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GLACIATION OF THE FRENCHMANS CAP NATIONAL PARK

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(With four plates and four text figures.)

ABSTRACT

Mapping of glacial landforms of the Frenchmans Cap National Park shows that an independent system of cirque and valley glaciers accumulated upon and caused considerable modification of this area during the last Pleistocene glaciation. Clearest evidence of the sequence of advance and retreat is seen in the Vera Valley where, during the maximum phase, the ice reached the Vera Creek gorge and overspilled a series of diffluence passes further upstream. The history of subsequent retreat can be traced with reference to the post-maximum moraines.

Some detailed work on the morphology of the lake basins makes possible a classification of lakes which shows clearly the nature of the glaciation and the influence of lithology. Careful consideration of the aspect, morphology, and elevation of the cirques in which many of these lakes occur provides interesting comment upon the factors governing accumulation and ablation in the maritime mountain environment of glacial and late-glacial times.

Radiocarbon assay of leaf fragments in rock-flour below peat in a bedrock depression on the lip of one of the high discrete cirques gives a minimum age of 8720 ±220 years B.P. for the final deglaciation of the Frenchmans Cap area in particular and the western Tasmanian mountains in general.

The present study is discussed with reference to some of the chronological problems of the last glaciation. The possibility of a sequence of glacials and interglacials is recognized. As the Pleistocene climate changes of the southern hemisphere form the broad context of the present study attention is drawn to analogies in the New Zealand sequence.

INTRODUCTION

The Frenchmans Cap area surrounds the prominent peak of that name in the west of Tasmania some 25 miles east of Queenstown. It is bordered to the west and north by the great bend of the Franklin River, while to the south and east lie the Lachlan and Loddon Plains. The present study, based on some 60 days field work, and on air photo interpretation, covers an area of about 70 square miles. Complete air photo coverage exists as well as maps at a scale of one mile to the inch (Frenchmans Cap National Park map and Franklin sheet), with contour intervals of 100 and 50 feet respectively.

The land varies in altitude from about 550 feet in the gorge of the Franklin to 4739 feet at the Cap summit. The peaks contained within the mountain area form a continuous range from the Clytemnaestra north to the Frenchmans Cap and the Lion's Head, then east by the razorback ridges at the head of the Livingston Valley, to Nicoles Needle and the Philps Peak massif. These mountains are formed of the resistant siliceous Precambrian metamorphics of the west and southwest of Tasmania, standing locally as ridges and peaks above the steep-sided valleys, which here are cut into softer schists and phyllites. To the east the quartzites drop away through dense temperate jungle to the Loddon Plains. These button-grass plains are base-levelled to rock-bars at the heads of gorges that eventually lead to the Franklin River, the master stream of the region. Marked by its incised character and structurally controlled sections and rapids, this river drops from 950 feet near the confluence with the Loddon to 550 feet at the confluence with the Livingston 25 miles downstream. The Cap area drains to the Franklin River in a more or less radial pattern, in places closely controlled by strike and fault directions. The area has a wet, cool, temperate climate and is largely covered with rain forest, which makes access to much of it a matter of difficulty.

No previous geomorphological work as such has been completed in the area although the physiographic problems of nearby areas to the north have been discussed by Spry (see below). Earlier publications include papers by Ward (1908a, 1908b, 1909) and papers by Finucane and Blake (1933), Blake (1936, 1937), and Ahmad, Bartlett and Green (1959). The only other references of note are the accounts of early explorers, e.g., Franklin, Philp and Rumney, and Calder.

Geology

The major contributions to the geology of the area are the papers by Spry (1957), Spry and Zimmerman (1959), Spry (1962, 1963) and Spry and Gee (1964). The first two accounts cover the geology of the nearby Mt. Mary and Mt. Mullens areas respectively, while the last three concern the Frenchmans Cap area (see map in Spry 1962, p. 112).

Spry describes the Precambrian metamorphics of the area as " strongly deformed metasediments with a low to moderate grade of regional metamorphism " (1962, p. 112). The gap area is mapped as Mary Group east of Franklin Group although Spry has recognised the presence of Scotchfire Group rocks (1963, p. 113). Further work in the area (Duncan, pers. comm.) shows the following:—

Franklin Group	Mica, garnet and albite schists with quartzites and amphibolites.	These rocks comprise the lower dissected area west of the Cap.
Mary Group	Massive and schistose quartzites. The massive quartzites are white, thickly bedded and concordant.	Comprise the main ridges and peaks of the Cap area.
Scotchfire group	Phyllites quartz-schists and dolomitic schists. The phyllites are softer than the Mary Group and have marked cleavage.	Generally underlying the ridges of Mary Group rocks and forming the saddles and gullies.

This succession is a structural one and carries no other implications. As in the adjacent Precambrian areas stratigraphic position and regional metamorphism are debatable (Spry 1962, p. 113).

Detailed mapping is often difficult due to structural complexity, strong faulting, and differential silicification which causes the rapid and irregular changes in lithology. There have been at least two periods of folding. Superimposed upon the effects of the Frenchman Orogeny is that of the later (Devonian) Tabberaberan Orogeny: it is the evidence related to the latter folding that is mapped by Spry (1962, fig. 4, p. 113).

The influence of lithology on the topography has already been outlined. The influence of structure on topography is evident in the drainage pattern. The area is in that part of Tasmania where " the folded Precambrian basement is uncovered and the rivers have evolved a trellis pattern with parallel ranges marking the strike of the harder rocks " (Davies in Gill, 1962, p. 244). This is true of the Frenchmans Cap area as a whole and of the Franklin Group rocks to the west in particular. The parts immediately surrounding the Cap itself show the influence of fault or joint control in the alignment of the valleys.

GLACIAL GEOMORPHOLOGY

It seems reasonable to suppose that the glaciers of the Frenchmans Cap area were superimposed on a fluvial landscape of dominantly joint- and fault-controlled valleys interrupted by small areas of karstic depression some of which are now occupied by glacial lakes. The resulting glacial landforms are typical of mountain glaciation in the highlands of Tasmania. Catenary troughs, cirques, cirque-lakes, comb ridge aretes and moraine ridges are common features above 2000 feet, while outwash valley trains occur at lower elevations (Fig. 1). Basically a complex and high cirques and radial valleys, the area contains eight cirque and valley glacier systems heading at altitudes between 3000 feet and 3400 feet, with lengths varying between three miles and less than half a mile, and areas from eight square miles to less than one quarter of a square mile. The following description should be read with reference to Figs. 1 and 2 and one of the published topographic maps.

The Vera glacier

Perhaps the most interesting valley is that once occupied by the Vera glacier and its tributaries. Fed from two cirques in the Philps Peak massif, the Marilyn and the Barron, this glacier left evidence of expansion to a valley glacier maximum stage and subsequent retreat to the cirques after a post-maximum still-stand.

The larger of the two cirques heads at a col known as the Barron Pass, a gap opened by headwall sapping in the comb ridge extending from the Nicoles Needle group eastwards to Philps Peak. The erratic blocks on the crest of the pass are much more likely to be protalus than moraine left by transfluence across the pass. The exact limits of inundation of the ridges is hard to assess from air photos because the overridden appearance of the high schist outcrops is due mainly to the geological structure. The hanging Marilyn valley has been selectively eroded out of dolomite. Below the 1000 foot backwall topped by Mary Group quartzites and quartz-schist, Lake Marilyn, which is 60 feet deep, is contained behind a rock lip of dolomite. Down valley a second and much smaller lake, now rapidly being filled by colonizing vegetation, hangs above a 400 feet drop to Lake Vera. Here, at the meeting point of the Marilyn and Barron glaciers, the half-mile long lake is 180 feet deep and very much " down at heel " (see Fig. 3).

The nature of the preglacial valley into which the glacier expanded has caused an interesting series of features. Like the Tahune valley to the west, the upper section is joint controlled. Both the Tahune and Vera valleys turn sharply to follow the strike direction in their middle sections. Unlike the Tahune glacier, the Vera glacier was severely restricted where its valley changed direction, so that the low strike ridges on the outside of the bend were overridden as the ice spilled out through five diffluence passes overlooking the Loddon Plains. Although there is no evidence of how far these outlet glaciers advanced towards the plains, they were responsible for the agglomeration of tillflow (Hartshorn 1958-footnote p. 477, Sharp 1960, p. 333) and poorly stratified outwash, the coarse nature of which is explained by the comparatively high gradient. Below these deposits, valley trains were laid down on the plains as may be seen where Philps Creek has incised five feet into outwash immediately upon reaching the flatter gradients of the Loddon Plains. The aerial photographs suggest that this valley train was braided.

The weathered state of the schist boulders in the till on the flanks of the hills separating the diffluence ice tongues above the upper Vera Plains indicates that the deposits are among the oldest surficial sediments in the region.

The lower Vera valley itself provided a further outlet. It was considerably modified, the preglacial gorge-like walls being overridden and a giant roche moutonnee formed at the valley bend, the walls of which exhibit vertical stoss and lee features (Lewis 1949). Ice from two north bank tributary cirques heading at the comb ridge west of Nicoles Needle may have joined the lower Vera glacier at this stage.

The maximum extent of the lower Vera glacier coincided with the Vera Creek gorge. Drift fills

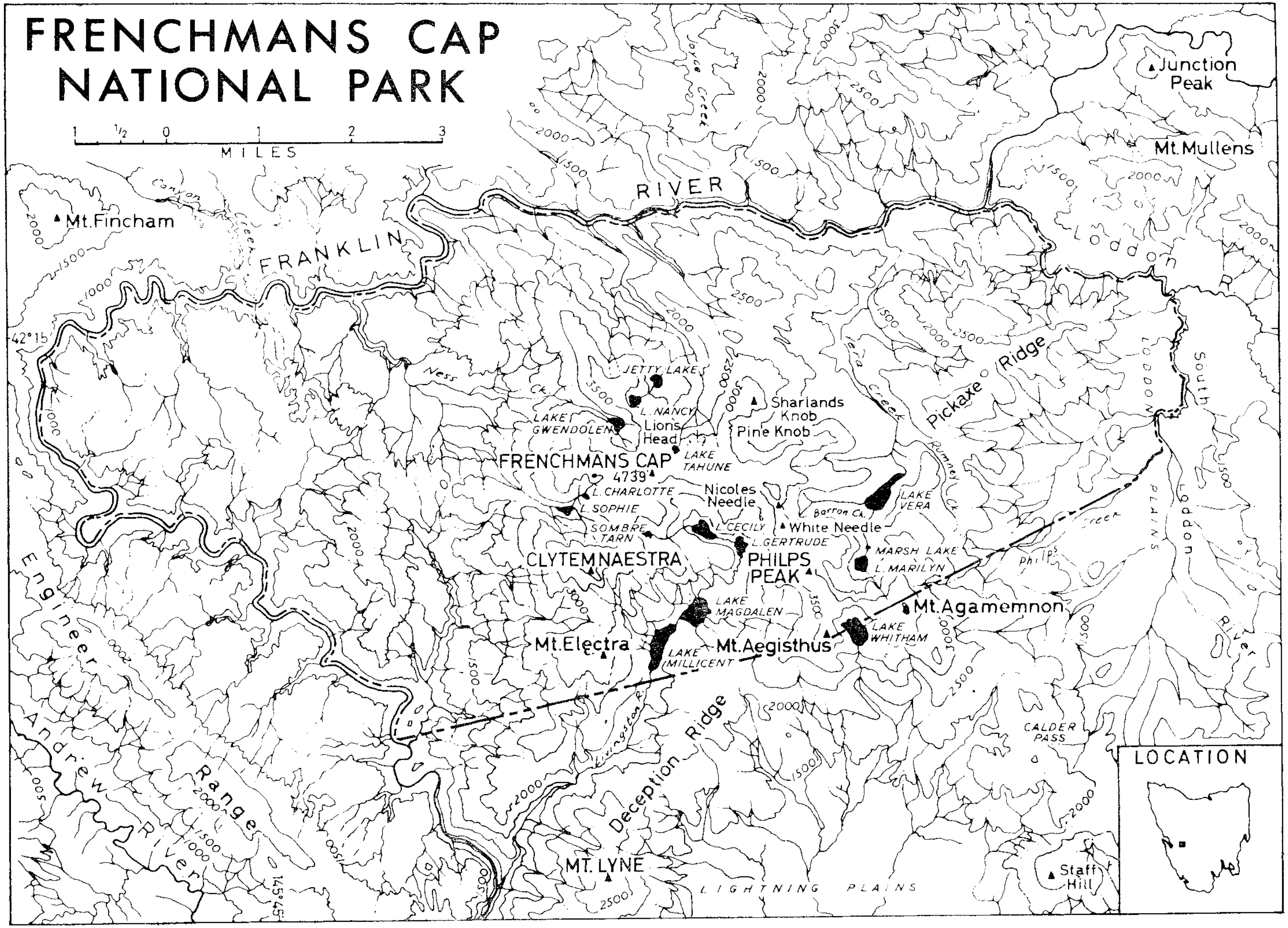


FIG. 1.—Location Map. Frenchmans Cap National Park (Nomenclature from the National Park Map and from the climbing notes of J. Middleton-Elliott).

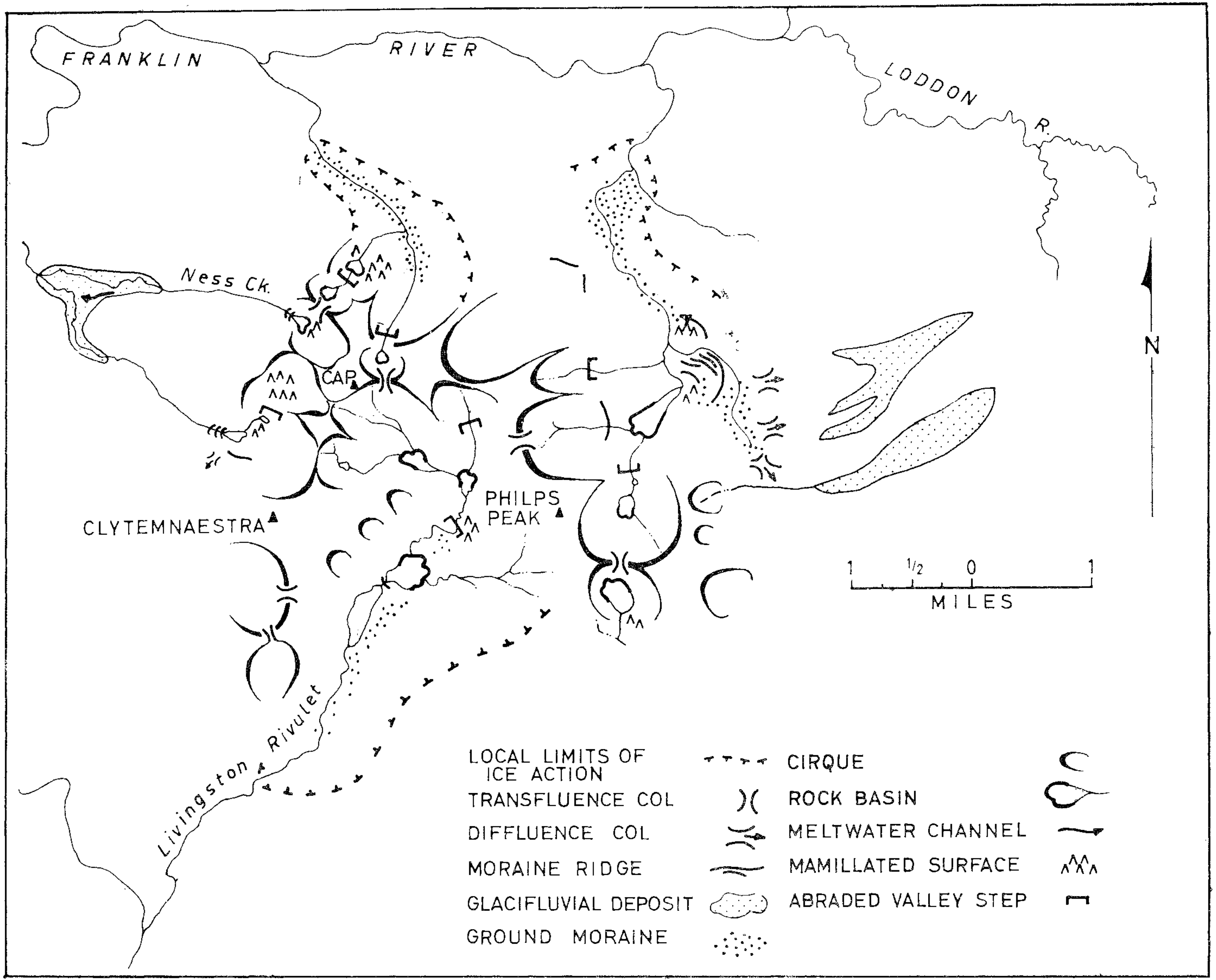


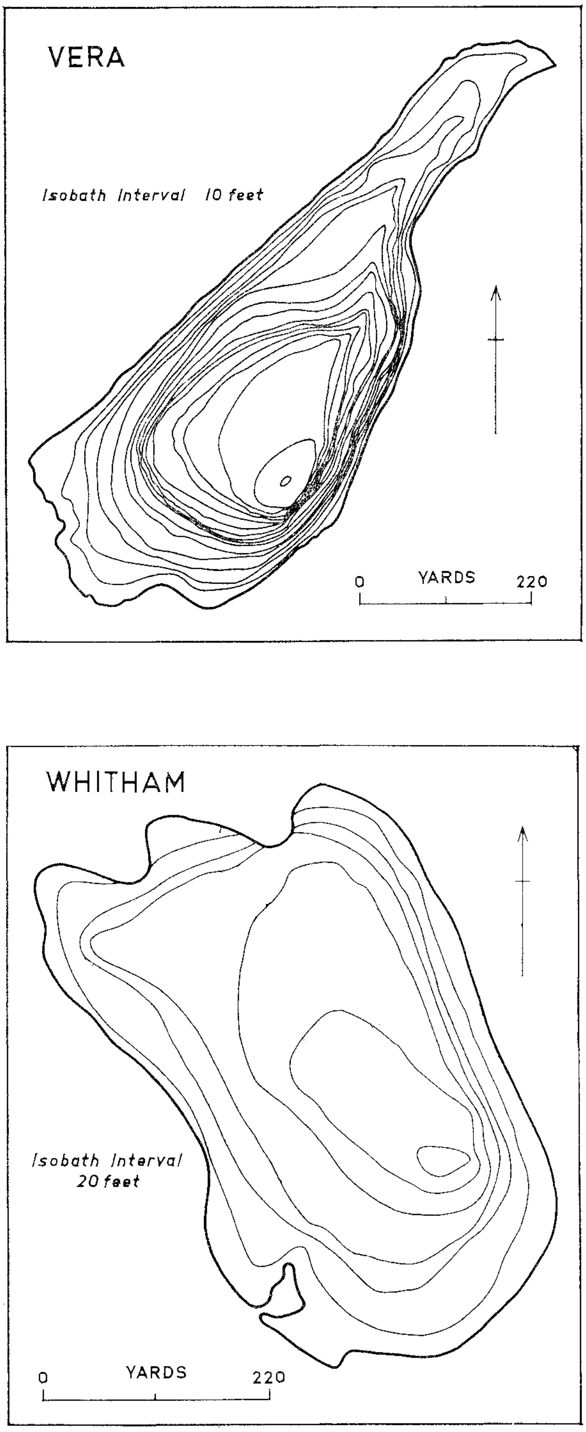
FIG. 2.—Glacial Landforms of the Frenchmans Cap massif.

the lower half of the valley above the gorge and has been terraced by postglacial river erosion, the present Vera Creek being incised into the till. In places the till extends to some 50 feet above the river level and must be at least 30 feet thick. It contains a large percentage of boulders, is coarse, and permeable, with a matrix of pronounced sandiness. The glacier snout may have extended a short distance into the gorge, but, as would be expected, there is no evidence for this. The narrow (20 feet to 30 feet at the base) 200-300 feet high gorge marks a distinct change of character in the valley. Above the gorge are glacial deposits; below, the valley is deeply incised and V-shaped. The same is true of the Tahune valley to the west. That glaciers from the Frenchmans Cap area were responsible for Pleistocene damming of the Franklin valley as suggested by Spry and Zimmermann (1959), to explain the presence of varved clays in the Artists Hill-Mt. Mullens area upstream is not immediately apparent.

At the glacial maximum ice in the upper Franklin and adjacent valleys may well have been coterminous with the ice of the St. Clair district and *ipso facto* with that of the central plateau area. Ice tongues from such a huge accumulation area would be pushed long distances. Ice in the Cap area on the other hand accumulated in a strictly limited area, from which the ice soon moved to the lower levels of the ablation zone. It is suggested then that although evidence is lacking it is most likely that the same ice which deposited till on Artists Hill and left large erratics at about the same height near the foot track below, and north of, the Mt. Mullens saddle, might well have placed a moraine dam downstream. The maximum extent reached by outer lobes of the main central ice cap of Tasmania is still a matter of debate. That glacial advance into the Franklin-Collingwood river junction area was a maximum stage feature is illustrated by the weathered state of the Artists Hill till (Spry and Zimmermann 1959, p. 3). It is important to realise that the location of this till is a better indication of its age than is the weathering state of dolerite boulders contained therein. The amount of weathering of dolerite till stones in a given time will vary from place to place due to differences in composition of the dolerite, composition of the till matrix and relation to ground water fluctuation. Moreover there is the possibility that drift containing weathered dolerite was transported en-masse, probably in the frozen state. This process has been suggested as a possible explanation for the presence of fragile lumps of Permo-Triassic sandstones in the bedded moraines at Lake St. Clair (Derbyshire 1965, p. 381). Perhaps these factors explain the anomalies of the weathering state of different tills of probable similar age, e.g., the Arm River till and the Artists Hill till (Spry 1958, Spry and Zimmermann loc. cit.).

The next marked stage in the history of the Vera glacier is indicated by the large recessional moraine at Lake Vera. This rises steeply 350 feet above the lake outlet and 200 feet above the Upper Vera Plains. A cover of ground moraine and scattered erratic boulders was left between the lake and the diffluence passes from which the glacier retreated. The Lake Vera moraine is aligned north-south while the glacier responsible for its

FIG. 3.—Isobathic maps Lake Vera and Lake Whitham.



formation came from the southwest. This alignment was probably governed by the rate of retreat of the various parts of the ice front from the diffluence passes (for some of the ice tongues had spread out more acutely to the main direction of flow than others), and also by the position of strike ridges which outcrop at several places on the northern end of the moraine. The retreat of the glacier at this stage can be traced in detail.

The differential withdrawal from the maximum extent at the diffluence passes was most rapid from the southernmost tongue. This left the eastern end of the moraine area open for meltwater escape over the ground moraine of the upper Vera Plain. This water escaped northward around the snout of the ice and thence down the lower Vera valley. At first this was accomplished along the now dry meltwater channel on the east side of the giant roche moutonnee. Subsequent retreat exposed lower escape routes until drainage very similar to that of the present was established.

This apparent stage of equilibrium of the Vera glacier, marked by the main moraine, was terminated by a slight advance which produced a fine push moraine capping the crest of the main feature. There followed a retreat in three main stages as shown by three small moraine ridges plastered aslant the northern end of the proximal slope of the Vera moraine (Plate 1). These ridges, clearly demarcated by their narrow tops and steep slopes, are emphasised by the meltwater channels between them. The orientation of these ridges reflects the position of the retreating ice tongue against the Vera moraine and also the bedrock strike. The alignment of the strike ridges of dolomite and quartz-schist which outcrop at this place suggest that the Vera moraine, although the largest in the whole region, is bedrock-cored. That these ridges should affect the alignment of the smaller moraines indicates that the ice was much thinned in comparison with the earlier time when drift was deposited regardless of outcrop orientation. The travertine deposit on one of the recessional moraines suggests the presence of dolomite bedrock close to the surface of the western end of the moraine ridges. Further retreat resulted in the exposure and channelling of the lower Vera Plain and in the building of the small moraines at the outlet of Lake Vera. At this stage the ice had thinned to such an extent that the truncated spur on the east side of the glacier diverted the snout so that the moraine here was built mainly on the western side. Meltwater escaped around its eastern end and became the forerunner of the present lake outlet. Several large dolomite erratics occur on the lake shore and on the slopes of the moraines. These, and the dolomite till-stone exposed in a section prepared for the collection of samples for particle-size analysis, had their origin in the dolomite of the Marilyn cirque.

The history of the glacier concluded with its retreat to the Marilyn and Barron cirques. A moraine in the Barron cirque may signify stillstand there, but no definite evidence was found of a similar stage in the Marilyn cirque. In general the high cirques of the Frenchmans Cap area lack moraines on their lips. Such accumulations as occur there are in most cases best ascribed to protalus formation.

This general lack of cirque lip moraines is interesting. On the one hand it suggests that the rock basins contained active ice mainly during the glacial maximum (see Dort 1957). The lack of evidence for re-occupation of overridden cirques after the retreat of the Central Plateau ice cap (Jennings and Ahmad 1957) may be taken as support for this. Certainly rotational slip of cirque glacier ice, postulated as the cirque forming agent elsewhere (Lewis 1960), is, following the calculations of Lewis (1949), much more likely in the Frenchmans cirques at a time of maximum ice thickness.

On the other hand active ice may have occupied the cirques until late glacial times without leaving a moraine, because of a lack of excavated material in a post-maximum stage.

The Livingston glacier

The bulk of this glacier accumulated in the deep cirque in the lee of the Cap that is now occupied by Lake Cecily. Bounded by the Clytemnaestra Ridge to the west and the Deception Ridge to the east, the glacier was constricted at its maximum at the lower end of Lake Millicent. The ice may well have overtopped the eastern trough shoulder while its western neve fields under the Clytemnaestra Ridge were carving out a series of cols, the main one of which is the South Col of the Cap.

Lake Cecily is at the base of the deepest cirque in the region—its backwall being topped by the 1500 feet face of the Cap itself. The glacial origin of the cliff is indicated by a number of things, although it is possible that the scarp is related to a fault. The great depth of the Cecily cirque is typical of lee-side accumulation areas in Tasmania and this is generally recognised as a function of aspect. The lack of buttressing common to cliffs resulting from periglacial solifluction suggests that the Cap face was immediately associated with the Cecily glacier. Although it is unlikely that active ice covered the top of the Cap it is probable that a thin carapace of neve contributed meltwater which aided headward sapping as would the moderately dense jointing of the quartzite. This glacier, together with the Tahune, Gwendolen and upper Cwm glaciers, was in the process of reducing the Cap to a hornpeak when climatic amelioration cut short the process.

The extent of the Livingston Valley glacier is problematical. Air photo interpretation suggests that glacial deposits extend into the Livingston gorge below Lake Millicent and that, as mentioned above, ice may have overtopped Deception Ridge. It is unlikely that glacial deposits would be preserved in the gorge of either the Livingston or Franklin Rivers. As for the Lightning Plains immediately southeast of Deception Ridge, Wells (1955), writing of the geology of the general area southeast of the Cap, states " no evidence of glacial physiography was found in the area and Pleistocene glacial deposits are lacking ".

Many of the apparent anomalies in the Livingston Valley can be understood if the shape of the valley cross section is considered. It is over one and one half miles from crest to crest across the valley at Lake Magdalen whereas below Lake Millicent this distance is less than one half mile. Considering the size of the accumulation area for

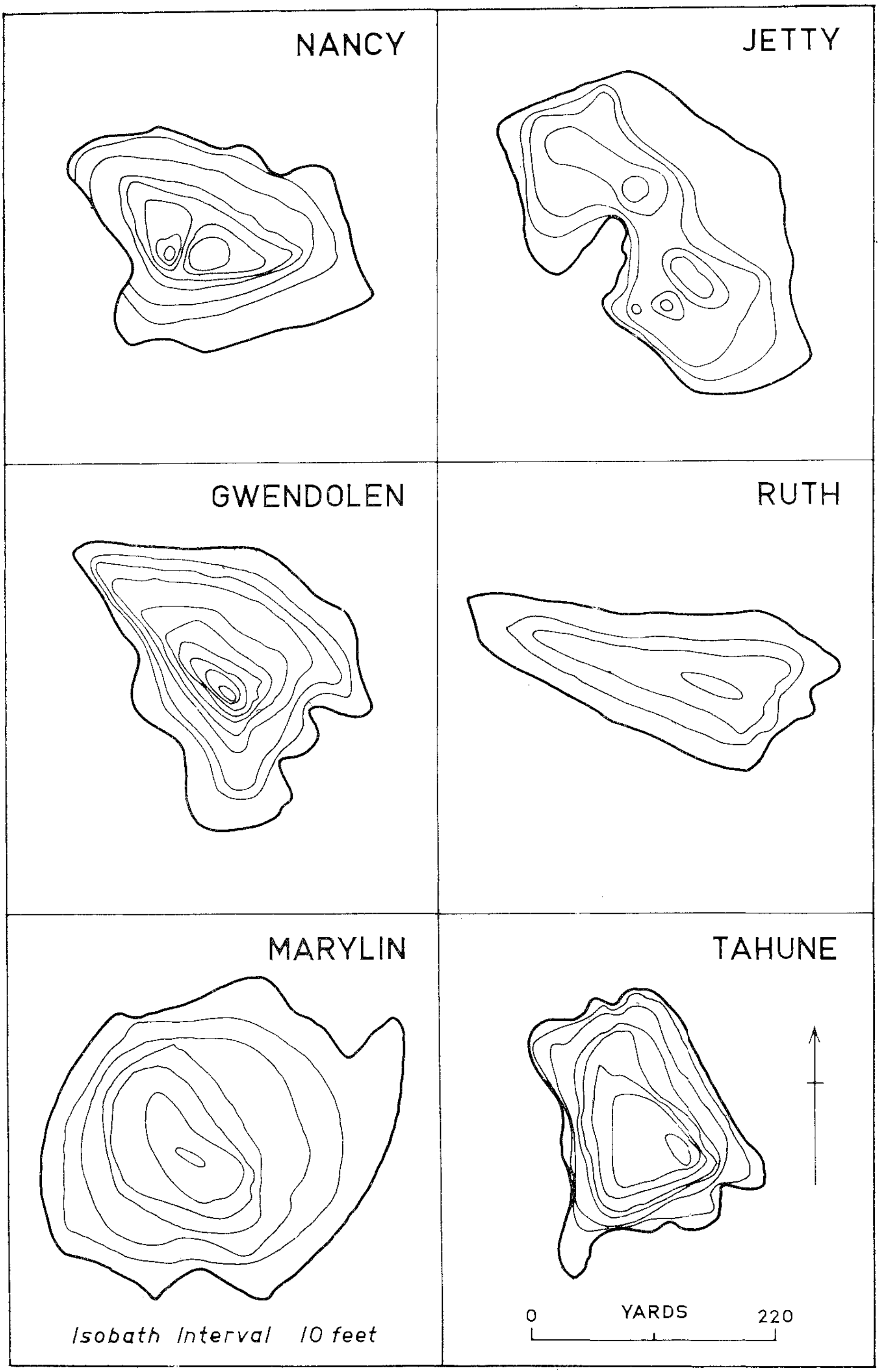


FIG. 4.—Isobathic Maps of the smaller cirque lakes.

this glacier it is justifiable to postulate penetration of the lower valley with only limited maximum stage transfluence of Deception Ridge.

A more detailed history of the valley would require lake sounding. The main lakes were too large for accurate single-handed work with sounding line and the depth sounder used was incapable of receiving signals from the muddy lake bottoms. The following comments are therefore tentative. Lake Millicent lies immediately upstream of the valley constriction and is probably partly moraine dammed. Lake Magdalen also is moraine dammed. However its situation in the widest part of the valley below a mammilated valley step which forms the low cliffed northern shoreline of the lake suggests it is a partial rock basin. Lake Gertrude, situated below a small cirque below the junction of softer Scotchfire and the Mary Group rocks has swampy shores on deep clay and silt detritus. The lake has been subject to infilling from both the Cecily and Gertrude cirques. The small cirques on the west wall of the trough probably represent the late cirque stage seen in the other valleys.

The Sophie and Upper Cwm glaciers

The trough of the Sophie glacier is located on the west side of the Clytemnaestra Ridge. This glacier was substantially augmented by ice from the Upper Cwm situated under the Cap summit on the west-southwest side. This ice may have been coterminous with the ice of the Gwendolen glacier (via the west col) and the Cecily glacier (via the south col), at the glacial maximum. It descended to join the Sophie glacier over two steps, between which lies the small rock-basin tarn, Lake Charlotte. Some of the finest mammilated rock surfaces to be found in Tasmania occur in the Upper Cwm. Smooth stoss and lee profiles in the roches moutonnees reflect north-south movement although inherent weaknesses in the quartz-schist bedrock have been exploited regardless of ice movement direction as seen by the small, comparatively deep, rock basins (see Plate 2).

Lake Sophie is a shallow lake (30 feet deep at most), dammed by a relatively extensive deposit of drift in the form of a series of recessional moraines. A section ten feet deep, exposed by the lake outlet, shows a bouldery till of quartzite, quartz-schist, and chlorite schist boulders and cobbles in a sandy matrix. The lateral moraine standing some 200 feet above the southern edge of the lake and running parallel to it is particularly well formed. Westward (downstream) it curves more and more sharply towards the lake, but is not coterminous with the recessional moraines.

Further south between the Clytemnaestra and Mt. Electra a large west-facing cirque parallels the Sophie valley. It is a comparatively poorly developed form which, nevertheless, probably overtopped the ridge to the southeast to become coterminous with the southernmost cirque overlooking the Livingston Valley. Considering the position of these cirques it is not unlikely that snow patches developed on the Deception Range in the lee of Mt. Lyne (3050′) several miles to the S.S.W.

The Gwendolen glacier

Parallel to the Sophie glacier, and to the north, the Gwendolen glacier occupied the trough overlooked by the west and north cols of the Cap. A lake basin, 90 feet deep, was eroded in massive quartzite and quartz-schist under the deepest part of the glacier, which, judging by the height of the Nancy col on the northern side of the trough, was over 400 feet thick at its maximum stage. At this time both the Sophie and Gwendolen glaciers were cascading down V-shaped valleys below their troughs, and melting rapidly among the outwash deposits, till and protalus deposited during the glacial advance to this maximum. These deposits are braided below the Gwendolen trough and several larger abandoned channels can be seen below the Sophie trough. Glacial deposits have buried low ridges of the softer Franklin Group rocks between the Gwendolen and Ness Creeks.

Subsequently the drift deposits at the end of the Gwendolen trough were laid down. This material is predominantly composed of gravel and boulders with sandy fines. Under the backwall (which is not as uniformly vertical as it might have been had the geological structure been favourable) lee ends of well formed roches moutonnees project into the lake.

The Tahune glacier

On the north side of the Cap the Tahune glacier headed beneath a steep backwall rising from 2200 to 4700 feet. This was broken at the 2700 foot level by the small, high cirque containing the rock basin of Tahune. The main valley, as mentioned above, is probably joint or fault controlled in its upper section, and strike controlled in its lower section above the Tahune gorge. At the maximum extent slight diffluence occurred from the outer part of the valley bend, while the glacial snout reached at least one and a quarter miles to the Tahune gorge. As with the Vera valley there is a distinct change of character of the river valley at the gorge. There is also a contrast between the upper and lower sections of the glaciated part of the valley. The broad catenary shape of the upper section shows that it was more susceptible to modification than the strike valley section, despite the additional contributions of ice from the Nancy and Jetty Lake areas to this part of the valley. These cirques were bounded to the east by the upper Tahune ice while they built small lateral moraines on their western sides. Air photo interpretation suggests strongly that the position of these cirques was governed by what is taken to be a well marked structural weakness there. The drop of 650 feet between the two cirques would have accommodated a cascading glacier. Lake depths, rock lips, and the mammilated bedrock of the platform surrounding Jetty Lake indicate similar abrasive action to that already described in the Upper Cwm. In neither of these places were any striations found. Other less spectacularly abraded areas likewise lacked such marks of ice movement direction. The conclusion is reached that these metamorphic rocks are characterised by so many structural foliations that it is impossible to separate lineations due to glaciation from those resulting from the surface expression of minor structural features. The Tahune cirque hangs nearly 1000 feet above the Tahune trough headwall. It is a small steep-sided and overdeepened cirque excavated in Scotchfire dolomite under the north face of the Cap. Lake Tahune, although little more than 100 yards across is over 70 feet deep and is contained behind a dolomite rock lip through which it drains by way of solution channels. The rock-lip has a thin cap of bouldery till and protalus.

The Whitham glacier

Lake Whitham lies at 3060 feet in an overdeepened cirque southeast of Philps Peak. Outcropping dolomite at several places around the lake shore indicates selective erosion of Scotchfire dolomite as in the Lake Marilyn cirque. The maximum extent of the Whitham glacier is unknown for the ice cascaded steeply after passing over the glacially moulded quartz-schist rock-lip. This lip is well smoothed to a height of at least 70 feet above the lake level (Plate 3).

The East Philp glaciers

The only other glaciers of the area were the small "precipitation shadow" cirques high on the eastern slopes of Philps Peak massif overlooking the Loddon Plains. One of these cascaded on to the southernmost diffluence tongue of the Vera glacier. The other was situated on the east side of Mt. Agamemnon. Neither were overdeepened enough to contain lakes and much of the material deposited in them is not till but protalus, including protalus ramparts. As with two other cirque-like features nearby they would have held only snow-patches for most of the last glacial period.

GLACIAL DEPOSITS

Thirteen till samples were analysed by sieving, decantation and hydrometer analysis in the C.S.I.R.O. Soil Laboratory in Hobart. The samples brought out for analysis vary in size from 40 lbs. to 2-3 lbs., the boulders having been measured in the field. The small clay content of most samples is explained by the fact that most of the bedrock from which the tills were derived comprises resistant quartzites and quartz-schists, and that most very fine particles are carried away by meltwater in areas of mountain glaciation. Tests for lime content of samples showed that any calcareous materials had been leached out, a condition expected in this area. The results in fact were typical of analysis of tills derived from conspicuously jointed rocks which usually show a high proportion of boulders and pebbles in a matrix of pronounced sandiness.

The differing proportions of quartzite and quartz-schist in tills makes differentiation based in weathering difficult as quartzites remain almost completely undecomposed as boulders in surficial deposits. As a result, the predominantly quartzite till of the lower Vera valley shows very little sign of weathering, while the Vera diffluence pass tills, which may reasonably be assumed to be of similar age, contain weathered schist cobbles and boulders. In contrast to quartzite, dolomite boulders may become completely decomposed as till stones, as found in pits excavated in the Vera retreat moraines.

In considering the significance of weathered till particles the question arises as to how much drift is derived from the preglacial weathered mantle and how much is derived from the mechanical action of ice on bedrock. While the ability of ice to abrade and erode unweathered bedrock is clearly evident in the powerful glacial moulding of stoss and lee features, there is little evidence on which to base an assessment of the amount of material contributed to the drift by pre-glacial weathering. The weathered dolomite boulders mentioned above almost certainly decayed after emplacement in the till as glacial transport in mountain valleys would be unlikely to allow the retention of cohesion in weathered boulders. The preglacial weathering of the till stones in question is rendered further unlikely by their presence in post-maximum stage tills.

GLACIAL LAKES

A classification of lakes (Table 1) of the Frenchmans Cap area is useful in illustrating the glaciation of the area. It should be used in conjunction with a topographic map and all figures.

TABLE I

Type	Name	Remarks
A. Cirque lakes due mainly to glacial erosion and overdeepening.	Whitham Marilyn Tahune	Definite roxen lakes all selectively eroded in Scotchfire Group rocks with a high or 100% dolomite content.
		Other rock basins
	Nancy, Jetty*	eroded in quartz-schists and quartzites
	Gwendolen	a valley head cirque with a low partial moraine dam.
	Cecily	partially moraine dammed—deeply eroded in softer Scotchfire Group rocks
	Sombre Tarn	small cirque lake on valley side now almost filled by sedimentation and protalus.
B. Valley lakes of compound origin due to erosion and deposition.	Vera	glacial overdeepening in Scotchfire rocks at the junction of two glaciers.
	Sophie	valley "cirque" lake—shallowed by debris from higher tributary cirque with greater accumulation area.
	Gertrude Magdalen Millicent	valley lakes possibly marking retreat stage of Livingston glacier. Lake Millicent may be a miniature glint lake (see below).
Other	Marsh Lake Lake Charlotte	Small rock basins on steps below cirques.

* Jetty Lake is eroded in Scotchfire quartz-schist and might be classed as a roxen lake.

Lake Millicent is thought to fit Hutchinson's type 29 (1957, p. 160), in that it is probably a rock-basin resulting from " glacial corrasion in specific sites where the ice flow is impeded by the pre-existing topography ", although the examples quoted (op. cit. pp. 68 and 160), are on a much larger scale than the one in question here. Attention is drawn to the different use of the term " glint lake " by Charlesworth (1957, pp. 250, 265, 285). The term " roxen lakes " (Charlesworth 1957, p. 265), has been employed here for lakes selectively eroded in the dolomite lenses of the Scotchfire Group rocks. It is possible that some other rock basins were initiated in depressions in dolomite which no longer outcrops.

The isobathic maps (Figs. 3 and 4) were plotted from soundings made with the aid of a rubber boat and sounding line. Plotting was done from bearings taken from the boat and from shore. They show that in some cases considerable overdeepening has occurred, especially in the high rock basins, most of which have their deepest points nearer the backwall than the lip, which is of course typical of cirque lakes in general.

ASPECT AND MORPHOLOGY

Although the glaciation of the area was imposed on a small, somewhat isolated massif, with a generally radial drainage pattern, the effects of advantages possessed by south and east facing valleys and cirques for the accumulation of glacial ice can be seen. These advantages are two in number. First, in areas of prevailing westerly winds, south and east facing valleys are likely to accumulate more snow due to both orographic precipitation and snow drifting. However, this advantage is also possessed by north facing valleys in the lee of north-south ridges. Secondly, south and east facing slopes are more protected from insolation (see also Charlesworth 1957, p. 298). It is significant that the largest glaciers accumulated in the lee of north-south ridges while the largest cirque faced south-east on the lee of the Cap itself.

Closely related to this effect is the asymmetry of form shown mainly by the north-south ridges, in particular the Clytemnaestra Ridge including the Cap (Plate 4). As previously observed by A. N. Lewis (1929) this profile of steep lee slopes is typical of most of the glaciated areas of Tasmania. In the Frenchmans Cap area the effect is emphasised in places such as the Clytemnaestra Ridge where the resistant quartzite and quartz-schist has been worn away on the lee sides to expose softer underlying rocks while the quartzite remains to cap the ridge and form the western slopes. Post-glacial processes have tended to preserve and in some cases emphasise this general asymmetry on the crests of the north-south ridges, the very tops of which take on a striking resemblance to the crests of mobile sand dunes due to the erosion caused by the growth of heavy snow cornices along the crests on the lee sides. Referring to the action of such snow cornices created by wind on ridge crests, Charlesworth (1957, p. 299), states that they " tore off astonishing amounts of rock when they broke away ". The prevailing westerlies over Tasmania caused not only the necessary drifting, but also the eventual deflation of fines exposed by melting or washed from under snow. Snow cornices form on the Clytemnaestra Ridge during the present winters and have been known to last regularly into October. They might reasonably be assumed to have been semi-permanent in late glacial and early post-glacial times.

MULTIPLE GLACIATION

The cirques can be divided into two groups on the basis of altitude. Half those mapped in Fig. 2 fall between 2100 feet and 2300 feet and the other half between 2700 feet and 3400 feet. Early work in Tasmania used cirque levels to differentiate between the separate glaciations (Lewis 1922, 1924, 1933). It is now recognised (for example, Jennings and Banks 1958) that there is no reason why cirques at different altitudes cannot contain ice at the same time, and, if they do not, then it is the high cirques that contain ice as a result of retreat of ice from all lower levels. Glacial features in these late stage cirques will therefore be younger than features down valley.

Cirque levels have also been used to determine orographic snowlines within different stages of the one glaciation. However, given the altitude necessary for the existence of solid precipitation the exact siting and level of any cirque is pre-determined by factors of accumulation (aspect, precipitation amount and slope angle) and lithology. Obviously to assume different ages for cirques on the basis of their altitudes is only an introductory exercise to more detailed study. The " roxen lake " cirques and the small high East Philp cirques of the Frenchmans Cap area illustrate the effect of local conditions on cirque glaciation.

Nor do the glacial deposits, which are in general a more promising indicator, provide any evidence of multiple glaciation. Higher tills are in some cases obviously less weathered than tills in lower areas. This indicates nothing other than the usual single phase sequence of cirque glaciers expanding to a maximum as valley glaciers, followed by retreat into the original cirques. No signifcant differentiation of the fluvioglacial sediments is evident. Although the evidence is consistent with with only one glaciation it is well to remember that proof of multiple glaciation in Tasmania may yet be forthcoming (Patterson 1965, Derbyshire et al. 1965, p. 3). The present investigation is essentially a reconnaissance one and, in an area such as this, evidence of multiple glaciation is not quickly available, it being generally recognised as more difficult to obtain in areas of alpine glaciation than in areas formerly covered by continental ice-caps. Ice streams are more likely to erode evidence of previous glaciations because they are confined to the one route of movement. Even where evidence exists, it is hard to distinguish unless by obvious differentiation of weathering because of a general lack of differentiation in texture, lithology and colour of deposits of different ages, arising from this confined flow and the fact that the same rocks are eroded each time. Weathering in mountain areas produces immature and incomplete soil profiles. Further, comparative weathering in mountain areas of high rainfall cannot be used as effectively as in areas of continental glaciation, because glacial materials are generally non-calcareous either in their original state, or because of high leaching rates. In the valleys of the Cap area it is unlikely that glaciers of penultimate and earlier glaciations extended beyond the well marked limits of the last at the Vera, Livingston and Tahune gorges. All drift belonging to these glaciers would then have been redistributed such that if it existed at all it would now be very difficult to distinguish.

Two possibilities remain: either previous glaciations were much slighter than the last or insufficient snow accumulated to form a significant quantity of glacial ice. Gentilli (1961, pp. 478-479) argued on climatological grounds for the existence of glaciations prior to the last while admitting the lack of relics from it. Probably older glaciations were no more extensive than the last, with periglacial processes predominating for much of the time. In this case the resulting landforms would have been destroyed in the interglacial periods by fluvial erosion. This would mean that the latest glaciers formed and advanced upon a fluvial landscape.

Therefore, bearing the difficulties in mind, the conclusion is reached that the Frenchmans Cap area shows evidence of only one glaciation. Present opinion (Jennings and Banks 1958) correlates this glaciation with the Wurm or Wisconsin glacial phases of the northern hemisphere. The suggestion is made that previous glaciations may well have occurred despite the lack of evidence.

Certainly the evidence in the area correlates readily with the most recent glacial phase. Preservation of moraines and other glacial landforms, the generally undeveloped state of drift weathering, the presence of illite in some deposits (Charlesworth 1957, p. 226), the depth and newness of the lakes despite active infilling, all strongly suggest this. The equally fresh evidence of glaciation in the nearby Linda Valley (where radio-carbon dating of fossil wood found in varves puts the initial advance there at approximately 26500 years B.P. (Gill 1956)) is ascribed to the latest glacial phase, the radio-carbon date corresponding with the beginning of the Wisconsin glaciation in North America.

Problems of the last glaciation

The systematic significance of the present study lies in relating it to (i) the general glacial history of Tasmania and (ii) factors affecting ice accumulation in alpine areas in order that the climate changes responsible may be correctly interpreted. To this end some major problems of more than immediate significance are discussed. These are the extent of the last glaciation, the sequence of events in that glaciation, and the dating of these events.

(*a*) *Extent*

The maximum extent of the last glaciation in Tasmania is at present undefined. Beyond the well worked glacial features of the individual glaciated areas lies a zone of uncertainty: cf. Jennings and Banks (1958), map p. 299 and Derbyshire et al. (1965). In mountainous regions like the Cap area, where valleys often descend into gorges, evidence in this outer zone is both scanty and problematical. On the central plateau the limits of the last ice sheet remain unmapped. Indeed some workers point to the possibility that features of the eastern plateau may be referable to an earlier glaciation (Jennings and Ahmad 1957). To the west the great gorges of the Fury, Mackintosh and Canning Rivers belie glaciation but the tills exposed on the Murchison Highway are unlikely to have been deposited by ice of local origin. To the north the picture is as yet far from clear despite its importance to current engineering projects in that area. To the south and east similar problems leave current thought in a state of flux. (Derbyshire 1963, p. 108; Jennings and Ahmad 1957, p. 74).

The uncertainty in defining the maximum extent of the Cap valley glaciers has been mentioned in previous sections. Accordingly Fig. 2 should not be regarded as final. In the light of these difficulties, future consideration of the maximum glacial advance here and in the rest of Tasmania should take account of the following possibilities:—

(1) If the last glaciation attained an early, comparatively brief and extensive maximum, the evidence for this might be obscured by the subsequent periglacial processes.

(2) A distinction must be drawn between the mountain glaciers and the Central Plateau ice cap, which at one time had a potential accumulation area of some 600 square miles or more. In contrast, the area above the firn line in the Frenchmans Cap area cannot, at any time, have been greater than six square miles.

(*b*) *Post-maximum stage sequence of events*

The various retreat stages of the valley glaciers have been dealt with above. The glaciation of the Cap area almost certainly involved a synchronous valley glacier maximum stage that must have grown from, and must have retreated to, the high cirques. It was not possible to date the valley retreat stage. Perhaps future work in this or other parts of Tasmania will provide details of the fluctuations of the last glaciation. If the New Zealand evidence can be taken as an indicator of Southern Hemisphere climate changes in general, a sequence of three glacial advances might be expected with the last retreat beginning some 14000 years B.P. after the Kumara 3 and Poulter advance of the Otira glaciation of South Island (see Suggate 1965, p. 84).

Present work so far in the Frenchmans Cap area has yielded only one datable deposit. A radiocarbon assay of wood and leaf fragments taken from rock flour (mainly (unweathered) quartz and mica particles according to X-ray analysis—McIntyre pers. comm.) and sediment beneath $1\frac{1}{2}$ metres of peat in a small bedrock basin on the lip of the Nancy cirque at 3400 feet gave a date of 8720 ±220 years (before 1950) (Kigoshi 1965 pers. comm.). This date, which can be regarded as a minimum age for the deglaciation of the high cirques of the Frenchmans Cap area in particular and the western Tasmanian mountains in general, is the closest to a true date for deglaciation of the western high cirques presently available. Among the vegetal fragments dated were branchlets of *Microcachrys tetragona* which today grows mainly in the higher more exposed parts of the Cap area.

CIRQUES, SNOWLINES, AND THE GLACIAL-AGE CLIMATES

Among the methods used for quantifying Quaternary climate change is that taking cirque levels as indicators of glacial age snowlines and temperatures. Davies (in Gill 1962, p. 245) and Galloway (1965, Fig. 2) have spoken briefly, and in general terms, of the Tasmanian snowline while at the same time mentioning the problems involved in defining it. The difficulties are due to the east-west climatic contrast across Tasmania and the complications inherent in assessing the inter-relationships of temperature and precipitation changes (Galloway op. cit., p. 609).

It may be assumed that snowline levels change with progressive deterioration and amelioration of climate during glacial times. It is important when suggesting former snowline elevations and the corollary climatic inferences to define the stage of glaciation to which it refers. Maximum stage snowlines might be deduced from climatic data derived using Ahlmann's curve (1948): evidence of former lower limits of solifluction and tree lines being used to help establish temperature while modern climatic records of temperature and precipitation (ablation season length, &c.) may also be used.

Cirque stage glaciation snowline levels on the other hand are orographic snowlines, and, as mentioned above, this is strictly a function of climate and lithology (page 126). Wind and precipitation shadow effects can be seen in both individual mountain ranges and across the whole Island. Cirque glaciation, except as part of an advancing hemicycle, is essentially a late-glacial or "post-glacial" event. Late-glacial cirque ice is very sensitive to climatic factors, and, at least as investigations to date are able to ascertain, existed mainly in the high discrete cirques which are forming the subject of further investigation.

SUMMARY OF GLACIATION

The following sequence of events for the last glaciation is suggested. Climatic deterioration resulted in the accumulation of snow and the formation of snow patches in the depressions of a fluvial landscape, probably between 30000 and 26000 years B.P. The early accumulation hollows were provided by larger solution features in dolomite areas and by structural weaknesses elsewhere. Cirque glaciers formed, ice continued to accumulate, and valley glaciers advanced to a maximum. All this time the highest peaks and ridges stood above the glacial ice as indicated by the evidence of prolonged frost shattering.

In the Vera valley, glacial ice accumulating in the Marilyn and Barron cirques advanced down valley and over the low dolomite and quartz-schist ridges where the Vera Plains now stand, and discharged over five diffluence passes at the east end of the plains and also down the lower Vera valley to the Vera gorge, before which it was joined by the Pickaxe and Davern glaciers flowing down short, steep gradients from heads on the north side of the Nicoles Needle Ridge.

At the maximum stage the Livingston glacier reached the gorge above the Franklin River junction possibly over-spilling the low section of the east trough wall (Deception Ridge). Small cirques formed above the west side of the glacier in the lee of the west trough wall (Clytemnaestra Ridge). The Tahune glacier had reached its maximum extent at the Tahune Creek gorge and may have discharged a small ice tongue over the diffluence pass beneath Sharlands Knob. Meanwhile, the Gwendolen and Sophie glaciers were cascading down on to the Ness Creek Plains. The Whitham glacier had overflowed its cirque and the east Philp cirques contained active ice at this stage. The valley constrictions and diffluence passes and the hanging valley snouts that characterised the glaciers at this stage were unfavourable for the building of well marked terminal moraines, and much of the debris was carried away by the constricted and fast-flowing meltwaters. These were responsible for the valley trains of the Loddon Plains and Ness Creek Flats.

The post-maximum stands are better marked in some valleys than in others. In the Vera valley the glacier thinned at the point of greatest constriction, so that retreat took place rapidly to this point. Ice retreated from the diffluence passes and a major stillstand was recorded in the main Vera moraine. After a slight advance that emplaced a push moraine on top of the main moraine, retreat with six halts took place until all that remained of the glacier was in the Barron and Marilyn cirques. The former cirque contains a moraine testifying to this halt in the retreat.

The retreat stages of the Livingston glacier are less obvious but it seems reasonable to postulate a retreat from the maximum extent to the lower end of Lake Gertrude with a major halt across the deposits separating Lakes Millicent and Magdalen. The retreat of the Tahune glacier is not marked by any one moraine ridge although there is ample drift on the floor of the valley. The final stages saw the ice retreating into the Tahune, Jetty and Nancy cirques. The moraine at the mouth of the Tahune cirque contains a large proportion of protalus from the north face of the Cap. The lack of moraine in the high cirque is as much due to abrasion having reached below the zone of open weathered joints as to a postulated brief late-glacial cirque phase.

The Sophie and Gwendolen glaciers retreated into their flat hanging troughs from their steeper lower valleys. The former, being smaller and therefore more sensitive than the other, built a series of low moraines near the trough mouth while the Gwendolen glacier left a more uniform, though also ridged, till cover. The Sophie glacier retreated into the upper Cwm from which limited outwash was derived for the deposit of a small delta in Lake Sophie immediately below Lake Charlotte.

The Marilyn cirque glacier left a shallow protalus cover on the lip. The Whitham glacier retreated to its cirque and melted without leaving a moraine. The absence of debris is explained by the lower headwall.

After the glacial maximum the East Philp cirques contained only stagnant ice or neve and behaved generally as nivation cirques or snow patch hollows. Protalus ramparts formed in the northern of these cirques as well as in the neighbouring nivation hollows.

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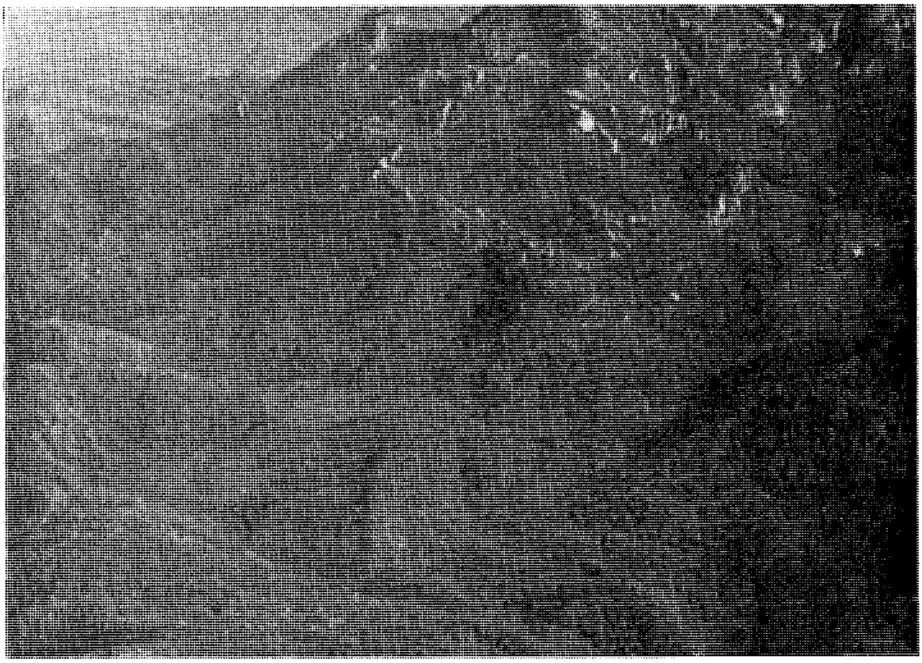

PLATE I.—Aerial view of the Vera Retreat moraines. Diffluence pass middle-left, main Vera moraine with push moraine crest lower centre, recessional moraines on proximal moraine edge. Hanging cirque containing Lake Marilyn upper right. Lake Vera middle right.

PLATE II.—Aerial view from west of Clytemnaestra looking north-northwest, across the Sophie cirque valley into the Upper Cwm. Cap summit upper right.

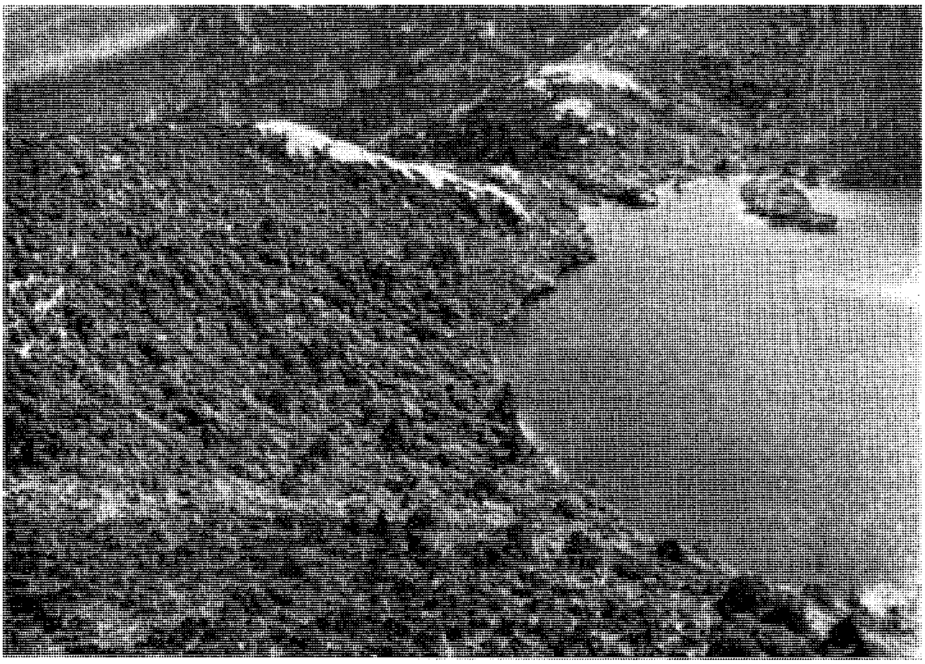

PLATE III.—Abraded rock lip of Lake Whitham cirque. Looking southwest.

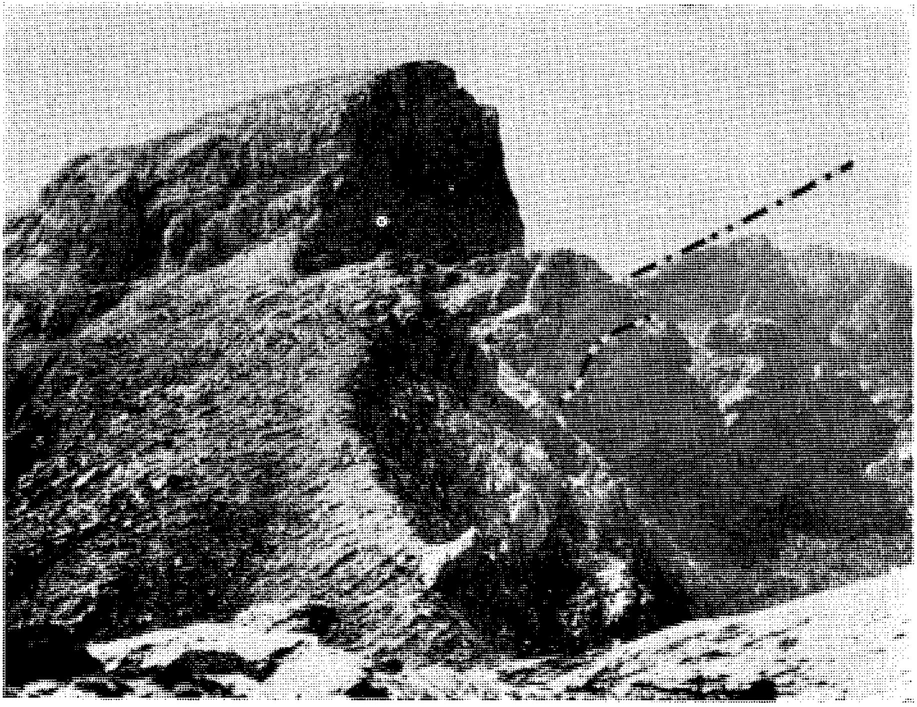

PLATE IV.—Asymmetrical ridge and lee side excavation. View from the summit of Clytemnaestra looking north to the Cap summit. Note quartzites overlying darker coloured dolomites, schists, and phyllites of the Scotchfire Group.

[Plate II.—Geography Department, University of Tasmania. All other photos by the author.]