CAINOZOIC VOLCANISM IN AND AROUND GREAT LAKE, CENTRAL TASMANIA

By

F. L. SUTHERLAND

(*Tasmanian Museum, Hobart*)

and

G. E. A. HALE

(*Hydro-Electric Commission, Hobart*)

(With five text figures and four plates)

ABSTRACT

Upper Cainozoic basaltic volcanism about Great Lake involved the eruption of a succession of mineralised entrail breccias, 215+ feet (65 m), aquagene tuffs and agglomerates, 40+ feet (12 m), unmineralised entrail breccias, 160+ feet (48 m), and massive flows and dykes, individually up to 200+ feet (60 m) thick with sequences up to four flows and 300+ feet (90m) thick. Associated with the volcanics are some lacustrine and fluviatile sediments, up to 88+ feet (27 m) thick.

The aquagene pyroclastics and entrail breccias are confined within the present Great Lake depression, and closely resemble hyaloclastites and bedded breccias in the upper parts of Icelandic intraglacial pillow lava piles. They probably represent emergent elongate fissure volcanoes that erupted into past high water levels in Great Lake.

The massive sub-aerial lavas erupted from centres both within and outlying the Great Lake depression; those within probably erupted during low or drained water levels.

Over twenty eruptive centres can be inferred on structural and petrological grounds and most are aligned along intersecting NW, NNW, N, NNE and ENE lineaments. There is some evidence of late or post-volcanic local tilting and jointing and of recent adjustment movements on lineaments.

The bulk of the volcanic rocks are tholeiitic olivine-basalt, but some tholeiite and alkali olivine-basalt occurs amongst the massive lavas. The Great Lake volcanic association is a typical example of the tholeiitic associations of Tasmania and falls within a general belt of such rocks extending from far NW Tasmania to the Derwent Valley. The Great Lake rocks resemble to some extent basalts of the Hawaiian province, and the known stratigraphy suggests a somewhat similar pattern of magmatic evolution and eruption.

INTRODUCTION

Drought conditions in Central Tasmanian catchments during 1967-68 lowered the level of Great Lake from above 3,378 feet to a low at 3,346 feet on 29 March 1968 (fig. 1), almost returning the lake to its original size and configuration prior to damming in 1923 (3,333 feet; Legge, 1904; Lewis, 1933). Large peripheral areas of the lake bottom became exposed, providing some excellent, cleanly washed outcrops of basaltic rocks. A study of these combined with detailed mapping of basalt outcrops inland, has deciphered much of the volcanic history of the region, and is reported in this paper.
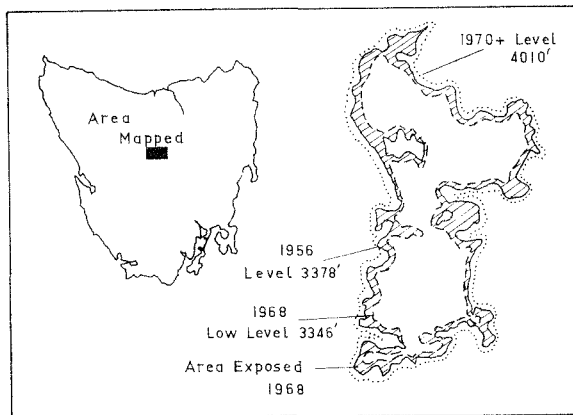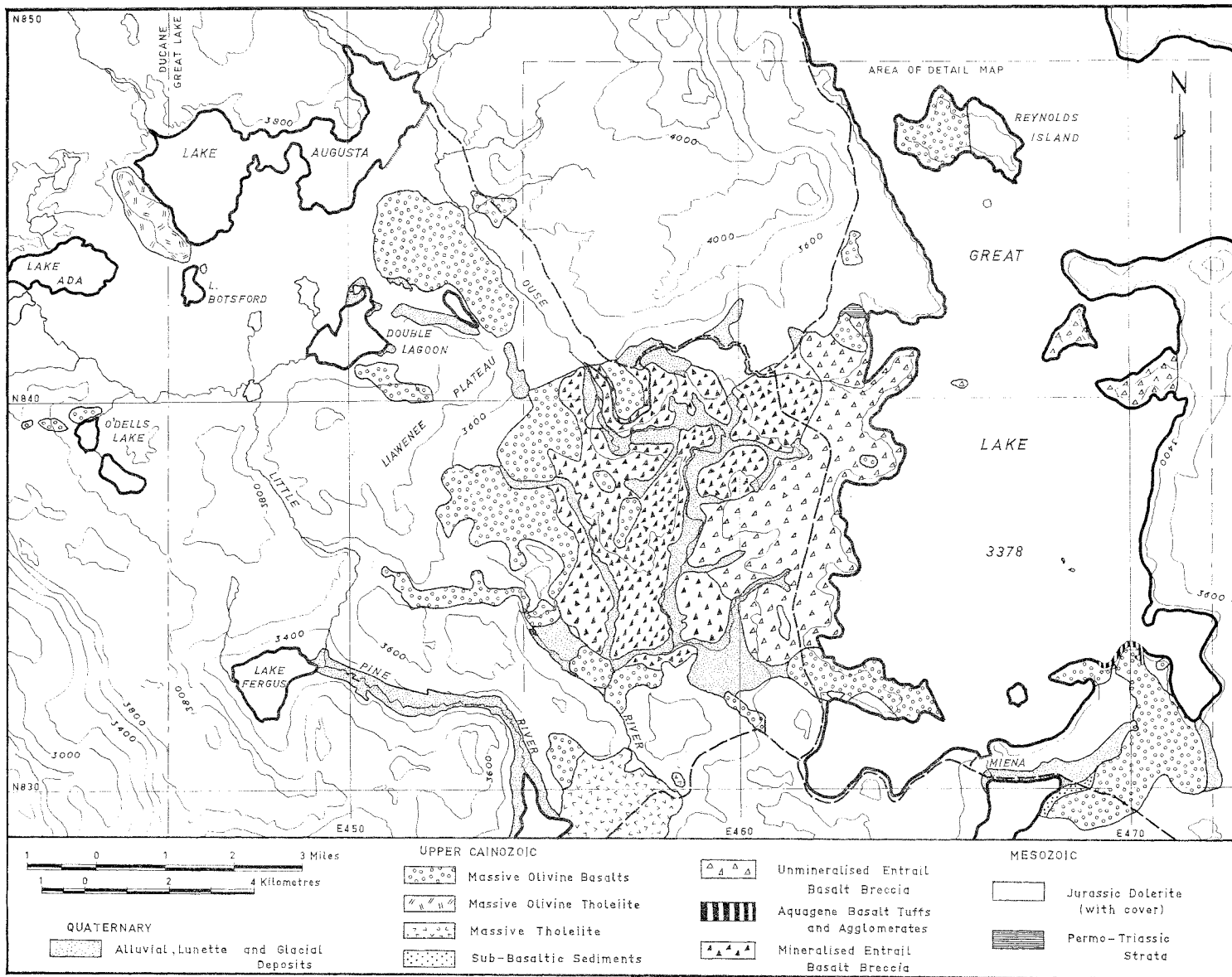

FIG. 1.—Great Lake locality and levels.

Following initial examinations of the new exposures by G. E. A. Hale in February 1968, detailed geological mapping, sampling and microscopic petrographic studies were commenced by F. L. Sutherland in March and extended inland by 1969 to cover all the basalt outcrops in the Great

Fig. 2.—General geology, Great Lake area.

UPPER CAINOZOIC

- Massive Olivine Basalts
- Massive Olivine Tholeiite
- Massive Tholeiite
- Sub-Basaltic Sediments

- Unmineralised Entrail Basalt Breccia
- Aquagene Basalt Tuffs and Agglomerates
- Mineralised Entrail Basalt Breccia

MESOZOIC

- Jurassic Dolerite (with cover)
- Permo-Triassic Strata

QUATERNARY

- Alluvial, Lunette and Glacial Deposits

1 0 1 2 3 Miles
1 0 2 4 Kilometres

Lake Quadrangle and eastern part of the Du Cane Quadrangle (fig. 2). Samples and sections of the rocks are housed in the Tasmanian Museum collections. Chemical analyses of selected basalts were carried out by the Tasmanian Department of Mines Laboratories, Launceston.

Great Lake is refilling to 3,391 feet and with the completion of current H.E.C. works its planned level will be 3,410 feet, re-covering many of the exposures described in this paper. Photographic plates of some of the outcrops fortuitously exposed by the drought low level are included here, and an aerial photographic coverage of Great Lake flown on 3.7.1967 by the H.E.C. shows much of the bared lakesides.

PREVIOUS LITERATURE

Brief geological observations were given by Legge (1904) in his physiographical account of Great Lake and he illustrated the basaltic cliffs at the Beehives. Lewis (1933) summarised the general geology of the area, referred to geological comments of earlier workers, photographed basaltic cooling columns at 'The Battery' and suggested a glacial origin for Great Lake. Edwards (1939) discussed the physiography and age of the basalts in the area and considered the basaltic plain on the Liawenee Moor to be a 'distinctly youthful' lava surface.

In the first comprehensive regional mapping and geological account of the area, Voisey (1949) gave some details on the basalt sequences and suggested that Liawenee Moor resulted from action of an ice sheet and was not an original lava surface.

Further regional mapping and sub-surface exploration by the H.E.C. and regional mapping by the Tasmanian Mines Department produced the 1 inch to 1 mile Great Lake and Du Cane Sheets (Blake, et al., 1956; Jennings, et al., 1961; MacLeod, et al., 1961). This provided the data for the discussions of the Jurassic dolerite and its structure, reported by Jaeger and Joplin (1955); Joplin (1957); Wiebenga and Polak (1957, 1961, 1962); Wiebenga (1958); Jaeger and Green (1958); McDougall (1958, 1964); Carey (1958); Jaeger (1964); Sutherland (1966) and most recently by Jones, Haigh and Green (1966) following a regional geophysical survey.

Amygdale minerals in the basalt breccias at Liawenee Canal were listed by Sutherland (1965). Sutherland and Corbett (1967) compared these with the amygdale minerals and basalt breccias of far NW Tasmania and suggested an aqueous origin for the Liawenee breccias.

Glacial features in the vicinity of Great Lake were mapped and discussed by Derbyshire et al. (1965, 1966), and Derbyshire (1968) presented evidence of older glacial deposits in the area.

STRATIGRAPHY

General

A flat basin-like intrusive sheet of Middle Jurassic dolerite underlies the Great Lake region, with minor roof remnants of Permo-Triassic strata (Jones, Haigh and Green, 1966). Further probable Permo-Triassic outcrop was mapped in this study above Canal Bay, W Great Lake, where cherty hornfels strata, at least 50 feet (15 m) thick, dip shallowly SW.

Great Lake occupies the dolerite depression, probably shaped largely by differential erosion, faulting and tilting, and glacial action to its present form.

The Cainozoic basalts and associated sediments are mostly confined to the Great Lake depression and its drainage along the Ouse and Shannon Rivers, but there are scattered basalt outcrops on the western plateau margin up to 7 miles (11 km) inland. Geophysical investigation indicated Quaternary sub-basaltic sediments at Tods Corner (Wiebenga and Polak, 1962), and H.E.C. drilling proved similar sediments in NE Great Lake. Other, scattered sedimentary deposits of probable Quaternary age within the Great Lake depression are described by Jones, Haigh and Green (1966).

The basaltic rocks include pyroclastics, entrail breccias and massive lavas, and the probable approximate succession in increasing age is:—

Massive flows and dykes 300+ feet (90 m).

Unmineralised entrail breccias 160+ feet (48 m).

Aquagene tuffs and agglomerates 40+ feet (12 m).

Mineralised entrail breccias 215+ feet (65 m).

Post-basaltic deposits include Quaternary glacial, peri-glacial and talus deposits, and fluvial and lacustrine alluvium (Legge, 1904; Lewis, 1933; Voisey, 1949; Derbyshire, 1968).

Sub-basaltic Sediments

Unconsolidated sediments underlie superficial surface deposits in a marshy depression SE of Tods Corner and occupy a trough in the dolerite bedrock, probably fault bounded to the SW (inferred from seismic profiles, Wiebenga and Polak, 1962). They lie below massive basalt S of Tods Corner, are baked near the contact and range to at least 88 feet (27 m) thick, extending from 3,367 to 3,333 feet in the NW and from 3,492 to 3,402 feet in the SW. They are considered to be mainly water-saturated clays with gravels, representing Cainozoic fluvio-lacustrine deposits. A small area of cherty hornfels near the basalt base SE of Maclanachans Point, at 3,350 feet, and baked conglomeratic beds below basalt at 3,400 feet with a marshy depression at 3,350 feet E of Shannon Lagoon, probably represent extensions of these beds.

Further sediments of similar age around Great Lake may include the siliceous conglomerate on Cramps Peninsula, considered a lithified river gravel by Jones, Haigh and Green (1966), and probable lacustrine beds intersected by H.E.C. drilling at Poatina Tunnel Inlet. The latter extend from 3,330 to 3,003 feet (41'-68' 3" depth, D.H. 5127) with a downward succession of:—

   5 feet (1.5 m), black, soft to very soft clay, with a few sandy (tuffaceous?) lenses;

   18 feet (5.4 m), light grey, soft, friable, laminated sandstone, with numerous plant remains and consistent 20° dip;

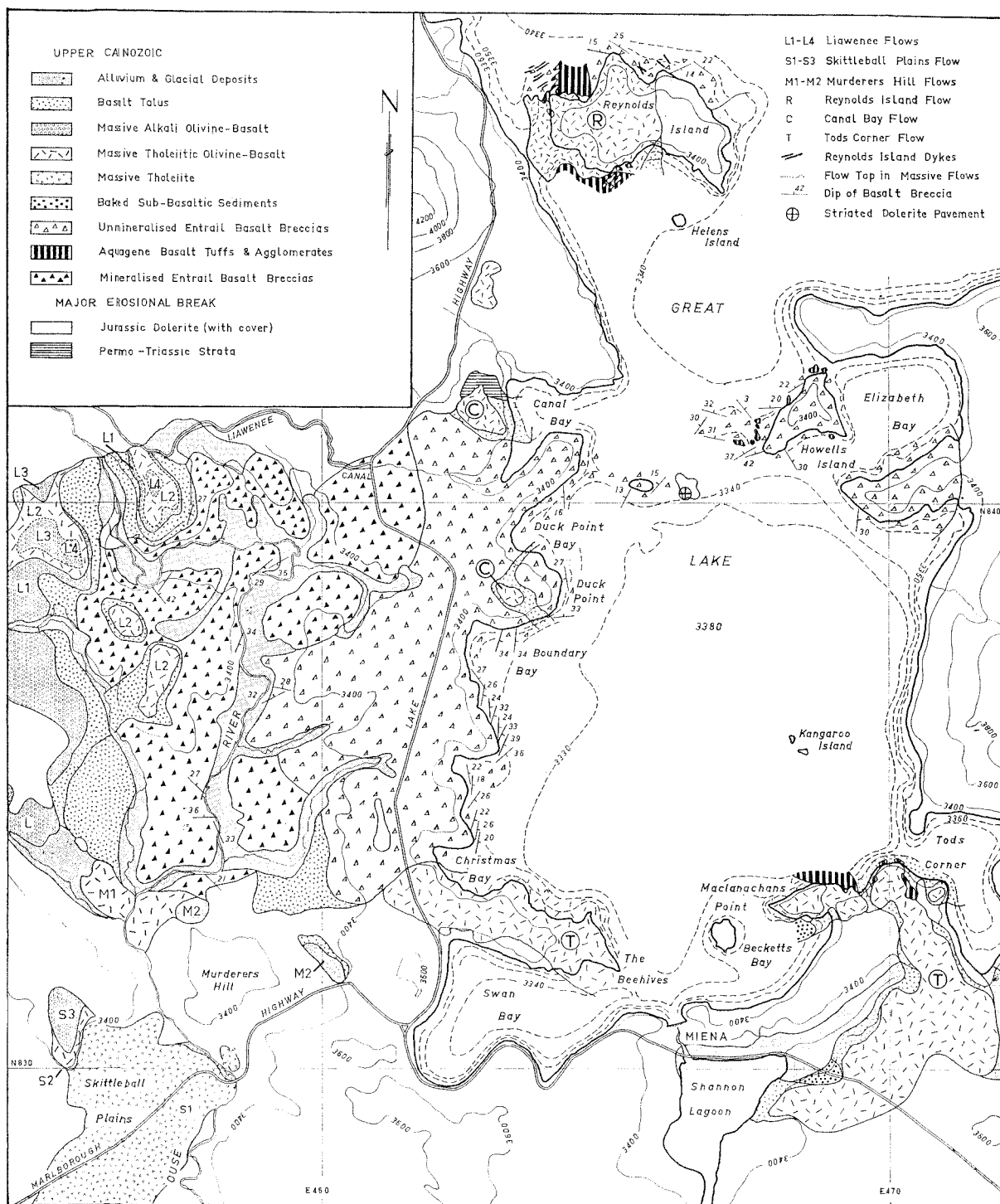   2 feet (0.6 m), light grey, friable, massive (tuffaceous?) sand;

FIG. 3.—Detailed geology, volcanic rocks, Great Lake area.

2 feet 3 inches, light yellow brown, very fine sandstone, with dips at 35°, and limonitic staining near base, on dolerite bedrock.

Mineralised Entrail Breccias

This series outcrops from Liawenee Canal in the north, W of Ouse River to Murderers Hill in the south (fig. 3). It is at least 215 feet (65 m) thick, marginally overlaps Jurassic dolerite and extends to unknown depth below Ouse River. Exposures are poor except in Liawenee Canal and in cliffs on Ouse River.

Successions of entrail lavas, largely brecciated, and embedded in a fragmentary rubbly and tachylitic matrix are heavily mineralised with amygdaloidal fillings and inter-fragmentary cement. Their structure is essentially similar to the unmineralised entrail breccias to the east, whose internal forms are unobscured by secondary mineralisation and are described in detail later. The secondary minerals (Sutherland, 1965) show a general paragenetic sequence, inferred from numerous partial amygdaloidal sequences, in depositional order of calcite (usually massive), phillipsite (rare), chabazite (Analysis 5, Table 1), an association of tacharanite, tobermorite, opal and nontronite, followed by apophyllite and finally further calcite.

Dips up to 42° in the breccias suggest a series of dissected cones, mostly between 1 to 2 miles (1.5-4.5 km) across, erupted from about six centres aligned dominantly NNE which coalesced to form elongate fissure volcanoes (figs 3 and 4). They are amongst the oldest extrusions in the area, being disconformably overlain by massive lavas at Liawenee and overlapped by unmineralised breccias E of the Ouse River, but their precise age is unknown.

Aquagene Tuffs and Agglomerates

Bedded pyroclastics occur around Reynolds Island, disconformably underlying unmineralised entrail breccias and massive lavas, around Howells Island, interbedded with and disconformably underlying the breccias, and between Maclanachans Point and Tods Corner, disconformably underlying massive lava (fig. 3, plate 2). Their base is nowhere exposed. They are almost wholly basaltic, with sporadic pieces of country rocks, and show structures consistent with shallow water deposition. Vitric tuffs predominate and in thin sections show palagonitised basalt glass (Appendix 1) suggesting a sub-aqueous eruptive hyaloclastite origin, but the agglomerates may include some sub-aerial ejecta.

Reynolds Island

Outcrops from 3,385 to 3,345 feet indicate a total pyroclastic thickness greater than 40 feet (12 m). Exposures are poor in the NW, but good outcrops up to 7 feet (2 m) thick lie under basalt cappings in the SW (plate 2, fig. 1). Here, alternations of dark to light grey, finer compact and coarser friable vitric tuffs, weathering yellow, contain rarer lapilli tuffs and agglomerate in the top 3 feet (plate 2, fig. 2). Bedding is commonly irregular and wavy, and ranges from laminae about 0.5 mm thick in the fine tuffs to 30 cm thick for the agglomerate. Small-scale current ripple and tabular cross-bedding is common in some of the beds, suggesting current flow from the north.

The agglomerate contains scoriaceous basalt in angular to sub-rounded fragments, up to 5 cm and in rarer rounded bombs up to 10 cm across, embedded and grading into a lapilli tuff matrix, weathering reddish yellow. It includes sporadic angular to sub-rounded pieces of Jurassic dolerite and massive basalt, with some larger basalt bombs, up to 45 cm across; similar pieces occasionally occur in the other finer beds.

The beds are gently folded and warped, and generally strike between 15°-150° with dips mostly between 6°-15°, but locally steeper. They are cut by prominent, wide-spaced steep cross-joints, with two main sets trending at 125°-130° and 25°-30°, and small dislocations show throws of several cm (plate 2, fig. 2). The eruptive vent was apparently located off the NW end of Reynolds Island, and probably also erupted the overlying entrail breccias and massive lavas.

Howells Island

Fine and coarse vitric tuffs, at least 7 feet (2 m) thick, form small isolated pockets around Howells Island from 3,380 to 3,350 feet. They resemble the Reynolds Island occurrence, but lack the coarser lapilli tuffs and agglomerate.

They interbed with entrail breccias SW of Howells Island, and show a general strike of 115°-120° and dips up to 12° NNE. Slump and intrastratal deformational structures are common in some of the finer beds, which have a glassy baked appearance and a brittle conchoidal fracture (plate 2, fig. 3). Further north, the beds underlie the breccia, show some inverse grading and scour structures, and are more strongly warped, contorted and dislocated (plate 2, fig. 4). At one place overturned beds dip sub-vertically east, with cross-faults throwing up to 7 feet (2 m), and probably represent slide structure (plate 2, fig. 5). Such structures indicate deformation and dislocation of unconsolidated beds by emplacement of entrail breccias. The general dips suggest an eruptive source to the south, probably the same centre that erupted the entrail breccias.

Maclanachans Point

Bedded tuffs and agglomerates, weathering yellowish or reddish ochre in colour, disconformably underlie massive basalt, inland from 3,455 feet to below 3,345 feet on Great Lake shore (plate 2, fig. 6) with a maximum exposed height of almost 25 feet (8 m). The stratigraphy is complicated by folding and lateral variations, but a measured sequence from top downwards gave:—

10 (+) feet (3 m)—agglomerate

10 feet (3 m)—tuffs and agglomerate,

grading from up to 3 feet agglomerate at base into lapilli tuffs interbedded with and passing into fine grained vitric tuffs at top.

3 (+) feet (1 m)—fine to medium grained vitric tuffs.

The vitric tuffs form compact, conchoidally brittle, dark grey to chocolate red, fine grained, irregular laminae, 0.5 mm to 1 cm thick, alternating with coarser, more friable beds up to 10 cm thick. They show small scale current ripple cross bedding and slump structures.
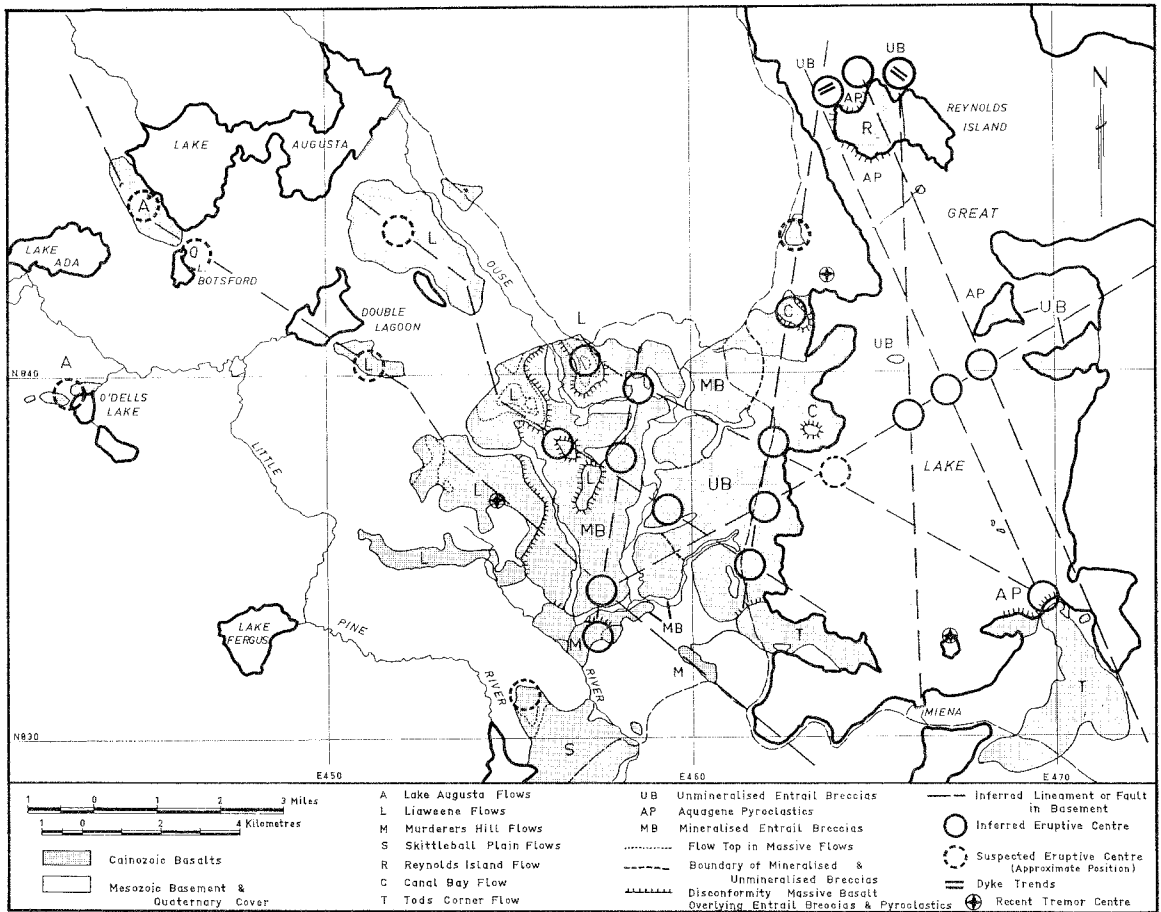
FIG. 4.—Structural geology, volcanic rocks, Great Lake area.

The lapilli tuffs contain numerous small fragments up to 0.6 mm and rare small bombs up to 4 cm across of scoriaceous basalt, in a finer vitric tuffaceous matrix. The agglomerates are similar, but scoriaceous basalt forms 20%-25% of the rock as fragments between 0.5-2.5 cm across and in larger bombs generally 5-10 cm across with hollow centres. In places basalt fragments form stringers along the bedding and there are sporadic angular fragments of Jurassic dolerite, gritty sandstone and baked cherty sediment (?), up to 10 cm across. The coarser beds show some tabular medium-scale cross-bedding.

The beds generally strike between 150°-200° and are gently folded about NNW to NNE axes; they dip at 8°-45°, but locally near-vertically. Dislocations in the beds sometimes contain reddish glassy clay, and sparse, steep blocky cross-jointing shows a dominant NNE trend. The coarse agglomerates suggest a source in the immediate vicinity.

Unmineralised Entrail Breccias

Crudely bedded entrail lavas and breccias outcrop around Reynolds Island, Howells Island and Elizabeth Bay and between Canal and Christmas Bays, W to the Ouse River (fig. 3). The series is at least 160 feet (48 m) thick, disconformably overlaps aquagene pyroclastics, mineralised entrail basalt breccias and dolerite bed rock, and disconformably underlies massive basalt flows at Reynolds Island, Canal Bay, Duck Point and Christmas Bay (fig. 3).

The beds are successions of flow units, individually up to 90 feet (30 m), but mostly between 20-60 feet (7-20 m) thick. Each unit has a base of rubbly, scoriaceous entrail basalt breccia 3-15 feet (1-5 m) thick, passing up into complete entrail lava forming the lower third to half of the unit. The upper and thicker part of the units grades into brecciated entrail lava, becoming increasingly rubbly and scoriaceous towards the top, and inter-

mingling with a finer fragmentary tachylitic matrix. Their bedding dips from near-horizontal to about 45°, but is mostly 10°-35°. Dips are difficult to measure accurately within the units and true regional dips are probably slightly less than the maximum local dips shown on the map.

As in the mineralised series, the measured dips suggest lava cones erupted from a number of centres, which coalesced in many cases into elongate fissure volcanoes. The existence of at least eight eruptive centres is inferred, aligned along intersecting lineaments, dominantly NNE and ENE in trend, with lesser NW and N trends (fig. 4). At some localities, swings of flow direction up to 90° up the succession, indicate effusion from elongate eruptive centres. Dips in the Reynolds Island exposures, although generally suggesting narrow elongate WNW to ENE trending fissure cones, show some complexities, possibly due to later dyke intrusion and massive basalt extrusion.

Sparse, steep open joints cut the breccias and are prominent SE of Canal Bay, where irregular sets trend mostly NW-WNW and NE-NEN.

The basalt in the breccias is tholeiitic olivine basalt with an abundant dark glassy matrix (Appendix 1). In the tachylitic parts the glassy base is generally palagonitised and alters to nontronitic clay. Similar basalt also forms the underlying aquagene pyroclastics and the mineralised entrail breccias.

Internal Lava Forms

The complete entrail lava consists of heaped, moulded, sinuous, sometimes anastomosing tubular basalt entrails, with dark, rough, irregular, cracked tachylitic outer crusts (plate 3, fig. 1). The entrails show slight swellings along their length and individually can be traced for tens of feet. They are circular to oval to elongately flattened in cross-section, ranging from about 9 inches (25 cm) across to 30 feet (10 m) wide in the flat, flow-like bodies. They show rough radial jointing and the tachylitic crusts are 1-8 cm thick, sometimes forming internal multiple sets.

The basalt in the entrails is dense and sprinkled with small rounded to elongate irregular vesicles. These rarely form radial or concentric trains and are mostly confined to and become increasingly abundant towards the outer margin of the entrails (plate 3, fig. 4). Central cavities in the entrails, sometimes multiple, commonly mimic in cross-section the form of the entrail, or may show flat-floored domical shapes; they form conspicuous caverns in the wider flattened bodies (plate 3, fig. 1). Vesicles and cavities are always empty. Ropy flow structure commonly forms within the larger cavities (plate 3, fig. 5), but is lacking or only incipiently developed on exterior surfaces. The basalt contains very rare inclusions of baked cherty sediment and quartz up to 10 cm across.

The brecciated lava (plate 3, fig. 2) consists of numerous, often disorientated, broken segments of entrail up to 6 feet (2 m) long. These are embedded in variable proportions of smaller, commonly angular entrail fragments, mostly 5-10 cm across, amidst a friable matrix of angular tachy-litic fragments up to 2 cm across set in a clay cement. The larger entrail pieces are rare towards the tops of the flow units which are largely rubbly, with smaller, often scoriaceous fragments in abundant fine matrix. Internal, commonly multiple, tachylitic cooling crusts are common in the broken entrails, associated with cracks in the outer crust (plate 3, fig. 4). Some cracks extend into interior cavities producing chilled tachylitic and scoriaceous internal zones around cavity margins. More rarely cracks may contain tachylitic tongues of lava welled up from entrail interiors. Vesicularity also generally increases in the brecciated entrails and may produce completely scoriaceous interiors. Isolated, bulbous bodies, mostly 1 to 3 feet (0.3-1 m) across, occur sporadically through the entrail breccia, and show cracked but finely striated tachylitic crusts (plate 3, fig. 3). These may represent broken and possibly tumbled entrail toes. Many of the above features suggest brecciation more or less contemporaneous with entrail extrusion.

Age

The fresh, unaltered appearance of the rocks, with empty vesicles and cavities, suggests a comparatively young age. A small remnant of a dolerite pavement dips 4°-5° W under the breccias SE of Canal Bay at 3,345 feet, and bears striations (trend 93°-97°, plate 1) suggestive of glaciation and hence a post-Tertiary age.

The degree of subsequent erosion of the breccias and capping lavas, with evidence of glaciation of some of the later massive basalts around Great Lake, suggests that they pre-date the last late Pleistocene glaciation (about 26,500 years, Derbyshire, et al., 1965). The underlying striated pavement, thus, may relate to deposits of an earlier glaciation, evident 3 miles NW and 10 miles SW, and probably significantly older than 30,400 years (Derbyshire, 1968).

Origin

The entrail bodies are identical to pillows in intraglacial Icelandic lava piles illustrated by Jones (1968, plates 1 and 4; 1969, plates 1 and 2) and considered by him to form by digital advance in the manner of slow-moving pahoehoe flow. In overall structure the entrail breccias correspond closely with the equivalent foreset bedded breccias and foreset flow foot breccias (Saemundsson, 1967; Jones, 1966, 1968; and earlier workers), considered by them to result from eruption of sub-aerial flows into water. Such an origin adequately explains sub-aqueous features in the Great Lake breccias, such as their pillow-like form and association with aquagene pyroclastics, as well as sub-aerial features they display such as strongly scoriaceous flow tops and bottoms and incipient external pahoehoe flow structures. Again, the entrails in their degree of vesicularity, general paucity in interstitial micro-vesicles and general absence of vesicle zoning, resemble type 2 pillows (Jones, 1969, plate 3), considered to result from partial sub-aerial degassing of lava prior to aqueous immersion. This, however, precludes the use of their vesicle size and density as an accurate indicator of water depth of emplacement.

The succession of aquagene pyroclastics, bedded breccias and massive flow cappings at Great Lake greatly resembles the upper parts of the Icelandic intraglacial volcanoes and probably those of many marine basaltic volcanoes (Jones, 1966, 1968; Saemundssen, 1967). During growth, these volcanoes erupted their products into shallow water producing phreatic bedded hyaloclastites, but on becoming partly emergent formed foreset bedded breccias, and finally the eruptions were sub-aerial yielding pahoehoe flows. The Great Lake aquagene pyroclastics and overlying entrail breccias were therefore probably erupted into a shallow water environment. The eruptions were presumably rapid, preventing accumulation of intercalated sediment as found in the pépérite lacustrine volcanoes (Jones, 1969). The capping sub-aerial flows at Great Lake, however, mainly erupted after considerable dissection of the pyroclastics and breccias and are a separate phase. The possibility that the Great Lake pyroclastics and entrail breccias represent eruption directly into glacial ice rather than a water body needs consideration, but their composite structure differs from that of described true sub-glacial eruptions (Walker and Blake, 1966; Saemundsson, 1967).

The origin of the entrail breccias is important in reconstructing the form and size of the original structures. Reconstructions based simply on bedding dips give cones built to heights mostly between 1,000 and 2,500 feet (300-800 m). This poses problems in the degree of subsequent erosion for such relatively young structures. However, if the breccias result from sub-aerial flow into water, then dips in the breccias cannot be extrapolated upwards, and the overlying lavas may have had much shallower slopes. Such table mountain structures (Jones, 1966; Saemundsson, 1967) need not exceed a few hundred feet in height.

Outcrops of the aquagene pyroclastics and entrail breccias to 3,450 feet level suggest aqueous inundation to at least this level, although folding in the pyroclastics presents some uncertainties. An aqueous stand in Great Lake near this height may correlate with the surface cut in the older mineralised breccia series and preserved under the younger massive lava cappings at about 3,500 feet. If so, then the unmineralised breccia piles probably stood little more than 250 feet (75 m) high. Change into a strongly emergent sub-aerial phase on continued eruption was probably prevented by sympathetic rises in general water level as eruptions filled the Great Lake depression.

The older mineralised entrail breccias presumably originated under similar shallow water conditions and may overlie aquagene pyroclastics. This suggests a previous stand in Great Lake with volcanoes built to heights of over 215 feet (65 m). The mineralisation presumably took place under conditions of immersion during subsequent water stand in Great Lake, accompanied by regional heating during eruption of the younger series.

Liawenee Moor, thus, was probably produced by post-volcanic erosion undermining relatively unconsolidated breccias in table mountain structures capped with solid flows. Past higher lacustrine erosion, possibly associated with seasonal surface ice action (Legge, 1904), may have been responsible as easily as action of an actual ice sheet.

Massive Flows and Dykes

Massive basalt lavas and associated dykes are widespread on the western and southern regions around and inland of Great Lake (fig. 2). Voisey (1949), Blake, et al. (1956) and Jones, Haigh and Green (1966) also map basalts on the S bank of Pine River Valley and on the north side of Pine Lagoon. However, detailed field and petrological microscopic examinations of these rocks suggest that they are merely locally prominent outcrops of Jurassic dolerite bedrock (Appendix 1).

The basalt flows are up to 200 feet (60 m), but mostly 20-100 feet (6-20 m) thick, either in single extrusions or successions up to four flows and 300 feet (90 m) thick. They are typically sub-aerial valley-filling and plain-flooding flows and some show strongly scoriaceous margins and/or well developed columnar cooling joints. A number disconformably overlie aquagene pyroclastics and unmineralised entrail breccias and include the youngest extrusions in the area. Generally empty vesicles, except for thin coatings of sublimate or alteration products, and flat flow tops in some lavas attest to their young age. However, some massive flows, stratigraphically isolated or overlying the older mineralised breccias, may possibly pre-date the younger unmineralised breccias and pyroclastics. The lower flow at Skittle Ball Plain has been dated Upper Cainozoic in age on palaeomagnetic evidence (Green and Irving, 1958).

In composition, the lavas range from tholeiite through tholeiitic olivine-basalt to alkali olivine-basalt. Individual flows commonly show different characteristic microscopic mineralogy and textures (Appendix 1), enabling easy delineation. The lavas erupted from at least nine separate centres and different sets of flows are described below.

Reynolds Island Flow and Dykes

This flow disconformably overlies aquagene pyroclastics and entrail breccias from 3,400 feet to below 3,350 feet, with maximum thickness of about 100 feet (30 m). It probably extends SW to Helens Island (Legge, 1904), and a poor exposure of similar basalt 1 mile NW of Canal Bay around 3,400-3,500 feet, may be a further extension, or a small separate extrusion.

The flow base shows a wavy, often irregular contact (plate 2, fig. 1). In places it develops a brecciated entrail structure, with subordinate rubbly and fine tachylitic fragmentary matrix, up to 6 feet (2 m) thick and grading into dense massive basalt. The entrail bodies are commonly cracked, with development of multiple cooling crusts, and the base is penetrated by clastic dykes and irregular inclusions of the underlying pyroclastic sediment, which may enter cracks in individual entrails (plate 2, fig. 2). This suggests basal digital advance of the lava, with localised brecciation and load casting due to flow over wet unconsolidated sediment.

A swarm of small dykes of massive basalt cuts the underlying entrail breccias NW of Reynolds Island below 3,380 feet and probably represents

the feeder system for the overlying flow (fig. 3). The dykes trend NE to ENE, range up to 20 feet (6 m), but mostly less than 5 feet (1.5 m), across, and can be traced over lengths of 130 feet (40 m). They are steeply dipping, with slightly sinuous, narrowly chilled tachylitic margins that become lobate and irregular in places. A few dykes bifurcate or send tongues up to 10 feet (3 m), long in the host rock. Dips of dyke margins suggest slight widening with depth. Jointing tends to parallel dyke margins, but there is also some cross and irregular joints, and some sidesteps in the dykes may represent small dislocations.

A few similar dykes cut entrail breccias on the NE side of Reynolds Island, but these show WNW to NW trends and may be associates of the breccias rather than the massive flow. One dyke tangentially cuts across, and shows chilled margins against, an earlier dyke.

The flow and dykes are tholeiitic olivine-basalt, with an abundant black glassy mesostasis, and contain rare inclusions of cherty and gritty sediment up to 10 cm across (Appendix 1).

Canal Bay Flow

Massive basalt with a scoriaceous margin occupies a valley at 3,450 feet W of Canal Bay, and overlies baked Permo-Triassic (?) strata to the north and unmineralised entrail breccias to the south. A small remnant of similar basalt caps the breccias above Duck Point and the flow probably originated from Canal Bay, where it attains its maximum height and thickness of about 100 feet (30 m).

The rock is tholeiitic olivine-basalt and microscopically characteristically contains glomeroporphyritic corroded augite (Appendix 1).

Liawenee Flows

Flow cappings are prominent on Liawenee Plateau, overlying dolerite at 3,850-3,700 feet, and below on Liawenee Moor, overlying mineralised entrail breccias at about 3,500 feet and infilling valleys cut in and along the dolerite scarp down to 3,350 feet at Armitage Rivulet. The flows range to 100 feet (30 m) thick and on Liawenee Moor a sequence of at least four flows reaches a maximum thickness of 250 feet (75 m) SW of Liawenee Hill. The flow SE of Lake Augusta has an extensive flat surface, but the other flows are generally more dissected.

A succession of three massive columnar and scoriaceous flows on Liawenee Hill is detailed by Voisey (1949); most of his scoriaceous zones belong to the middle flow, and the second from the top flow to the SW on Liawenee Moor has wedged out. Quarrying at the base of the top flow has exposed an intervening friable, brownish grey, fine tuff (Appendix 1). This is generally less than 1-2 feet (0.5 m) thick, but at one place extends 7 feet (2 m) upwards into a narrow irregular channel in the base of the overlying basalt, suggesting load casting. Silicified peaty material and pieces of silicified wood (mostly banksia; Dr J. Townrow, pers. comm.) were found at the base of the tuff.

The Liawenee flows are mostly alkali olivine-basalts, but some are transitional towards tholeiitic types, and the second flow in the Liawenee Hill succession is subophitic to ophitic tholeiitic olivine-basalt. Intergranular textures are typical in the large flow SE of Lake Augusta and in the lower Liawenee horizon, but glomeroporphyritic and subophitic to poikilitic textures are more common in basalt at Double Lagoon and in the upper flows on Liawenee Moor (Appendix 1).

The eruptive centres for the flows are not precisely known. The large flow on Liawenee Plateau apparently issued just SE of Lake Augusta and possibly flowed down to form the lowest flow on Liawenee Moor. The basalt at Double Lagoon apparently erupted from another local centre and may have flowed down to form the top lava on Liawenee Moor. Other possible centres may exist around Liawenee and NW of Armitage Rivulet.

Tods Corner Flow

This flow, at least 120 feet (40 m) thick, extends from Tods Corner to Shannon Lagoon, disconformably overlying aquagene pyroclastics, baked gravels and clays, and Jurassic dolerite bedrock, from 3,450 feet to below 3,340 feet on Great Lake shore (plate 2, fig. 6). The Beehives—Christmas Bay promontory is an isolated arm of the flow.

The basalt is dense with prominent cooling columns, commonly large scale but becoming small scale at The Beehives. It reaches its highest summit above 'The Battery', W of Tods Corner, where a striking change from the normal vertical and sub-vertical columns to horizontal columns (plate 10, Lewis (1923), p. 22) suggests a NW trending feeder dyke, associated with the eruptive centre for the immediately underlying pyroclastics.

The rock is tholeiitic olivine-basalt with ophitic-poikilitic texture (Appendix 1).

Skittle Ball Plain Flows

Three flows outcrop here from N of Little Pine Lagoon, W to Ouse River and S to Monpeelyata Canal. The succession from top to bottom is:—

3,500-3,450 feet—50 feet (15 m), massive columnar to scoriaceous alkali olivine-basalt,

3,450-3,4500 feet—50 feet (15 m), massive to scoriaceous tholeiitic olivine-basalt,

3,400-3,200 feet—200 feet (60 m), massive columnar to scoriaceous tholeiite.

The higher flows outcrop only N of Little Pine Lagoon, but the lower tholeiite fills an old valley of the Ouse and outcrops extensively with a strongly scoriaceous top forming much of the flat surface of Skittle Ball Plain.

The alkali olivine-basalt is intergranular type, the tholeiitic olivine-basalt is sub-ophitic to ophitic, and the tholeiite contains orthopyroxene (Appendix 1). The precise locations of the flow feeders are uncertain, but an eruptive centre may occur N of Little Pine Lagoon where the basalt reaches its highest elevation.

Murderers Hill Flows

Poorly exposed massive to scoriaceous basalt outcrops N of Murderers Hill, disconformably overlying mineralised entrail breccias and dolerite bedrock. The basalt between 3,350 and 3,250 feet fills an old valley of the Ouse and is intergranular tholeiitic olivine-basalt. This is capped by an

intergranular alkali olivine-basalt between 3,350 and 3,400 feet, extending E in an isolated arm towards Miena Hill.

These flows possibly erupted from the centre for the underlying breccias, but they also resemble the upper two flows of the Skittle Ball Plain sequence, and alternatively may have flowed upstream to Murderers Hill.

Lake Augusta Flows

Massive to scoriaceous basalts, at least 50 feet (15 m) thick, outcrop on the dolerite plateau around 3,800 feet on the W side of Lake Augusta as far as O'Dells Lake. The main outcrop at Lake Augusta is olivine-tholeiite containing ortho-pyroxene, and the small outcrop at Lake Botsford is alkali olivine-basalt with glomeroporphyritic augite (Appendix 1). The eastern outcrop at O'Dells Lake is tholeiitic olivine-basalt with inter-granular to sub-ophitic texture and the western outcrops are alkali olivine-basalt resembling the Lake Botsford rock. These flows suggest eruptive centres on the west side of Lake Augusta and on the north side of O'Dells Lake.

STRUCTURE

The dolerite structure of the Great Lake base-ment is summarised by Jones, Haigh and Green (1966). Structures imposed on this basement include post-Jurassic epeiorogenic faulting, tilting and jointing, and Cainozoic volcanic edifices (already described in detail).

The major post-dolerite faulting belongs to the late Mesozoic-early Palaeocene epeiorogeny of Tas-mania (Banks, 1962) and some of it may represent movements along original Jurassic structures (Sutherland, 1966). A number of structural trends can be inferred around Great Lake from field work, aerial photographic study, geophysical data and volcanic alignments.

Faults and prominent lineaments are suspected:—

1. NNW through Tods Corner, based on seismic data (Wiebenga and Polak, 1962); eruptive centres lie on this line S of Howells Island and N of Reynolds Island.

2. NNW through 'The Battery', bounding the dolerite of Becketts Bay and the straight steep margin of Great Lake N of Canal Bay; eruptive centres lie on this line at 'The Battery', E of Duck Point and NW of Reynolds Island.

3. N from Shannon Lagoon, bounding dolerite at Maclanachans Island, SE Canal Bay, Helens Island and E Reynolds Island; seamed weathered dolerite was excavated on this line at Miena Dam (Lewis, 1933, p. 24) and eruptive centres lie on it E of Duck Point and NE of Reynolds Island.

4. NW from Christmas Bay to E Lake Augusta, marked by four eruptive centres and a topographic break on Liawenee Plateau.

5. NW from Swan Bay to W Lake Augusta, marked by four eruptive centres and topo-graphic breaks S of Miena Hill and on Liawenee Plateau.

6. NNE from Murderers Hill to Liawenee, marked by four eruptive centres.

7. NNE from Christmas Bay to Reynolds Island, marked by five to six eruptive centres.

8. ENE from Cramps Bay to Armitage Creek, marked by five eruptive centres and a strong topographic break from Sandbanks Tier.

The structural pattern (fig. 4) suggests Cainozoic volcanism located mainly along intersecting NW, NNW, N, NNE and ENE trending fissures. These trends are further reflected in fault lines mapped by Blake, et al. (1956) to the NE, and on the small scale by structure lines in the dolerite base-ment (Voisey, 1949) and dyke trends at Reynolds Island (fig. 3).

Structural trends post-dating volcanic rocks include steep sparse blocky jointing in the entrail breccias and pyroclastics, mainly as two main intersecting sets trending NWN-WNW and NEN-ENE. Dips up to 5° W on flat floors of vesicle cavities in entrail lava SE of Canal Bay, suggest slight post-eruptive westerly down-tilting, pre-sumably from local movements on nearby NWN and N trending lineaments, possibly during later massive basalt extrusion. Tremors have been detected around Great Lake by the Tasmanian University seismic network since higher lake filling under HEC works. The tremor centres (fig. 4) lie mainly on the steep gravity gradients of Jones, Haigh and Green (1966), near suspected lineaments, and suggest minor adjustment movements on faults.

PETROLOGY AND PETROGENESIS

The Great Lake basalts range from saturated to undersaturated types (petrographic descriptions, Appendix 1; chemical analyses 1-5, Table 1; plate 4). Tholeiitic olivine-basalts, just saturated with about 50%-51.5% SiO_2 predominate and typically contain variable amounts of dark glassy mesostasis (Analyses 2 and 3). Rare, more saturated basalts contain orthopyroxene and are termed tholeiite and olivine-tholeiite (Analysis 1). Undersaturated alkali olivine-basalts form about one-quarter to one-third of the extrusions. They commonly contain titaniferous augite and are mostly near-saturated types, with some grading transitionally to the tholeiitic olivine-basalts (Analysis 4).

Although compositionally fairly uniform, the tholeiitic olivine-basalts vary considerably in tex-ture from porphyritic hyaloophitic (Ouse and Bridgewater types) through intergranular (Jordan type), sub-ophitic (Pontville type), ophitic (Mid-lands type) and rarer poikilitic and glomeropor-phyritic types, dependent on the cooling history (McDougall, 1959). Porphyritic hyaloophitic rocks (often palagonitised) are typical in the quenched basalt of the aquagene tuffs and entrail breccias, and also occur (non-palagonitised) in dykes and more chilled parts of massive basalts. Other textures are typical of massive basalts, with sub-ophitic to ophitic varieties typical in scoriaceous parts. Amongst the alkali olivine-basalts, por-phyritic intergranular to glomeroporphyritic inter-granular textures are common and some show
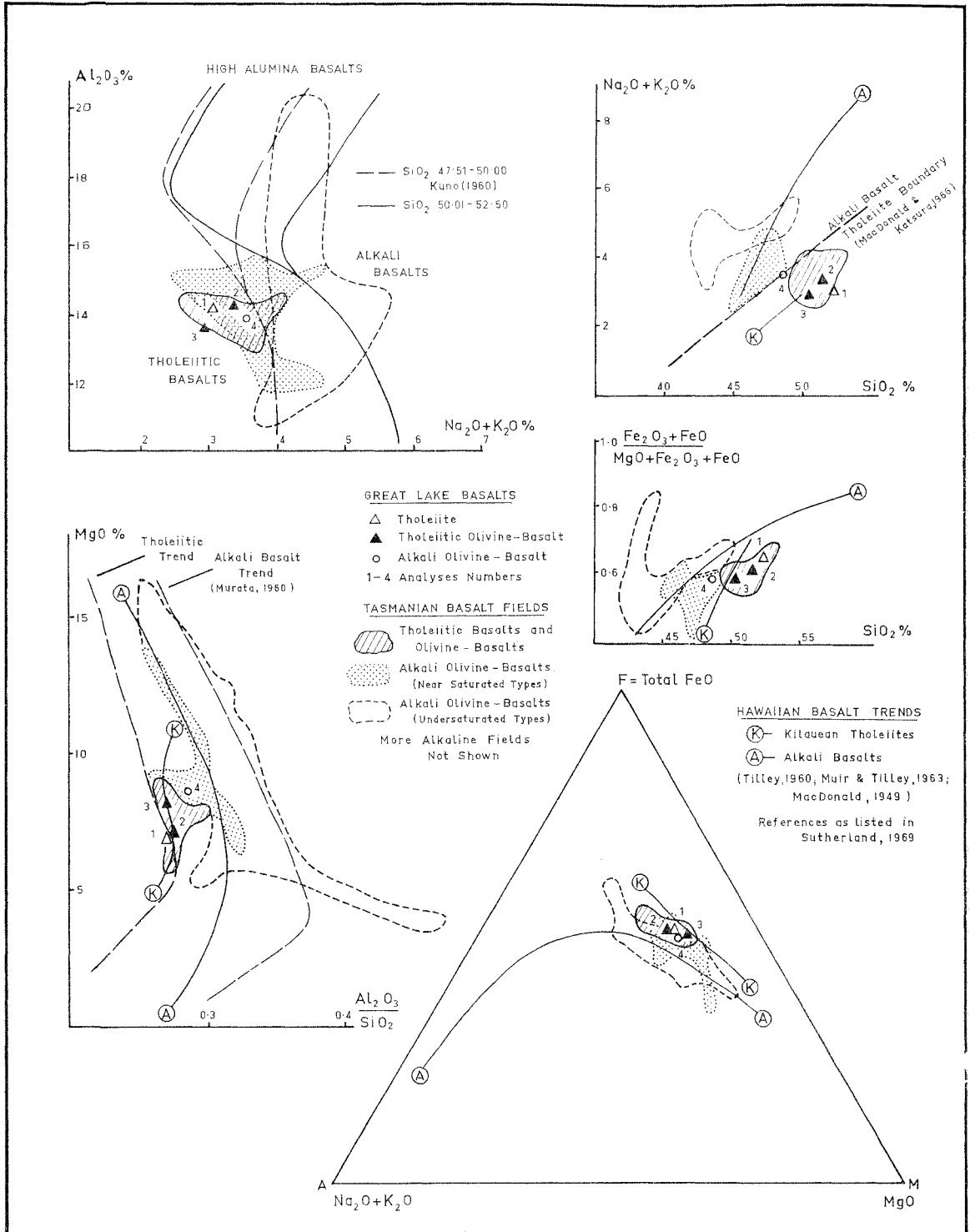
FIG. 5.—Chemical variation diagrams, Great Lake basalts, with Tasmanian basalt fields and Hawaiian basalt trends (modified from Sutherland, 1969).

sub-ophitic, ophitic or poikilitic textures, corresponding to the Mersey-Branxholm, Deloraine, and Burnie types of Edwards (1950).

The major element chemistry of the Great Lake basalts is plotted on variation diagrams (fig. 5) to show their relationships with the Tasmanian basalt fields and the Hawaiian tholeiitic and alkali basalt trends (Sutherland, 1969a). The Great Lake basalts resemble to some extent those of the Hawaiian province, interpreted as arising from segregation of parent olivine-tholeiite magma at depths of 35-70 km, with some fractionation at shallower depths to more saturated and aluminous compositions, and trends to alkaline olivine-basalt magma in waning stages of volcanism (Green and Ringwood, 1967).

The known volcanic stratigraphy at Great Lake superficially resembles a Hawaiian pattern of volcanism. The bulk of the basalts, including known older horizons, are olivine-tholeiitic types. Alkali olivine-basalts occur amongst massive flows, capping tholeiitic entrail breccias and pyroclastics, and represent later stages of extrusion. In massive flow successions that include tholeiitic lavas, the capping flow is always alkali olivine-basalt. Early minor extrusion of alkali olivine-basalt is also a possibility under initial more restricted mantle melting, prior to build up of full scale melting and tholeiitic volcanism. Most of the Great Lake alkali olivine-basalts would not fit here, but some few small stratigraphically isolated remnants on the Liawenee Plateau may do so.

SUMMARY AND DISCUSSION

Variations in form of the Great Lake lavas are attributed to extrusion under conditions ranging from sub-aqueous through partly emergent aqueous to sub-aerial, giving hyaloclastite pyroclastics, entrail breccias and pahoehoe to massive lavas respectively. The hyaloclastite pyroclastics and entrail breccias, in the absence of true pillow lavas, suggest extrusions into comparatively shallow water. Their restriction to, but widespread distribution within, the Great Lake depression implies eruption into a lacustrine environment and consequently the existence of major water stands at Great Lake at least as far back as the early stages of volcanism, probably in the Upper Tertiary. Massive sub-aerial flows overlying dissected pyroclastics and entrail breccias cut to below present lake level suggest eruptions during generally lower or completely drained water stands. Such conditions apparently existed at least during the later stages of volcanism, probably in the Pleistocene.

Lava successions, with entrail breccias resembling the Great Lake examples, occur elsewhere in Tasmania (Sutherland and Corbett, 1967; Sutherland, 1969a) and suggest extrusions into a variety of aqueous environments. Thus, the Great Lake breccias probably represent lake extrusions, possibly including intraglacial eruptions, the breccias in far NW Tasmania probably represent coastal sea extrusions, and breccias at Mersey-Forth-Sheffield probably represent extrusion into large rivers or dammed river lakes. Recent mapping by F. L. Sutherland at Bridgewater on the Derwent shore revealed a dissected bank of entrail breccias, elongate NW 1½ miles (2.4 km) and (0.8 km) wide, standing 120 feet (36 m) high and extending at least 30 feet (9 m) further below river level. This probably represents eruption into the Derwent River estuary in a past higher stand. Restricted developments of entrail lava and breccias at the base of massive flows, as at Reynolds Island in Great Lake and elsewhere in Tasmania, are thought to be lavas that flowed over small or shallow water bodies, such as small rivers and creeks, ponds, lagoons, marshes and flats of water saturated sediments.

The dissected coastal breccias in far NW Tasmania also pose difficulties in reconstruction, as do the Great Lake structures, unless a partly emergent sub-aerial structure above a confined water level is envisaged. Without this, their reconstruction requires cones 5,000-10,000 feet (1,500-3,000 m) high and corresponding water depths, for which there is no evidence on the Tasmanian coast. With this, breccia heights and water depths are at least 300 feet (90 km) but need not greatly exceed this, and Tertiary marine beds on the Tasmanian coast imply sea-levels to such heights. The NW structures were probably coastal island volcanoes, erupting lavas into the sea, but some more complete entrail lavas within them may be entirely submarine extrusions, in the lower parts or from submerged flank eruptions. The breccias at Lorinna are still largely preserved under capping massive lavas and reach about 400 feet (120 m) thick (Spry, 1958); this may indicate the approximate heights of some of the breccia structures.

The Tasmanian breccias are mineralised in far NW Tasmania (Miocene-Oligocene), in the Mersey-Forth area (probably Lower Miocene) and in the older series at Great Lake, and are unmineralised at Bridgewater and in the younger series at Great Lake (Pleistocene?). The difference is considered one of age, with the older breccias having undergone suitable mineralising conditions, such as sufficient burial by later lavas, prolonged ground or free water immersion and raised regional temperatures due to subsequent volcanism. All the mineralised breccias contain very similar suites of secondary minerals, except the NW coastal breccias in which soda-rich minerals are more prominent; this may be due to subsequent immersion or percolation by freshwater compared with seawater (Sutherland and Corbett, 1967; cf. Analysis 5, Table 1).

The Tasmanian Cainozoic entrail lavas and breccias are composed of uniform tholeiitic olivine-basalt. Poor development of such bodies amongst the equally prevalent alkali olivine-basalts of the State may be coincidental, but some factors affecting pillow formation (Solomon, 1969) were possibly optimal for the tholeiitic extrusions. Hyaloclastite pillow lavas are also known in the Tasmanian Cambrian, e.g., at King Island (Solomon, 1969), but differ in their general form from the Tasmanian Cainozoic structures, presumably due to environmental differences.

Tasmanian Cainozoic basalts tend to group geographically into distinct associations (Sutherland, 1969a). The Great Lake basalts, with those of

far NW Tasmania (Sutherland and Corbett, 1967) and Upper Mersey-Forth (Spry, 1958), can be taken as detailed examples of the tholeiitic associations, and they fall within a general belt of such rocks, extending from far NW Tasmania through Waratah, Middlesex and the Central Plateau to the Derwent Valley. These tholeiitic associations are considered to represent areas of greater mantle melting under Tasmania during Cainozoic volcanism. They contrast with the alkaline basalt associations, exemplified by the Tamar lavas (Sutherland, 1969b, 1969c), considered to represent areas of lesser mantle melting. The two associations are transitional through mixed tholeiite-alkaline basalt areas which may pass into a further mugearitic association, at present being mapped and investigated in detail.

Cainozoic tholeiite lavas, lacking olivine, such as the Skittleball Plains flow, are almost identical in mineralogy and chemistry to the Tasmanian Jurassic dolerites. Recent work indicates their presence in a number of areas (St Patricks Plains, Upper Mersey-Forth, Noland Bay, Georges Plain, Upper Macquarie River, Crown Lagoon, Woodsdale and Lune River). Thus, distinction of Jurassic or Cainozoic age is obviously unsafe on routine petrological grounds in the absence of clear field evidence or dating; this particularly applies to isolated dyke or plug bodies.

The Great Lake lavas appear to include some of the youngest in Tasmania and are comparable with the Newer Volcanics of Victoria (Upper Pliocene-Recent; Singleton and Joyce, 1969). Volcanism is now extinct at Great Lake, but measurements of present heat flow in the region, in comparison with some in the older volcanic and non-volcanic areas in Tasmania, would be of interest.

ACKNOWLEDGMENTS

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APPENDIX 1—PETROGRAPHIC DESCRIPTIONS, GREAT LAKE BASALTS

Quoted numbers refer to thin sections catalogued in the Tasmanian Museum collection. Quoted optical properties and mineral compositions were determined as in Sutherland (1969b).

JURASSIC DOLERITE

Basalt outcrops of Voisey (1949), Blake, *et al.* (1956) and Jones, Haigh and Green (1966); S bank Little Pine River (694, 708-709, 712-718) and N bank Little Pine Lagoon (691, 711, 719, 737a).

THOLEIITE (Skittleball Plains, 695, 710, 720-724, 736, 741, 742)

Fine grained varieties contain glomeroporphyritic zoned bronzitic orthopyroxene (1%-5%, 3 mm max., $2V_x \sim 81°-69°$, $\sim$ Fs 17-23) in an intergranular to sub-ophitic groundmass of zoned labradorite laths (40%-45%, 1.2 mm max., mostly $<$0.3 mm, Ab $\sim$ 35-58), clinopyroxene grains (30%-40%, mostly $<$ 1 mm), blades and irregular grains of opaque titan-iron oxide (3%-6%, 0.3 mm max.), and an intersertal dark to pale brown or grey glassy mesostasis (2%-30%) charged with iron oxide globules.

These grade into fine to medium grained rocks containing glomeroporphyritic zoned orthopyroxene (5%-17%, 2.5 mm max., $2V_x \sim 80$-59, $\sim$ Fs18-36) and zoned labradorite (up to 20%, 2.5 mm max., Ab $\sim$ 30-52) in finer grained groundmass, with brownish to pale grey mesostasis and sometimes interstitial quartz, green opal, chalcedony and clay (plate 4, fig. 4).

The orthopyroxene shows corrosion riddling, marginal alteration and replacement by zoned augite ($2V_z \sim 60°$-47°) or pigeonite ($2V_z \sim 3°$-15°, o.a.p. $\perp$ 010, but may $\parallel$ 010 marginally). The groundmass clinopyroxene is zoned augite ($2V_z \sim 60°$-51°), or pigeonite ($2V_z \sim 20°$-30°, o.a.p. $\perp$ 010, but may $\parallel$ 010 marginally) sometimes mantled with augite ($2V_z \sim 59°$-45°). Rare, corroded and altered olivine ($<$3%, 2 mm max., mostly $<$0.3 mm) is sometimes present. The rocks are dense or scoriaceous, commonly with empty vesicles, but sometimes with linings of chalcedony and clay.

OLIVINE-THOLEIITE (W Lake Augusta, 810, 808)

Corroded glomeroporphyritic zoned olivine (5%-8%, 2 mm max., $2V_z \sim 92°$-106°, $\sim$ Fa 19-48) and bronzitic orthopyroxene (3%-5%, $2V_x \sim 80°$-70°, $\sim$ Fs 18-26), with marginal reaction alteration to clinopyroxene (mainly augite, $2V_z \sim 60$-54), are set in an intergranular to sub-ophitic groundmass of augite (30%-35%, 0.3 mm max., with some pigeonite) and zoned labradorite laths (40%-45%, 0.7 mm max., Ab $\sim$ 32-52), with a dark intersertal

mesostasis (10%-15%) containing thin blades (3%, 0.3 mm max.) and granular globules of opaque titan-iron oxide (plate 4, fig. 3).

THOLEIITIC OLIVINE-BASALTS

Vitric Basalt Tuffs (Reynolds Island, 684, 685; Howells Island, 679-681; Maclanachans Point, 686-688).

Abundant angular to subrounded shards and fragments of pale yellow sideromelane or palagonitic glass (plate 4, fig. 1) may contain euhedral to strongly corroded fresh zoned olivine (1.2 mm max., mostly $<$0.7 mm; $2V_z \sim 91°$-97°, $\sim$ Fa 17-29) up to 12% in coarser bands and usually 5% in finer bands. Some olivines are broken and fragments appear in the matrix. Very rare augite and plagioclase (0.3 mm max.) and angular quartz and interlocking quartz fragments (to 1 mm) may be present.

Fragments range to 0.1 mm in the finest bedding laminae and between 0.3-3 mm across and greater in the coarser bands. Banding down to 0.5 mm thick, may be marked by incipient development of limonite and sometimes by grading with finer olivine-poor material passing into larger olivine-rich fragments. Rare, larger fragments show a halo of finely comminuted shards (plate 4, fig. 2). There are sporadic scoria fragments (to 4 mm), with smaller pieces largely altered to iron hydroxides, and rare pieces of hyaloophitic basalt. The base (10%-35%) is an opaline and chalcedonic cement with irregular pore spaces.

Entrail Basalt Breccias (Mineralised Series, Liawenee Canal, 482; Unmineralised Series, Ouse River, 677, Reynolds Island, 678, SE Canal Bay, 683).

The unaltered basalt consists of porphyritic to glomeroporphyritic corroded zoned olivine (5%-15%, 1.8 mm max., $2V_z \sim 89°97°$, $\sim$ Fa 13-29) and zoned partly skeletal plagioclase laths and crystallites (25%-55%, 0.8 mm max., Ab $\sim$ 40-52) in a black opaque to dark brown hyaloophitic glassy base containing clinopyroxene and small patches of opal and chalcedony. Rare plagioclase, with resorbed cores may be present.

The altered basalt of the tachylitic entrail crusts and fragmentary matrix is pale greenish yellow sideromelane or palagonite showing perlitic cracks and sparse vesicles, and containing scattered glomeroporphyritic olivine (12%), plagioclase laths (5%), and numerous crystallites. Matrix fragments are angular to subrounded, commonly 0.2-10 mm across, in a dark opaque to brownish limonitic and clayey cement. In the mineralised series this

cement is replaced with suites of secondary minerals (see text).

Porphyritic Hyaloophitic Basalts (Reynolds Island, 673, 673a, 682; N Canal Bay, 674; W Murderers Hill, 730, 731).

These contain corroded to euhedral glomero-phenocrysts, phenocrysts and grains of zoned olivine (8%-15%, 1.8 mm max., $2V_z \sim 88°$-$98°$, $\sim$ Fa 11-31), laths of zoned labradorite (30%-45%, 1 mm max., Ab $\sim$ 38-52), grains and crystallites of augite (15%-25%, 0.3 mm max.), and a dark glassy hyaloophitic to intersertal mesostasis with patches of opal, chalcedony, clay and minor carbonate. The Reynolds Island rocks may contain feldspathic quartz sandstone and quartzitic inclusions, showing some fused grain boundaries.

Porphyritic Intergranular Basalts (Canal Bay, 675; Duck Point, 676, 759).

Corroded zoned olivine (5%-12%, 1.5 mm max., $2V_z \sim 92°$-$107°$, $\sim$ Fa 19-50) is interspersed in an intergranular to sub-ophitic groundmass of zoned labradorite laths (35%-45%, 2.4 mm max., Ab $\sim$ 37-58) and zoned augite (25%-35%, 1.2 mm max., $2V_z \sim 63°$-$49°$), with an intersertal brown glassy mesostasis (15%-25%). Augite, and occasional labradorite, shows corrosion riddling, and glomero-porphyritic augite sheaves and aggregates may form to 3 mm across.

The mesostasis associates with thin elongate ilmenite blades (3%-4%, 1.2 mm max.) and contains acicular crystallite sheaves, apatite needles, grains and globules of opaque titan-iron oxide and patches of chalcedony and clay. The rock grades marginally into sub-ophitic basalt.

Porphyritic Sub-ophitic Basalt (Skittleball Plains, 738-740; Liawenee Moor, 754, 754a, 1150, 1154, 1155, 1161, 1162; N O'Dells Lake, 799, 807).

Corroded zoned olivine phenocrysts and grains (2%-17%, 3.6 mm max., $2V_z \sim 88°$-$101°$, $\sim$ Fa 11-37), sometimes show marginal reaction alteration to granular augite, and are scattered in an intergranular to sub-ophitic-ophitic groundmass of zoned labradorite laths (35%-45%, 1 mm max., Ab $\sim$ 37-58), zoned augite (35%-40%, 1 mm max., $2V_z \sim 62°$-$42°$), titan iron oxide blades and grains (3%-6%, 0.8 mm max.) and brown to grey glassy mesostasis (2%-15%) with interstitial greenish opal (up to 15%).

Porphyritic Ophitic Basalts (Beehives, 692, 693; Tods Corner, 689, 728; Shannon Lagoon, 733, 734).

Corroded zoned olivine phenocrysts and grains (8%-12%, 1.8 mm max., $2V_z \sim 91°$-$100°$, $\sim$ Fa 17-35) and zoned labradorite laths (35%-40%, 0.6 mm max., Ab $\sim$ 37-52) are intergrown with ophitic to poikilitic zoned augite (30%-45%, 3 mm max., $2V_z \sim 62°$-$51°$). The mesostasis passes from abundant hyaloophitic opaque black glass (25%) containing augite (plate 4, fig. 5), to ophitic augite ($2V_z \sim 61°$-$44°$), blades and grains of titan-iron oxides (3%-6%, 0.6 mm max.), intersertal greyish brown glass (5%), and patches of opal, chalcedony and clay.

Rare inclusions of fused quartzitic sandstone show corroded quartz grains in a clear to brown glassy base with flow lines. Streamed drops and patches

of devitrified greenish opaline glass, numerous small cordierite crystals with opaque iron oxide concentrated in the cores, sporadic sheaves of sillimanite with granular haloes of opaque iron oxide, rare sanidine, and tridymite (?) occur in the glassy base.

ALKALI OLIVINE-BASALTS

Porphyritic Intergranular Basalts (Liawenee Moor, lower levels below 3,500 feet, 725, 727, 748, 1151, 1156, 1158-1160; Liawenee Moor, upper levels above 3,500 feet, 726, 1153, 1157, 1163, 1164; Liawenee Plateau, S Lake Augusta, 749, 750, 752, 752a, 753, 753a; S O'Dells Lake, 797; Lake Botsford, 798; Murderers Hill, 696, 732; Skittleball Plains, 690, 735).

Euhedral to strongly corroded zoned olivine phenocrysts, grains and rare granular aggregates (10%-25%, 2.4 mm max., $2V_z \sim 86°$-$100°$, $\sim$ Fa 7-36), sometimes show part alteration to serpentine, translation lamellae or strain polarisation. The groundmass is intergranular, with zoned, sometimes fluidal, labradorite laths (30%-45%, 0.8 mm max., mostly <0.3 mm, Ab $\sim$ 34-62), colourless to pinkish titaniferous augite (20%-35%, 0.8 mm max., mostly <0.3 mm), blades and grains of opaque titan-iron oxide (3%-5%, 0.6 mm max.), and interstitial material including zeolites (up to 15%, mainly chabazite, rarely with natrolite), green to yellow serpentinitic clay, chalcedony, opal and rarely carbonate. Amygdales may contain zeolites, clays or carbonates.

Some rocks show scattered to glomeroporphyritic microphenocrysts and phenocrysts of zoned titaniferous augite, best developed in the Lake Botsford rock (up to 5%, 1.5 mm max., colourless to faintly pleochroic from pale fawn to pinkish mauve, with normal, hourglass and oscillatory zoning, $2V_z \sim 64°$-$40°$). Rare plagioclase phenocrysts (1.2 mm max.) show corrosion riddled cores and narrow overgrowths. Fused sediment replaced by clinopyroxene, quartzitic fragments and rare small peridotite inclusions may be present.

Porphyritic Sub-ophitic Basalt (Liawenee Plateau, Double Lagoon, 750a, 751).

Corroded zoned olivine phenocrysts and grains (20%-25%, 2.4 mm max., $2V_z \sim 88°$-$100°$, $\sim$ Fa 11-36) occur in a groundmass of zoned labradorite laths (30%-40%, 0.6 mm max., Ab $\sim$ 36-56), intergranular to sub-ophitic elongate prisms and grains of zoned titaniferous augite (20%-25%, 0.6 mm max., X faint mauve, Y mauve, Z pale fawny mauve, Y > Z $\backsimeq$ X, $2V_z \sim 72°$-$51°$), and intersertal dark glassy mesostasis (15%-20%).

Porphyritic Ophitic Basalt (Liawenee Moor, top flow, 747, 747a, 1152).

Corroded zoned olivine phenocrysts (5%-12%, 2 mm max., $2V_z \sim 88°$-$98°$, $\sim$ Fa 11-31) and rare plagioclase phenocrysts (2.4 mm max., with corrosion riddled cores infilled with clinopyroxene and iron ore) are set in ophitic to poikilitic groundmass of small zoned labradorite laths (35%-45%, 0.6 mm max., mostly <0.3 mm, Ab $\sim$ 36-58) enclosed in large plates of faintly pleochroic zoned titaniferous augite (25%-35%, 1.5 mm max., $2V_z \sim 67°$-$42°$), with interspersed blades and grains of opaque

titan-iron oxide (4%-6%, 0.6 mm max.) and interstitial serpentinitic clay (up to 15%).

The rock (plate 4, fig. 6) appears to represent a more completely crystallised variety of the Double Lagoon rock, and is transitional and closely similar to the tholeiitic ophitic olivine-basalt of the Tods Corner flow.

Vitric Tuff (Liawenee Hill, below top flow, 1146).

This contains small angular to subrounded fragments (<0.2 mm) of basalt glass (95%), rare titaniferous augite and very rare olivine. It probably represents a fine explosive phase of the overlying porphyritic ophitic basalt.

TABLE 1—CHEMICAL ANALYSES AND C.I.P.W. NORMS,
GREAT LAKE BASALTS

Analysis				1	2	3	4	5
SiO_2	..	....	....	52.3	51.5	50.1	48.6	47.3
Al_2O_3	.	....	....	14.2	14.2	13.6	13.9	18.2
Fe_2O_3	.	....	....	1.4	1.1	2.5	2.4	——
FeO	....	....	....	9.5	10.2	9.1	9.9	——
MnO	....	....	....	0.16	0.17	0.17	0.17	——
TiO_2	..	....	....	1.7	1.6	1.8	1.8	——
P_2O_5	.	....	....	0.23	0.24	0.27	0.29	——
CaO	....	....	....	9.3	8.5	8.6	8.2	9.0
MgO	....	....	....	6.9	7.1	8.3	8.6	——
Na_2O	.	....	....	2.5	2.5	2.4	2.7	0.38
K_2O	....	....	....	0.56	0.87	0.52	0.84	1.6
H_2O-	.	....	....	0.83	0.42	1.2	1.2	——
H_2O+	..	....	....	0.30	0.69	1.0	0.71	23.8
CO_2	....	....	....	0.20	0.92	0.20	0.26	——
Total	..	....		100.08	100.01	99.76	99.57	100.28

C.I.P.W. Norm				1	2	3	4	5
Q	..	....	....	4.10	3.09	2.40	——	——
Or	....	....	....	3.31	5.14	3.07	4.96	——
Ab	....	....	....	21.16	21.16	20.31	22.85	——
An	....	....	....	25.87	24.96	24.80	23.33	——
Di	....	....	....	*14.37*	*8.10*	*12.06*	*11.26*	——
	En	....	....	3.92	2.13	3.71	3.36	——
	Fs	....	....	3.14	1.87	2.16	2.13	——
	Wt	....	....	7.31	4.10	6.19	5.77	——
Hy	....	....	....	*23.85*	*29.13*	*26.74*	*20.71*	——
	En	....	....	13.26	15.55	16.95	12.70	——
	Fs	....	....	10.59	13.58	9.79	8.01	——
Ol	....	....	....	——	——	——	*6.35*	——
	Fo	....	....	——	——	——	3.74	——
	Fa	....	....	——	——	——	2.61	——
Mt	....	....	....	2.03	1.60	3.63	3.48	——
Il	..	....	....	3.23	3.04	3.42	3.42	——
Ap	....	....	....	0.55	0.57	0.64	0.69	——
Ca	....	....	....	0.46	2.09	0.46	0.59	——
$H_2O\pm$	..	....	....	1.13	1.11	2.20	1.91	——
Total	..	....		100.06	99.99	99.74	99.55	——

1. Tholeiite.   Massive flow, S bank Ouse River, Marlborough Highway, Skittleball Plains,   .

2. Tholeiitic olivine-basalt.   Massive flow, Shannon Lagoon, quarry E side Lake Highway.   .

3. Tholeiitic olivine-basalt.   Entrail breccias, SE of Canal Bay, Great Lake   .

4. Alkali olivine-basalt (transitional type).   Massive flow, quarry in top flow, Liawenee Hill, Great Lake.

5. Chabazite.   Mineralised entrail breccias, Liawenee Canal, Great Lake.   .

Analyses by Tasmanian Department of Mines Chemical Laboratory,
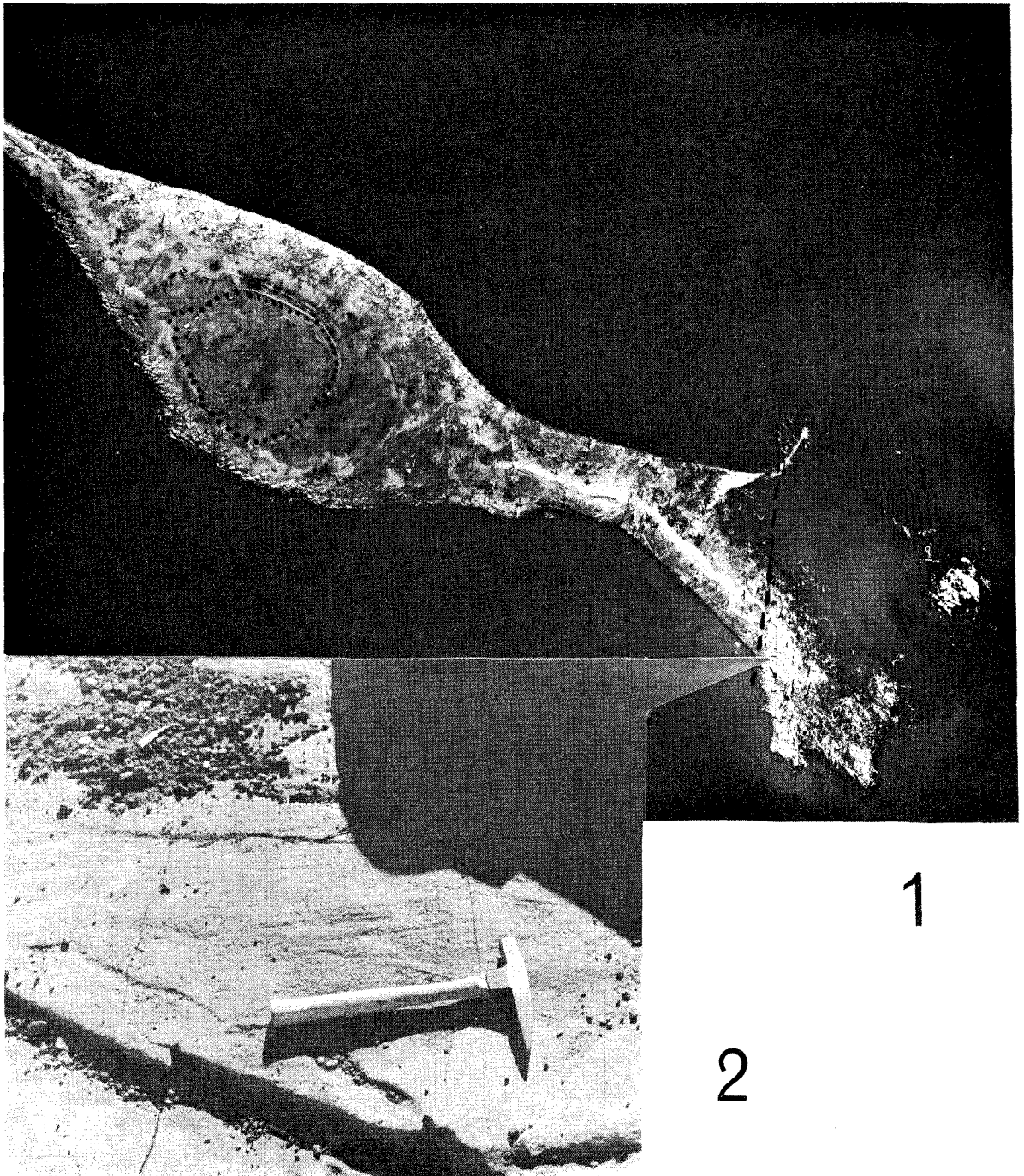*J. Furst, Analyst.*

PLATE 1—

FIG. 1.—Peninsula SE of Canal Bay, uncovered by lowered level in Great Lake (Little Neck of Legge, 1904). H.E.C. aerial photograph. Great Lake, Run 4, 71842, 3 July 1967. Previously submerged land is indicated outside dotted line. Cliffs of basalt breccias are exposed on SW side and the underlying dolerite basement is exposed on E side, marked off by dashed line.
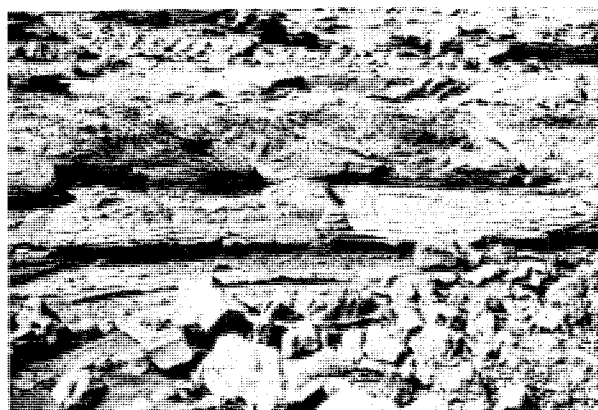
FIG. 2.—Pre-basaltic striated dolerite pavement remnant, at basalt-dolerite contact on peninsula, SE of Canal Bay, looking N. Its position on the aerial photograph (fig. 1) is indicated by arrow tab.

1

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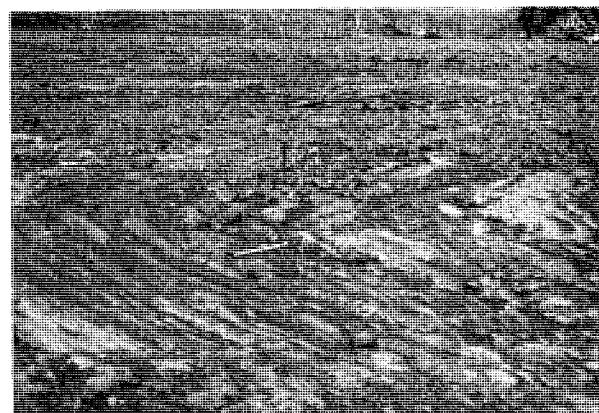
4
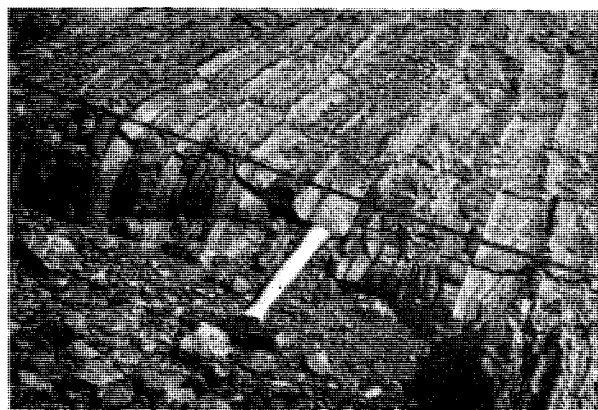
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PLATE 2

FIG. 1.—Massive basalt (Reynolds Island flow), disconformably overlying bedded aquagene tuffs and agglomerate, emerged shore, S side of Reynolds Island, looking SE.

FIG. 2.—Basal contact of massive basalt (Reynolds Island flow), overlying bedded aquagene basalt tuffs and agglomerate, showing penetration of clastic dykes (top right) into the basalt by load casting and dislocation of the beds. Locality as above.

FIG. 3.—Bedded aquagene tuffs, showing intrastratal deformation, emerged shore, SW of Howells Island, looking N.

FIG. 4.—Eroded dome in folded bedded aquagene tuffs, emerged shore, N.W. of Howells Island, looking SE.

FIG. 5.—Steep, overturned (?) beds of aquagene tuffs forming slide structure in basalt breccias, emerged shore, W of Howells Island, looking E.

FIG. 6.—Massive, columnar basalt (Tods Corner flow), disconformably overlying folded aquagene tuffs and agglomerates, emerged shore, E of Maclanachans Point, looking S.
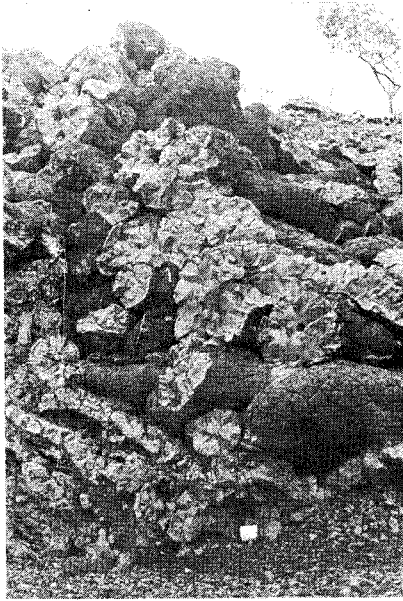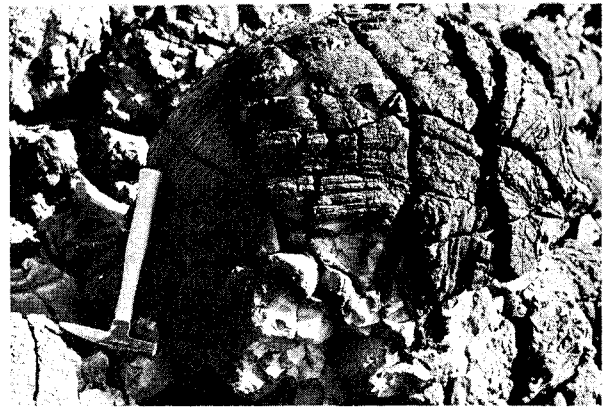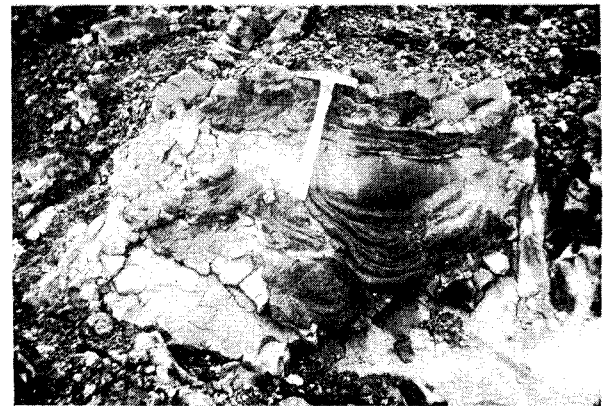
PLATE 3—
FIG. 1.—Entrail lava horizon in unmineralised basalt breccias. Note passage to entrail breccia at extreme top right and bottom left, and empty vesicles and cavities. Emerged shore, W side Howells Peninsula, looking E.
FIG. 2.—Entrail breccia horizon. Note multiple cooling crusts at top right and unmineralised matrix. Locality as above.
FIG. 3.—Lobate broken flow toe in unmineralised basalt breccia, showing striated tachylitic cooling crust. Locality as above.
FIG. 4.—Cross-section of entrail in unmineralised basalt breccia, illustrating development of multiple cooling crusts by cracking. Emerged peninsula, SE of Canal Bay.
FIG. 5.—Internal ropy flow structure, within elongate entrail cavity. Locality as above.

1MM.