering evidence that the preceding Beginners Luck Alluvial Phase dated from between 75 and 50 ka BP. Two uranium-series dates on

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speleothems from Niggly Cave, south of the plateau, indicate minimum ages of 15 ± 5 ka BP and >350 ka BP for underlying gravel deposits, while gravels in nearby Sesame Cave underlie speleothems dated at 325 ka BP and 350 ka BP (Eberhard 1997).

METHODS

Reconnaissance mapping of Quaternary landforms and sediments proximal to karst areas was undertaken (Figures 1–3). Patterns of ice flow were determined from the topography, distribution and orientation of moraines and the distribution of erratics. Morphostratigraphic relationships between moraines and glaciofluvial gravel deposits provided the primary means of differentiating ice advances of different ages (Figure 4). Extraglacial slope deposits and alluvial fans were also examined. To assess the relative age of surficial sediments the condition of glacially abraded rock surfaces, moraines and sediments was studied. The thickness of weathering rinds formed on subsurface dolerite clasts was recorded, together with the depths in the weathering profiles at which these occur, and the nature of the soil profiles, following the typology employed by Birkeland (1984) (Table 1). Weathering rinds were measured on samples of 20–50 clasts from 95 field locations (Figures 1–3) following the procedures of Kiernan (1990b). Fieldwork was focused in the Humboldt and Lawrence valleys, but results were checked against similar data from the Broad Valley and Mt Field East obtained in 1983–89 (Kiernan 1989a) and compared with similar studies from elsewhere in Tasmania to evaluate likely age relationships (Table 2). The glacial climatic stages during which ice advances occurred were interpreted from these data.

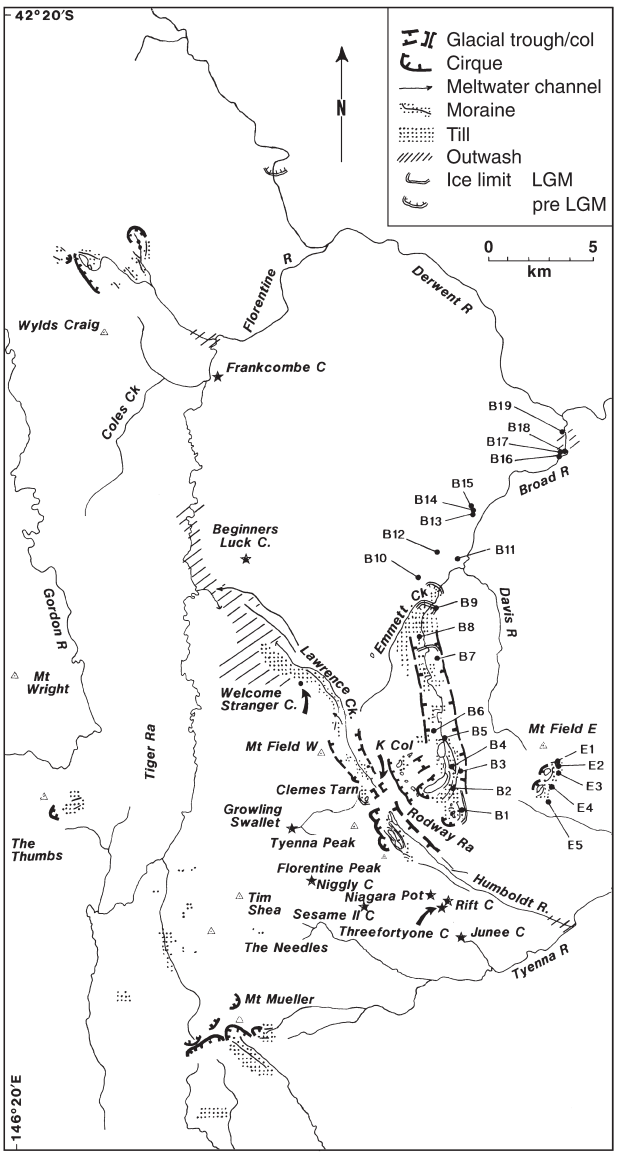

Figure 1 Glacial features in the the Mt Field area and localities mentioned in the text. The sites of weathering studies in the Broad Valley and around Mt Field East are indicated (see Table 1) together with glacier limits in those localities. See Figures 4 and 6 for details of the Humboldt and Lawrence valleys. R, river; Ck, creek; C, cave.

Mapping former glacier limits revealed the extent of glaciation in karst catchments. Glaciofluvial, alluvial and colluvial sediments (including reworked tills) were traced from the mountains to karstic stream-sinks. Cave-sediment sequences were examined to gauge the potential for speleothem dates to constrain the ages of associated clastic sediments. Speleothems were dated by the uranium-series disequilibrium method (Table 3), which can be applied to speleothems that contain sufficient U (>0.02 ppm) provided the system was initially free from non-authigenic ^{230}Th, as monitored with the ^{230}Th/^{232}Th index (Latham & Schwarcz 1992). Speleothems from two caves proved suitable for dating. The samples were digested in excess nitric acid, spiked with ^{228}Th/^{232}U and equilibrated by H_2O_2 oxidation and boiling for several hours. Uranium and Th were pre-concentrated by scavenger precipitation on ferric hydroxide. Iron was then removed by ether in 9 mol L^{-1} hydrochloric acid, and U and Th separated by ion exchange chromatography on 'Dowex 1x8' resin. The purified, carrier-free fractions of U and Th were then electroplated onto stainless steel disks and counted for alpha particle activity *in vacuo* on an Ortec Octéte unit with silicon surface barrier detectors for 2–4 days. Each spectrum was corrected for background and delay since chemical separation, and processed using tailored software (Lauritzen 1993). All analyses were performed at the Uranium-Series Geochronology Laboratory at the Department of Geology, University of Bergen.

OBSERVATIONS AND RESULTS

Evidence for glaciation of the karst beyond Global Isotope Stage 2 ice limits

Two prominent moraines beneath Tyenna Peak imply that a minor niche glacier previously formed at the head of the stream that enters Khazad-Dum Cave (Figure 2). Other moraines and glacial erratics imply invasion of the karst catchment upslope from Junee Cave by diffluent ice from the Humboldt Valley. Between 1120 m on Tyenna Peak and the floor of the Tyenna Valley, the Humboldt Ridge, which consists of sandstone and mudstone, forms a barrier to entry into the karst catchment of dolerite-bearing colluvial or fluvial sediments from the deep Humboldt Valley. *In situ* dolerite in the valley system upslope from Junee Cave is confined to ~0.2 km^2 at the summit of Tyenna Peak, but moraines composed almost entirely of dolerite can be traced southwards from Lake Belton over Humboldt Ridge into the Junee Cave valley. Humboldt Ridge is mantled by dolerite boulders to below 500 m. Due to this ice advance the eastern half of the Junee Cave valley is mantled by many metres thickness of dolerite-rich colluvium reworked from till, and dolerite-bearing alluvial fans can be traced as low as 285 m, where their distal margins are overlain by more recent gravels deposited by the Tyenna River.

Another ridge of sandstone and mudstone, informally termed 'Lawrence Ridge', forms the western margin of the Lawrence Valley from ~500–1160 m. This ridge, which rises up to 120 m above the valley floor, is mantled by erratic dolerite boulders up to 5 m long that locally form a pronounced moraine crest dissected by meltwater channels. Reworked fragments of weathered compact till, probably basal till, occur on the ridge crest. While the summit of Mt Field West provides an alternative source for some

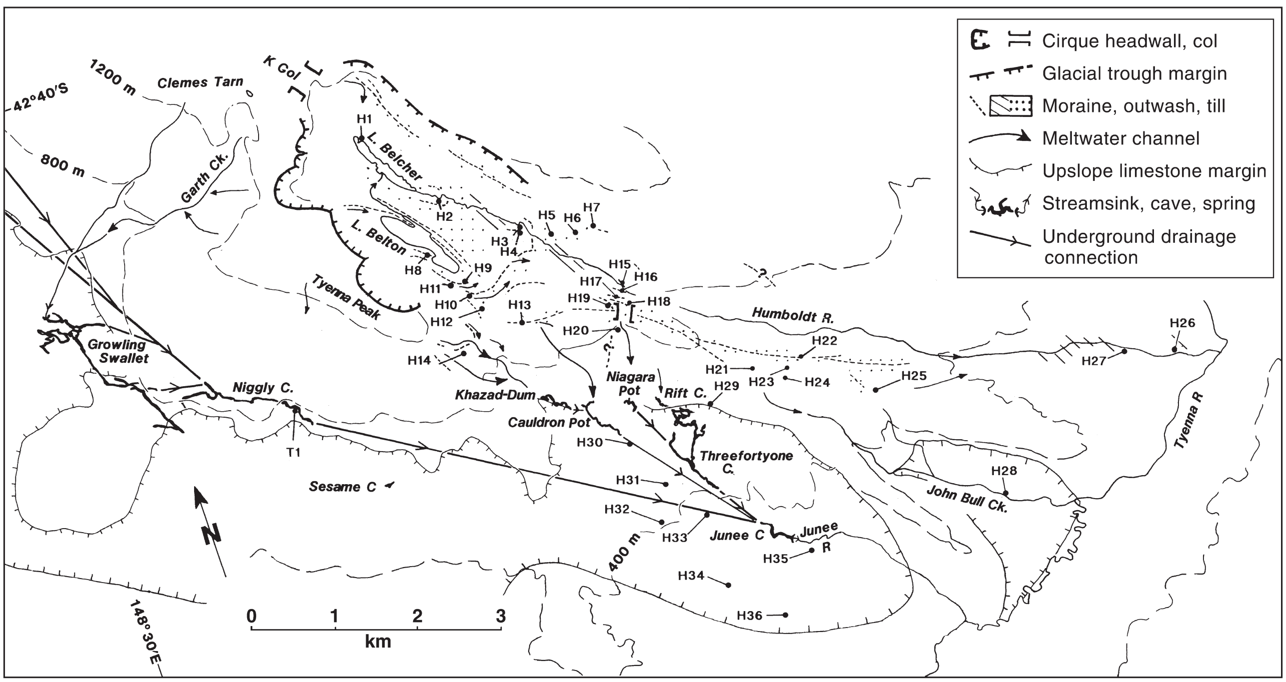

Figure 2 Glacial and karst landforms and sampling sites on the southern margin of the Mt Field plateau. R, river, Ck, creek, C, cave.

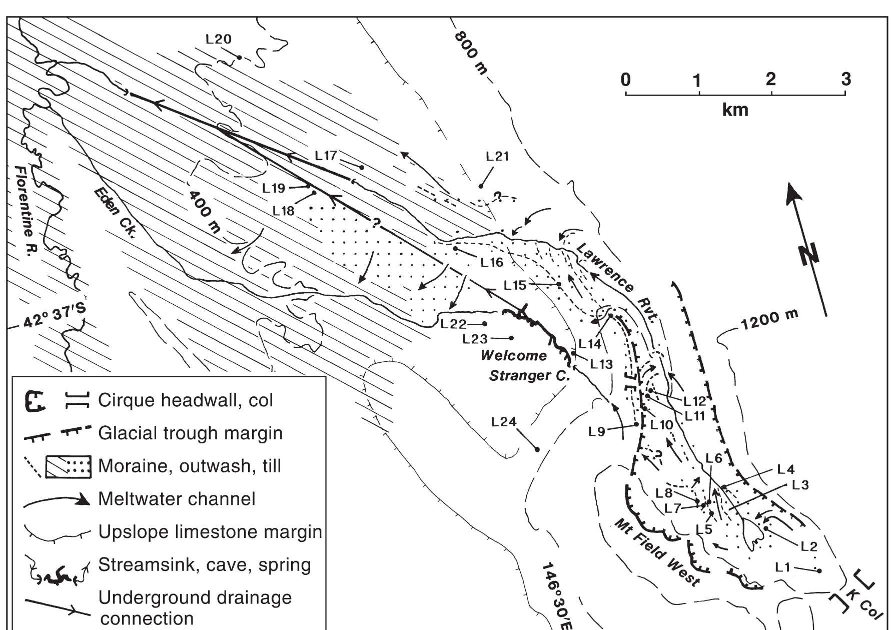

Figure 3 Glacial and karst landforms and sampling sites in the lower Lawrence Rivulet area on the western side of the Mt Field plateau. R, river, Ck, creek, C, cave.

dolerite-bearing deposits on the floor of the Florentine Valley, the thick dolerite mantle on the western slopes of Lawrence Ridge requires that it was overridden by ice that overflowed the Lawrence Valley. A lobe of heavily weathered diamicton west of the ridge is interpreted as till and glaciofluvial sediment, although no unequivocal moraine crests are discernible below ~600 m. Tributary streams on the eastern side of Lawrence Valley that are deflected from their downslope course at altitudes comparable to moraine crests on Lawrence Ridge suggest

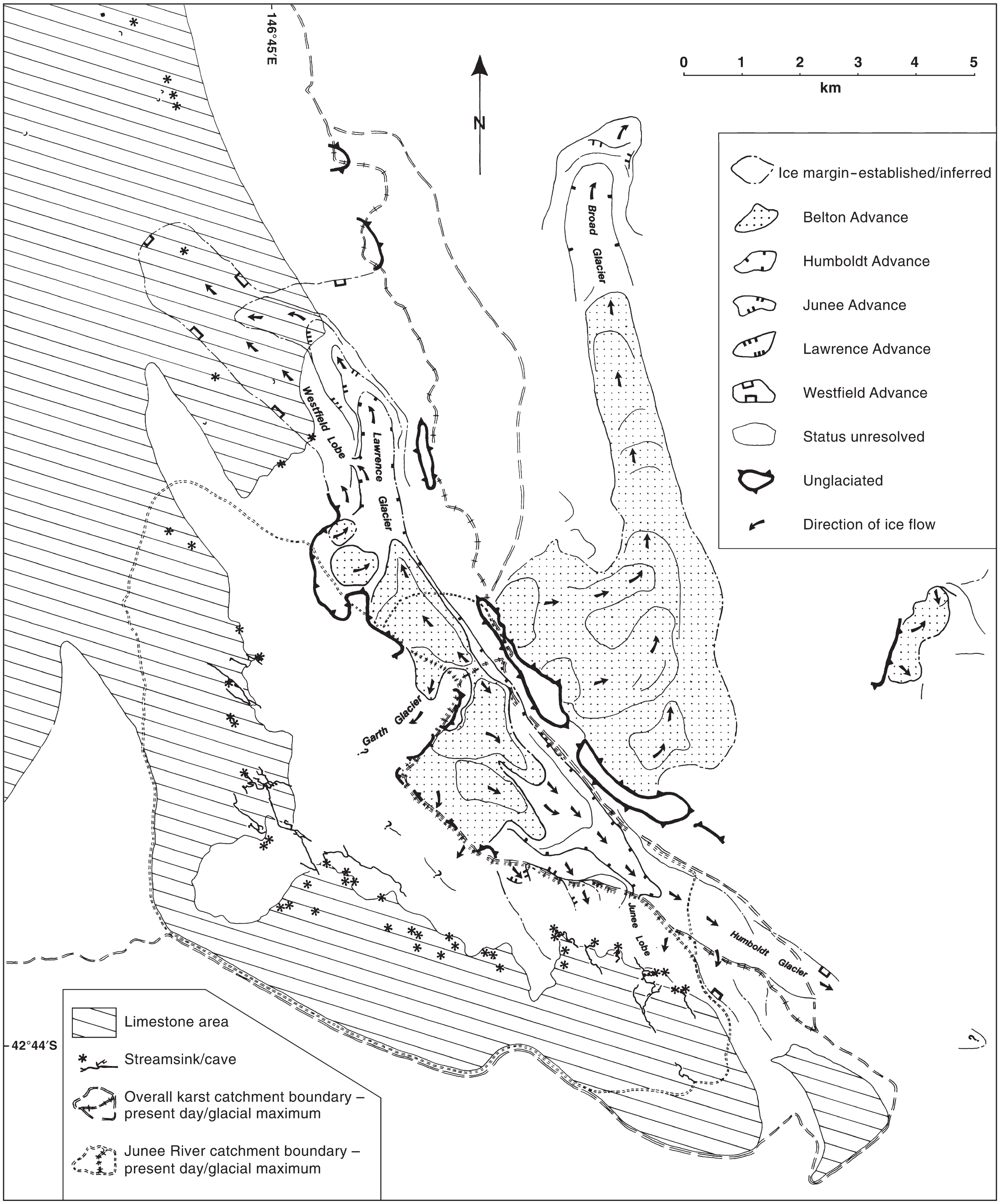

Figure 4 Glacier limits and extent of invasion of karst areas and catchments. The contemporary karst catchments and the extent of additional areas that drained to the karst at glacial maximum are indicated.

inheritance from meltwater channels around a former ice margin.

Differentiation of glacial stages

Morphological evidence permits identification of a series of distinct ice advances or pronounced stillstands during retreat (Figure 4). The weathering status of the sediments suggests that five stages of glaciation are represented (Table 1).

During the Belton Advance a small lateral moraine was constructed ~200 m south of Lake Belton in the Humboldt Valley. Glacially abraded bedrock surfaces inside this ice limit remain unweathered. The moraine is well-preserved and weathering rinds on subsurface clasts in the till are thin (mean 1.4 ± 0.7 mm, range 0.2–3.4 mm). They occur only in the upper 20–30 cm of a profile in which a thin A horizon directly overlies an oxidised C horizon, with no texturally enriched B horizon present (a profile of A Cox Cu-type). Weathering rinds on the associated outwash gravel are thinner than in the till (mean 0.6 ± 0.2 mm, range 0.2–1.2 mm). Except for limited landslide and rockfall deposits where ice-contact moraines were left unsupported or glacially oversteepened rock slopes have failed, colluvium is absent from inside the ice limits. Outside the Belton ice limits colluvium is more abundant and subsurface clasts exhibit weathering comparable in thickness to the Belton till. Comparably weathered till occurs in the upper Lawrence and Broad valleys, and at Mt Field East.

Tills of the Humboldt Advance occur beyond those of the Belton Advance and are bounded downstream by moraines on the floor of the Humboldt Valley ~4.8 km from

Table 1 Weathering status of tills around the Mt Field plateau.

Criteria	Belton	Humboldt	Junee	Lawrence	Westfield
1. Abraded rock surfaces	Fresh	Fresh	Moderately degraded	Heavily shattered	Heavily shattered
2. Moraine condition	Fresh	Fresh	Moderately degraded	Often buried	Rarely preserved
3. Glacial sediments					
Surface Jdl. rinds (mm)	1.4 ± 0.7	2.9 ± 0.9	1.9 ± 0.7		
sites	H8,9	H10,16	H19, 22–24		
Surface Jdl. pits (mm)	0.9 ± 0.4	2.9 ± 0.7	1.8 ± 0.7		
sites	H1	H16	H19, 22, 24		
Profile type	A Cox Cu	A Cox Cu/thin Bt	Bt	Bt	Bt
Depth with Jdl. rind (m)	0.1–0.2	0.4–0.6	>1.0	>2.0	>5.0
Subsurface Jdl. rinds					
mean (mm)	1.3 ± 0.7	3.7 ± 0.9	4.7 ± 1.6	13.5 ± 2.6	33.0 ± 16.1
min–max (mm)	0.2–4.8	0.1–7.6	0.6–13.2	6.8–20.7	11.5–50.7
range (mm)	4.6	7.5	12.6	13.9	39.2
sites	H1, 2, 8, 9; L1–8 B1–6; E2–4, 6	H3, 6,10,11,12,16, 18, 20; L10–12	H7, 13, 14, 19, 21–24 B9; E1	L14	H25; L16, 20
Subsurface hornfels rinds (mm)	1.8 ± 0.8	1.6 ± 0.5			
sites	H8, 9	H6, 11			
4. Glaciofluvial and fluvial sediments					
Subsurface Jdl. rinds (mm)	0.9 ± 0.2	1.7 ± 0.5	2.3 ± 0.6	16.6 ± 5.3	
sites	H4, 5, 27; T1; L19; B16;	H15, 17, 19, 28; B7, 8, 17	L23; B18	H26, 34, 35, 36; L17, 18, 22, 23	(B19?)
5. Colluvium					
Abundance inside ice limits	Negligible	Limited	Extensive	Abundant	Dominant
Subsurface Jdl. rinds (mm)	1.2 ± 0.5	2.0 ± 0.5	4.0 ± 1.3	15.8 ± 4.9	38.5 ± 16.1
sites	L11, 12, 24	H25; L21	H30; L9, 13, 15 B10, 11, 13, 15; E5	H31, 32, 36; B12, 14	H33; L16

Jdl., Jurassic dolerite.

Table 2 Uranium-series age determinations from Threefortyone Cave and Welcome Stranger Cave.

Lab. No. (ULB)	ID	U (ppm)	$^{234}U/^{238}U$	$^{230}Th/^{234}Th$	$^{230}Th/^{232}Th$	Age (ka)
Threefortyone Cave						
390	TF I	51.8	1.59	0.746	>1000	131.8 + 4.8/–4.6
400	TF II	0.59	2.46	0.145	>1000	16.7 ± 2.1
Welcome Stranger Cave						
393	WS III	1.9	1.68	0.322	250	41.0 ± 1.2
398	WS II	0.18	1.87	0.121	>1000	13.9 ±0.8
399	WS I	0.23	1.76	0.162	8.8	19.0 ±0.8
				(corrected age)		16.0 ± 1.0

Table 3 Suggested correlation of Tasmanian glaciations and thickness (mean ± standard deviation and range in mm) of subsurface dolerite weathering rinds in tills.

Area (Reference)					Brunhes Chron. <--		--> Matuyama Chron.	
Mt Field	Belton	Humboldt	Junee	Lawrence	Westfield	–		
(This paper)	1.3 ± 0.7	3.7 ± 0.9	4.7 ± 1.6	13.5 ± 2.6	33.0 ± 16.1	–		
	0.2–4.8	0.1–7.6	0.6–13.2	6.8–20.7	11.5–50.7	–		
Mt Anne	Judd	Timk	Weld	–	–	–	Eliza	
(Kiernan 1990a)	1.2 ± 0.1	4.1 ± 0.91	–	–	–	–	61.0 ± 29.7	
	0.2–1.4	3.6–6.9	–	–	–	–	?–170.0	
Nive catchment	Phase A	Phases B/C	Phases D/E	Phase F	–	–	–	
(Kiernan 1999)	1.3 ± 0.4	2.3 ± 0.7/2.5 ± 0.6	3.2 ± 1.3/5.1 ± 2.0	12.7 ± 3.8	–	–	–	
Derwent Valley	Cynthia Bay	Beehive	Powers Creek	Clarence	Wayatinah	–	–	
(Kiernan 1985, 1991)	1.5 ± 0.5	3.0 ± 1.6	5.1 ± 2.2	12.7 ± 3.8	–	–	–	
	0.1–4.6	0.5–6.9	1.0–15.8	2.8–37.0	–	–	–	
Gordon/Guelph	Cynthia Bay	Beehive	–	–	–	–	–	
(Kiernan 1985)	1.2 ± 0.9/1.5 ± 0.7	3.1 ± 1.4/5.1 ± 1.7	–	–	–	–	–	
	0.6–1.4/0.3–4.3	1.6–6.1/1.0–8.8	–	–	–	–	–	
Franklin Valley	Dixon	Beehive	Taffys Creek	Wombat Glen	Stonehaven	–	–	
(Kiernan 1985, 1989b)	1.6 ± 0.5	4.1 ± 1.2	7.4 ± 3.8	12.0 ± 5.9	41.2 ± 11.3	–	–	
	0.4–3.8	1.9–9.2	1.9–15.1	1.7–30.5	14.8–89.0	–	–	
King Valley	Dante	Chamouni	Bull –David	Cableway	Moore	–	Thureau	
(Fitzsimons *et al.* 1993)	1.5 ± 0.2	(no till data)	–	7.6 ± 1.8	14.3–17.3	–	14.5–75.5	
Central West Coast Range	Dante	Comstock III[a]	Linda III		–	–	Linda 1	
(Kiernan 1980, 1983b)	<1[b]	4.8 ± 1.6	6.8 ± 12.9		–	–	74.6 ± 25.9	
	–	2.0–10.0	?–21.0		–	–	–	
Henty/Pieman	Margaret	Julia Creek	Boco	Bobadil	–	–	Bulgobac	Que
(Augustinus 1999)	–	–	7.1 ± 1.8	13.1 ± 3.7	–	–	–	–
	–	–	–	–	–	–	130.0–240.0	–
Forth Valley	Oakleigh	Borradaile	–	***[c]	–	Patons	Lorinna	
(Kiernan & Hannan 1991)	1.5 ± 0.5	3.1 ± 0.8–4.6 ± 2.0	–	9.8 ± 3.6	–	21.5 ± 5.9	–	
	?–4.5	?–5.0	–	?–20.1	–	?–34.0	?–200	
Mersey	Rowallan	Arm	–	–	–	–	Croesus	
(Hannan & Colhoun 1987)	0.3 ± 0.1	1.1 ± 0.5	–	–	–	–	50.8 ± 29.5	
	0.1–0.6	0.5–2.2	–	–	–	–	6.0–190.0	

[a] Revised to eliminate two samples now considered older.
[b] Samples from <30 cm depth.
***[c] Tributary glacier originally interpreted as being of Lorinna age.

K Col. Again, abraded rock surfaces inside the Humboldt ice limits remain little weathered and moraines are well-preserved with little infill of meltwater channels by till reworked from moraine slopes. Subsurface dolerite weathering rinds in the till average twice the thickness of those of the Belton Advance till (mean 2.7 ± 0.8 mm, range 3.2–6.1 mm) and occur to ~60 cm depth in soil profiles that exhibit Bt-type profiles with a thin texturally enriched B horizon, although some profiles have not developed significantly beyond A Cox Cu-type. The associated outwash deposits exhibit weathering rinds that are again significantly thinner than those in the till (mean 1.4 ± 0.5 mm, range 0.5–3.4 mm). Slope deposits inside the Humboldt stage glacier margins are of limited extent and are not weathered significantly more than those inside the Belton limits, but colluvium outside the Humboldt limit is more weathered (mean subsurface dolerite rinds 2.0 ± 0.5 mm, range 1.3–3.2 mm). Till with comparable weathering characteristics has been recognised in the Lawrence Valley.

The maximum ice limits during the Junee Advance are defined by moraines south and west of Humboldt Ridge upslope from Junee Cave. Glaciated dolerite surfaces between the Humboldt and Junee limits are shattered and pitted, solution pans up to 40 cm wide and 10 cm deep occur on sandstone pavements, and the moraines are significantly degraded. Subsurface dolerite weathering rinds in the till are well-developed (mean 3.5 ± 1.4 mm, range 0.8–8.2 mm) and occur to depths of over 1.0 m in profiles invariably of Bt-type. Comparably weathered till comprises part of a major moraine complex in the Broad Valley near its confluence with Emmett Creek and forms a degraded moraine east of Lake Nichols in the Mt Field East area. Colluvium is thick, widespread and well-weathered both inside and outside the Junee Advance ice limits.

The Lawrence Advance is defined from a lateral moraine crest at 830 m on Lawrence Ridge that is highly degraded and exhibits a profile of Bt-type at least 2 m thick in which subsurface dolerite weathering rinds are very thick (mean 13.5 ± 2.6 mm, range 6.8–20.7 mm). Weathering rinds in glaciofluvial sediments downstream from the interpreted glacier terminus are also thick (mean 8.8 ± 3.4 mm, range 2.2–24.2 mm). Lithified till in the lower Broad Valley (Thrush & Sharples 1998) may be of Lawrence age or older.

The Westfield Advance is defined from deeply weathered till at the toe of Lawrence Ridge and glacigenic sediment on the floor of the Florentine Valley. Little moraine topography survives. The till is very heavily weathered, with thick weathering rinds on subsurface dolerite clasts (mean 67.1 ± 25.9 mm, range 11.5–95.0 mm) in horizons of Bt-type at least 5 m deep. Weathering rinds in the associated outwash are also very thick (mean 19.3 ± 5.5 mm, range 9.2–34.0 mm). Comparably weathered till occurs beneath weathered slope deposits at 520 m on the crest of Humboldt Ridge.

Karst landforms and sediments

Fieldwork was focused on two localities where stream-sinks received meltwater discharged from glacier margins. One major concentration of large inflow caves occurs in the valley system upslope from Junee Cave that was invaded by a diffluent lobe of the Humboldt Glacier. The caves most recently affected by glacial meltwater are Niagara Pot, Rift Cave and Threefortyone Cave, which lie respectively 1.3, 1.6 and 2 km southeast of the Humboldt Ridge saddle to which ice extended during the Humboldt stage (Figure 1). Abundant coarse gravels on the surface and in some cave passages are beyond the competence of the present streams and indicate earlier torrential discharge from the ice margins. Present-day drainage into both Niagara Pot and Rift Cave flows underground to Threefortyone Cave (Eberhard 1994).

A complex of semihorizontal stream passages, both active and fossil, occurs ~150 m below the Threefortyone Cave entrance. Speleothem samples from a fossil stream passage that has largely been filled by gravels were submitted for uranium–thorium dating (Figure 5). Sample ULB 390 is from a broken stalactite buried at ~20 cm depth in these gravels and gave an age of 131.8 +4.8/–4.6 ka BP (Table 2). Sample ULB 400 is from the base of a stalagmite 90 mm in diameter that formed on the surface of the gravels and dated to 16.7 ± 2.1 ka BP. These dates indicate the gravels were deposited during the Last Glacial Stage.

A meltwater channel, cut through Lawrence Ridge during the Humboldt Advance, discharges into the Welcome Stranger Cave (Figure 1) in which at least two phases of gravel aggradation have occurred. Three speleothem samples were obtained for uranium–thorium dating (Figure 6, Table 2). Sample ULB 399 from the base of a stalagmite that formed on gravels in an upper level passage gave an age of 16.0 ± 1.0 ka BP. Sample ULB 398 from a broken stalagmite found sandwiched between the gravel and an overlying rockfall deposit gave a result of 13.9 ± 0.8 ka BP. Sample ULB 393 was obtained from the base of a column near the upstream limit of the cave. Gravels embedded in the base of this and other nearby speleothems indicate that they grew on a former gravel surface that has

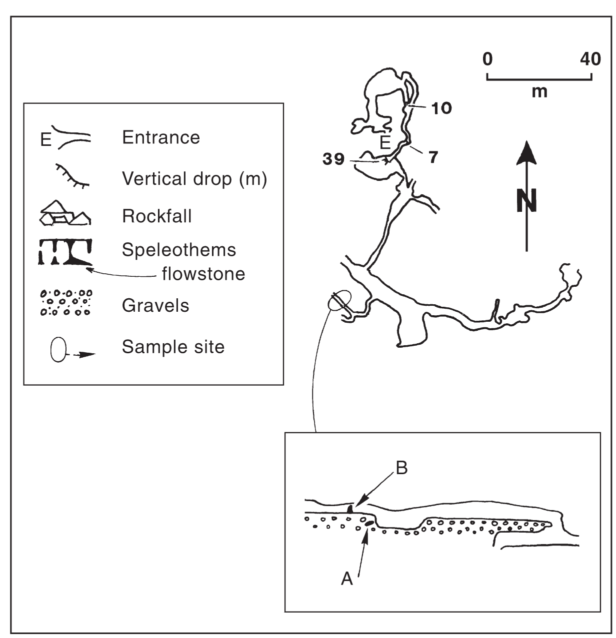

Figure 5 Map of part of Threefortyone Cave and stratigraphic context of dated samples. A, 131.7 +4.8/–4.6 ka BP (ULB 390); B, 16.7 ± 2.1 ka BP (ULB 400). See Figure 2 for location.

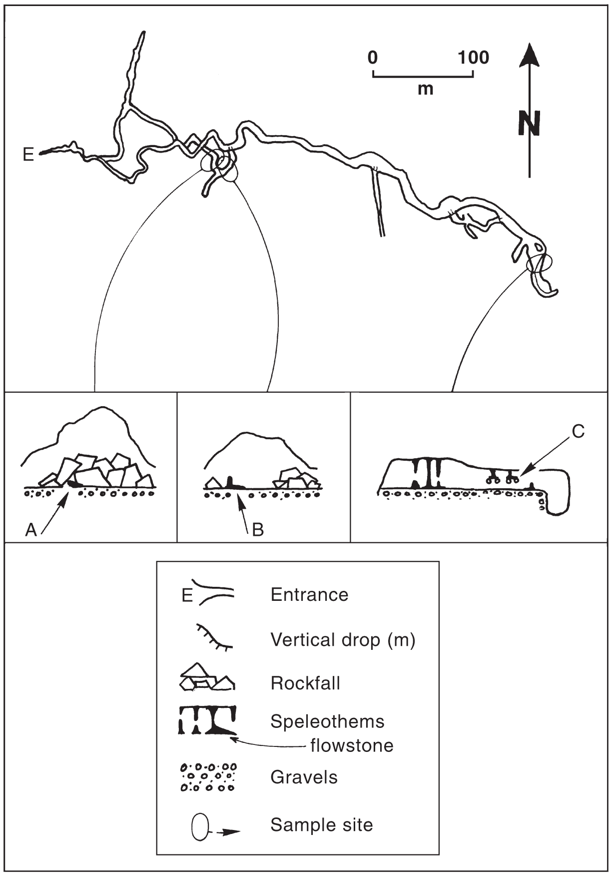

Figure 6 Map of Welcome Stranger Cave and stratigraphic context of dated samples. A, 13.9 ± 0.7 ka BP (ULB 398); B, 16.0 ± 2.1 ka BP (ULB 400); C, 40.9 ± 1.2 ka BP (ULB 393). See Figure 3 for location.

since been eroded away, leaving the columns suspended from the roof. This sample gave a significantly older age of 41.0 ± 1.2 ka BP.

DISCUSSION

Extent and pattern of glaciation

BELTON ADVANCE

During the Belton Advance the ice at Lake Belton was partly tributary to a small glacier ~2 km long at the head of Humboldt Valley (Figure 4). At least one substage is identifiable in the Humboldt Valley. Advances in several other parts of the Mt Field plateau can be correlated with the Belton Advance on morphostratigraphic and weathering evidence. A well-preserved moraine formed from lightly weathered drift occurs at the head of the Garth Valley west of K Col, but the maximum ice limits there have not been determined. Similarly fresh moraines and minimally weathered drift indicate that the Lawrence Glacier extended 2.8 km downvalley. Two small cirque glaciers formed on the adjacent slopes of Mt Field West. A much larger glacier arose from well-defined cirques on the eastern slopes of the Rodway Range and flowed down the Broad Valley, while small cirque glaciers constructed moraines inside earlier glacial deposits below Mt Field East.

The lithology of the tills is consistent with the glacier flow patterns suggested by erosional and depositional landforms. Belton Advance tills are characteristically dominated by dolerite. In the Humboldt Valley, sandstone and mudstone clasts are abundant only near the valley head (38% at H1; 13% at H8) and are diluted rapidly downstream (6–8% at H9, 10). This confirms erosional evidence (Peterson 1969) for the Humboldt Glacier having been nourished more by ice from west of lakes Belton and Belcher than by ice that accumulated at the head of the Humboldt Valley beneath K Col, which is eroded into sandstone and mudstone. The most recent glaciofluvial sediments contain a significant proportion of sandstone and mudstone clasts (13% at H4) due to late-stage meltwater discharge after the main Humboldt glacier had retreated towards its headwall at K Col. In the Lawrence Valley, where sandstone and mudstone only crop out close to Lake Hayes, sandstone and mudstone clasts comprise ~7% of till around the lake (L3, 4) but are absent from the higher moraines.

HUMBOLDT STAGE

The Humboldt Glacier was nourished partly by ice that descended from the Lake Belton shelf and constructed a lateral moraine 0.5 km south of that lake (Figure 4). Humboldt Ridge was not overrun, but ice penetrated a short distance through the saddle at ~815 m. A high proportion of dolerite in the till (87–96%, sites H3, 6, 15–18) again reflects erosion predominantly at high altitudes. A local increase in Parmeener Supergroup clasts just over the Humboldt Ridge saddle (23%, H20) probably represents reworking of older moraines. A glacier terminus ~5.5 km from the head of the Lawrence Valley is interpreted from morphological evidence. Till ~50 m above the valley floor ~3.5 km downstream from K Col and a marginal meltwater channel 400–500 m beyond Lake Hayes suggest ice ~140 m thick; together they imply a downglacier ice-surface gradient of ~86 m/km. A saddle at ~880 m towards the top of Lawrence Ridge, the source of the present Welcome Stranger Cave stream, was cut through an older dolerite moraine by west-flowing meltwater that scoured to bedrock. In the Broad Valley, slightly weathered glaciofluvial sediments are associated with an ice limit ~2.2 km beyond the ice limit correlated with the Belton Advance.

JUNEE ADVANCE

Moraines south and west of Humboldt Ridge indicate that during the Junee Advance a small niche glacier descended to ~1030 m in the headwaters of the Khazad-Dum stream. Ice flowed at least 1 km south from Lake Belton and deposited till at ~980 m altitude on Humboldt Ridge and into the catchment of Cauldron Pot (Figure 4). The Humboldt Glacier east of Lake Belton was ~170 m thick. Its downvalley extent has not been determined, but diffluent ice extended through the Humboldt Ridge saddle at 815 m. At least three substages can be recognised on morphostratigraphic evidence. A significant proportion of sandstone and mudstone clasts (8–21%, sites H21–24) is consistent with effective glacial erosion much further downvalley than during the Belton and Humboldt advances.

LAWRENCE ADVANCE

The Lawrence Glacier was ~120 m thick ~5.5 km downstream from K Col. A total length of ~6.2 km is inferred from a lateroterminal moraine on the left bank and a deflected stream on the right bank interpreted as a marginal meltwater channel. The moraine crest beyond the Humboldt Ridge saddle may be of equivalent age and implies that the Humboldt Glacier was over 100 m thick at this point, amply sufficient to flood into the karst.

WESTFIELD ADVANCE

The Lawrence Glacier extended at least 8.5 km and perhaps 10 km downstream from K Col and overran Lawrence Ridge. Reworked till above the meltwater-scoured col at the head of Lawrence Ridge implies the Lawrence Glacier was at least 120 m thick 3.5 km from the valley head. It is difficult to envisage this volume of ice in the Lawrence Valley without invoking tributary ice from the plateau to the east. This implies glaciation well beyond the ice limits accepted by most previous researchers, with the exception only of Lewis (1945). Surficial sediments further east on the Mt Field Plateau that are generally regarded as periglacial deposits (Davies 1967) are likely to include reworked tills. Water-rounded sandstone and mudstone predominate among clasts <10 cm in the low-altitude drifts on the floor of the Florentine Valley, but form <10% of clasts >30 cm (L18, 22, 23). The till of this Westfield Advance recognised in the Humboldt Valley lies 7.8 km downstream from K Col and 60 m above the valley floor. Dolerite colluvium in sinkholes above Junee Cave is much more heavily weathered than the Lawrence till, but is less weathered than the Westfield till.

Implications of past glaciation for karst evolution

The moraines and glacial sediments along Humboldt Ridge imply invasion of the eastern end of the karst by glaciers, meltwater and glacigenic sediment. Hence, cold-climate geomorphic processes on the karst here have not been confined to periglacial effects as has previously been suggested (Goede 1973; Eberhard 1994, 1996) and the evolution of the Quaternary landforms, sediments and soils have been at least partly preconditioned by glaciation. Similarly, there has been significant invasion by ice of the karst along and south of Lawrence Creek and a wide area has been buried by glaciofluvial sediment.

Reconstruction of the equilibrium-line altitudes for the glaciers is impeded by incomplete knowledge of glacier margin positions. Equilibrium-line altitude estimates and palaeotemperature reconstructions based upon them are particularly complicated where alpine topography interferes with the mesoscale climate (Soons 1979; Meirding 1982). Meirding (1982) has demonstrated considerable differences in reconstructed palaeotemperatures depending on the technique employed. Given these uncertainties only broad estimates of palaeotemperature during the stages identified in the Humboldt and Lawrence valleys seem warranted. Altitude ratios of the valley glaciers and the maximum altitudes of lateral moraines suggest equilibrium-line altitudes of ~1100 ± 100 m for the Belton Advance and ~900 ± 100 m for the Westfield Advance. The modern mean summer temperature 0°C isotherm lies at 2301 m (Mackintosh 1993). Assuming a lapse rate of 0.6°C/100 m, summer temperatures were ~7.2°C lower than now during the Belton Advance and ~9.1°C lower during the Westfield Advance. Mackintosh (1993) suggested summer temperatures were ~10.8°C lower during Global Isotope Stage 2, a conclusion he based partly on temperatures he assumed necessary to allow formation or reactivation of blockfields on the plateau. It is difficult to date the blockfields directly using surface-weathering rinds, but weathering rinds formed on subsurface dolerite clasts suggest that matrix-supported periglacial colluvial deposits contiguous with the blockfields were probably mobile during earlier cold stages (Table 1).

When the glaciers were most extensive the catchment of the underground Junee River was greatly increased because ice overran present-day fluvial divides. In particular, the catchment of the small valley system upslope from Junee Cave was nearly doubled through invasion by a diffluent lobe of the Humboldt Glacier. Ice and meltwater from at least the eastern half of the upper Humboldt Valley was deflected towards the present cave systems. Not only was the catchment extended, but it incorporated more elevated terrain where precipitation totals were higher. Only ice that accumulated on the eastern side of the Humboldt Valley and basal meltwater in its thalweg can be assumed to have been retained in the Humboldt Valley. Runoff into the cave catchments would not have been confined to supraglacial and marginal meltwaters because karst-like drainage systems typically exist in temperate glaciers, the hydraulic gradients in which are governed by ice-surface contours rather than bedrock topography. Hence, meltwater under pressure in englacial and subglacial tunnels can flow uphill over bedrock ridges (Ford *et al.* 1983; Kiernan 1984). When the glacier was most extensive such karst-like glacier cave systems formed in ice may have been superimposed on the limestone karst. During lesser advances similar englacial and subglacial conduit systems would have permitted meltwater from the Humboldt to be lifted into the karst catchments. The concentration of large accessible inflow caves in this part of the karst belt contrasts with areas further west where the invasion of karst catchments by glaciers and meltwater did not occur. This concentration may reflect glacigenic stimulation of cave elaboration.

On the southern flanks of Tyenna Peak west of the Khazad-Dum stream are low bouldery ridges that were interpreted by Kiernan (1989a figure 1) as small moraines formed by a minor cirque glacier upslope from Sesame Cave. However, the very extensive glaciation of the Humboldt Valley implied by the present study necessitates very substantial ice input from the small plateau between Tyenna Peak and Florentine Peak. Shallow saddles southwards from this plateau provide obvious pathways for overspill of ice and/or meltwater towards the boulder ridges and that part of the karst that contains Sesame and Niggly Caves.

Given the maximum extent of the Humboldt and Lawrence glaciers it is likely that their present fluvial divide at K Col was inundated beneath a considerable thickness of ice at glacial maximum, sufficient to have nourished a glacier able to extend considerably further down the Garth Valley than the lightly weathered Belton Advance moraine at ~1080 m that marks the effective limit of investigation to date. Another concentration of major caves occurs in the lower Garth Valley, and while this is predictable solely on the basis of the large fluvial catchment, glacial influences are again likely. The relationship between cave entrances and former glacier margins in this valley is unresolved, but considerable volumes of meltwater would have reached Growling Swallet, the largest tributary stream-sink of the underground Junee River, and probably some other caves.

The catchment of the karst further north in the Florentine Valley was also more extensive than today during maximum glaciation due to ice flowing into the Lawrence Valley from some areas now drained by the Broad River system. The close proximity between the edge of the Westfield Lobe of the Lawrence Glacier and stream-sinks along the northern extremity of the underground Junee River catchment in the Florentine Valley raises the possibility that past glaciation may have played a role in defining the position of this drainage divide.

The low altitude of the limestone area and the limited descent of glaciers from the plateau means that even when the ice was most extensive the karst lay below the equilibrium-line altitudes where glacial erosion is most effective. Hence, glacial erosion of the karst around Mt Field is likely to have been far less pronounced than in some other Tasmanian karst areas, such as Dante Rivulet and Mt Anne (Kiernan 1982, 1990a). Depositional effects were more pronounced around Mt Field, and the large volumes of glacial and glaciofluvial sediment may have impeded karstification in some areas. The likely partial blocking of subsurface conduits in the lower Lawrence Rivulet Valley bears comparison with the burial of preglacial caves in the Lake Sydney karst beneath glacigenic sediment and their progressive postglacial re-excavation (Kiernan 1989c).

Dating

Around Mt Field weathering data provides the best means of estimating age differences between successive ice advances. The differences in soil profile development and the thicknesses of weathering rinds formed on subsurface dolerite clasts are consistent with the geographical pattern of ice advances. Soil profiles of A Cox Cu-type in tills of the Belton Advance contrast with profiles in the older tills that exhibit progressively thicker textural B horizons. Clasts in the Belton Advance tills are virtually unweathered below 10–20 cm depth. Weathering rinds in the Humboldt Advance tills are only slightly thicker, but occur to greater depths (20–40 cm). Erosion or reworking of the upper part of the weathering profile can result in weathering rinds still being relatively thin near the surface even on ancient moraines. However, these older tills may be distinguished using the depth in the profile at which rinds occur, and a trend towards thicker rather than thinner rinds at depth. The advanced weathering of the slope deposits that overlie the Westfield Advance drift further implies that it is of considerable age. Differences in matrix texture, likely precipitation and type of vegetation cover on the tills of the different stages are relatively minimal, suggesting that environmental differences do not account for differences in the observed thickness of subsurface weathering rinds.

Comparison of the weathering status of the Mt Field tills with glacial sequences elsewhere in Tasmania allows tentative correlations to be attempted (Table 3). Weathering of the Belton Advance deposits compares with tills attributed to Global Isotope Stage 2. Weathering rind thickness in the Humboldt Advance tills closely matches that of some tills attributed to Global Isotope Stage 6. Weathering of the Junee and Lawrence tills suggests that they are of at least Middle Pleistocene age, while the deeply weathered Westfield till is probably of Early Pleistocene or Pliocene age (Kiernan 1983b, 1989b).

Several lines of evidence suggest that the Humboldt Phase need not be as old as Global Isotope Stage 6. Post-depositional change to the morphology of the Humboldt moraines is very limited and relatively little colluvium occurs inside the Humboldt ice limits. Soil profiles have developed no further than shallow Bt-type and some profiles remain essentially of A Cox Cu-type. These characteristics suggest that the age difference between the Belton and Humboldt phases is not great. If weathering rind thickness is assumed to be a linear function of time and the mean thickness of ~1.3 mm in the Belton till is assumed to have taken ~20 000 years to develop, then weathering rinds in the Humboldt till that average 3.7 mm thick could have formed since an advance during Global Isotope Stage 4. The same conclusion can be drawn from maximum rind-thickness values. However, numerous field and laboratory studies have shown that the rate of chemical weathering is not linear, but slows over time as the buildup of residues impedes the evacuation of solutes. Colman and Pierce (1981) argued that the rate of rind thickening is a square root function of time, which, if true, would favour a Global Isotope Stage 6 age for the Humboldt Phase. Chinn (1981) suggested modal rind-thickness values may allow greater resolution than mean values. Modal rind-thickness for the Belton Advance tills in the Humboldt Valley decreases progressively from 2.4 to 3.2 mm in the outer moraines (H2, 9) to 1.2–1.6 mm nearer the valley heads (H1, 8). The suspicion of a younger age for the Humboldt Advance is reinforced by the modal rind-thickness values of 2.1–6.0 mm from the Humboldt Valley (H3, 6, 10–12, 16, 20) and 3.0–6.2 mm from the Lawrence Valley (L7, 10–12, 13, 19). If modal values increase by a factor of two between the youngest and oldest Belton Phase drifts despite the probability that they differ in age by no more than ~10 000–15 000 years, then modal values in the Humboldt tills that are only twice as thick as those in the Belton Advance would allow the Humboldt Advance to date to Global Isotope Stage 4.

Glacial meltwater entered the catchment of Threefortyone Cave during the Humboldt Stage. The date of 131.8 +4.8/–4.6 ka BP (ULB 390) obtained from a speleothem fragment in the gravel in this cave, combined with the date of 16.7 ± 2.1 ka BP (ULB 400) obtained from a stalagmite that formed on the surface of the same gravel unit, indicate that these gravels post-date Global Isotope Stage 6 and pre-date the latter part of Global Isotope Stage 2 (Figure 5, Table 2). For conservation reasons only a small volume of speleothem could be removed for dating, but substantial carbonate precipitation over the gravel sequences elsewhere in Threefortyone Cave hints that their surfaces have been stable for a long time.

The catchment of Welcome Stranger Cave was also invaded by glacial meltwater during the Humboldt Phase. The date of 40.96 ± 1.2 ka BP (ULB 393) obtained from the basal stalagmite component of a column that formed on gravels in this cave implies that they are no younger than earliest Global Isotope Stage 3 (Figure 6, Table 2). Dates of 16.0 ± 1.0 ka BP (ULB 399) and 13.91 ± 0.77 ka BP (ULB 398) obtained from stalagmites that formed on the surface of gravels further downstream in Welcome Stranger Cave indicate only that these gravels pre-date late Global Isotope Stage 2. Hiatuses in speleothem development may be explicable by environmental conditions and hence they may allow only minimum age estimates for underlying sediments (Ford & Williams 1989).

Hence, the weathering evidence combined with the speleothem dates suggest that the Humboldt Advance pre-dates Global Isotope Stage 2, but is younger than Global Isotope Stage 6 and thus may have occurred during Global Isotope Stage 4. Advances of this age have not yet been clearly demonstrated from Tasmania. Fitzsimons *et al.* (1993) suggested that the Chamouni Formation outwash gravels in the King Valley in western Tasmania were deposited during Global Isotope Stage 4. They cited as evidence: (i) a radiocarbon date of 48 700 +2900/–2100 ka BP (SUA 2599) from within overlying silts interpreted as glacial lake sediments; (ii) the presence inside the Chamouni ice limits of the Dante Formation that is believed to date from Global Isotope Stage 2 (Kiernan 1983a); and (iii) a suggestion that the weathering status of the Chamouni Formation is comparable to the Dante till. The radiocarbon date and morphostratigraphic relationship to the Dante Formation provide only a minimum age for the Chamouni Formation. Fitzsimons *et al.* (1993) indicated that dolerite weathering rinds in the Chamouni Formation outwash gravels have a mean thickness of 1.5 ± 0.7 mm, which compares with 1.5 ± 0.2 in the Dante till. However, weathering rinds in outwash gravels are typically 11–20% thinner than in their associated tills (Colman & Pierce 1981; Kiernan 1990b), while weathering rind figures provided in Fitzsimons *et al.* (1993) indicate that rinds from outwash gravels in the older Cableway Formation a few kilometres further south in the King Valley are ~40% thinner than in the Cableway till. Hence, it is difficult to interpret age relationships from weathering data when dissimilar sediment types are being compared. Weathering rinds from outwash gravels of probable Chamouni age that occur directly beneath the Dante Formation at its type section are 3.3 ± 1.1 mm thick, even though weathering of these gravels is likely to have been retarded or halted since the Dante Formation was deposited over them at *ca* 20 ka BP (Kiernan 1980). Hence, the Global Isotope Stage 4 age suggested for the Chamouni Formation by Fitzsimons *et al.* (1993) is equivocal.

However, the evidence from Mt Field increases the likelihood that glaciation occurred in Tasmania during Global Isotope Stage 4 and raises questions regarding some other similarly weathered Tasmanian glacial sequences that have been assumed to date from Global Isotope Stage 6. Just as the Belton and Humboldt drifts at Mt Field cannot be discriminated statistically, mean rind-thickness values in tills elsewhere in Tasmania that have been attributed to Global Isotope Stage 2 and Global Isotope Stage 6 also overlap at one standard deviation (Table 3). For example, tills in the Franklin Valley interpreted by Kiernan (1989b) as dating from the Beehive Glaciation (Global Isotope Stage 6) exhibit mean rind-thickness values of 2.9–6.7 mm (maxima 3.7–9.2 mm) that would allow more than a single glaciation to be involved. Tills deposited during the Dixon Glaciation, which has been assumed to date from Global Isotope Stage 2, exhibit subsurface dolerite weathering rinds that are only slightly thinner (mean 1.5–1.6 mm, maxima 2.3–3.8 mm) than those in the youngest Beehive tills. Hence, while it is probable that the more lightly weathered till was deposited during Global Isotope Stage 2, a greater age is conceivable for some distal parts of the large Dixon moraine complex and for similar sequences elsewhere.

While it is reasonable to interpret the dated gravels in Threefortyone and Welcome Stranger Caves as glaciofluvial outwash based on known trajectories and implied timing of meltwater influxes, not all gravel fills in the caves of this area are glacigenic. Weathering evidence suggests many colluvial mantles and alluvial fans around Mt Field are of comparable age to the tills (Table 1). If glaciers formed in the Mt Field area during Global Isotope Stage 4 then periglacial slope instability and alluviation is also likely to have occurred, as suggested by Goede and Harmon (1983). Their suggested Global Isotope Stage 4 age of the Beginners Luck Alluvial Phase rests partly on the basis of weathering rinds that average 0.9 ± 0.2 mm in thickness from 1.5 to 1.8 m depth in gravels outside Beginners Luck Cave (Goede & Harmon 1983). Again this dating is not secure, because although these rinds were obtained from well below the typical depth of any Global Isotope Stage 2 weathering profile, if they are from gravels that lack a moisture-retaining matrix their weathering status would allow a considerably greater age than Global Isotope Stage 4. The gravels in Niggly Cave and Sesame Cave shown by Eberhard (1997) to pre-date *ca* 350 ka BP suggest recurrent phases of gravel filling and re-excavation of cave sediments that extend back much further than Global Isotope Stage 4. They also confirm that at least some caves in formerly ice-marginal positions are likely to be sufficiently old to contain a legacy of early glacial advances. The young dates obtained from speleothems on the gravels in the upper levels of Welcome Stranger Cave and in Threefortyone Cave are suggestive that these gravels formed during Global Isotope Stage 2 although they do not demand it. These dates are comparable to the result of 15.5 ka BP obtained from a speleothem from Niggly Cave (Eberhard 1997) that formed on gravels now shown to be potentially glaciofluvial.

Regional comparisons

The long and complex history of glaciation now evident from Mt Field is consistent with the glacial record from elsewhere in southern temperate latitudes.

Clapperton (1993) grouped 15 moraines in the Lago Argentino and Lago Buenos Aires in South America into five complex zones to which he assigned probable ages of 2.3–1.2 Ma, *ca* 750 ka, <750 – >120 ka and <120 – >14 ka BP.

The oldest drifts are heavily weathered like Westfield drift, but the available dating is inadequate to allow correlation of the Mt Field and South American deposits. Later South American advances peaked at >33.5, 29.6, 26.94, 23.06, 21.0 and 14.89–13.9 ka BP (Lowell *et al.* 1995). This is broadly consistent with multiple advances having occurred during the last Glacial Stage at Mt Field, but weathering and radiocarbon evidence provide only equivocal evidence for glaciation in South America during Global Isotope Stage 4 (Porter 1981).

The earliest known drift in New Zealand is that of the Ross Glaciation, which is overlain by Pliocene Old Man Group conglomerates that may also be glacigenic. Elsewhere in New Zealand the Moutere Gravel and overlying Porika drift predate any other terrestrial evidence for glaciation by *ca* 1 Ma. The very deep weathering of the Porika drift again invites broad comparison with the Westfield drift. Four glaciations have occurred in New Zealand during the last 350 thousand years, the Nemonan, Waimaungan, Waimean and Otiran (Suggate 1990), but again precise correlation of the New Zealand and Tasmanian drifts is not possible. Williams (1996) has demonstrated advances in Fiordland at *ca* 92, 66–49, 45–41, 39–21, 18–17 and 15–14 ka BP, compounding previous evidence for multiple advances in New Zealand during the Last Glacial Stage and consistent with the possibility that the Humboldt drift dates from Global Isotope Stage 4.

CONCLUSIONS

(1) Only small glaciers formed on the Mt Field Plateau during Global Isotope Stage 2, but valley glaciers descended to lower altitudes during at least four earlier glaciations when ice and meltwater was discharged into karstic catchments.

(2) Some caves have formed in ice-marginal positions and have functioned as underground meltwater channels. Till weathering and uranium–thorium dating of speleothems suggest that glacial invasion of karst catchments may have occurred as recently as Global Isotope Stage 4.

(3) Tasmanian weathering evidence is inadequate to allow discrimination between multiple short-lived glaciations such as those suggested by Williams (1996) in New Zealand. Coupled with the equivocal discrimination between Global Isotope Stage 6 and Global Isotope Stage 4 tills, some uncertainty remains as to the maximum extent of ice during Global Isotope Stage 2 in Tasmania.

(4) Further integrated research into the alpine and karst geomorphology of this area may facilitate improved understanding of the evolution of karst under cold-climate conditions, while the dating of cave deposits by uranium-series methods offers the potential for better constrained chronologies of Tasmanian glacial and periglacial events.

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REFERENCES

Augustinus P. C. 1999. Dating the Late Cenozoic glacial sequence, Pieman River basin, western Tasmania, Australia. *Quaternary Science Reviews* **18,** 1335–1350.

Birkeland P. W. 1984. *Soils and Geomorphology*. Oxford University Press, New York.

Brown A. V., McClenaghan M. P., Turner N. J., Baillie P. W., McClenaghan J. & Calver C. R. 1989. Geological Atlas 1:50 000 Series, Sheet 73 (8112N) Huntley. *Geological Survey of Tasmania Explanatory Report.*

Chinn T. J. H. 1981. Use of rock weathering-rind thickness for Holocene absolute age-dating in New Zealand. *Arctic & Alpine Research* **13,** 33–45.

Clapperton C. 1993. *Quaternary Geology and Geomorphology of South America*. Elsevier, Amsterdam.

Colman S. M. & Pierce K. L. 1981. Weathering rinds on andesitic and basaltic stones as a Quaternary age indicator, western United States. *US Geological Survey Professional Paper* **1210**.

Davies J. L. 1967. Tasmanian landforms and Quaternary climates. *In*: Jennings J. N. & Mabbutt J. A. eds. *Landform Studies from Australia and New Guinea*, pp. 1–25. ANU Press, Canberra.

Derbyshire E., Banks M. R., Davies J. L. & Jennings J. N. 1965. *A Glacial Map of Tasmania.* Royal Society of Tasmania Special Publication **2**.

Eberhard R. 1994. *Inventory and Management of the Junee River Karst System, Tasmania.* Forestry Tasmania, Hobart.

Eberhard R. 1996. *Inventory and Management of the Florentine Valley Karst, Tasmania.* Forestry Tasmania, Hobart.

Eberhard R. 1997. Age constraints for clastic sediments from two caves in the Junee–Florentine karst, Tasmania. *Papers and Proceedings of the Royal Society of Tasmania* **131,** 67–72.

Fitzsimons S. J., Colhoun E. A., Van Der Geer G. & Pollington M. 1993. The Quaternary geology and glaciation of the King Valley. *Geological Survey of Tasmania Bulletin* **68**.

Ford D. C., Smart P. L. & Ewers R. O. 1983. The physiography and speleogenesis of Castleguard Cave, Columbia Icefields, Alberta, Canada. *Arctic & Alpine Research* **15,** 437–450.

Ford D. C. & Williams P. 1989. *Karst Geomorphology and Hydrology*. Unwin Hyman, London.

Goede A. 1973. Hydrological observations at the Junee Resurgence and a brief regional description of the Junee area, Tasmania. *Helictite* **11,** 3–24.

Goede A. & Harmon R. 1983. Radiometric dating of Tasmanian speleothems—evidence of cave evolution and climatic change. *Journal of the Geological Society of Australia* **30,** 89–100.

Hannan D. G. & Colhoun E. A. 1987. The glacial stratigraphy of the upper Mersey Valley. *Australian Geographical Studies* **25,** 36–46.

Hume N. G. 1991. Hydrological and speleological investigation of the Junee–Florentine karst, south west Tasmania. BA(Hons) thesis, University of Tasmania, Hobart (unpubl.).

Jennings J. N. 1967. Some karst areas of Australia. *In*: Jennings J. N. & Mabbutt J. A. eds. *Landform Studies from Australia and New Guinea*, pp. 269–279. ANU Press, Canberra.

Jennings J. N. & Banks M. R. 1958. The Pleistocene glacial history of Tasmania. *Journal of Glaciology* **3** (24)**,** 298–303.

Kiernan K. 1971. Caves and karst of Junee–Florentine, Tasmania. *Australian Speleological Federation Newsletter* **53,** 4–10.

Kiernan K. 1980. Pleistocene glaciation of the Central West Coast Range, Tasmania. BA(Hons) thesis, University of Tasmania, Hobart (unpubl.).

Kiernan K. 1982. Glaciation and karst in Tasmania: review and speculations. *Helictite* **20,** 11–14.

Kiernan K. 1983a. Relationship of cave fills to glaciation in the Nelson River Valley, central western Tasmania. *Australian Geographer* **15,** 367–375.

Kiernan K. 1983b. Weathering evidence for an additional glacial stage in Tasmania. *Australian Geographical Studies* **21,** 197–200.

KIERNAN K. 1984. Morphological karst on some New Zealand glaciers. *Journal of the Sydney Speleological Society* **24,** 257–273.

KIERNAN K. 1985. Late Cainozoic glaciation and mountain geomorphology in the Central Highlands of Tasmania. PhD thesis, University of Tasmania, Hobart (unpubl.).

KIERNAN K. 1989a. Soil and water degradation in carbonate rock terranes. *Australian Journal of Soil and Water Conservation* **3,** 26–33.

KIERNAN K. 1989b. Multiple glaciation of the upper Franklin Valley, Western Tasmania Wilderness World Heritage Area. *Australian Geographical Studies* **27,** 208–235.

KIERNAN K. 1989c. Drainage evolution in a Tasmanian glaciokarst. *Helictite* **27,** 3–12.

KIERNAN K. 1990a. The alpine geomorphology of the Mt Anne massif, southwestern Tasmania. *Australian Geographer* **21,** 113–125.

KIERNAN K. 1990b. Weathering as an indicator of the age of Quaternary glacial deposits in Tasmania. *Australian Geographer* **21,** 1–17.

KIERNAN K. 1991. Glacial history of the upper Derwent Valley, Tasmania. *New Zealand Journal of Geology and Geophysics* **34,** 157–166.

KIERNAN K. 1999. The southern margin of the late Cainozoic ice cap on the Central Plateau of Tasmania. *Australian Geographer* **30,** 5–33.

KIERNAN K. & HANNAN D. 1991. Glaciation of the upper Forth River catchment, Tasmania. *Australian Geographical Studies* **29,** 155–173.

LATHAM A. G. & SCHWARCZ H. P. 1992. Carbonate and sulphate precipitates. *In*: Ivanovich M. & Harmon R. S. eds. *Uranium-series Disequilibrium: Applications to Earth Marine and Environmental Sciences*, pp. 423–459. Clarendon Press, Oxford.

LAURITZEN S.-E. 1984a. A symposium: arctic and alpine karst. *Norsk Geografisk Tidsskrift* **38,** 139–143.

LAURITZEN S.-E. 1984b. Evidence of subglacial karstification in Glomdal, Svartisen. *Norsk Geografisk Tidsskrift* **38,** 169–170.

LAURITZEN S.-E. 1993. *'Age4U2U'. Program for reading ADCAM energy spectra, integration peak correction and calculation of ^{230}Thorium/^{234}Uranium ages*. Computer Program Turbo Pascal Code, 5000 lines Department of Geology, University of Bergen.

LEWIS A. N. 1922a. A preliminary sketch of the glacial remains reserved in the national park of Tasmania. *Papers and Proceedings of the Royal Society of Tasmania (1922)***,** 16–36.

LEWIS A. N. 1922b. A supplementary note to a preliminary sketch of the glacial remains preserved in the national park of Tasmania. *Papers and Proceedings of the Royal Society of Tasmania (1922)***,** 177–179.

LEWIS A. N. 1945. Pleistocene glaciation in Tasmania. *Papers and Proceedings of the Royal Society of Tasmania (1944)***,** 41–56.

LOWELL T. V., HEUSSER C. J., ANDERSEN B. G. *ET AL.* 1995. Interhemispheric correlation of Late Pleistocene glacial events. *Science* **269,** 1541–1549.

MACKINTOSH A. 1993. Late Quaternary glaciation of Mt Field Plateau, southern Tasmania. BSc(Hons) thesis, University of Newcastle, Newcastle (unpubl.).

MEIRDING T. C. 1982. Late Pleistocene glacial equilibrium-line altitudes in the Colorado Front Range: a comparison of methods. *Quaternary Research* **18,** 289–310.

PETERSON J. A. 1969. The cirques of southeastern Australia: studies in morphology and distribution; history and significance. PhD thesis, Monash University, Melbourne (unpubl.).

PORTER S. C. 1981. Pleistocene glaciation in the southern Lake District of Chile. *Quaternary Research* **16,** 262–292.

SHACKLETON N. J. 1987. Oxygen isotopes, ice volume and sea level. *Quaternary Science Reviews* **6,** 183–190.

SOONS J. M. 1979. Late Quaternary environments in the central South Island of New Zealand. *New Zealand Geographer* **35,** 16–23.

SUGGATE R. P. 1990. Late Pliocene and Quaternary glaciations of New Zealand. *Quaternary Science Reviews* **9,** 175–197.

TAYLOR G. 1922. Some geographical notes on a model of the national park at Mt Field, Tasmania. *Papers and Proceedings of the Royal Society of Tasmania (1921)***,** 16–36.

THRUSH M. & SHARPLES C. 1998. A Reconnaissance of the glacial geomorphology of the Broad River State Forests of South Central Tasmania. Consultants report to Forestry Tasmania, Hobart (unpubl.).

WILLIAMS P. W. 1996. A 230 ka record of glacial and interglacial events from Aurora Cave, Fiordland, New Zealand. *New Zealand Journal of Geology and Geophysics* **39,** 225–241.

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