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PERIGLACIAL PHENOMENA IN TASMANIA

Abstract

Talus, solifluction debris, rock glaciers, nivation cirques, protalus moraines and bedded screes of late Quaternary age and some presently-active periglacial forms are described from several localities in the mountains of central Tasmania, and their climatic implications considered. The Quaternary forms are ascribed to the last pleniglacial and the succeeding deglacial hemicycle. The pleniglacial climate of the highest unglaciarized mountain summits appears to have been cold and relatively dry with severe wind-drifting of snow but without general permafrost. A depression of the mean annual temperature of at least 6.5°C relative to the present is suggested.

While deposits and landforms due to the action of cryonival processes during the Pleistocene period are widespread in the mountains of central and eastern Tasmania, and common in the alpine tracts of the highlands of the south-eastern mainland of Australia, relatively few descriptive accounts of field data are at present available. In the maritime west of Tasmania, steep slopes, slow weathering of the dominantly siliceous rocks and the relatively small size of the weathering products combine with the high rainfall to render both production and preservation of substantial periglacial landforms unlikely (Davies, 1967). It is for this reason that the bulk of our knowledge of the nature and distribution of Pleistocene periglacial activity is based on observations made in the centre and east of the island, where the Jurassic dolerite is the dominant bedrock. Congeliturbates are widespread on the dolerite uplands. Their precise extent is unknown, although a map showing areas of predominantly periglacial processes during the Pleistocene has been published by Davies (1965). Field occurrences are described by Jennings (1956), Davies (1958), Galloway (1965), Caine (1968) and Derbyshire (1968), and their broad climatic framework considered at some length in a recent valuable survey by Davies (1967).

In the course of field work and aerial photographic interpretation undertaken during the compilation of the Glacial Map of Tasmania (Derbyshire, Banks, Davies and Jennings, 1965) and the Glacial Map of North-west-Central Tasmania (Derbyshire, 1968a), some attention was given to the examination of landforms resulting from cryonival processes. This was done in an attempt to gauge the general nature of the cold period climates which affected the extra-glacial areas during glaciation and the glaciated slopes during and since deglaciation. Attention was directed towards nivation cirques, and to deposits on slopes steeper than 5 degrees, especially rock glaciers, protalus moraines and rhythmically-stratified screes.

TALUS AND SOLIFLUCTION DEBRIS

Periglacial deposits, locally very thick (e.g. over 54 m (180 feet) in an old channel of the Mersey River at Parangana [1] (Paterson, 1966), mantle most hillsides above 450 m (1,500 feet above sea level) in central Tasmania. Where, due to steeper slopes, the deposits are transitional in nature from solifluction to talus or snowmelt deposits, the accumulations may be very bulky and the modification of glacial slopes considerable. In the Lake St. Clair district, for example, massive dolerite boulder deposits set into a weathered matrix occur on slopes as gentle as 10 degrees (Derbyshire, 1963). The size of the included boulders is such that solifluction must have played a part in the transportation of some of this material. Encroaching upon it is found a very coarse talus made up of large, little-weathered dolerite columns with no observable matrix, resting at angles of between 15 and 45 degrees. Typically, this coarser talus occurs as a series of fans or aprons below dolerite cliffs, the small source areas of the talus usually being quite obvious (Pl. 1).

Despite the great accumulated volume of this steeper talus, therefore, much of it appears to have grown as a succession of localized though often large landslides, the processes of slump, subsidence and rockfall all playing a part. Enlargement of joints, sapping and cambering, followed by rock--slide and large rock-falls, has taken place on the dolerite summits of mountains particularly in response to mechanical weathering. Frost-wedging along the dolerite joints has assisted the sliding of enormous joint-bound masses of dolerite, especially where adjacent mountain slopes have been oversteepend by glacial erosion. Several stages in this process have been preserved on the summit of Mt. Olympus (west-central Tasmania), evidence including deep rifts in the dolerite summit (up to 24 m: 80 feet deep), potential landslips bounded by these deep rifts, landslipped rock masses which have preserved their initial form and, finally, large landslips which have spread themselves down the mountain sides (Fig. 2 and Pl. 2).

[1] Localities mentioned in the text are shown on Figures **1 and 5.**

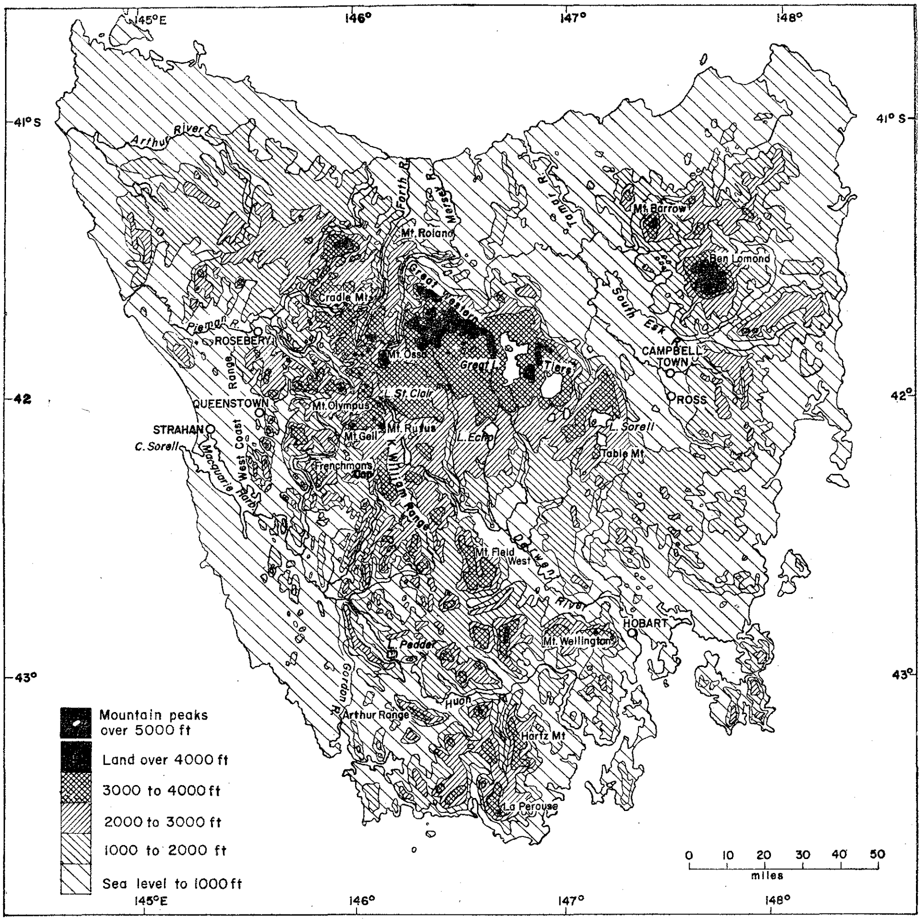


Fig. 1. Tasmania: orography

ROCK GLACIERS

The steep, coarse dolerite talus on the south-western slopes below the summits of Mt. Olympus and Mt. Gell (Figs. 2, 3, and Pls. 2, 3) has been fashioned into the steep fronts and transverse arcuate ridges of typical rock glaciers. Two types of rock glaciers have been recognized in the literature: the first type marks the site of decaying ice-tongues „qui sont enterrées sous les debris provenent des versants" (Tricart, 1963, p. 266), and the second is a mixture of rock-fall debris and interstitial ice derived from precipitation and snow avalanching. The two types are closely related and the second may, in favourable circumstances, develop from the first (Lliboutry, 1953;

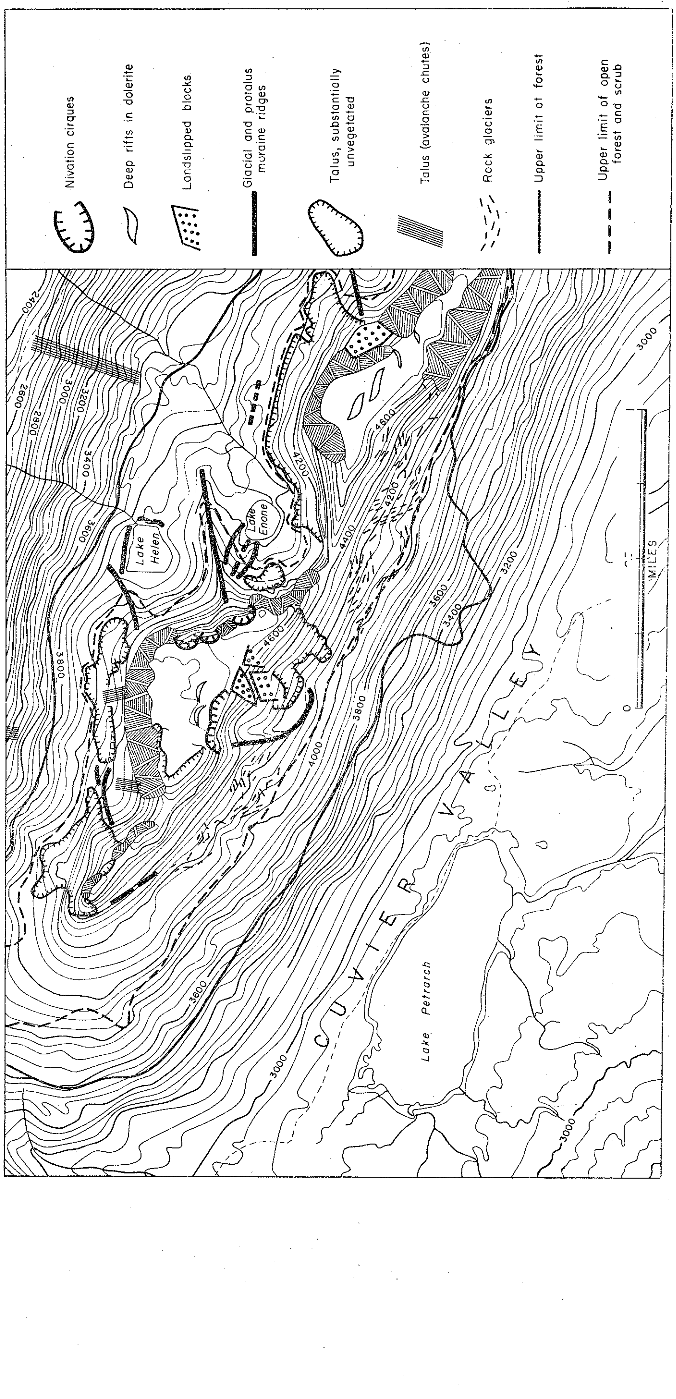


Fig. 2. Periglacial landforms on Mount Olympus, west-central Tasmania

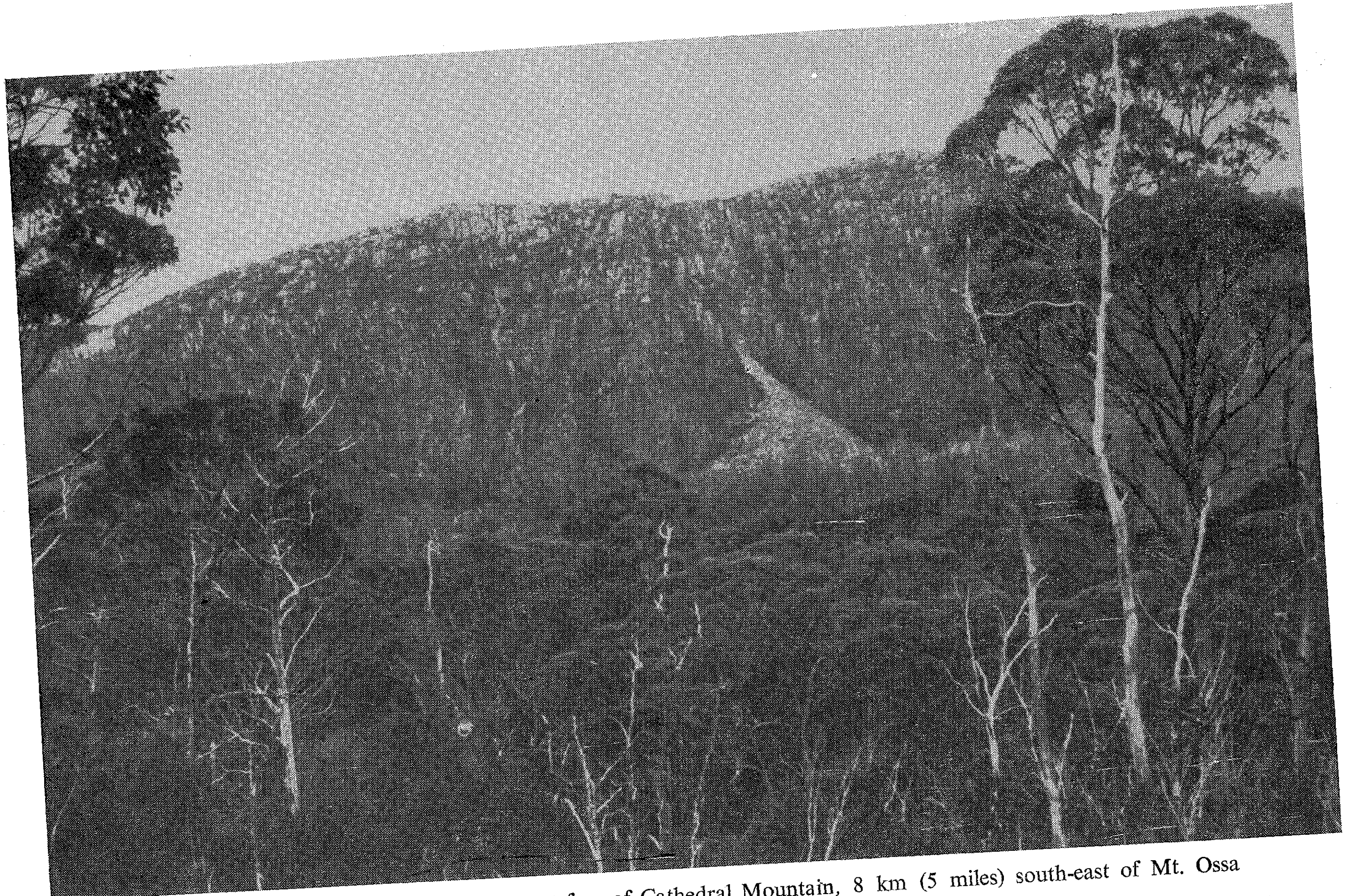
Wahrhaftig and Cox, 1959). Apart from Goldthwait's (1913) discussion and the composite eluviation-freezing processes proposed by Lliboutry (1961), there appears to be rather general agreement that interstitial ice is responsible for the mobility of rock glaciers. Notwithstanding the evidence assembled by Thompson (1962), the present distribution of active rock glaciers in North America and Europe suggests that general permafrost is not a prerequisite for their active development (*cf.* Kesseli, 1941).

The morphological and site characteristics of rock glaciers have been clearly distinguished by Outcalt and Benedict (1965), the first type characterizing cirque floors and the second occurring typically beneath free faces and steep valley walls where they are readily supplied with talus. The Tasmanian rock glaciers appear to be of the second type. They are backed by straight or only weakly indented free faces and they pass laterally into normal talus slopes. Moreover, the distribution of rock glaciers and glacial cirques appears to be matually exclusive, there being only one weakly developed cirque, probably of nivational origin, on the western face of Mt. Olympus, and none at all on the west face of Mt. Gell.

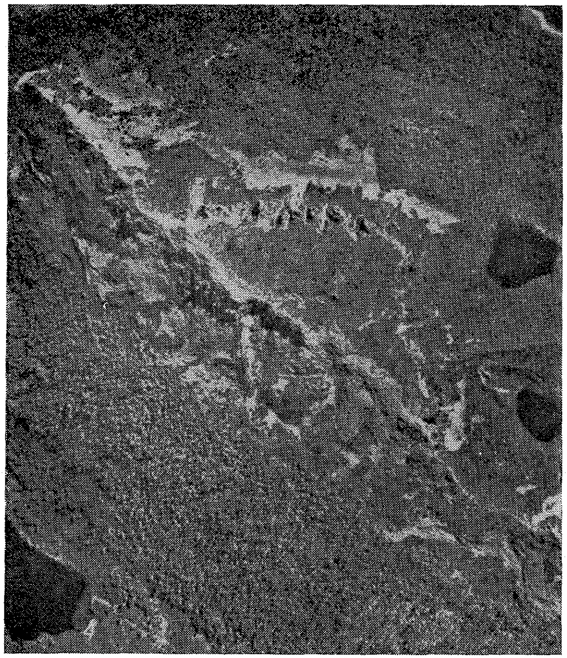
The rock glaciers of Mt. Olympus (Fig. 2 and Pl. 2) lie on the south-west facing slopes windward of the plateau-like summit and between approximately 1370 m (4,500 feet) and 1,188 m (3,900 feet) above sea level. They extend for a distance of some 3.1 km (2 miles) and cover an estimated area of 0.84 km^2 (0.33 square miles). Average over-all surface gradients range between a minimum of 14 degrees on the lower, now vegetated parts, to 26 degrees on the unvegetated parts.

The rock glaciers consist of large blocks of dolerite, for the most part joint-bound prisms broken into a range of sizes varying from $3 \times 2 \times 2$ m ($16 \times 6 \times 6$ feet) down to small boulder size. Only near the foot of the rock glaciers is there any evidence of accumulation of fine debris. Here, and at various locations and altitudes on their surfaces, patches of scrubby alpine vegetation (mainly dwarf southern deciduous beech: *Nothofagus gunnii*) is well established. Toward the south-eastern and north-western ends of the rock glacier tract, this vegetation covers a notable proportion of the surface. This fact, together with the observed stability of the component boulders on steep slopes suggests that these rock glaciers are all relic forms and are not active under the present climatic regime. A qualitative impression was gained of greater weathering of the blocks on the lower rock glacier slopes where locally thick red-brown clay silts are present, in contrast to the apparently fresh joint faces of the dolerite columns higher up the rock glaciers. Moreover, as may best be seen on the south-west facing slopes to the south of Lake Enone, the talus falls into three parts: steep, partly vegetated rock glaciers (A on Fig. 2); more gently-sloping (c. 16–20 degrees), almost wholly vegetated rock glaciers (B); and weathered talus now forested and extending down to 975 m (3,200 feet) (C). Whether the latter talus has the characteristic form of rock glaciers is not known due to the forest cover: if it does, then the total area covered by rock glaciers on Mt. Olympus is enlarged by 0.41 km^2 to 1.25 km^2 (0.49 square miles). Neither is it clear whether the rock glaciers constitute a single system, the vegetated parts merely representing the lower tongues which have become vegetated due to accumulation of fines by wash during and following their formation. On the other hand, the two-fold (or possibly three-fold) character of the talus on the south-west face of Mt. Olympus suggests the possibility of more than one period of formation of the rock glaciers. Given that the rock glaciers and cirques are mutually exclusive in areal terms, the major formative phase of rock glacier growth seems likely to have coincided with the last pleniglacial. More limited activity may have occurred throughout the last deglacial hemicycle, but re-activation of the rock glaciers in the middle post-Glacial cool phase seems unlikely in view of the generally rather fine grain textures of the associated colluvia found on low slopes to the east (Derbyshire, 1968 a, p. 37). The relatively low altitude of the weathered talus (C) on Mt. Olympus is below the level attained by the Cuvier valley glacier during the last glaciation (Derbyshire, 1963), although there is no reason to assume that maximum glacierization coincided with the climatic pleniglacial (Derbyshire, 1968 unpub. mss.). Accordingly, even this lowermost member of the Mt. Olympus talus may be no older than the first half of the last deglacial hemicycle. Closely related periglacial deposits in the form of blockstreams and blockfields have been described from several Tasmanian localities. On the Ben Lomond plateau in north-east Tasmania, the blockfields of Talus Valley are, at least in part, late-Pleistocene glacial deposits which have been re-worked by frost action and solifluction rather than rock glacier creep (Caine, 1968), possibly during the last deglacial hemicycle. On that part of the Central Plateau formerly covered by an ice-cap, blockstreams are more widespread on the northern margins than on the south (where the glacial forms are fresher), again suggesting formation during the retreat from the last glacial maximum (Jennings, personal communication).

The Mt. Gell rock glaciers (Fig. 3 and Pl. 3) lie windward of the summit and adjacent to discrete and valley-head cirques which radiate eastward and southward from the summit ridge of the mountain. The rock glaciers run for about 1.6 km (1 mile) between the summit ridge and 1,173 m (3,850 feet) above sea level. Analogy with Mt. Olympus suggests that weathered talus (? forested rock glacier) may extend as low as 1,097 m (3,600 feet) above sea level, however. The Mt. Gell rock glaciers are smaller than those of Mt. Olympus, covering only 0.64 km^2 (0,25 square miles), although the inclusion



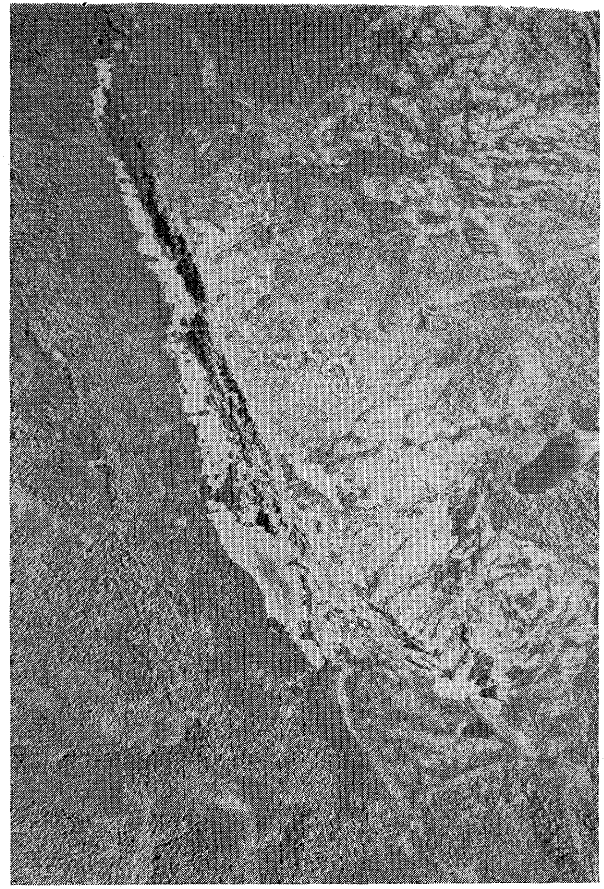
Pl. 1. Talus cone of dolerite debris on west face of Cathedral Mountain, 8 km (5 miles) south-east of Mt. Ossa



Dept. of Lands, Tasmania, Photo

Pl. 2. Vertical aerial photograph of the western half of the summit of Mt. Olympus

For interpretation, see Figure 2



Dept. of Lands, Tasmania, Photo

Pl. 3. Vertical aerial photograph of the summit of Mt. Gell

For interpretation, see Figure 3

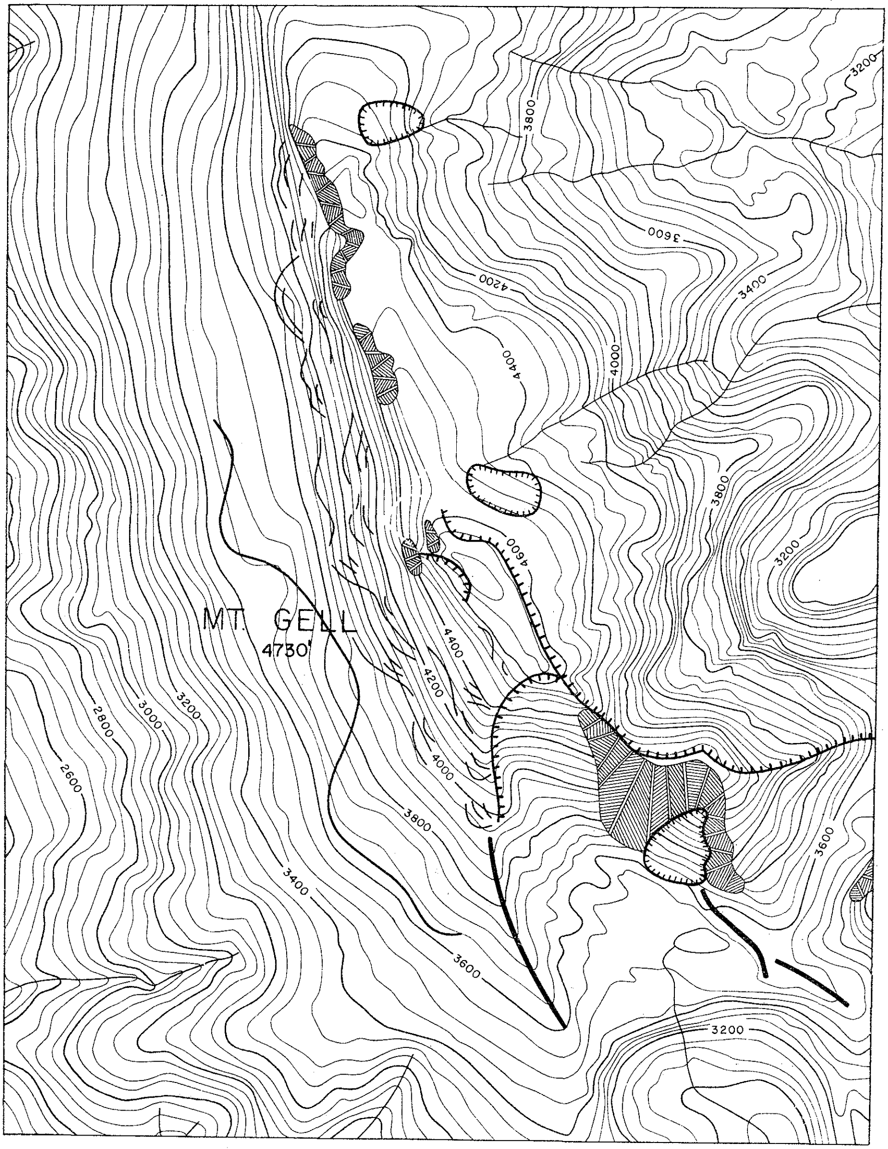


Fig. 3. Rock glaciers and associated landforms about the summit of Mt. Gell

Legend as for Figure 2, except that cirques include those of glacial origin

of the forested talus would almost double this figure. The smaller extent of undoubted rock glacier on Mt. Gell may reflect heavier precipitation and greater cloud amounts toward the west during their formation (as at present), conditions generally considered inimical to widespread rock glacier development. Alternatively, it may reflect merely the rather lower altitude of Mt. Gell (1,432 m) compared with Mt. Olympus (1,463 m). It may be significant that the slightly lower dolerite summit of Mt. Hugel (1,386 m: 4,550 feet), lying between Mts. Olympus and Gell, has not developed rock glaciers. It would appear that the critical minimum altitude for rock glacier initiation lay at between 1,372 m (4,500 feet) and 1,432 m (4,700 feet) in this part of Tasmania during the last glaciation. Blockstreams moving largely by rock glacier creep have been reported from 1,680–1,780 m above sea level (5,200–5,500 feet) in the Toolong Range of southern New South Wales (Caine and Jennings, 1968) and in eastern Victoria where they extend down to 1,188 m (3,900 feet) locally (Talent, 1965). They appear to date from the last glaciation (younger than c. 35,200 yr. B.P.).

NIVATION CIRQUES AND PROTALUS MORAINES

Nivation has been important in the development of minor landforms and the modification of glacial and periglacial slopes in many mountain areas during and since the maximum of the last glaciation. The process remains mildly active under the present climate in Tasmania and in the alpine tract of the south-eastern mainland of Australia to the extent of preserving the form and sharpening the headwalls of some nivation cirques and niches. Nivation

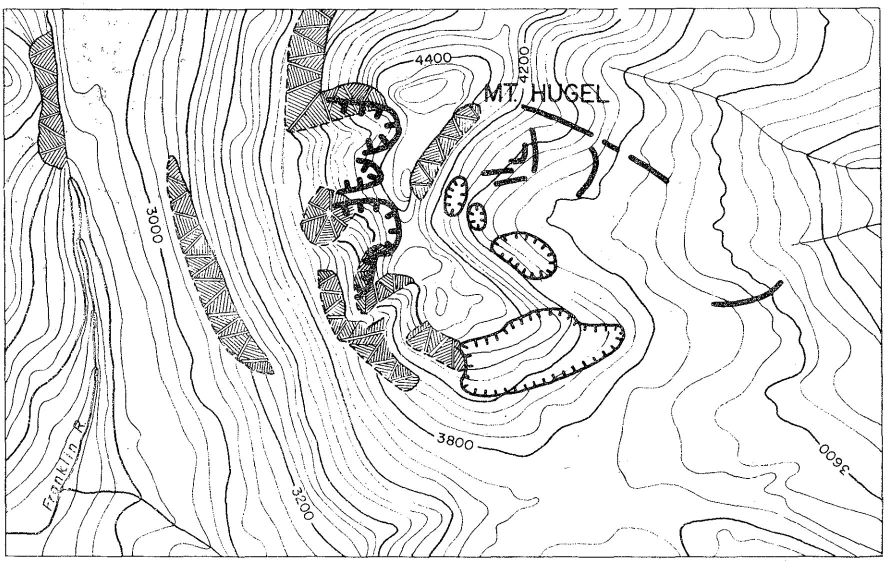


Fig. 4. Some landforms of the Mt. Hugel summit

Legend as for Figure 2

cirques are found set into glaciated slopes in west-central Tasmania e.g. into the glaciated summit plateau of the King William Range (Pl. 4), as well as on the backwalls of some glacial cirques. The nivation cirques of Tasmania share the markedly asymmetrical distribution on east-facing slopes and low depth-area ratios of the glacial cirques of the western centre and may be explained in the same terms, i.e. severe wind-drifting of snow from south--west and west, exposure to the northerly sun having been a control secondary to the direction of wind-drift accumulation (Derbyshire, 1964, 1968a).

Protalus moraines are known from a few localities in south-eastern Australia. All are relic forms. Many Tasmanian cirque moraines of pleniglacial age have a large protalus element (e.g. the moraine in the northernmost cirque on Mt. Olympus: Fig. 2) although late-Glacial and postglacial forms are small and poorly formed. An example is provided on Mt. Hugel, where poorly formed protalus ridges occur well up the backwell of a glacial cirque (Fig. 4). The change in aspect from the due easterly orientation of the glacial cirque to an east-south-easterly orientation in the retreatal moraines and, finally, a south-easterly aspect in the protalus moraines is striking, and demonstrates the progressive change in the relative importance of the wind-drift accumulation and shade factors in favour of the latter as deglaciation progressed.

BEDDED SCREES

Bedded screes (*éboulis ordonnés*) consisting of alternately stratified coarse and fine angular to subangular talus with the stratification lying parallel to the surface of the steep (28–40 degrees) uniform slopes, occurs on east--facing slopes in the Mersey River valley and on northerly and north-westerly facing slopes of the Great Western Tiers. Their origin is due to a complex of factors not properly understood (*cf.* Tricart, 1963; Embleton and King, 1968). It appears that the debris derived from frost-shatter moves downslope under gravity and may be sorted seasonally by meteoric waters including snowmelt (*cf.* Tricart, 1956).

The correlation of bedded screes and leeward slopes, noted in Poland and France, led Dylik (1955) to stress the role of drifted snow in their formation. Tasmanian examples are known from both leeward and windward slopes (Fig. 5). The coarse bedded screes of the Mersey River valley of northern Tasmania lie between 426 m (1,400 feet) and 457 m (1,500 feet) above sea level. They have been exposed by engineering works at points 1.6 km (1 mile) north of Walter's Marsh, where bedded screes on a slope of 40 degrees overlie moderately weathered, loose-textured glacial till associated with the drift suite of the last regional glaciation (Pls. 5 and 6); 4.8 km (3 miles) south-

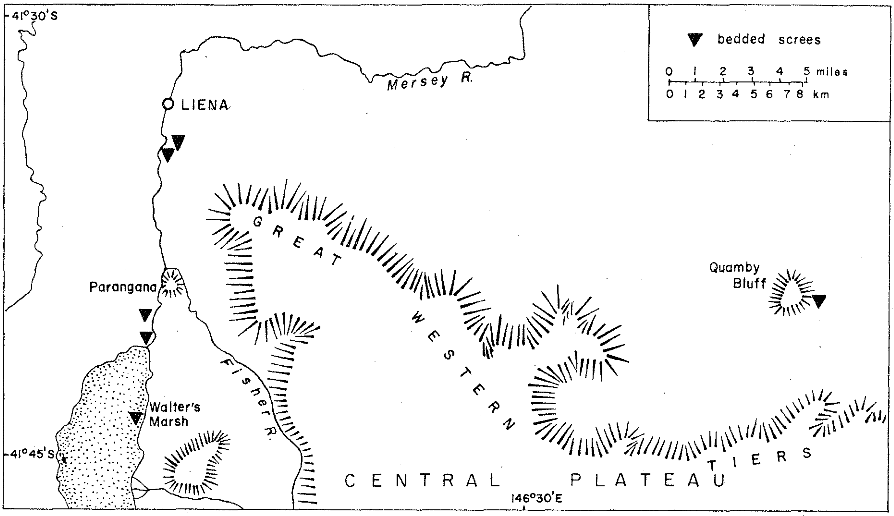


Fig. 5. Exposures of bedded screes in north-western Tasmania

-west of the Mersey-Fisher confluence, where the over-all surface gradient is 30–32 degrees (Pl. 7); and 3.2 km (2 miles) south-west of the same river confluence. At the latter location, the screes (with a surface slope of 28 degrees) are composed predominantly of the underlying Precambrian bedrock, but also contain some large dolerite cobbles and boulders with decomposed rinds over 5 cm (2 inches) thick. These boulders can only have been derived from the Central Plateau by ice transportation. Their weathering characteristics are similar to those of the older of two tills already recognized in this region (Derbyshire, 1968a), and the possibility must be entertained that these bedded screes were active during the last regional glaciation of Tasmania. The bulk of the deposits may prove to be broadly contemporaneous with the last pleniglacial, for bedded screes have not been seen to overlie the fresh lodgment till of the last regional glaciation.

Bedded screes with finer constituents are exposed on the northern and north-western slopes of the Central Plateau, $2\frac{1}{4}$ km ($1\frac{1}{2}$ miles) south of Liena at c. 457 m (1,500 feet) above sea level and between that altitude and 609 m (2,000 feet) on the flanks of Quamby Bluff, where talus of mean diameter 5–6.5 cm (2–$2\frac{1}{2}$ inches) is interbedded with fragments of 0.6 cm ($\frac{1}{4}$ inch) and smaller within a silty matrix (Pl. 8).

In view of the absence both of a preferred orientation and of nivational erosion forms, the drifting of snow and the presence of permanent snow patches which have been invoked to explain some bedded screes on leeward slopes (Guillien, 1951; Dylik, 1955) are hardly indicated for this part of Tasmania. Differential drifting of snow would in all probability have been inimical to the development of these steep, broad, uniformly-bedded screes (*cf.* Watson, 1965). The degree to which bedded screes are climatically diagnostic remains debatable, although any hypothesis must take account of the strongly bi-modal texture of the material and the alternation of some remarkably regular bedding in which the amount of silt matrix and the mean diameter of the angular fragments are the most striking characteristics. The periodicity of the bedding is a central problem. While there is no direct evidence to support it, an annual rhythm is worthy of consideration on the grounds of simplicity of process. For example, Czeppe (1968) has shown that cryonival processes in Spitsbergen follow a distinct seasonal pattern. Solifluction with some pipkrake action is at a maximum in the first half of the summer, pipkrake action reaches a maximum at the end of summer and in early winter, while frost shatter and associated rock-fall occurs in winter with maxima at the beginning and end of that season. On slopes varying from 18 to 40 degrees, and particularly on favourable lithology such as fissile siltstones or schists, such a seasonal pattern of dominant processes might well create an annual cycle in stratified slope deposits. The thickness of all the bedded screes noted in northern Tasmania is known to exceed 7 m (23 feet): maximum depths are not known. Measured sections have shown that 20–30 paired strata per metre is typical. While this throws little light on the question of an annual periodicity[2], it does suggest for slopes below about 600 m (c. 2,000 feet) in north-west central Tasmania a prolonged period of active slope wasting in a moist, periglacial environment (*cf.* Watson, 1965). There may have been some local re-activation of the screes in postglacial time, the only dated material from the Tasmanian bedded screes yielding a date of c. 3,000 yr. B. P. (T. N. Caine, pers. comm.).

PRESENTLY-ACTIVE PERIGLACIAL FORMS

Cryonival processes are active under the present climate above 1,828 m (6,000 feet) in the Snowy Mountains of the mainland (Galloway, 1963, 1965) and above 1,220 m (4,000 feet) in north-west central Tasmania. Limited frost shattering, notably frost-wedging in jointed dolerite, takes place in the Tasmanian highlands as may be seen from occasional freshly-moved blocks and rare rock-falls, e.g. at c. 1,250 m (4,100 feet) near the headwall crest of the Lake Rufus trough-end in the King William Range. Active development of a small nivation cirque has also been described by Peterson

[2] Field assessment of this possibility is in progress.

(see Figure 13 in Davies, 1969) at 1,500 m (4,690 feet) near the summit of Frenchman's Cap, western central Tasmania.

Frost sorting of dolerite debris producing patterned ground has been observed in Tasmania on the Central Plateau at 914 m (3,000 feet) where, however, it may have been stimulated by firing of the vegetation (Jennings, 1956); near the summit of Mt. Wellington and on the Ben Lomond plateau at 1,356 m (4,450 feet: Davies, 1967), and at 1,371 m (4,500 feet) on Mt. Rufus (Derbyshire, 1963). Frost sorting of mountain-top detritus derived from Precambrian quartzites and schists is active on the rounded summit of Mt. Campbell (1.6 km – 1 mile – north-east of Cradle Mountain) at 1,250 m (4,100 feet). That coarse sandstones and conglomerates within the Permo-Triassic sequence provide grain-size combinations particularly favourable to frost sorting is suggested by the development of patterned ground at rather lower altitudes on such rocks than is characteristic of the dolerite. For example, active patterned ground occurs on Permo-Triassic conglomerates as low as 1,005 m (3,300 feet) on the Murchison-Forth interfluve, 14.5 km (9 miles) south-south-east of Cradle Mountain, in particularly favourable conditions of lithology, groundwater and wind-removal of snow.

Under the present climate, congeliturbation appears less important than nivation in the mountains of south-eastern Australia. Frost action and periglacial solifluction occur on a limited scale only above 1,980 m (6,500 feet) in the Snowy Mountains (Galloway, 1965) and also on the Bogong High Plains of eastern Victoria where frost sorting is rudimentary, even above 1,860 m (6,100 feet). In Tasmania, solifluction is not occurring at present except on a very small scale and in particularly favourable circumstances and limited areas, none of them below 1,340 m (4,400 feet: *cf.* 4,200 feet of Galloway, 1965)[3]. This is 883 m (2,900 feet) above the often-quoted average lower limit of Pleistocene solifluction in Tasmania (457 m: 1,500 feet) and over 914 m (3,000 feet) above the lowest recorded occurrence of a Pleistocene congeliturbate in western central Tasmania (411 m: 1,350 feet).

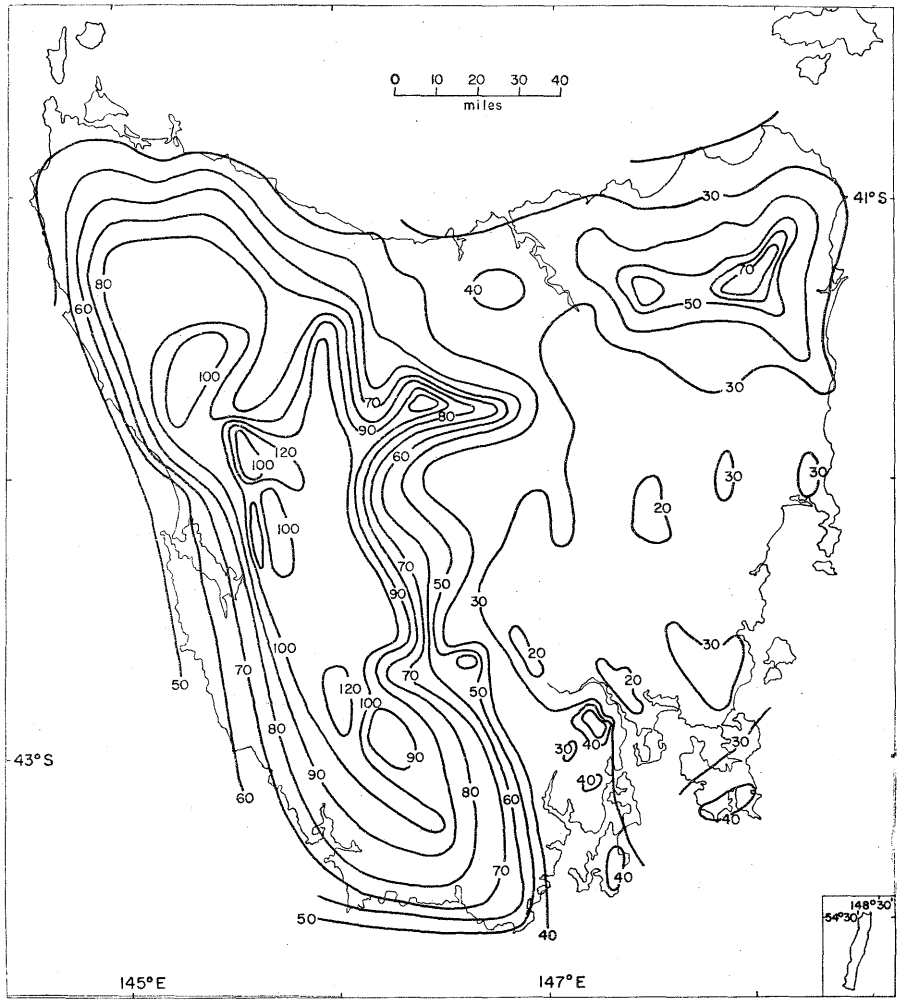
No actively-forming bedded screes have been reported from Tasmania. This is not necessarily to be explained in purely climatic terms, for it undoubtedly reflects also the absence of suitable lithology at the highest topographic levels where such action might be expected at present, the summits of all Tasmanian mountains above 1,525 m (5,000 feet) being of dolerite.

[3] The turf and stone-banked terraces at c. 1,220 m (4,000 feet) adjacent to moraine ridges in a cirque south of the summit of Mt. Rufus (Pl. 9) are relic forms.

CLIMATIC IMPLICATIONS

The climatic environments conducive to the development of the landforms and deposits described here are imperfectly defined, and none can be considered narrowly diagnostic in terms of specific temperature conditions. Nivation cirques, for example, have been described from a variety of snowy climates including continental and maritime lower-middle latitudes, maritime sub-arctic, and maritime and continental arctic (*cf.* Derbyshire, 1968b). Bedded screes have been described from quite a wide range of Pleistoenec environments (*cf.* Dylik, 1955; Soons, 1962; Embleton and King, 1968) while rock glaciers are considered to form under variable local conditions provided the mean annual temperature of the site remains below 0°C for a sufficient number of years.

However, all these forms appear to be potentially definable in terms of the amount of solid precipitation relative to the temperature, in any given locality. Judging from the literature cited above, it appears that rock glaciers are favoured by a cold, relatively dry climate with low snowfall amounts. There is accumulating evidence that high altitude cirques of low latitude mountains characterized by strongly-localized wind-drift accumulation of snow, high sun angles and low air temperatures tend to possess a distinctly low depth-area ratio (Garcia-Sainz, 1949; Derbyshire, 1968a). Ablation due to radiation rather than to conduction *via* advected warm air, firnification by recrystallization rather than compaction (Lliboutry, 1956 ; Shumskii, 1964) and the production of very little meltwater is characteristic. The shallow glacial cirques on Mt. Olympus are considered to have arisen under such a regime, as are cirques to the eastward and in the Snowy Mountains of the mainland (Peterson, 1968). The close association of these glacial cirques with shallow nivation cirques, protalus moraines, well developed rock glaciers and widespread evidence of severe congelifraction on the summit surface of Mt. Olympus argues in favour of broad contemporaneity of formation and hence a similar formative climate. With rapid decline in precipitation amounts and mean cloudiness eastwards of Mt. Olympus (Fig. 6), and with the wider altitudinal range of glacierization to the west (together with the change to more resistant and more finely-grained siliceous rocks), the area in which favourable lithological and orographic conditions coincided with appropriate climatic conditions for the development of rock glaciers was limited to the upper slopes of a few unglaciarized mountains in the north--west centre of the island. This region lies almost exactly at mid-point on a scale of continentality calculated for a zone about 42°S, using Johansson's formula (see Landsberg, 1958), Port Sorell on the west coast being the



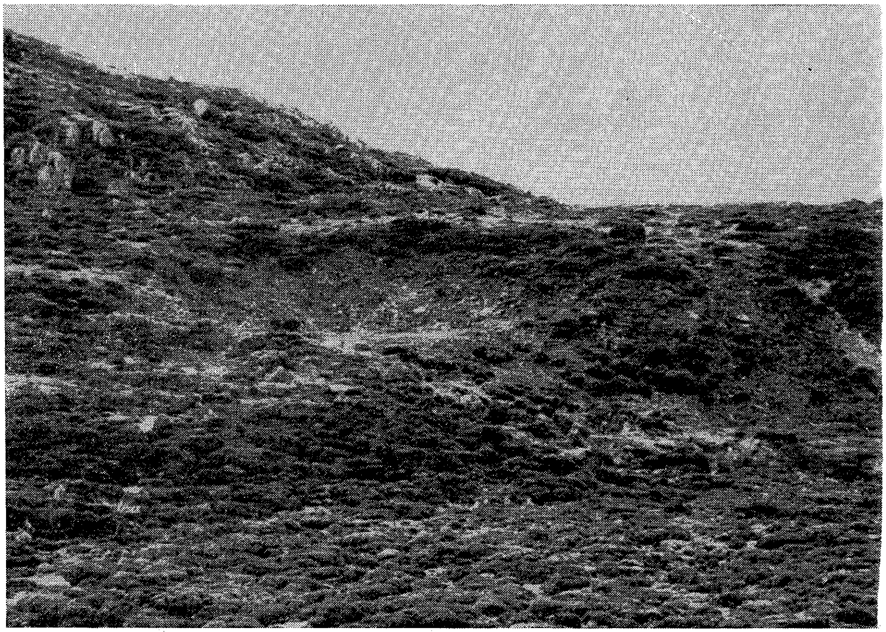
by H. E. C. Hydrology Section, July 1965

Fig. 6. Mean annual precipitation in inches for Tasmania

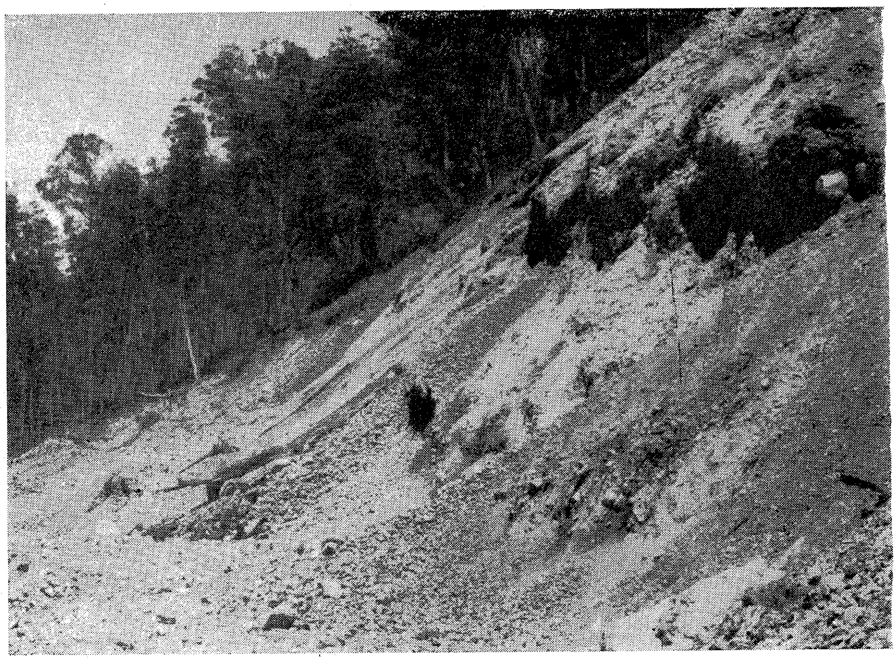
Isohyets at ten inch intervals

maritime extreme and Ross in the eastern midlands (long. 147°30'E) the continental extreme.

None of the periglacial phenomena so far described from the Tasmanian mountains appears to require the former general presence of permanently frozen ground. Frost wedge casts have yet to be recognized with any certainty. However, some estimate of the degree of severity of the temperature climate may be made on the basis of the evidence presented above. Estimates of

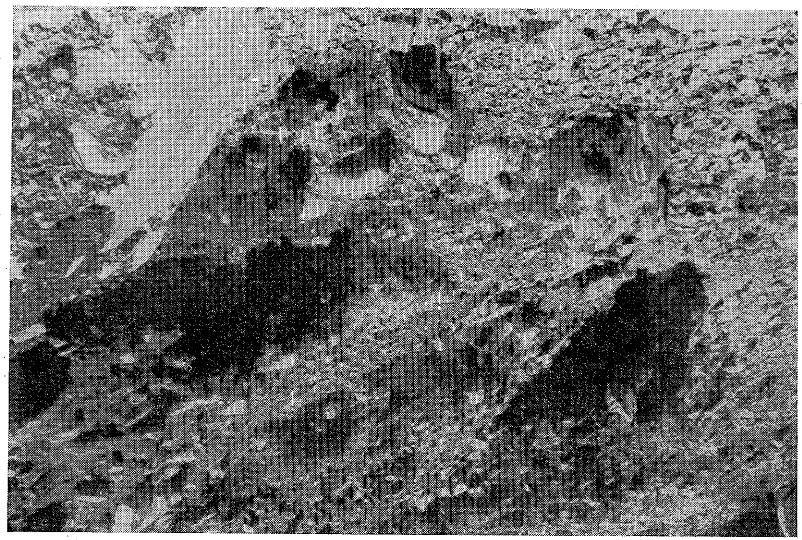


Pl. 4. Small relic snowpatch hollow at approximately 1,200 m (3,950 feet) on the plateau surface of the King William Range. Backwall is 10 m (33 feet) high



Pl. 5. Bedded screes overlying moderately weathered, loose-textured glacial till

Over 7 m (23 feet) of bedded scree is exposed on the right of the photograph



Pl. 6. The loose-textured till beneath the bedded screes shown in Pl. 5

Pl. 7. Exposure of over 9 m (30 feet) of bedded screes, 4.8 km (3 miles) south-west of the Mersey-Fisher confluence

Pl. 8. Over 7 m (23 feet) of bedded scree on Quamby Bluff, developed in fissile siltstones

Pl. 9. Partly vegetated stone-banked terraces at about 1,220 m (4,000 feet) in a cirque on the south-facing slopes below the summit of Mt. Rufus

the maximum annual mean temperatures for the maintenance of permafrost vary from −1.1°C (Black, 1954) to −5.0°C (Jenness, 1949; Mortensen, 1952), with −2°C being a commonly recurring figure (summary in Tricart, 1963). The movement mechanism of rock glaciers demands ground conditions approximating that of permafrost, at least locally, and Wahrhaftig and Cox (1959) have suggested a mean annual temperature of less than 0°C for rock glacier formation. In view of the well developed forms on Mt. Gell and Mt. Olympus, it would seem reasonable to assume a mean annual temperature for these summits in the zone of rock glacier formation (over 1,220 m: 4,000 feet) of at least −1°C or even −2°C or −3°C during the last pleniglacial. In the absence of any knowledge of the present temperature climate of these summits, no direct estimate of the amount of temperature depression during the pleniglacial can be offered. However, a mean annual temperature of 8.0°C is suggested for the Lake St. Clair pumping station (elevation 738 m: 2,420 feet) immediately to the east by the Bureau of Meteorology (pers. comm.). Assuming a normal (free air) lapse rate, this yields a mean annual temperature for the present of about 5.3°C at the 1,220 m level, suggesting a temperature depression at the pleniglacial of at least 6.5°C relative to the present. This is rather greater than that suggested by Galloway (1965) on the basis of average tree-line and solifluction limits, but compares closely with the temperature depressions for the highlands of southern New South Wales estimated from similar evidence (Caine and Jennings, 1968).

It seems likely that in ice-free areas of north-west central Tasmania during the last pleniglacial, small areas of permafrost may have developed, for substantial annual freeze—thaw cycles appear necessary to account for the deep cleaving of the dolerite of some unglaciated summits, including Mt. Olympus and the northern half of the King William Range. This, together with the few but large rock glaciers on windward slopes and the complementary distribution of shallow glacial or nivation cirques on leeward slopes, suggests a climatic regime on the higher summits characterized by severe winds between south-west and north-west and abundant local accumulation of drifted snow, the lower temperatures yielding lower absolute humidities relative to the present climate (*cf.* Derbyshire, 1969). Such a climate appears transitional between the B2 and A climates of Tricart (1963).

The altitudinal range of these conditions in west-central Tasmania may have been relatively small, however, for vigorous trunk glaciers existed in the valleys west of the Central Plateau adjacent to steep slopes which bore cirque and rock glaciers. The evidence of meltwater action in the drifts of these trunk glaciers suggests that there existed precipitation and temperature gradients in this part of Tasmania at least as marked as those of the present. The progressively lower altitude at which severe glacial erosion is evidenced, the relatively greater proportion of washed drift, and the increasing rarity of cryergic phenomena with distance westward of the Central Plateau are consistent with this view. So too is the rapid transition from severe glacial erosion to apparently unglaciated low plateau country north-east of Cradle Mountain (Derbyshire, 1966), and the distribution of stratified slope deposits on the north-western margins of the glaciated region. These northerly-facing slopes, with gradients averaging at least 20°, experience incident sun-angles of between 45° and 90° (mid-winter and midsummer, respectively). During a pleniglacial such high sun-angles would have produced marked variations in ground-surface temperatures in all seasons, but particularly in any winter spells of low cloudiness, thus facilitating the accumulation of frost-shattered debris on steep slopes, followed by colluvial action and eluviation of the talus in the warm season. A climate in the B2 range (Tricart, 1963) is suggested for this area above the 450 m (1,500 feet) level.

In conclusion, it may be said that there is accumulating evidence that the pleniglacial climate of the higher summits of the mountains of south-eastern Australia was not only cold (mean annual temperature below 0°C) but also dry relative to the present (see also Derbyshire, 1969; *cf.* Caine and Jennings, 1968). Present evidence suggests that this condition may have been limited to the highest unglacierized peaks only, at least in Tasmania (for a broader interpretation, see Galloway, 1965). However, it may prove possible to determine the altitudinal extent of such dry, cold conditions by further work on the bedded screes described above and on the inland dunes of eastern Tasmania.

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