THE ENVIRONMENT

OF THE PERMIAN GLACIAL SEDIMENTATION

OF TASMANIA,

With Particular Reference

To The Base Of The Permian.

BY

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Preface

The material submitted for the Degree of
Philosophy consists of the following:-

1. A dissertation, and the record of the field
and laboratory investigations covered by the candidate.
This is entitled "The environment of the Permian glacial
sedimentation of Tasmania with particular reference to
the base of the Permian".

2. A supporting paper by S.W.Carey and N.
Ahmad entitled "The environments and nomenclature of
glacial sedimentation". Although this is a supporting
paper, it might be appropriate to read it first, since
the theoretical study suggested a number of possible
lines of field and laboratory investigations of the
Permian sediments to determine their environments of de-
position, some of which were followed up by the candid-
ate and reported in 1. above.

3. Three joint papers by the candidate in
collaboration with other workers, and one paper by the
candidate alone, each on a different kind of terrain in-
volving the Pleistocene glaciation of Tasmania. These
were undertaken to broaden the candidate's knowledge
and understanding of glacial phenomena generally. These
papers are:-

> a. Jennings, J.N. and Ahmad, N.
>   "The legacy of an ice cap".
> b. Ahmad, N. and Baker, W.
>   "The re-examination of fjord theory of
>   Port Davey, Tasmania".

c. Ahmad, N. , Bartlett, H. and Green, D.
   "The Glaciation of the King Valley,
   Western Tasmania."

d. Ahmad, N.
   "Mersey Valley Glaciation".

4. Two joint papers in collaboration with
M.R.Banks, were undertaken to study the Permian and
Tertiary Stratigraphy of Western Tasmania.
   a. "Permian system in Western Tasmania".
   b. "Contribution to Cainozoic History of
      Western Tasmania".

With respect to the papers written jointly,
it is difficult, if not impossible, to apportion
authorship rigorously. However they are truly joint
papers, and the candidate has contributed materially
to each of them.

The author wishes to thank the staff of the
Geology Department, University of Tasmania, for the
courtesies shown to him during his stay. Special thanks
are due to Prof. S.W.Carey for his guidance in this
study and to Messrs A.Spry, M.R.Banks, and Dr. E. Wil-
liams, who fully discussed the various problems and
guided the writer in the field. The author is also
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and the members of the Mines Department for their help
and assistance. The author also wishes to thank Prof.
K.E.Caster and N.G.Lane for a great deal of encourage-
ment, advice and criticism.

Abstract

In Tasmania during the Lower Permian Epoch glaciers not only reached sea level but in places extended on the sea as shelf ice. Where-ever glaciers rested on the land tillites were deposited but under the shelf ice deposits of various types were laid down.

During this time icebergs were quite common and large iceborne erratics were deposited side by side with marine fossils. Conditions both on land and sea were very similar to those that exist today in the Antarctic region.

The glaciers flowed in part from a land mass lying west of the present west coast of Tasmania, which lay within $40°$ of the South Pole.

Waxing and waning of the glaciers is indicated by differences in the sediments, and by increase and decrease in the number and size of the erratics.

The gradual retreat of the ice sheet was followed by warm water sediments in the Triassic Period.

Introduction.

The glacial origin of some of the Permian deposits of Tasmania has been known since the middle of the last century, since Permo-Carboniferous rocks of the same character had already been studied in Victoria and New South Wales, and the Tasmanian beds were correlated with them.

Stephens in 1869 first reported the occurrence of the tillite in the Wynyard area but its age was not confirmed until 1907 when David after detailed study, correlated it with the tillite at Bacchus Marsh. (Johnston (1888) in his "Geology of Tasmania" had assigned this tillite a Pleistocene age). Since the discovery of coal bearing beds in the northern part of the state these deposits were studied in detail by the staff of the Mines Department. In the Mersey district and along the east coast bore holes had been sunk as early as 1887 in search of coal and much information about these deposits was gained. Lewis in 1924 reported the occurrence of tillite from the Styx River and later studied Permian deposits from several other parts of the state. Voisey in 1938 was the first person after Johnston to study the Permian System of Tasmania as a whole and he attempted to correlate the deposits from each area, and to correlate them as a whole with the Kamilaroi system of New South Wales. Since the inception of the Department of Geology in the University of Tasmania much study of these deposits has been carried out and Banks (1955,1957) has published a detailed section and correlation.

It has long been recognised that there were in Tasmania beds with both abundant glacial erratics and

marine fossils. Hitherto the environments of the deposition had not been studied since there were no other definite marine glacial deposits which could be used as standard. With the advance of the knowledge of the marine deposits from the Arctic and Antarctic regions the occurrence of similar deposits was reported.

The development of methods of mechanical analysis of the sediments facilitated comparison, but because of lack of detailed data of analyses conducted by other workers, this method could not be used as profitably as was expected.

Due to lack of any other suitable word the term 'large particles' has been used in this thesis to denote particles larger than 10 mm.

LABORATORY STUDY OF GLACIAL SEDIMENTS FROM THE
BASE OF THE PERMIAN.

Methods Used

Basal Permian rocks, which were accessible and which could be brought back to the laboratory in bulk were studied in detail, but other exposures were studied in the field only. The detailed study of these rocks included analysis of size distribution, roundness, sphericity, soling of the rock particles over one centimetre, and the till fabric analysis of part of Wynyard tillite.

In this study it was decided to use only those methods which has been used by other workers, so that the results could be compared, but in some instances, but in some instances, due to the differences in the nature of rock, those methods have been modified.
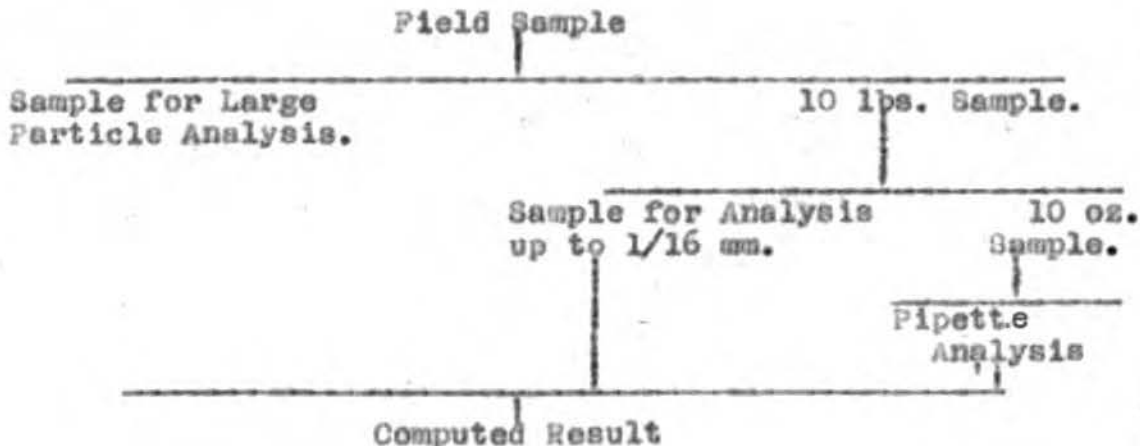
The weight of the samples disaggregated for this study was between 300 and 400 lbs. This quantity was taken, as it was found that the weights recommended by Krumbine and Pettijohn (1938, p 32) and Twenhofel (1941, p 46) were insufficient for such a coarse grained and badly sorted rock. Mechanical analysis was carried out with various weights, and it was found that when smaller quantities were analysed the results varied considerably, samples with some large boulders gave low percentage of clay and silt size, whereas samples with small pebbles gave a different proportion of the silt size particles. The large samples had some disadvantages; for example it was not possible to disaggregate the whole sample at the same time since available sieves were not large enough; thus it had to be done in parts.

Preparation of the sample.

A field sample of between 300 and 400 lbs. was taken and all the big pieces were broken with a wooden hammer, so that the large particles were not crushed. In this process if any large particle was broken, it was immediately cemented together with cellulose cement.

After this operation the sample was quartered several times, a final sample of 10 lbs. was taken for small grade analysis and the bulk of the sample was disaggregated for large particles. The 10 lbs. sample was further quartered and a 10 oz. sample taken for clay, silt and sand analysis, whereas the rest of the sample was sieved, and the sand fraction upto 16 mm. collected. The weight of any pebbles bigger than 4 mm. was subtracted from the weight of this sample and added to the large size particles.

The 10 oz. sample was completely sieved and all the grades were collected; fractions smaller than 1/16 mm. were analysed by the pipette method, but the weights of other fractions were added to the 10 lbs. sample. The results from the whole analysis were computed together.

```
                          Field Sample
                               |
_____
Sample for Large                         10 lbs. Sample.
Particle Analysis.                              |
                                                |
                    _____
                    Sample for Analysis              10 oz.
                    up to 1/16 mm.                   Sample.
                            |                           |
                            |                _____
                            |                     Pipette
                            |                     Analysis
                            |                         |
_____
                    Computed Result
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Disaggregation

1. Wetting: Samples were left in water for 24 hours, but this method did not disaggregate the sample.

2. Boiling: Samples were boiled in water for eight hours but this had no effect.

3. Heating and quenching: This method was partially effective but some of the pebbles were also broken so the method was not used.

4. Kerosene Treatment: At the suggestion of Mr N.G.Lane small pieces of the specimen were immersed in kerosene for one hour and then left in water for an hour; all samples except the King William Saddle specimen, were completely disaggregated by this technique.

Several other methods as roto-shaking and boiling in highly concentrated solution of sodium thiosulphate were used to disaggregate the King William Saddle specimen but all of them failed so it was decided to study it in thin section under the microscope.

As most of the published results of the mechanical analysis have used the Wentworth scale of the size classification (1922) it was decided to distribute the sieving and pipette results according to that grade.

Sieve Opening

In these analyses C.O.M. Standard sieves have been used. This standard very nearly coincides with ^{4}2 mm. sieve scale, and it was standardised with Wentworth scale for comparison with other published results.

Wentworth Scale mm.	4_2 mm.	Tyler mm.	Screen No.	C.O.M. grade mm.	Scale Sieve No.
4.	4.0.	3.96.	5.	4.0.	4.
2.	2.0.	1.98.	9.	2.0.	8.
1.	1.0.	0.99.	16.	1.0.	16.
1/2.	0.50.	0.495.	32.	0.50.	30.
	0.297.	0.295.	48.	0.295.	50.
1/4.	0.250.	0.246.	60.		
	0.149.	0.147.	100.	0.149.	100.
1/8.	0.125.	0.124.	115.		
	0.074.	0.074.	200.	0.074.	200.
1/16.	0.062.	0.062	250.		

Sieves Nos 50, 100, and 200 are slightly bigger than
Wentworth scale but as they cover the next range of the
scale, so they were included in the standard, and for
comparison with other analyses may be added into one
grade.

Sieving.

The completely disaggregated sample was poured
into the set of sieves and washed with water so that
only those particles that could not pass through a part-
icular sieve were left in it. This washed sieve residue
was transferred to a metal pan and brushed with a soft
brush while still wet. After brushing, the samples were
again transferred to the sieve and washed, this process
being repeated till clear water came out of the sieve,
even after washing and brushing. After the last washing
the residue was left in the sieve and dried in the oven,
and then was dry sieved on the mechanical shaker for
40 minutes.

After sieving had been completed all the rock
particles which were oversized for the No. 4 sieve were

hand picked and their weight was subtracted from the weight of that sample and added to the weight of the respective size sample.

At the end of the sieving the residue from each of the sieves was weighed and its percentage weight calculated.

Sphericity and Roundness.

Sphericity and roundness are here used according to Waddell's definitions (1934 and 1935).

The simplified formula for determination of roundness is as follows :

$$\text{Roundness} = \sum r/n.R.$$

Where r = the radius of a corner.

$\sum r$ = the sum of the radii of the corners.

n = the number of corners.

R = the radius of the maximum inscribed circle.

The formula for the sphericity is as follows:

$$\text{Sphericity} = D_n/D_s.$$

Where D_n = the diameter of the sphere having the same volume as the pebbles.

D_s = the diameter of the minimum inscribed circle.

The sphericity and roundness of all the boulders and cobbles from all the specimens were studied; pebbles from King William Saddle and Wynyard specimen were quartered four times so that they were about 100 in number, but all the pebbles from the Maydena and Palmerston specimens were studied as there were less than 100 in each. The total number of the large particles studied

in each case was between 100 and 150.

To measure sphericity and roundness,

Large particles were placed on a transparent glass plate with a source of light just below it. A second glass was fixed in an adjustable stand, so that the upper glass plate almost touched the top of each pebble. By this method a sharp image of the pebble was obtained which was traced on the paper. A celluloid sheet on which 130 concentric circles differing from one another by 1 mm. in radius had been drawn, was placed over the pebble outline, and hence the radius of the curvature of every corner, the radius of maximum inscribed circle and the radius of minimum circumscribed circle were measured.

Soling.

The soling of a pebble has been determined as a function of the flatness of the faces. A perfectly soled face is defined as that face which is flat so that it will as a whole rest on any other plane surface. The soling of the face is obtained by dividing the area of soled surface by the area of soled face. The soling of the perfectly soled face will be one, as the area of the soled face and the area of the soled surface will be equal; less soled faces will have soling in decimal fractions less than unity.

The outline of the soled face was drawn on the tracing table, and its area measured with planimeter. The soled surface was next pressed on an inking pad and its impression taken on absorbent paper placed on a hard surface; its area was also measured by a planimeter. Sometimes it was found that the soled surfaces were developed on more than one place on the same face, so several separate impressions developed on the paper;

in these instances areas of all these impressions were
determined and added to find the area of the soled sur-
face.

The formula for determination of the soling
is as follows:

Soling   = a/A
Where a   = the total area of the soled surface.
      A   = the area of the soled face.

Pipette Analysis.

In order to determine the proportions of the
clay, silt and colloidal particles (less than 1/16 mm)
the pipette method was used. All the residue finer than
1/16 mm. obtained from sieving was dried in the oven
at 50° C. and left at room temperature for 24 hours.
This sample was quartered and 25 gms. of the powder was
mixed with water and 20 cc. N/10 sodium oxalate and left
to stand for 24 hours. After soaking the suspension
was shaken to disaggregate all the lumps and more water
was added to make the volume up to 1000 c.c.

The suspension was allowed to settle in 1000
c.c. graduated cylinder. At intervals determined accord-
ing to Stoke's Law of settling 20 c.c. samples were drawn
off in a pipette, and evaporated to dryness in a beaker.

The weight of the residue in each beaker was
determined by subtracting the weight of the sodium ox-
alate in 20 c.c. (0.67 gms.)

Calculations.
A   = weight of the residue finer than 1/16 mm.
B   = weight of the residue finer than 1/32 mm.
A-B = weight of the particles between 1/16 and
       1/32 mm.

These weights were computed into percentages
with respect to the total weight.

Several weeks were spent to standardise results
for the finer fractions using a Gallenkamp sedimentation
balance (Plate 1), but the reliable size range proved to
be too restricted, and several attempts to extend into
the finer size ranges failed to give constant and re-
produceable results. This failure is attributed to pro-
cesses of physico-chemical reactions (such as the
absorption of water by the clay, and absorption of the
flocculant) during the two days required for the measure-
ments.

Till Fabric Analysis.

In order to determine the direction of the
ice movement at Wynyard the pebble orientation of the
tillite on the shore platform was determined. For this
analysis two areas were chosen; one within six feet of the
wall of the Permian glacial valley and the other in
middle of the exposure about two miles from the valley
wall.

The direction of the longest axis and its
inclination from the horizontal was measured for 100
pebbles from each locality, and the data was plotted on
stereographic paper. As the tillite in the area (middle
of the exposure) dips at an angle of fifteen degrees
towards the west, a conpensation for the inclination was
made after the results were plotted.

Graphs.

All the results from the sphericity, round-
ness and mechanical analysis study were plotted on ordin-
ary graph paper of 1 0 squares to the centimetre, and
where-ever possible intensity curves were drawn.

Quartile Measurements.

The measure of the average spread with respect to the median (Md) was taken as the quartile deviation. The quartiles (Q_1 and Q_3) on either side of the median with the frequencies of 25 and 75 percent was taken from the cumulative curves and the arithmetic quartile so calculated $= (Q_3 - Q_1)/2$. The geometric mean quartile has been calculated as $Q_g = \overline{Q_3/Q_1}$.

The skewness has been computed as $Sk = \overline{Q_3 Q_1/Md^2}$.

The log of the quartile has been calculated as Log_{10} So, where So is the geometric mean quartile.

Location of Deposits Studied.

The Permian System unconformably overlies the Devonian and older rocks, and is overlain disconformably by Triassic rocks. It is exposed at several places on the north coast between Table Cape and Tamar Head, along the east coast between Ansons Bay and South Cape, in the Mersey and Derwent Valleys and along the west coast at Point Hibbs, Mt. Dundas, Mt Sedgwick, Mt Read and Eden. Permian sediments probably occupy a wider area but in other parts they are covered by younger rocks.

The base of the Permian is exposed in several places and the deposits are of three types, tillite, conglomerates, and a sequence of sandstones, shales and mudstones. Tillite is exposed at Wynyard,(David 1907; Hills, 1922), Preolenna, and the Inglis valley (Hills, 1922; Banks et al. 1955), Dairy Plains (41° 35', 146° 31'), Mersey District (Reid, 1924), Deloraine (Wells, 1957; McKellar 1957), King William Saddle (42° 13' 146° 7'), Maydena (42° 45' 146° 35'), Weld River (Lewis 1923), Mt Sedgwick (Banks and Ahmad 1958), Mt Read (Banks and Ahmad 1958), and Point Hibbs (Hills, 1914, Banks and Ahmad, 1958).

The basal conglomerate is exposed in the Mersey valley
south of Mole creek (41°19' 146° 28'), in the Deloraine
area, (Reid, 1924) and in the Cradle Mountain area
(Reid, 1919). Occurrence of basal tillite is also re-
ported from Shoemakers Point (Twelvetrees, 1915), but
it needs checking. The basal tillite reported by
Lewis (1938) at Maria Island is not included with other
tillites since it is probably an ice-berg deposit.

For detailed study specimens were collected
from 1. Wynyard, 2. Palmerston, 3. King William Saddle,
and 4. Maydena, but the other exposures were either
in inaccessible places or very near the localities
from which specimens were taken, so the succession only
was studied in these places.

Wynyard

In this area the basal Permian tillite is
exposed along the shore line (David, 1907), at Preo-
lenna (Hills 1922), Oonah (Banks and others, 1955)
and in Hellyer Gorge. A specimen for detailed study
was taken from the shore line about 400 yds. east of
the Doctors Rock at the unconformity between between
the Permian and the Precambrian.

Palmerston

About half a mile east of Palmerston home-
stead in a limestone quarry in Mr. Jackson's paddock
mudstone with erratics unconformably overlies Pre-
cambrian dolomite. The unconformity is very irregular
and the contact disappears under the hill. A specimen
for detailed study was taken from the contact.

King William Saddle.

An exposure of the basal tillite occurs about

two miles east of King William Saddle on the Hobart-
Queenstown road in the road metal quarry. Tillite
overlies the Precambrian quartzite unconformably and
the basement is striated indicating a direction of
movement from west to east. In this locality there are
no other known exposures of the Permian tillite, but it
is expected that other deposits occur on Mt King William I.

Maydena.

This exposure is about three miles south-
east of Maydena township on the A.N.M. road. Here the
tillite overlies the Owen Conglomerate unconformably,
it is interbedded with sandstone and varves and it is
overlain by outwash beds.

Description of the Deposits Studied.

Wynyard

Basal tillite is exposed along the shore from
Table Cape to Doctors Rock. It rests unconformably on
Precambrian schists, which have been eroded by the
glaciers (Plate 27). Some traces of the striations
are visible on the contact but they are so poorly devel-
oped that they do not help in determining the direction
of the ice movement. David (1907, p 277) has reported
some intraformational striations on three pavements,
indicating that the direction of the ice movement was
from S.30°W. to N.30°E. These striations could not be
seen anywhere in the area. Either they have been eroded
away by the action of the sea or they are covered by
sand. The direction of ice movement determined from
the orientation analysis of pebbles was from S.20°W.
to N.20°E.

The tillite and the overlying beds dip from horizontal to 15° and the direction of the dip changes from east to west. East of Doctors Rock a fault has cut the rocks at right angles to the strike and it has displaced the beds by 50 ft; another fault has cut the beds west of Doctors Rock and the sequence is repeated from here west-ward; another fault has possibly cut the rocks about half a mile east of Wynyard, and the sequence is again repeated from the fault westward. David (1922) has estimated the thickness of these beds as 1220 ft., but from all evidence it appears that the thickness is not more than 600 ft. and is possibly between 400 and 600 ft. David may not have noticed the rolling dip and the faults. It is very difficult to determine the exact thickness of these deposits as the real base is not exposed anywhere, and the contact between the Permian and Precambrian is probably along the valley walls in which the deposits were laid down.

Lithology

The lowermost beds which can be seen at the contact are sub-glacial moraines or tillites. The section east of the Doctors Rock is as follows:

> Outwash
> Tillite
> Alternating beds of shale and sandstone
> Varves
> Outwash
> Tillites with lenses of outwash.
> Tillites
> Outwash with pebbly mudstone, grit and
> conglomerate beds.
> Tillite

Banks et al(1955, p 205) have called these shale and sandstone beds varves. (Plate 2). These

beds have two sets of rhythms; one set is represented
by thin laminae in the shale and the other is represented
by shale and sandstone bands. The sandstone bands have
current bedding and ripple marks and possibly represent
the long ice free periods, while the shale bands repres-
ent the summers and the winters of cold years. Similar
deposits develop in small shallow lakes very near the
ice front and in that case the whole set can represent
one year, the shale bands with laminae daily variations
in winter, and the sand bands summer periods. Without
any other evidence it is rather difficult to distinguish
between the two types of varves. Sometimes the tillite
bands are interbedded with pebbly mudstones (Plate 3
A & B), which have a very high percentage of clay, but
as they do not show any bedding they were possibly de-
posited in shallow water in front of the glaciers. The
grit (Plate 5 A & B) and conglomerate beds (Plate 5 A
& B) are mostly composed of coarse grains and they do
not have much cementing material.

On both sides of the Inglis River several
rolled pieces of sandstone lie in between two layers
of tillite (Plate - 7). These pieces have their longer
axis at right angles to the direction of ice movement,
and sandstone layers are rolled one around the other;
sometimes the two ends are snub nosed and in others the
ends are open like cut sausage. These pieces of sand-
stone have probably been rolled in between two layers
of tillite when the ice was advancing on an unconsol-
idated sandstone layer, and the orientation of their
axis with respect to ice movement also supports this
view.

Palmerston.

This sample was taken from a dolomite quarry,

where the Permian beds overlay Precambrian dolomite
unconformably; the contact is irregular, and the base-
ment does not show any traces of glacial action. This
sample is a thinly bedded mudstone with striated peb-
bles. No fossils were found in these beds, but they
are reported from adjacent areas. The pebbles did not
show any orientation but they had distorted the beds,
and the beds were interrupted along the side of the
pebbles. Some pebbles were resting with their longer
axis vertical or inclined to the bedding plane (Plate 8).
These beds were probably deposited in water either very
near the ice sheet or below it.

King William Saddle.

This sample was taken from a small exposure
in the road metal quarry beside the Lyell Highway about
two miles east of King William Saddle. The deposit
overlies unconformably a glaciated and striated pavement
of Precambrian quartzite (Plate 9). The outcrop is
seven feet high but the thickness of the deposit could
not be determined as no other bed overlies it. The
sample was taken from the contact between the Permian
and Precambrian beds. The deposit has a bedded appear-
ance, and lenses of light coloured sandy material are
enclosed in fine grained dark coloured layers.

The tillite contains pebbles of fossiliferous
Ordovician sediments and Cambrian tuff (Plate 10) which
indicates that these beds could not be older than Ordo-
vician, and it does not contain any Triassic or dolerite
boulders, which suggests that it is not younger than
Triassic. Since there are only three recorded glacial
ages in the island, namely Precambrian, Permian and
Pleistocene, and since the first and the last may be

eliminated on the pebbles evidence, the deposit is possibly Permian. Pleistocene deposits occur within 200 ft of this exposure but those tills have a very large number of Permian and dolerite boulders, which are absent from this tillite. Although the absence of these erratics cannot be taken as conclusive proof of its age they indicate a difference between the two tills. The Permian tillite is more compact than the Pleistocene, which also indicates that it is older than the Pleistocene one.

Maydena.

This specimen was collected from the road cutting, on the A.N.M. road. Here the tillite overlies the Owen Conglomerate but the contact could not be seen, and it is overlain by sandstone, followed by two layers of tillite interbedded with varves, and outwash beds. (Plate 28). The tillite has marine fossils but the sandstone and varves are unfossiliferous. Although none of the fossils found in the tillite were whole, at least the generic names of some of them could be established. (Plate 11). Mr N.G.Lane has very kindly examined these fossils and his report is as follows:

"Fossils found in the tillite include specimens of the following: Fenestella cf. dispersa ; ? Fenestella sp. (?Fenestella cf. expansa Crockford 1945); Streblotrypa sp.; a small crinoid stem; an internal mould of a pedicle valve of a small brachiopod and external mould of a ? pelecypod fragment; and an external mould of a small spirifer.

"All the specimens are small and were apparently broken up prior to incorporation in the sediment.

"There are 5 specimens of a small, fine Fenestella with three to four zooecia to fenestrule; oval

fenestrules about 1.0 mm. in length and 0.48 mm. wide;
branches 0.32 mm wide with small zooecial apertures
0.10 mm. in diameter and with centres of zooecial aper-
tures   about 0.28 mm. apart longitudinally. These spec-
imen have been referred to F.dispersa Crockford 1943,
but the material is not well enough preserved to permit
positive identification.

        "There is one specimen of much coarser ? Fen-
estella species with unusually long fenestrules. This
specimen has been referred to F. expansa Crockford 1945,
although it could also be cospecific with F.cinta Crock-
ford 1948 from the Upper Carboniferous of New South Wales.
The specimen exhibits fenestrules averaging 2.55 mm. in
length and 0.42 mm wide. The branches are approximately
the same width as the fenestrules. No information could
be obtained concerning the zooecia. F.expansa has been
recorded only from Lakes Creek Quarry, Queensland.

        "There are several small specimens of a thin,
ramose species of Streblotrypa. One specimen shows the
typical arrangement of longitudinal rows of zooecia
with elliptical apertures and areas of mesopore pits
between the apertures. This genus has been recorded
from the Permian of Western Australia, the Lower Carbon-
iferous of New South Wales and from the Satellite Silt -
stone of Woody Island, Tasmania (Banks et al 1955)."

            Measurements on the Wynyard sample.

        The specimen was collected east of Doctors
Rock from the lowermost tillite at the contact with
the Precambrian at least 20 ft. below any exposed out-
wash layer.

Mechanical analysis

The particle size of the specimen varies from below 0.00098 mm. (less than 1/1000 mm.) to above 128 mm., but the largest boulder seen in the field was more than 3 ft. long and 1 1/2 ft. wide. The sorting of the till-ite is very poor, the value for So being 13.2 and for Log So 1.13. As the figures suggest, this rock is almost unsorted, with all grades of the sediments deposited together. The association of shales and conglomerate did not have any effect on the sorting of the tillite and it was probably deposited by the glacier in the absence of water.

The size distribution of the rock is also not uniform, the quartile deviation is 0.5716 with a skewness of 3.48, and there are over 56.5% of particles smaller than 0.062 mm. (1/16 mm).

The size histogram (Fig. 2) has three peaks between 1/128 and 1/64, 1/16 and 1/8, and 64 and 128 mms. Between 1/8 and 4 mm. the sediments are quite uniformly distributed, but between 4 and 64 mm. they show a rise which is due to the third peak. The sediments in the silt size show a sudden drop which is also reflected in the colloid size.

The cumulative curve (Fig 3 W) is very irreg-ular, the rise and the fall in percentage being quite pronounced. Although it flattens a little near the top, it is very steep near the bottom which is due to high percentage of the boulder size, and the low percentage of the silt and colloid sizes.

The size and frequency histogram of the large particles (Fig. 4.) has two peaks between 20 mm. and 40 mm. and suddenly drops to 80 mm. after which it

drops gradually to 320 mm.; more than 60% is concen-
trated between 20 and 40 mm.

Roundness.

Roundness of the large particles (Fig. 5)
varies between 0.1 and 0.8, the least rounded being
with the boulders and the most rounded the small pebbles,
but the cobbles (80 - 120 mm.) show an increase in round-
ness after a slight drop between 75 - 85 mm. size
pebbles. The intensity contour also shows the relation-
ship between the size and roundness; largest number
of pebbles between 20 and 30 mm. have 0.4 roundness,
but on the whole there is no significant relationship
between size and the roundness since small pebbles also
have low roundness. The frequency and the roundness
graph (Fig. 9) has a peak between 0.3 and 0.4, but the
roundness of the largest percentage of the erratics
varies between 0.3 and 0.6.

Sphericity.

The large particles have sphericity (Fig. 7)
between 0.3 and 0.9 but show no relationship between
size and sphericity; smaller pebbles have the lowest
as well as the highest sphericity. The largest number
of particles have 0.6 sphericity.

The sphericity and the frequency graph (fig.
7) of the large particles shows a peak at 0.61 with
two minor peaks at 0.53 and 0.76. This graph and the
histogram shows a big decrease in the number of pebbles
below 0.5 and above 0.8 sphericity.

The comparitive roundness and sphericity
contours show that the largest number of large particles
have 0.4 roundness and 0.6-0.7 sphericity. This graph

also shows that for a wide range of roundness there is a limited range of sphericity.

Soling.

Out of 112 large particles examined 24 had soling; the soling varies between 0.02 and 0.52, the majority of the erratics having 0.2 - 0.3 soling.

Field Association.

Stephens (1869) first recognised the tillite on the shore line. Hills (1913) reported the occurrence of Permian beds from Preolenna and has described the full section from the area :

F.	Yellow and brown sandstone, unfossiliferous	550 ft.
E.	Sandstone, pebbly sandstone and mudstone with marine fossils	50 ft.
D.	White, yellow and black sandstone with coal.	140 ft.
C.	Mudstone with marine fossils	140 ft.
B.	Blue grey mudstone with bands of mudstone conglomerate unfossiliferous.	300 ft.
A.	Glacial Conglomerate.	1220 ft.

He has also reported the occurrence of Kerogenite (Kerosene Shale) interbedded with coal. The beds overlying the tillite are probably equivalent to the Palmerston Drift beds (p   ) and the Quamby Mudstone beds of Deloraine district. Banks et al (1955) have reported five occurrences of varves between the shore line and Oonah, and have come to the conclusion that the varves in the north were farther from the ice

front than the ones in the south, provided varves from
several of the occurrences were deposited in the same
lake. They have also reported the occurrence of oil
shales 23-43 ft. above the tillite.

Measurements of Palmerston Sample.

The specimen was collected at the contact
with the Precambrian dolomite and the lower most Permian
bed exposed. McKellar who has mapped the area immediately
to the west (1957) considers that the base of the Permian
in the quarry belongs to the McRae Mudstone Formation of
the Quamby Group (Personal commu.). This view implies
that the base here rests on a buried hill on the Permian
landscape at least 600 ft. high.

Mechanical analysis.

This sample has a wide range of particle
size, the smallest being colloid grade and the largest
over 128 mm. The largest boulder seen in the area was
over 16 cm. long and 9 cm. wide. The sample has quite
good Sorting Coefficient; So is 3.70 and Log So 0.56.
Considering the nature of the deposit this coefficient
is very low but the large percentage of sand, silt and
colloidal size particles have affected it. The sorting
of the sediment is also better than the tillites, the
quartile deviation being 0.0372 and the skewness 1.98.
Sediments below 1/16 mm. comprise more than 61% of the
total.

The sediments have three modes in the histo-
gram (Fig 2), the maximum is between 1/64 and 1/32 mm.,
and the other two modes are between 1/16 and 1/8 mm.
and below 1/1024 mm. respectively. The drop in the

percentage towards smaller size is gradual from 1/64 mm.
but the drop towards bigger sizes from 1/32 mm. is sud-
den, and the grade between 64 - 128 mm. is nearly ab-
sent.   The mode below 1/1024 mm. is due to a rise in
the total amount of colloidal size.

The cumulative curve (Fig 3) rises gradually
from the bottom but it is quite steep near the top, this
behaviour being due to the high percentage of medium
and fine sizes.   This tendency of the curve is quite
evident from the histogram.   The deposit has a fairly
large quantity of fine grade particles and a smaller
quantity of coarse grade.

The large particle cumulative curve (Fig. 4)
rises steeply, and is virtually a straight line up to
40 mm; above that it takes a slight bend, which shows
that the largest percentage lies in the pebbles size and
the number in the boulder size is very low.

Roundness

The roundness of the large particles (fig. 5)
varies between 0.15 and 0.8, there being no relationship
between roundness and size, as the lowest roundness lies
between 20 and 30 mm. and highest roundness between 30
and 40 mm. with the largest boulders showing 0.27 round-
ness.   The intensity contours show marked elongation
and concentration between 10 and 40 mm. with the highest
number of pebbles in any roundness at 20 mm.   The round-
ness of the particles drops below 0.3 between 60 and 80 mm.

The roundness and frequency graph (Fig. 9)
shows that the largest number of particles have 0.43
roundness, and the histogram shows that the largest
number have 0.4 to 0.6 roundness.

Sphericity.

Sphericity of the specimen (Fig.6) ranges from 0.32 to 0.9, with the highest concentration between 0.5 and 0.8. The sphericity and the frequency histogram (Fig 7) shows a mode between 0.7 and 0.8 but this peak is only slightly above the percentage for 0.6 and 0.7. This graph shows a sudden drop towards the higher sphericity but only a gradual fall towards the lower sphericity. The sphericity and the frequency curve is very irregular with the maximum at 0.7 with a second peak at 0.63.

The sphericity and the roundness contour (Fig 8) is most concentrated round 0.7 sphericity and 0.4 to 0.5 roundness. This graph also shows that for a certain grade of sphericity there can be a wide range of roundness.

Soling.

Of the 100 pebbles examined 16 showed soling, the highest being 0.62 and the lowest 0.02. From this outcrop all the particles bigger than 40 mm. were hand picked for farther study; a very high percentage of these show good soling and striations.

Field association.

Permian deposits from this area had been studied by Wells (1957) and McKellar (1957), and a full section of the deposits is reported by both of them.

McKellar is of the opinion (p 19) that these beds are possibly equivalent to McRae Mudstone of the Quamby Mudstone, and Wells has called similar beds Stocker's Tillite. The origin and the age of these deposits will be discussed in another section.

Measurements of King William Saddle Sample.

This specimen was collected at the contact between the Permian tillite and Precambrian quartzite, the basement is well striated and the striations indicate 5° north of west to 5° south of east direction of ice movement.

Mechanical analysis

The size distribution of the large particles (Fig 4) shows a rise between 20 and 40 mm., but another peak has developed between 80 and 160 mm. size. The largest boulders observed in the area were 320 mm. long and 100 mm. wide. The micro-mechanical analysis shows a rise towards finer grade and more than 58% particles are below 1/256 mm. size.

Microscopic Study in this section.

The slides are composed (Plate 12) of authigenic mica, angular and rounded quartz grains, quartzite, mica schist, and graphitic schist pebbles. Every slide shows bedded structures, sometimes containing fine and coarse grained matrix alternate. The authigenic mica and the pebbles show orientation along the bedding. Mica crystals have flow structure which surrounds the quartz particles.

Roundness.

The highest roundness (Fig. 5) is 0.9 and the lowest is 0.08. Roundness does not have any relationship to the size since both boulders and pebbles have a low degree of roundness. Between 80 and 120 mm. size the roundness and the number of the pebbles has increased but on both sides of this size grade the number drops,

although it increases below 60 mm. size. The intensity
contours show a concentration between 0.4 and 0.5 round-
ness and 30 mm. size. The frequency and roundness graph
(Fig.9) has several peaks but the maximum lies at 0.5
roundness, the second peak is at 0.4 but it is less than
half of the maximum. The histogram has two modes at 0.3 -
0.4 and 0.5 - 0.6 but the 0.4 - 0.5 grade is also quite
high. A minor mode has also developed at 0.1 - 0.2,
with more than 60% pebbles having roundness between 0.3
and 0.4.

Sphericity.

The lowest sphericity in this sample (Fig 6)
is 0.39 and the highest 0.89. There is no relationship
between sphericity and size. The intensity contours show
the highest concentration between 0.6 - 0.7 sphericity and
30 mm. size, with another increase between 90 and 100 mm.
size and 0.5 - 0.6 sphericity.

The sphericity and frequency graph (Fig. 7)
is very irregular and it has a maximum at 0.65 with minor
peaks at 0.61. On both sides of the peaks the number
drops gradually. The histogram has more than 46% large
particles between 0.6 and 0.7 while more than 70% are
concentrated between 0.5 and 0.7.

Soling.

Out of 137 large particles examined 52 showed
soling, the highest soling being 0.7 and the lowest 0.03.
The least soled particles were that of pink coloured
quartzite.

Field Association.

Basal Permian beds occur at Mt.Sedgwick, and
Mt. Read (Banks and Ahmad, 1958), and the nearest Permian
deposits are reported from Lake Meston area (Jennings and

Ahmad 1957). The thickness of the deposits at Mt Sedgwick
is comparitively less than other places and due to absence of
any overlying beds at King William Saddle, it is difficult
to estimate the extent and thickness of the deposits.

Measurements of Maydena Sample.

This specimen was collected from a road
cutting; here the tillite overlies the Owen Conglomerate
unconformably, and it is overlain by outwash beds. The
till has marine fossils which have been examined by Mr
N.G. Lane (p 15).

Mechanical Analysis

The largest boulder in the sample was 80 mm.
but the largest boulders seen in the area are about a foot
long and 10 inches wide. The specimen shows very good
sorting, So is 2.9, Log So is 0.46. The sorting co-
efficient of the deposit is 0.04375 and the skewness is
1.47. These results indicate that this rock is very
well sorted but the spread of the grain size does not
agree with these calculations. The size frequency
histogram (Fig. 2) has a mode at 1/64 -1/32 mm. and
two smaller modes at 1/16 - 1/8 mm. and below 1/1024 mm.
sizes. Particles below 1/16 mm.(Fig. 3) comprise more
than 71% of the sediment and only 12% are distributed
in the large particle grade. The size and intensity histo-
gram for large particles (Fig. 4) rises to a maximum at
20 mm. and gradually drops down to the larger sizes.

Roundness

The maximum roundness (Fig 5.) is 0.9 and
the minimum roundness is 0.14, size and roundness show-
ing no relationship. In the intensity graph the maximum
concentration is at 20 mm. and at 0.5 roundness. The
roundness and frequency graph (Fig.9) has a maximum at

0.5, with three other equal peaks at 0.42, 0.35 and 0.28 respectively. The histogram has a maximum between 0.4 and 0.5 with gradual drop on both sides.

Sphericity.

The highest sphericity (Fig. 7) is 0.92 and the lowest sphericity is 0.42, with maximum concentration between 0.6 and 0.7 sphericity and 30 mm. size. The sphericity and frequency graph has two equal modes at 0.7 and 0.65 and another pair of peaks at 0.62 and 0.6 sphericity. The histogram has a mode between 0.6 and 0.7, and it gradually drops towards both sides. The sphericity and roundness contour is elongated between 0.4 and 0.9 sphericity but this limited range of sphericity has a wide range of roundness, ie from 0.1 to 0.9.

Soling.

Out of 100 large particles examined, 7 showed soling, the least being 0.15 and the highest being 0.58.

Field Association.

The basal Permian tillite is reported from Styx and Weld Rivers (Lewis, 1923), while younger Permian deposits occur both east and west of this area, and are also reported from the Hobart district (Banks, 1957) and Florentine Valley.

In this area the tillite is overlain by Woody Island Siltstone beds (Banks, 1957) which have some fossils and glendonites.(Banks, 1955). Outwash beds with crude bedding and low percentage of clay and silt overlie the till deposits.

DISCUSSION OF THE RESULTS FROM THE SAMPLE STUDIED

The size distribution of the sediments under study is very wide and ranges from over 160 mm. to less than 1/1000 mm. The sorting coefficient of the rocks is between 2.9 and 13.2 and the median diameters are between 0.0156 and 0.0468. All the sediments have more than 50% particles smaller than sand size. The graphs are either bimodal or trimodal, with one maximum at 1/8 - 1/16 and the second at 1/32 - 1/64 or 1/64 - 1/128 mms., whilst the third peak in the Wynyard sample is in the boulder size, and in the Palmerston sample in the less than 1/1000 mm. size.

A comparison of the basal Permian deposits with other sedimentary rocks shows that these deposits are similar only to other glacial deposits. Shales, sandstone and other fine grained rocks differ from the basal Permian glacials in the following points :

1. They have a very small range in the sediment size.
2. All particles are smaller than 1/8 mm.
3. They are generally unimodal.

Gravel deposits and conglomerates have a wide range in the sediment sizes, but they are generally lacking smaller sized sediments, (ie. sand silt and clay). Sometimes the percentage of less than 1/16 mm. size is below one (Potter, 1955; Krumbein, 1942). The beach gravel analysis (Emery, 1955) shows a 750 - 10 mm. mid point (M), which again indicates a very low percentage of fine grained particles. Niino (1950) has described the submarine boulder-bearing beds from Wasaka Bay. These rocks have a wide range in sediment size but the distribution of the boulders is limited to selective places such

as on the top of the ridges; they are rarely found in depressions, and the boulders do not have any striations. Niino, after considering all other evidence has come to the conclusion that these deposits are derived from the deposits round Wasaka Bay. He has not given the detailed results of his mechanical analysis, so these cannot be compared with other deposits. Shumway (1953) has described the kelp rafter pebbles from off the coast of Baja, California; these are occasional pebbles occurring in fine sediments, but they have a thin coating of manganese which indicates the area as one of very slow deposition. In the Alaska Basin Menard has also reported similar deposits of ice-rafted pebbles with manganese coating, but the Baja, California, ones have Pholad-type holes and are found with red clay deposits whereas the Alaskan occurrence is in grey to yellow-grey mud. Emery (1955) has reported the occurrence of drift wood transported erratics from various parts of the earth. These erratics will possibly differ from glacially transported erratics in the following characters:

1. They will not have striations unless they are picked up from a till as described by Fairbridge (1952, p 141).

2. Generally these pieces will be angular unless they are from older deposits.

3. Generally these pieces will not be fresh, and will show effects of chemical weathering.

Hanglomerate deposits and the mud flows have a wider range of sediment sizes than the gravels. They have lenses of clay and sand and the boulders are rounded to sub-rounded, so similarities with the glacial deposits are suggested. The Big Horn Mountain conglomerate beds have been called glacial beds by various workers, but Sharp (1948) has come to the conclusion that this is

not so because :

    1. Only one possibly striated boulder was found.

    2. Only one boulder with possible chatter marking was found.

    3. The deposits are limited to mountain slopes.

    4. The age of the deposit is Pre-Pleistocene.

    Further evidence which denies a glacial

origin is:

    1. The lithology of the boulders in any one of the beds is predominantly the same.

    2. The beds rest on an angular unconformity.

Sharp has said that it is just possible that these deposits might have derived some glacial outwash but they were not of glacial origin.

        The glacial deposits contain large numbers of striated erratics and in the subglacial deposits, the proportion of fine sized particles is quite high. Tills do not contain rocks belonging to only one age or formation, unless they move on an area which was entirely covered by the same rock. Even in the Pleistocene tills of Tasmania where the ice mostly moved on a dolerite surface some erratics of other rocks are quite common in every deposit. Mudflow and fanglomerates are generally associated with diastrophic movements so they have limited distribution, but the glaciers are not related directly to any such movement and their deposits are widely distributed. The mudflows and other similar deposits have rounded sand-sized and smaller particles but the glacial deposits have angular particles in this fine fraction. Some deposits which may not be of glacial origin may resemble glacial deposits in some feature but the absence

of striated and facetted pebbles and rounded particles
of sand and smaller sizes coupled with limited distrib-
ution of these sediments will always help to distinguish
them.

The characteristic diagnostic features of
tills are as follows :

1. Heterogeneous composition and a very wide
range of sediment size.

2. Large particles are striated and facetted.

3. The rock particles remain fresh even a long
time after the deposition.

4. The large particles are generally rounded
but the fine grade particles are mostly angular.

5. The subglacial tills have oriented peb-
bles (Holmes, Till Fabric).

6. Some of the large particles have the
characteristic shape of the glacially transported errat-
ics. (Von Engeln).

7. The deposits may have bedding but the
grading will be absent.

8. The erratics will have a wide range of
lithology.

Although the above-mentioned characters
will help to establish glacial origin, some exceptional
deposits, especially reworked till, may also satisfy
these conditions, so it is possible for other deposits
to be confused with tills. Portmann (1956, p 421) has
said that moraines differ from aeolian, fluviatile and
lacustrine or marine deposits on the basis of their
heterogenous composition and lack of sorting, but they
may to some extent resemble deposits produced by mudflow,
landslides and similar phenomena. Flint (1947, p 111)

has said that other deposits will not have striated
pebbles unless they are derived from glacial rocks, which
will make the identification easy.

Basal Permian deposits of Tasmania have a
heterogeneous composition which is shown by the wide range
of sediments sizes and some of the deposits show poor
bedding similar to that of drift beds (Flint, 1947, p 102).
At least 20% of the large particles have striations,
facets and chatter marks (Plate 20), some of the large
particles have a good glacial pebbles shape (Plate 13),
(Von Engeln, 1930). The lithology of the erratics is
reasonably constant although sometimes the relative
number of the different types of erratics alter with the
changes in terrain. Microscopic study of the rock shows
that the sand, silt and clay size particles are quite
fresh and angular. The erratics are also fresh and even
the shale and mudstone pebbles show no effect of weath-
ering. The King William Saddle and Mt. Sedgwick deposits
rest on glaciated and striated pavements, and Wynyard
deposits have till fabric. From this evidence the glacial
origin of these beds is undoubtedly proved, but the
difference in their character suggests that they have
been deposited under different environments.

Trask (1930) has determined that a sorting
coefficient So. equal to 3.0 indicates normally sorted
sediments, and with So. above 4.50 shows a poorly sorted
sediment. The So. value of the Wynyard and King William
Saddle deposits are above 4.50 but the So. value of the
Palmerston sample is 3.7 and that of the Maydena sample
is 2.90. These values indicate that Maydena and Palmer-
ston sediments are normal to poorly sorted sediments,
but the range of sediment sizes is definitely wider than

even poorly sorted sediments. Krumbein (1933, p. 388)
has published the results of 48 mechanical analyses of
tills. The sorting coefficient of these was calculated
from the quartiles by the author and found to be as low
as 2.4. These results are not in accordance with the
accepted concept of the coefficient of sorting, and the
values of So. as given by Trask cannot be taken as final
for all tills. In the tills, there is generally a max-
imum in the middle grades. If this maximum is high in
comparison with other points it affects the value of the
coefficient.

The mechanical composition of the tills
depends on the terrain over which the glaciers pass,
and the environment in which they are deposited, (Flint,
1947, p 115)(Portmann, 1956, p 422). In the hard rock-
country the till will have more coarse grade material
but in soft-rock country it will have more fine grade
material. The second factor which can increase or de-
crease the proportion of the finer grade particles is
the sorting of the tills. Results of mechanical analysis
of tills published by Krumbein, (1933), Holmes (1952),
Shepp (1953) and others show consistency in the size
distribution and in the amount of material of any
particular size, when it is deposited under similar
conditions and in the same area. The mechanical analysis
results of the Permian beds in Tasmania shows a gradual
increase or decrease in the proportion of the fine grade
sediments with possible changes in the environment. The
Wynyard deposit has 45.8% of silt and smaller size part-
icles, Maydena has 58.8%, the Palmerston deposit has
71.3% and an analysis of the Woodbridge beds (type area)
carried out by Mr. Woolley has 61% particles smaller than

1/16 mm.

Comparing the other marine and land glacial deposits with the Tasmanian Permian deposits it appears that those at Wynyard are of the ordinary till type deposited as subglacial moraines in contact with the ice; the Maydena tillite, with broken marine fossils, was possibly deposited in the water but very near the shore; the Palmerston beds were possibly deposited under the ice but in deeper water and the Woodbridge beds (other than type area) with a larger number of fossils but higher proportion of sand were possibly deposited in the outer part of the pack ice zone. Evidence in support of the above conclusions will be left until other factors have been discussed in the next section.

The till fabric analysis of the Wynyard deposits (Plate 14) near Doctors Rock, shows a concentration of pebbles with their longer axis pointing from 30° west of south to 30° east of north but from the stereogram (Plate 14 B) of the pebbles near the mouth of the Inglis River the concentration is from south to 15° west of south to 15° east of north. Although there is no obvious difference in the till from the two areas, the apparent discrepancy in the direction of the ice movement may be explained as follows:

1. These are two different tills belonging to two different periods with two distinct directions of ice movement, or

2. This altered direction of the ice movement is due to local variations in the topography. As previously stated, the till from the two areas does not differ lithologically, and other evidence also indicates

that probably both belong to the same age, but there is
reason to believe that the difference in the orientation
of the pebbles is due to dissimilarity of topography.
The Doctors Rock analysis was made within five feet of
the unconformity and possibly it was the walls of the
valley in which the ice moved, that affected the direct-
ion of the ice movement. The Inglis River stereogram
has a girdle between 20° and 40° in the south west corner,
and the beds in this area dip with an angle of 15° towards
south west. The Doctors Rock sample does not have any
pronounced concentration of pebbles so it lacks a girdle.

Holmes (1941) in his paper on till fabric
has come to the conclusion that most of the pebbles in
the ice are rotated with their longer axis at right angles
to the direction of the ice movement but along the ice
till contact, this movement is changed from rotation to
gliding so that the largest percentage of the particles
have their longest axis parallel to the direction of ice
movement since this position offers the least resistance.
He has also postulated that generally the till is stat-
ionary and it is the ice layers only which move. If
this is true it will mean that once a thin layer of till
has been deposited at the base of the glacier, there will
be no movement on the basement, so there is little chance
of its being striated or fluted unless this is done by the
glacier before the deposition of the till. The layers
of till and ice are not separate bodies which can be de-
marcated in the glacier. From the top to bottom the
proportion of debris gradually increases; sometimes ice
and rock debris are in layers, so that at the base of the
glacier it has very little ice. The speed of the ice
layers also decreases from the top to the bottom, so that

at the base it is at a minimum (Seligman, 1947, p 14,
Portmann, 1956, 424).

If the inclination of the beds is compen-
sated in the stereogram of the till fabric at Inglis
River (Plate 14 ᴿᵇ.), the girdle shifts to between 8^o and
28^o, but the concentration of the elongation of the
pebbles will remain in the south west. This stereogram
indicates that at least in this case the most stable
position for the pebbles during transport was 10^o to 20^o
dipping against the direction of the ice movement.

If we postulate four pebbles in the ice
with their longest axis parallel to the direction of the
ice movement, we will see that (Fig. 1) pebble A is in
the most unstable position, and pebble D is in the most
stable position, whereas pebbles B and C are in an inter-
mediate position. Of the pebbles B and C, C is the more
stable, as the ice moving forward will push pebble B
into the A position, whereas the ice will flow over
pebble C and it may be left in that position for a longer
time. On the whole, pebbles C and D will stay longer in
that position than pebbles in A and B positions. All the
time the pebbles will have a tendency to rotate along
their intermediate axis, but the pebbles in the adjacent
layers will exert a force on the pebbles rotating along
any other axis, as they will be moving at different speeds,
and bring it to a new position so that its longer axis
points in the direction of ice movement, as this will be
the position of least resistance with respect to the ad-
jacent pebbles. The Doctors Rock stereogram does not
show any concentration of pebbles in any one position,
except for the two in the south-west and north-east mar-
gins and one in the centre. A very vague girdle in the
southern semicircle represents the concentration of the
pebbles in this stereogram which may be due to:-

1. This stereogram represents a different till, which was deposited by ice moving in different directions from the previous one.

2. Due to the proximity of the valley walls some pebbles were orientated to the vertical shearing of the wall and they took up the vertical position.

3. Some of the pebbles are from the super glacial moraines.

It is very difficult to say which of the three is the true explanation but possibly the first does not hold good in this case as possibly both the studies were made on the same till.

The roundness of the erratics varies from 0.05 to 0.9, and this factor is not related to the size or lithology of the pebbles. The large particles smaller than 60 mm. show all degrees of roundness. Between 60 and 80 mm it is below 0.4, between 80 and 130 mm it increases to 0.5, and above 130 mm. the roundness drops to 0.3. The largest number of erratics with high roundness are between 20 and 40 mm. size. An analysis conducted by Mr Lane on the large particles from the Permian beds of Lindisfarne also shows nearly the same characteristics in the size and roundness of the pebbles (Fig.11).

These roundness results suggest that during transport the erratics were not only rounded but also crushed under glacial stress. If the erratics were subjected to the process of rounding only, the roundness should have increased with the decrease in size, but this does not appear to be the case. The roundness has stayed fairly uniform as long as the large particles were bigger than 130 mm. This indicates that the large

particles are more susceptible to crushing when they
are bigger than 130 mm. The large particles between 80
and 130 mm. have mostly less than 0.5 roundness; this may
be due to:

1. The erratics between 80 and 130 mm. are
more effectively crushed when their roundness is above
0.5.

2. The erratics larger than 80 mm. are not
crushed when their roundness is between 0.3 and 0.5.

Either or both of these reasons may be
responsible for crushing and there is no particular
reason to favour either of them.

The erratics between 80 and 60 mm. are less
rounded because they represent the freshly broken pieces
of the higher sizes and possibly large particles of this
size are very quickly rounded, and at the same time re-
duced in size so that none of the higher roundness are
left in this size. Below 60 mm. the roundness probably
increases to the maximum which in the glacial deposits
possibly does not greatly exceed 0.9, and in these
grades, pieces with lower roundness are possibly added
due to crushing of the larger particles.

As all these samples were taken from very
near the basement it indicates that possibly the movement
of the glacier is not limited to the layers of the ice,
but the moraine layers also move to a certain extent,
otherwise the process of crushing would not be effective
in these layers.

During the analyses, it was found that some
erratics could be reassembled by cementing the pieces to-

gether. Some of these erratics had possibly been crushed
just before lodgement and had not been deformed or abraded
very much. In each case there were clearly fragments of
one original boulder broken in the ice by the shearing
movement of the ice and deposited separately but near
enough together to be included in the same quarter ton
sample (Plates 15-20). The shape of these boulders in-
dicated that they had been well rounded (Plates 18-20)
before they were crushed and some of them showed very
good striations (Plate 16) on the original exterior,
and sometimes the new faces were rounded (Plates 15 & 17).
None of these boulders was less than 80 mm.(after assemb-
ly). It is rather difficult to say whether the erratics
over 80 mm. were the only ones which were crushed in
this way or whether the smaller ones were also crushed,
but it may be possible that erratics larger than 80 mm.
have been more affected by stress, as is also indicated
by the roundness study of these sediments.

The effect of the glacial stress on erratics
of different composition varied. Mudstone erratics were
fractured in three directions (Plate 16) with two planes
of fracture at right angles to each other and the third
making an angle between 30° to 40° with the two planes.
Quartizitic sandstone erratics were fractured along a
plane diagonal to intermediate and shorter axes (Plate 21),
so these pieces took on a half moon shape. Sometimes
the quartzite erratics were fractured along planes parallel
to the intermediate and shorter axes, but in most cases
these were sheared in any direction. The shale, slate
and schist erratics were sheared along bedding plane
and plane of schistosity respectively, so most of the

erratics were flat and discoidal in shape.

This selective effect of shear on the er-
ratics of different composition may be due to inherent
lines of weakness in them or it may be due to the sel-
ective effect of the glacier on the different types of
rocks. It is not easy to see how a boulder embedded in
ice flowing by laminar flow can develop sufficient stress
to cause elastic fracture, since both the viscosity and
the rupture strength of the boulder would normally be
greater than that of the ice. It is true that the shear
force involved in laminar flow rises (not linearly)
with the rate of flow, and also true that as the proport-
ion of boulder, sand and clay detritus in the ice increases
so does the average viscosity of the ice. Nevertheless,
the shear force in the ice should not normally ever reach
the rupture strength in the boulders. Perhaps the rupt-
ure mechanism involves the interaction of two or more
boulders, all under shear load from the flow pressing on
a third to the extent that the third suffers elastic
failure.

The sphericity of these erratics is higher
in comparison to the roundness and it varies between 0.3
and 0.92. The sphericity like roundness, is not related
to size and composition. The largest number of the
erratics have sphericity between 0.6 and 0.8 and the most
spherical erratics are between 20 and 40 mm. The spher-
icity of the erratics was not affected during transport,
and even large erratics have high sphericity, but the
highest sphericity is among pebbles of 10-20 mm. size.
It appears that in the glacially transported material
the sphericity is not related to roundness but the

erratics with the highest roundness will also have highest sphericity, and the inverse relationship does not hold.

The soling analysis of the erratics did not prove as profitable as it was expected. The areas on all sides of the exposures from which the specimen were collected are covered with younger rocks, so it was difficult to establish the relationship between the distance travelled and the extent of soling. The soft rocks are definitely more soled than the hard ones, but again it could not be established that the softer rocks had travelled less than the harder ones.

CORRELATION OF BASAL PERMIAN DEPOSITS.

The basal Permian deposits are diverse and
are of different ages. They vary from tillite to con-
glomerate, shale and sandstone.

Earlier authors who have described these
beds have sometimes used the same terms for different
types of deposits, so it has been found necessary to
discuss the regional section of the basal part and to
establish its position with respect to time of depos-
ition.

Three sections from three different parts
of Tasmania are reproduced for the correlation and the
deposits in the different areas are matched with the
nearest section reproduced.

There are some differences in the correlation
of these sections so they are discussed below and in the
absence of any other method the deposits have been cor-
related on a lithological basis only.

Voisey (1938, p 329) has correlated the
fresh water beds from Elephant Pass with the pelionite
and tasmanite deposits of the Mersey areas. This cor-
relation is not accepted for the following reasons:

Voisey correlated the tasmanite deposits
with coal beds on the basis of Reid's (1924, p 43) state-
ment that tasmanite deposits are the marine facies of
the coal bearing beds and consequently they do not occur
one over the other. From Bore Hole 15 Reid reported
that the coal seam occurs 446 ft. above the tasmanite
deposit. Wells (1957, 9-10) observed that tasmanite
beds occur 300 - 400 ft. below the coal-bearing beds.
Banks (1955, p 209) reported the occurrence of tas-

manite 50 ft. above the tillite at Oonah.

All this evidence indicates that the tasmanites are about 300 - 400 ft. below the coal seam; perhaps Reid failed to notice this.

Reid (1919, p 76) reported the pelionite bed from the unfossiliferous beds above the tillite on Barn Bluff, which are probably equal to Quamby Mudstone beds of the Deloraine area. Voisey had correlated these beds with the Mersey Coal Measure beds on the basis that no carbonaceous beds were reported, from rocks younger than Lower Coal Measures. Hills et al(1922, p 232), proposed that the Preolenna Coal seam is equivalent to the Mt. Pelion coal seam which according to Reid is about 600 ft. above the pelionite beds.

Wynyard Area.

Stephens (1869 p 17) first reported the occurrence of the tillite from this area. Twelvetrees (1906) and David (1907) studied the deposits on the shore platform at Wynyard and identified them as Permian in age. Hills (1913) and Banks et al. (1955 b) studied the deposits inland south of the shore platform and reported the occurrence of tillites, varves and other younger Permian beds.

On the shore platform the basal Permian beds overlie the Precambrian schist unconformably and are overlain by outwash beds. These deposits have a rolling dip with a maximum of 15^{o}, and the direction of the dip changes from west to east. David noted that the base of the tillite is exposed east of the Doctors Rock and the thickness of the beds is /220 ft.

Hills reported a full section of Permian from Preolenna (p 17) and has followed David in proposing that the tillite at the base is / 220 ft. thick.

Banks noted that at Oonah the tillite (250+ ft.) is overlain by 50 ft. of sandstone and siltstone followed by tasmanite beds. The top of the beds is not exposed so the section is not complete.

Tasmanite shales	2 ft.	
Siltstone	20-40 ft.	
Sandstone	18 in.	Rare
Siltstone	1 ft.	Marine Fossils.
Sandstone	2 ft.	
Tillite	250+ ft.	

North of Oonah and South of Doctors Rock varves and tillites occur at several places.

David (1950 p. 314) proposed an upper Carboniferous age for the tillites at the shore platform and Banks proposed an Artinskian or Sakmarian age for the tillite at Oonah. Probably the age assigned by David is a little older and there is no direct evidence to support this age. If the stratigraphic position of the tillites at Wynyard proposed by Hills (1913) and its correlation with the Mersey District tillite proposed by Voisey (1938, p 329) are accepted the age of the Wynyard tillite can be correlated with the tillite beds at Oonah. Banks (1955, b) reported a tasmanite bed 50 ft above the tillite at Oonah, and Wells (1957) noted one 50 ft. from the base of the Permian beds, and near the base of the Quamby Mudstone which conformably overlies the tillite in the Deloraine area.

Due to inadequate description of the beds
from Preolonna, it is not possible to correlate these
with other areas but it can be suggested that the fossil-
iferous mudstone below the Coal Measures is equivalent
to the Golden Valley Group, the non fossiliferous mud-
stone is equivalent to Quamby Mudstone and Palmerstone
marine tillite, and the basal tillite in the Oonah area
and on the Wynyard Shore platform is equal to Stockers
Tillite.

In the Wynyard area on the east side of the
exposure the basal tillite rests on the Precambrian beds
which gradually rise towards the east. The rise in the
height of these rocks is more than the western dip of
the Permian beds. In the absence of any post-Permian
faults it appears that in the east the Permian beds are
exposed along a wall of the Permian valley which was re-
corded by David as the basement.

Deloraine, Palmerston and Western Tiers Area.

The geology of this area was described in
detail by Wells (1957), McKellar (1957) and Voisey (1948).

Voisey reported the occurrence of the
basal tillite at Park Nook and McKellar called the lower-
most Permian beds in the Western Tiers area "Tillitic
Conglomerate". Wells called the lowermost beds "Stock-
ers Tillite", but he has mentioned that these beds do
not have any striated or facetted pebbles.

A problem of nomenclature arises here. The
formation which Wells has called the Stockers Tillite is
not all tillite, but it "grades imperceptibly into the
Quamby Mudstone" (Wells, 1957, p 8). Likewise the form-

ation which McKellar has called Stockers Formation is not
all tillite. The author has examined the bore cores
drilled through this formation and described by McKellar,
and found that the upper 5 ft. has the characters of the
Palmerston beds described on pages 13-14, whereas the
lower part of the formation is tillite, which is strictly
comparable with the Wynyard Tillite, and which is foll-
owed upwards by formation which overlies the Wynyard
Tillite at Preolenna. The author therefore proposes
to divide the Stockers formation of both Wells and Mc
Kellar into Wynyard Tillite below, and the Palmerston
Marine Tillite above, and to include in the latter the
lowermost 10 or 20 feet of the Quamby Mudstone formation,
which appears to be more appropriate so included. The
sequence then becomes :

      3. Quamby Mudstone.

      2. Palmerston Marine Tillite.

      1. Wynyard Tillite.

         Glaciated Pavement or unconformity.

     A complete section of the Permian deposits as
reported by Wells and McKellar is reproduced and it is
accepted as a whole except for the above discussed change.

Mersey District.

     The geology of the area has been studied
from the bore core logs published by Reid (1924), and
due to lack of sufficient data, it has not been possible
to correlate the beds very satisfactorily.

     In this area tasmanite occurs about 50 ft.
above the base of the Quamby Mudstone and it is so far
reported from only one horizon. The coal-bearing beds
occur about 300-400 ft. above the tasmanite, and have

been correlated by Banks (1957) and Voisey (1938) with other areas. Therefore these two beds have been used as markers for the correlation of the deposits.

In Bore Cores, 12, 13, 14, 15, 22, and 27, tasmanite beds occur at various heights above the unconformity. In bore holes 12, 14, and 22, the fossils occur at very nearly the same depth from the tasmanite beds, but the second band of fossils in Bore 12 is absent in core 22 and core 14 does not reach to that depth. The rhythmic nature of the beds, with mudstone and pebbly mudstone alternating in all the cores indicate that possibly up to some depth below the tasmanite the same beds are present in all the bore cores. At the base some of the cores have conglomerates, others have mudstone, but core 27 has tasmanite at the base.

These beds below the tasmanite indicate that the difference in the sediments at the base is possibly due to differences in the horizon which rests at the base, otherwise they can be correlated with one another.

The beds below the tasmanite are correlated with Quamby Mudstone and Palmerston Drift beds. The variations in the thicknesses could be due to difference in the rate of sedimentation at different times and in some cores tillite and Palmerston Marine Tillite may also be present but cannot be identified. The beds below the coal bearing beds and above the tasmanite are correlated with lower Liffey beds and Golden Valley Group of the Wells section.

Northeast Coast.

The basal beds of this area have been named by Voisey (1938, p 323) the St. Mary's Basal Stage, and

this is reported from St Mary's District, Avoca, Fingal, Dalmayne, Picanini Creek, Seymour, Llandaff, and the Mt. Paul area.

The basal stage has conglomerate, which contains some striated and facetted pebbles. It is 10-15 ft. thick and it is overlain by current bedded sandstone, (Plate 26), which passes into coal-bearing beds. The total thickness of the basal stage is 100 ft. and it is overlain by Gray's Stage which has limestone, sandstone, mudstone and shales, with Aviculopecten sprenti, Spirifer strzelecki, S. tasmaniensis, Terrakes fragile, and other fossils. Voisey has correlated the Gray Stage with the Upper Marine beds of the Mersey area and Banks (1957) also correlated the carbonaceous beds with the freshwater beds of the Quamby Brook area.

The grading of the conglomerate at Elephant Pass into carbonaceous beds indicates that they are younger than the basal tillites of the other areas, but they may have been derived from some glacial deposits. However, Voisey has correlated the basal stage at Elephant pass with the basal tillite but expressed his doubt about their age due to the difference in the nature of the beds.

On the basis of the above evidence it is proposed that the basal beds at St Mary's and Elephant Pass are possibly younger than tillite but older than the freshwater beds. They are probably equal to the lower beds of Liffey Sandstone or the Upper beds of the GoldenValley Formation.

Lilydale-Beaconsfield Area.

The occurrence of the Permian deposits here

was first reported by Twelvetrees (1918), and later by
Voisey (1938). Banks (1957) noted that the basal tillite
is overlain by unfossiliferous mudstone followed by
Limestone with Eurydesma cordatum, and Calcitornella
stephensi. Voisey (1938) reported that the basal beds with
mudstone conglomerate (Quamby Mudstone) are 100 ft.
thick, and they have a conglomerate at the base. This
section suggests that either the basal tillite is absent
and Quamby Mudstone rests on the base or it overlies a
conglomerate and it is overlain by beds of Quamby Age.
The beds of the Quamby age are followed by Darlington
Limestone beds.

Twelvetrees noted that a bed of tasmanite
overlies the freshwater beds, but Nye (1924) expressed
his doubt about its being tasmanite.

Green (1957) reported that at Beaconsfield
Quamby Mudstone beds have glendonites and the base is
not exposed.

Central Plateau.

Dairy Plains.

A Tillite is exposed in a limestone cave on
the property of Mr Clark at Dairy Plains where it over-
lies the limestone unconformably but the overlying beds
have been removed by erosion. On lithological basis,
this tillite is correlated with Stockers Tillite in the
Deloraine area.

Upper Mersey Valley.

In the Upper Mersey Valley a white con-
glomerate with quartzite pebbles overlies the Precamb-
rian schists and quartzites unconformably. This bed is
about 10 - 20 ft. thick and is exposed near Snake Creek
and under Clumner Bluff. In the Snake Creek area it is

probably overlain by beds of Quamby Mudstone age but in the Clummer Bluff area it overlaps the beds equivalent to Liffey age.

The age of this bed is perhaps different in different areas but it is younger than the Wynyard tillite.

King William Saddle

At King William Saddle a tillite rests on Precambrian Quartzite. Its age and lithology have been discussed on page 14 and it is of Wynyard tillite age.

Western Tasmania

Mt Sedgwick.

The occurrence of the Permian deposits here was first reported in the last century and the section has been studied by Banks and Ahmad (1958). Here the basal tillite rests on porphyries of the Dundas Group. The basement is striated and the direction of ice movement is indicated from west to east. A roche moutonnée-like structure with the plucked face towards the east is seen in the middle of the exposure. Towards the north and the south the Cambrian is exposed about 200-300 ft. above the base of tillite and it appears that in this area the tillite was deposited in a small valley.

The complete section of the Permian beds is reproduced in a separate paper, but the basal section is reproduced here :

      4. Limestone beds (possibly Darlington.)

      3. Mudstone without fossils at the base, erratics present.

      2. Conglomerate (outwash).

      1. Basal Tillite.

The tillite is possibly equivalent to
Wynyard Tillite and the conglomerate bed may be equiv-
alent to the Palmerston Marine Tillite. The beds between
the Darlington Limestone and the Marine Tillite are poss-
ibly equivalent to Quamby beds.

Mt Read.

The basal Permian beds were first reported
by Montgomery (1896) and during a recent study of the
west coast Permian deposits, a mudstone bed with crinoid
stems was found on the top of the tillite (Banks and
Ahmad, 1958). At one place a set of striations running
from north to south was found along a logging track. At
this place the basement was dug but the quartzite at the
base was so decomposed that no trace of any striations
could be seen. North and south of this point the tillite
is overlapped by the mudstone beds and it appears that
the valley in which the ice moved extended from west to
east or vice versa. If either of the two was the dir-
ection of ice movement, it appears improbable that the
north - south striations would have been carved by the
glaciers. At some places such striations have developed
where the logs have been dragged on the same surface
and it is possible that these also may be the drag marks.

The tillite in this area is correlated with
the tillite at Mt Sedgwick. ie. with Wynyard Tillite and
the mudstone bed with Quamby Mudstone beds.

Mt Pelion Area.

Benson (1916, p 31), Reid (1919, p 27),
Hills et al (1922, p 236) reported the occurrence of
Permian deposits in this area.

The section as constructed by Hills (1922)
is as follows:

1. Freshwater Beds with coal seams   ?
2. Mudstone with marine fossils   300 ft.
3. Mudstone with pelionite
   (unfossiliferous)   ?
4. Conglomerate and tillite   100 ft.

Banks (1957) suggested that the freshwater
beds are equivalent to the freshwater beds of Quamby
Brook area. The beds between the freshwater beds and
the pelionite is possibly equivalent to the Lower Liffey,
Golden Valley and Upper Quamby beds of Quamby Brook,
and the beds below the pelionite is equivalent to Lower
Quamby beds and tillites. The conglomerate beds with
the tillite may be equal to outwash beds at Mt Sedgwick
or the Conglomerate at the base of the Permian deposits
in the Mersey area.

Point Hibbs.

Hills (1914) reported the occurrence of
Permian beds at Point Hibbs, and the detailed section
was studied by Banks and Ahmad (1958). Permian rocks
with vertical dip are in contact with the Devonian (?)
Limestone dipping at 45° to the east. Permian beds
strike north-south and face west. The contact between
the Permian and the limestone may be a fault contact,
or a steeply tilted unconformity.

The section at Point Hibbs is as follows:

4. Thin bed of Limestone (May be Darlington)
3. Mudstone with fossils and alternating
   bands of erratics and mudstone.
2. Mudstone with erratics grading into
   tillite towards the base.
1. Outwash Conglomerate.

The outwash conglomerate at the base is non-fossiliferous and grades into tillite. The tillite grades into the mudstone which is non-fossiliferous at the base, but the fossils appear near the top and there is a large number of fossils in the erratic bands. Glendonites are fairly common near the top of the mudstone but they were not observed near the base.

Although the limestone band did not have any index fossils it is very near the position which would have been occupied by the Darlington Limestone. Glendonite-bearing beds are reported from Woody Island, and Banks et al (1955, a) suggested that they are near the base of the Permian beds. In other areas they are reported only from the Woody Island siltstone beds. The tillite and outwash beds are possibly equal to the Wynyard tillite, and the beds with mudstone and erratics are equivalent to Woody Island Siltstone. The Limestone is equal to the Darlington Limestone but in the absence of any other fossils evidence, this correlation is only tentative.

Zeehan and Pieman Area.

Moore (1896) reported a tillite from Zeehan and this has been studied in detail by Spry (Perso.Commu). The tillite overlies Precambrian rocks and the rocks above it have been removed by erosion.

Spry (Person.Comm.) reported the occurrence of a tillite from the Lower Pieman Valley. This tillite is cut by a dolerite dyke so it must be older than Mesozoic dolerite, but due to lack of any other evidence its age is not definitely established.

Due to lack of knowledge of the ages of

these two tillites they are not discussed with these
deposits, and it is possible that they may be older
than Permian.

Maydena and Woody Island.

Maydena.

Banks (1957) reported a tillite resting on
a Lower Paleozoic base. The tillite is about 200 ft.
thick and it is interbedded with three groups of varves
near the top. The varve groups are 3-5 ft thick and
they do not show any slumping or dragging. The section
is as follows :

        10. Woody Island Siltstone.
        9. Outwash Conglomerate.
        8. Varves.
        7. Tillite.
        6. Varves.
        5. Tillite.
        4. Varves.
        3. Tillite.
        2. Sandstone.
        1. Tillite.

        Lower Paleozoic.

The tillite is overlain by Woody Island
siltstone but the contact could not be seen and it is
possible that a fault runs between the top of the out-
wash beds and the Woody Island Siltstone so that the
beds between the two are missing.

The tillite is correlated with the Wynyard
tillite. Banks (1957) correlated the glendonite-bearing
beds with the Woody Island Siltstone beds, which makes
the outwash beds equivalent to Palmerston Marine Tillite.

Similar tillite and outwash beds have been reported by Lewis (1924) from the Styx and Weld river Valleys and they are tentatively correlated with the basal beds at Maydena.

Woody Island.

Banks et al (1955, a, p 219-229) reported the Permian section from this area as follows :

8. Eurydesma Limestone (Darlington Limestone).
7. Dreamy Bay Tillitic Sandstone.
6. Johnston Point Sandstone. (Glendonites).
5. Lewis Point Siltstone and sandstone.
4. D'Entrecasteaux Tillite.
3. Satellite Siltstone.
2. Sunset Bay Sandstone (Glendonites).
1. Woody Island Siltstone.(Glendonites).

The occurrence of Darlington Limestone on top of the Dreamy Bay Tillitic Sandstone indicates that the beds below it are equal to the Lower Golden Valley Formation of the North. Banks (1957) correlated the Woody Island Siltstone beds with the Quamby Mudstone beds.

Southern Tasmania.

Shoemakers Point.

Twelvetrees (1915) reported the occurrence of mudstone conglomerate resting on Precambrian rocks in this area and observed that the deposits have local boulders.

"Mudstone Conglomerate" has quite often been used by earlier writers to describe basal tillite, the rocks of the Palmerston Marine Tillite beds and Quamby Mudstone, so from this description it is not

possible to identify the origin of the deposit. In view
of the report that it has locally derived boulders and
pebbles it is suggested that it may be of either age
but since it is non fossiliferous, it may be of Palm-
erston Marine tillite or Wynyard tillite age.

Ida Bay.

Two miles inland from Ida Bay on the old
tram line of the Lune Timber Company, a Permian bed is
exposed under the fifth bridge.

This deposit has a few unidentifiable fossils
and a high percentage of rounded pebbles which are less
than 5 cms in length. The cementing material has more
than 15% of Ca CO_3 . The overlying rocks are not exposed
due to thick vegitation.

Because of lack of any evidence the age of
these beds cannot be determined but on a lithological
basis it is proposed that they are of Lower Permian age.

Woodbridge.

At Woodbridge a non-fossiliferous mudstone
containing abundant soled, striated and grooved pebbles
is exposed along the Channel Highway. The overlying
beds have been removed by erosion and the base is not
visible. Lewis (1924) correlated this deposit with what
is known as the "Woodbridge Glacial Formation" of other
areas.

The relation of the group of strata known
as the "Woodbridge Glacial Formation" to these beds at
Woodbridge which are allegedly the type for the formation
is being investigated by Mr N.G.Lane, and there is con-
siderable doubt whether they are the same. The "Wood-
bridge Glacial Formation" contains Fenestella, Avicul-
opecten, Spirifer and other fossils in large numbers

but the deposits at Woodbridge are non-fossiliferous.
The bedding in the "Woodbridge Glacial Formation" is
thick (plate 22) but the beds at Woodbridge are thin.
Lithologically, also, the deposits at Woodbridge are
different from those of the "Woodbridge Glacial Form-
ation". The mechanical analysis results (Woolley) have
more than 61% of sediments finer than clay, which suggests
a similarity with Palmerston Marine Tillite.

The stratigraphic position of the deposit
at Woodbridge cannot be suggested but in the absence of
fossils and the difference in the lithology it is pro-
posed that possibly the beds are older than the "Wood-
bridge Glacial Formation" and perhaps equal to Palmer-
ston Marine Tillite.

Maria Island.

Lewis (1938, p 433) reported the occurrence
of basal tillite at Reidle Bay but Banks (1957) observed
that in this area the marine beds overlie an arkose
cliff-breccia which in turn overlies a granite. Banks
also reported that in this area the granite shows a re-
lief of 50 ft. and it is quite irregular. In the north
of the Island near Darlington the lowermost Permian
bed is an erratic bearing marine conglomerate with
bryzoans, Eurydesma and other fossils. From the
Reidle Bay area Lewis also reported the occurrence of
Eurydesma and Strophalosia within 50 ft. of the base
of the Darlington Limestone on top of the tillite.
The presence of the Darlington Limestone on top of
the tillite indicates a younger age for the deposit. This
would place it somewhere in the age of the Quamby Mud-
stone age.

The local evidence with respect to Maria
Island can be interpreted in two ways:-

1. The glaciers did not reach Maria
Island, which was an ice-free island or peninsula in
the sea beyond the terminus of the glacier during the
time of deposition of the Stockers Tillite. Subsequently
(during Quamby Mudstone times) the general subsidence of
the shelf led to the transgression of the sea with pro-
duction of shore-line arkoses followed by marine beds
with fossils and erratics from icebergs derived from
the now reduced glaciers.

2. Maria Island was a buried hill on the Lower
Permian landscape which was either completely overwhelmed
by the ice sheet or which was a nunatak projecting
through the ice. Till may or may not have accumulated
on its flanks according to the regimen of the local
glacier movement. The pulsating warming of the climate
which characterised Stockers-Quamby-Darlington times
caused the ice barrier to retreat back to the west of
Maria Island, which with the isostatic uplift due to
ice load reduction, became an island in the sea, sub-
jected to subaerial erosion which destroyed the evidence,
if any, of the earlier glacial cover. During later
Quamby times the progressive regional subsidence led
to the gradual submergence of the island, with basal
shore-line breccias and conglomerates and arkoses,
followed by conglomerates with fossils, then Quamby and
Golden Valley type marine sedimentation with abundant
marine fossils and erratics dropped from drifting
icebergs. At this time the Coles Bay granite mountains,
and the granite mountains at Llandaff were still islands
on the Permian Sea. Their foothills were progressively
submerged, but Llandaff at least remained an island
until Triassic times.

No evidence can be won from the Maria
Island area to choose between these two interpretations,
and only circumstantial evidence can be found from
outside. The fact that during early Permian times the
glacial conditions are found as far north as northern
New South Wales (Lochinvar horizon) and that the direc-
tions of ice movement on the pavements at Inman Valley,
Bacchus Marsh, and in Tasmania are consistent with a
single large cap with radial movement, lend colour to
the view (but do not prove) that the icesheet of the
Stockers Tillite probably completely overwhelmed Maria
Island.

Again if we compare the kind of sediment-
ation over the Maria Island buried hill when the effect
of the hill on the sedimentation was no longer signif-
icant, with the kind of sedimentation in adjacent regions
where the basal tillite is present, some pointer to the
climate at Maria Island at the time of till deposition
should be obtained. Thus by the time of the Darlington
Limestone, the effect of the Maria Island buried hill
seems to have ended. This limestone contains erratics
from floating icebergs just as does the equivalent lime-
stones at Hobart and Golden Valley, so the climate
could not have been appreciably different. However
there is a significant increase in lime content in the
sediments towards Maria Island not only on the Darling-
ton horizon but in the succeeding formations. This
probably implied that the water at any rate was some-
what warmer there.

Ross Island off the present Antarctic ice
sheet may be cited for comparison. David and Priestley

(1914) have described remnants of the former extension of
the main Antarctic ice sheet over this island leaving
a deposit of till which contains erratics derived from
the mainland. Subaerial erosion is at present actively
destroying this till and remoulding the landscape. If
after a lapse of time this island became submerged and
then buried in periglacial marine sediments, the sequ-
ence would be shoreline breccias and shingles, then
near-shore deposits, then marine drift with fossils and
iceberg transported pebbles and sand, then perhaps lime-
stone - exactly as we find on Maria Island today.

We must conclude that there is no definite
evidence to decide whether the basal Permian glaciers
covered Maria Island or not. We must reject as unproven
the claims of those who have recently argued that the
glaciers could not have extended to Maria Island. The
present writer considers that it is rather more prob-
able that they did, but he recognises that this too is
not proved.

THE ENVIRONMENT OF THE PERMIAN GLACIAL DEPOSITS

OF TASMANIA

The Permian basal deposits include all those deposits which are at the base of the Permian System, irrespective of their horizon or their mode of deposition. These beds are overlain by marine beds with Spiriferids, Fenestellids, Bryozons, and other fossils, so their age is easily established, but due to lack of index fossils their stratigraphic position in the Permian System is sometimes not easy to establish, as they overlap the older beds, near the margin of the basin.

In Tasmania the basal beds can be divided into at least two types.

1.   The basal tillites.
2.   The basal conglomerates.

The basal tillite or the Wynyard Tillite occurs only west of 147° E. Long. (Map 1), with the exception of the U. Mersey Valley, the N. Western part of the state up to Waratah, the area round Mt. Roland and the south western part of the state round Port Davey. The occurrences of the tillites are not continuous but they can be correlated because of the overlying rocks, and their distribution can be found although they are sometimes overlain by Mesozoic rocks.

At Maydena, and in the Styx and Weld River valleys the Permian tillite is exposed (Lewis, 1924; Banks, 1957). It overlies the Owen Conglomerate unconformably and has broken pieces of marine fossils of Permian aspect (p 15)

At Maydena four bands of tillites are separated by three bands of sandstone and varves. The fossils occur only in the tillites and even a thorough search so far has not yielded any fossils from the varves or the sandstone.

The occurrence of these fossils in the tillite can be explained in the following ways:

1. The glaciers extended into the sea, and incorporated older sediments so that during transport they were crushed and included in the tillite.

2. The ice extended over a sea bed which had recently risen above the sea, and incorporated the unconsolidated sediments in the tillite.

3. The ice extended over the sea, animals lived under it, the shells were deposited with the tillite as it was falling in the water, and they were crushed during consolidation.

Reade,(1874, p.32; 1883, p 90), Jamieson,(1882, p. 150), Lamplugh,(1884, p. 312, 1891, p. 419), and Gregory,(1926, p. 363), have reported the occurrence of crushed marine fossils in the drift beds of England. Reade,(1876, p.33), has proposed that these fossils were crushed by the wave action of the stormy seas. Lamplugh has proposed that the deposits from the floor of the North Sea were incorporated by the ice and transported to the land, so the fossils were crushed, and Jamieson (153) has proposed that these shells have been incorporated from older beds. Gregory has reported the occurrence of the striated and broken fossils, but he has said that they were not transported and they were in situ. He has explained the breaking of

the fossils by the shrinkage of the clay, since some of them had an epidermis and others were whole. The difference in the explanations put forward by Reade and others is mainly due to the fact that he discusses the deposits from the west coast of England, and Lamplugh and Jamieson discuss the deposits from the east coast of England. Whereas on the east coast the crushing can be easily explained as the ice was moving from the east, this is not so simple on the west coast. The other evidence that all the authors, except Gregory, have put forward, indicates that the crushing was possibly due to the transport of the shells.

Miller (1953, 28), and Armstrong and Brown (1954, p. 355) have reported the occurrence of complete fossils in the tills, from N. America. These fossils are found in the glacial deposits which according to them had been laid down by a floating ice sheet, in the sea.

The varves and the sandstone may have been deposited in fresh water or in brackish water so they cannot be used as definite evidence for either of the two conditions. Wallace (1927) and De Geer (1940, p. 104) have suggested that varves can be deposited in both fresh and brackish waters, and De Geer has reported the occurrence of Portlandia (Yoldia) arctica from the brackish water varves of the Stockholm region.

The occurrence of the end-moraines on top of tillites may have been responsible for the formation of the lake in which the varves were deposited, but similar gravel beds are reported from coastal glaciers, and sometimes they are deposited below sea level (Garwood and Gregory, 1898, p 210; Field, 1932).

The Maydena succession at first sight poses a mechanical problem since a true till overlies waterlaid sandstones, and another true till overlies waterlaid varves without producing any shearing, slumping or drag in the sediments. Twenty feet of till would seem to imply a substantial glacier. Its weight on the sediments must have been considerable and one might expect slippage to have occurred in the soft underlying sediments. We can't postulate that the glacier was floating, for then we would expect some evidence that the pebbles had been dropped through water, which is not the case. It is difficult to assume that the depth of water was just sufficient to float the glacier, for such depth depends on the glacier thickness, and this would vary during the complete waxing and waning cycle recorded in the sediments, so that there should be signs of drag when the moving ice pressed on the sediments and signs of dropped pebbles when there was water below the ice. However the sequence can be interpreted without anomaly.

The Pre-Permian basement in this area stood somewhat higher than the basement regionally. This is not surprising since the basement here is Owen Conglomerate, perhaps the most resistant physically of all the Pre-Permian rocks. Although the early Permian glacier over-rode this area, sediments did not remain here until the waning stages of the glaciation when sediment had accumulated to some depth in surrounding areas, including basal tillite and some strata with marine fossils. When continuing regional subsidence combined with sedimentary filling of the depression,

finally led to accumulation of sediments at Maydena
the glaciation was well on the wane, fossiliferous
marine beds had already been deposited west of this
point.

The basal tillite represents normal till depos-
ited on the Ordovician basement at the sole of a glacier
during a cold phase of the general retreat. The tillite
contains broken fossils picked up from Permian beds
further west. The fact that the glacier was working
over earlier Permian marine glacial drift west of this
point explains the higher proportion of clay fractions
found in the petrological study of this till, as com-
pared with other tills (See Carey and Ahmad, 1957, for
discussion of composition of tills).

This ice advance was followed by a warmer phase
when the glacier retreated out of this area, which
was left as a lagoon on a plain in front of the glacier.
This received meltwater-borne sand forming a sandstone
which filled the lagoon. No marine fossils have been
found. If they are found, it alters the picture very
little.

Next came a period of refrigeration. The first
result in this area was the freezing of the soil as
permafrost at least down to the earlier till if not to
the basement. Subsequently the re-developing ice-cap
extended over the area. There was no particular tend-
ency to shear the sandstone during the over-riding of the
ice-flow since it lay in a depression, and the simplest
shear planes passed above it, and the shear strength
of the permafrost sandstone would be substantially
greater than that of the glacier ice, so that shear was

confined to the latter. Another possibility is that the Owen Conglomerate floor rises immediately west of this area (The regional geological map suggests that this was very probably so). Hence the gouging action of the glacier on the underlying soft Permian sediments, might dig deeply west of this rock ridge but fail to gouge immediately east of the barrier. A layer of till was deposited over sandstone, which again contains fragmental Permian fossils derived from marine Permian beds being carved by the glacier somewhere to the west.

Next followed a further similar cycle of warming and refrigeration with deposition of lacustrine (?) varves and another till. Again the varves were protected from glacial scour either by freezing or by virtue of a high ridge of Owen Conglomerate to the west. The next cycle brought varves and marine outwash with abundant fossils and dropped pebbles.

It is clear that these warm-cold cycles are superimposed on a general warming, since the tillite sequence is 120 ft., 20 ft., 1 ft., and the interglacial sequence is coarse meltwater deposits, fine meltwater deposits (varves) and general marine drift and outwash.

The tillite deposits of northern and western Tasmania are very similar to each other so they are dealt with together.

These deposits are sub-glacial moraines and their thickness varies from a few feet to over three hundred feet with a probable upper limit of six hundred feet. Mechanical analysis of the tillite shows a heterogeneous composition, and poor sorting with angular

pieces of quartz and striated and facetted erratics.
These beds are exposed at Mt. Sedgwick, Mt. Read, King
William Saddle, Dairy Plains, Mersey District, Point
Hibbs, Northern and Western margin of the Central Plat-
eau, Wynyard and its environs in the Inglis Valley, and
at Mt. Dundas its occurrence is suggested. Tillites
have also been reported from some other areas but on
examination they have been found to be other than sub-
glacial tillites, so they are not included here. Twelve-
trees (1915) has reported a bed of tillite from Shoe-
maker's Point, but this has not been checked.

These tillites are overlain by outwash beds,
varves, mudstones with erratics, and marine beds in
different areas. The only uninterrupted complete
section so far recorded is from the bore cores in the
Western Tiers, north of the Great Lake, and here the
tillite is overlain by a mudstone conglomerate with
erratics (Marine Tillite beds). The tillite grades
into the Palmerston beds, which in turn grade into Quam-
by Mudstone.

The results of the mechanical analysis, the lack
of bedding, the presence of striated and facetted
boulders (plate 23), and striated pavements at the base,
leave no doubt about the glacial origin of these beds,
but the overlying beds indicate that these may have
been deposited below sea level. In the Western Tiers
area the tillite grades into fossiliferous marine
deposits. The sequence in this area indicates that as
the ice started to retreat, it thinned down and start-
ed to float, so the tillite grades into a water-deposit-
ed bed. At Oonah (p 43) the tillite is overlain by
marine beds and there may have been a slight break in

the retreat of the ice and encroachment of the sea but the break does not show any unconformity or traces of erosion, so possibly the time lapse was not great. In the Mt. Sedgwick area (p 49), the tillite is overlain by a conglomerate bed which passes into mudstone with fossils, and is followed by a mudstone with erratics. At Mt. Read the tillite is directly overlain by marine beds, and here there is no indication of a break. At Point Hibbs (p.5) the lowermost bed is an outwash-like bed with conglomerate and is overlain by tillite which grades into fossiliferous marine beds and beds with glendonites. The contact at Point Hibbs may be a fault contact but the presence of the tillite on top of the outwash beds suggests that the tillite did not rest completely on the base, so outwash was not incorporated in the tillite. Fuller (1914, p. 133), and Shaler (1896, p 972) have reported the occurrence of gravel beds overlain by sediments which have been deposited by a floating ice sheet so that the bottom beds were not incorporated by the ice. Possibly the conditions at Point Hibbs were similar so that the bottom beds were not incorporated in the tillites but the depth of water was possibly not very great because the erratics do not show any sign of having been deposited in water. Here the uppermost part of the tillite grades into mudstone and siltstone with erratics which pass into fossiliferous beds.

The occurrence of marine fossils on top of the tillite, in the Oonah area and at Mt.Read without any sign of erosional break and the gradual transition of the tillites into marine beds at Point Hibbs and in the Western Tiers, indicate that within a short time

of the retreat of the ice the tillite was covered by
the sea. Possibly in some places there was no time
lapse between the retreat of the ice and advance of the
sea and the tillites were deposited below sea level
but in the absence of the water.

The differences in the nature of the overlying
beds may be due to differences in the rate of retreat
of the ice or to the position of the ice front with
respect to the water, or to different conditions of
isostatic recoil following glacial unloading. If the
rate of retreat is slow the ice will gradually thin
down and start to float, so the till will grade into
beds like the Palmerston Drift Bed and fossiliferous
strata with numerous erratics. On the contrary if the
retreat is swift, large quantities of meltwater will
be released and may wash the tillite so that an out-
wash bed will be laid down on top of the tillite. If
the ice front is protected it will melt gradually and
slowly so that a bed like the Palmerston Drift Bed will
be deposited on top of the tillite, but when the ice
front faces a stormy sea carving will be very fast and
the waves will wash the deposit so an outwash-like
deposit will be laid on top.

Basal Conglomerate

This is mostly a white conglomerate with a large
proportion of well-rounded quartzites pebbles (Plate 24)
and the thickness varies between 10 & 50ft. These de-
posits overlie the Precambrian and Lower Paleozoic
rocks unconformably and they are overlain by Permian
beds of different horizons. They are present in only

those areas where basal tillite is absent, and are re-
ported from Mersey Valley, at some places in Mt. Pel-
ion area, and from the north east coast east of Tamar
Valley. The horizon of these beds is not known and they
can be of any age older than the deposits which over-
lie them. In all cases they are overlain by fossil-
iferous marine beds belonging to Quamby and Golden
Valley formation or coal bearing beds of Liffey form-
ation.

The absence of the lowest Permian beds on the
unconformity presumably means that these areas in quest-
ion were hills on the unconformity surface. There is
no local way of deciding whether they were outside the
glaciated area or were over-ridden by the ice or were
nunataks above the ice-sheet. Subsequent subaerial or
subaqueous erosion before they commenced to receive sedi-
ments through the general subsidence of the region,
could destroy all evidence of the earlier glaciation.
The basal conglomerate might be derived locally from
shore line process during submergence, or might be
largely made up of reworked pebbles winnowed from till-
ite shallow enough to be within reach of wave attack.
If the first marine beds deposited above the obviously
shoreline deposits have similar glacial features (errat-
ics etc.) to the contemporaneous beds overlying basal
tillites in surrounding areas, the reasonable presumpt-
ion that these areas may have been similarly glaciated.

In the Mersey valley this conglomerate has
rounded quartzite pebbles which are sometimes flat. Their
size varies from cobbles to pebbles and they have very
little cementing material. Sometimes the lithology of
the pebbles is affected by the lithology of the under-

lying rocks. In places they show crude sorting and current bedding but in other places they show neither of these features. The pebbles and the boulders are quite fresh and the surface of the older rocks on which they lie are also quite fresh and unweathered.

Although ice-free areas are not unknown from glaciated regions, the occurrence of tillites east and north of the Upper Mersey Valley (Map No. 1) suggests that possibly the glaciers crossed Cradle Mountain and the Upper Mersey Valley. If this was an ice-free area the tillites which were deposited on the east side of it could have come from the north, east or the south, but this does not agree with the known directions of ice movement. Thus it appears improbable that this was an ice free area, especially since tillites were deposited east of it.

Debenham (1921, p. 940) and Taylor (1922, p. 63) reported the absence of till from the Antarctic land surface at some places. Debenham has explained this by assuming that the till carried at the base is sometimes not deposited on the land but is carried to the sea. If the conditions in Tasmania during the Permian were similar to those in the Antarctic region today, the absence of the tillites from the Upper Mersey Valley and Cradle Mountain area may be due to non-deposition.

David (1914, p. 262) has reported end moraines from Ross Island and Mt Erebus; these moraines have erratics from the mainland of Antarctica, which indicate that at one time the ice from the mainland extended up to Ross Island and deposited the till.

The erratics, which have been transported a great distance in the moraines are small and rounded and some of them have no striations. In most places these deposits are still covered with ice which protects them from erosion. If the Antarctic ice cap melts causing a rise in sea level, and this island is exposed to weathering, possibly these deposits also will be reworked and redeposited on the shore line as a conglomerate, which will be very similar to the conglomerates at the base of the Permian. The reworked moraines as reported by Ferrar (1907) will possibly form similar deposits if the sea level gradually rises, and the waves are able to resort the deposits.

Either of the explanations advanced above can account satisfactorily for the absence of tillites at the base of the Permian.

The presence of the conglomerates in the eastern part of the state which was perhaps glaciated may be explained in the same way, although the only evidence of the glaciation in this part is the presence of some striated boulders in St. Mary's and adjacent areas, and the tillites in Maria Island.

Shales, Sandstone and Tasmanites.

So far these deposits have been reported from only the Mersey District and they have been found in the bores that were sunk in search of tasmanite beds. In some bore logs the description is not clear so it is uncertain whether tillites or some other rocks occur at the base.

The logs of the cores indicate that the beds below the tasmanite beds were possibly deposited contemporaneously with the Quamby Mudstone but in some cores older beds also may be present.

Bore core 13 has water-worn pebbles at the base, which appears similar to the basal conglomerates. Bore cores 12, 14 and 22 have pebbly mudstone at the base which may be the tillite or it may be marine drift of the Palmerston type. Bore Hole 27 has tasmanite at the base, and the bore hole 15 has sandstone at the base.

If bore holes 12, 14 and 22 have tillites at the base, this is the easternmost reported occurrence of tillites in the area. Otherwise, in the absence of the tillites, it is suggested that the lowermost beds might be correlated with the Palmerston Marine tillite and they were perhaps deposited towards the end of the glaciation. The occurrence of marine fossils higher up in these beds helps to correlate them with beds in other cores and possibly they were deposited at the same time as the Quamby Mudstone. The difference in the lithology of the basal beds may be due to the differences in the times when they were deposited.

Palmerston Marine Tillite facies

Beds of this facies are transitional between the basal tillite and the overlying Quamby Mudstone. Their stratigraphic position is known only from the bore cores in the Western Tiers area. On Mt. Sedgwick and at Point Hibbs, beds very similar to these and stratigraphically in the same position are recorded. Their correlation is however doubtful since they are not exactly similar.

The Palmerston marine tillite beds have been described on page 13-14, and the features like lamination, bedding, vertical orientation of the pebbles with respect to the bedding plane and the relationship between the bedding and the pebbles indicate that these beds were probably deposited in the water. The gradual transition into marine fossiliferous beds as shown in cores from Blackwood Creek indicates that these beds were probably deposited in the sea. The vertical orientation of the pebbles and the truncation of the laminae on the sides indicates that these beds were deposited in very calm water or below the wave base.

The transition of the tillite into these beds with crude bedding, the absence of marine fossils, the high percentage of the clay and finer size particle in the deposit (61%), and the high roundness and sphericity of the erratics indicates that these deposits were laid down in sub-glacial, but near the point where the ice begins to float.

The absence of the marine fossils in these deposits may be due to wide fluctuations of salinity owing to melt-water currents under the ice or from the open sea, with consequent restriction of light and circulation.

Quamby Mudstone and Woody Island Siltstone

Banks (1957) has correlated these two formations and is of the opinion that they are younger than the basal tillites.

Quamby Mudstone.

This is a fine grained mudstone with some facetted

boulders and pebbles. Reid (1924, p. 47) reported that
the tasmanite beds have boulders up to 2 ft. in size.
Wells (1957, p.8) observed that the Quamby Mudstone
has pebbles up to 3 cms., and McKellar (1957, p.9)
reported that erratics are entirely absent from the
lower half of the beds. Possibly Wells and McKellar
did not observe larger erratics because of poor expos-
ures , as the tasmanite beds occur 50 ft. above the
base and from them, erratics of all sizes are recorded.
These beds show crude bedding near the bottom but the
bedding improves above the tasmanite beds. Reid re-
ported that the tasmanite beds contain Eurydesma,
Aviculopecten and Keeneia and Crespin (1947) has id-
entified marine foraminifera in the Tasmanite beds.
McKellar observed that bands with many fossils and
erratics alternating (Pl. 25) with bands with few
fossils and erratics occur in the upper half of the
formation; these bands are wide apart at first but
they come closer together near the top. In the top-
most part of the formation, banding is absent but the
fossils and erratics occur continuously.

The pattern of these rhythmic beds recalls the
kind of change which is often found in varves. Common-
ly winter laminae are thicker near the bottom of a
glacial retreat sequence and become thicker upwards,
whereas the summer laminae are thin at first but thick-
en upwards. The retreat sequence of varves may be re-
placed upwards by thicker continuous beds of varvoid
material.

In the rhythmic beds reported by McKellar the
bands with fewer fossils and erratics are thicker near

the bottom but become thinner towards the top. This
suggests that like the thick winter layers of the
varves, these represent periods of ice advance. Groups
of cold years are always more common and of long durat-
ion near the maximum of the glaciation, but their
length and frequency decrease with waning glaciation.

The high concentration of fossils and the
erratics in the second set of layers may be due to:

1. At the end of the glaciation a large quant-
ity of fresh water was added from the melting ice, thus
causing the death of the animals while the erratics
were added because of the increase in the intensity of
glaciation.

2. Large quantities of water from the melting
ice were added to the sea which washed the upper layers
of all fine material.

It is very difficult to select either of the
two as the explanation for the occurrence of the large
number of erratics and fossils in one band, since other
factors may also have played a part. It is also poss-
ible that both of the suggested conditions may have been
responsible.

Armstrong and Brown (1954, p. 351) described a
similar deposit which grades down into till and which
is overlain by till, and they have concluded that it
was deposited in the interglacial period. In that
case there is no doubt about the time of deposition
but the Quamby beds do not have any layer of tillite
on the top to indicate the time of the deposition.
Possibly they are post-glacial deposits which may have

been laid down in the pack ice zone where the retreating and the advancing glaciers had a great effect on the sediments but the water was, perhaps, open for long stretches of time so that algae and other marine life could exist in the water.

The environment and origin of the tasmanite has been discussed separately by Carey and Ahmad (1957) which is included herewith. Reid (1924, p.47) reported that the thinly laminated tasmanite beds are interbedded with mud bands which are almost devoid of spores in both sets of rhythms. The sets of thin laminae in the tasmanite beds are possibly due to annual variations in production of the spores by algae, while the second set of bands may be due to bad weather years in which the spores could not develop. The occurrence of erratics in these beds is evidence of icebergs in this area.

Hough (1950, p 257) described laminated bands of clay from the Antarctic Sea, and postulated that they are possibly due to seasonal variations, and are similar to varves. Perhaps these tasmanite beds can also be treated as marine varves in which spores have played a major part.

The Quamby Mudstone in the Mersey District and the Mudstone Conglomerate beds of Preolenna and other areas were possibly deposited in small basins or narrow inlets. The boundaries of these basins are as yet not very clear but to a certain extent they can be traced from the outlines of the deposits and the areas where they overlap the younger beds.

In the Wynyard district the basal tillite is reported from the coast line, from the Preolenna area, from Oonah and from Hellyer Gorge. In the Preolenna area the Quamby Mudstone overlies the tillites. In the south beyond Hellyer Gorge Permian beds are not known. In the east the tillite is overlain by outwash beds, while further eastward Precambrian rocks are encountered so that possibly on this side also the Permian deposits were overlapping. The boundaries of this basin are not very clear but they appear to run from the coast in the north-east to Hellyer Gorge in the south-west. The eastern side started somewhere east of Doctor's Rock and the western boundary was beyond Preolenna.

In the Native Plains and Mersey district, the lower Permian beds are overlapped by younger beds west of the Don River, but west of Dairy Plains the younger Permian beds rest on lower Paleozoic rocks and tillite is absent. In the north the tillite is overlapped by Quamby Mudstone. In Bore Holes 21 and 24 (Reid 1924) the younger Permian beds rest on lower Paleozoic and the Quamby beds are absent. In Bore Hole 26, the Mersey Coal Measures (Liffey Group) is present but the Golden Valley Group is absent and in Bore Hole 28 Permian deposits are absent. In the south, in Bore Holes 12, 14, and 22, the tasmanite beds rest on older beds but in Bore Hole 27 and in Oliver's 50-Acre Block they rest on Precambrian deposits. In the south little information is available as most of the area is covered by dolerite, but Wells (1957, p. 9) reported that in the Quamby Brook region the tillite is overlapped by Quamby Mudstone. The existence of the basin in this

area is indicated by these occurrences but its bound-
aries cannot be drawn until the area is surveyed in
detail. Reid (1924, p.43) also suggested that the
tasmanite beds were deposited in small basins. The
beds themselves indicate to a certain extent that they
could have been deposited only in an area which was
fairly closed and little disturbed by waves and currents.
In an open sea these spores probably would have been
carried away, or may not have settled because of their
light weight.

From all the available evidence it appears that
by the end of maximum glaciation the ice extended to
the sea and subglacial moraines were being deposited
below sea level but in the absence of sea water. As
the intensity of glaciation decreased the ice thinned
and the glaciers started to float. In this period
the depth of the water below the ice was possibly very
small and deposits transitional between the tillite
and the Quamby beds were laid down (Palmerston marine
tillite facies). As time passed the ice retreated, the
depth of the water increased and the crudely bedded
Lower Quamby Mudstone beds were deposited seawards of
the ice front. By this time marine life had migrated
to these areas and this is represented by a small num-
ber of fossils. The retreat of the glaciers continued
but some minor advances and retreats also took place and
caused changes in the conditions in the sea. These
variations sometimes killed large numbers of marine
animals and added many erratics to the deposits, so
that bands with many fossils and erratics alternate
with bands which have few fossils and erratics.

Probably the Quamby Mudstone can be classified

as the inner zone deposit of the glacial marine sed-
iments described by Stetson and Upson (1937, p.58).
These deposits have poor sorting and a wide range
of sediment sizes, and they compare very well with the
characters of the Quamby Mudstone deposits. Inner
Zone deposits are laid down in front of the ice sheet
and due to the addition of sediments from the melting
ice sheet and the ice bergs they have very poor sorting
and a very wide range of sediment size. Carsola (1954,
pp 1566-67) reported similar sediments from the North
Coast of Alaska. In all areas this type of sediment
has not extended beyond 300-400 miles from the end of
the glaciers. From the occurrence of the Permian
tillites in Tasmania, the distance of the Quamby beds
from the front of the ice sheet can be determined pro-
vided the horizons of the tillites in the different
areas are found with some accuracy

Woody Island Siltstone and Higher Formations at Woody Island.

These beds are exposed in the Woody Island,
Maydena, Florentine Valley, and Beaconsfield areas
and have been correlated by Banks (1956 (a)) with
the Quamby Mudstone of the Quamby area, on the basis
of fossil evidence and the occurrence of glendonites.
The beds are composed of siltstone with erratics and
calcareous nodules up to 10 ft. in length, and
glendonites (Banks et al 1955 (b)). Eurydesma cordatum
is the most common fossil in the lower beds but in the
higher formations, Stenopora, Fenestella, Spirifer and
other fossils are quite common. At Woody Island some
of the beds show current bedding and other evidence

of shallow water deposition. The siltstone beds and the Sunset Bay and Alonnah Sandstones have glendonites which are pseudomorphs after glauberite (David et al 1905). On Woody Island the base is not exposed and the lowermost beds are siltstone with glendonites. At Maydena the siltstone overlies the tillite and the outwash beds but the contact is not seen.

David et al (1905) proposed that the original glauberite, which has been replaced by calcite to form the pseudomorph glendonite, was introduced post-depositionally. Ragatt (1938) postulated deposition in isolated basins in which the glauberite was a primary deposit due to changes in temperature. Debenham (1921, p.79; 1954, p.500) and Taylor (1922, p. 137) reported the occurrence of mirabilite from the Antarctic and it is suggested that the over-saturation of the $CaCO_3$ and the high concentration of the mirabilite might have played some part in the development of the glauberite. At the present time Banks and Hale are working on the problem and a detailed discussion of the matter is soon to be published.

The development of the calcium carbonate concretions may be due to rise in temperature when the sea was over-saturated (Raggatt, 1929). Bradley and Bramlette (1940, 3) recorded the increase in the $CaCO_3$ content of the Atlantic cores with decrease in the percentage of clastic material and they have interpreted this as being due to the decrease in the intensity of glaciation. In post-glacial times the

temperature must have been rising and the $CaCO_3$ may
have been deposited, as proposed by Raggatt.

Milner (1952, p. 373) said that siltstone is
deposited in shallow waters and may be associated with
glacial deposits. Washburn (1947, p. 63) described
siltstone beds with marine fossils and small erratics
overlying a gravel deposit on Victoria Island, Arctic
Canada.

The features recorded by Banks and others from
Woody Island indicate that these beds were deposited
in a shallow sea. The occurrence of erratics in them
and the outwash below at Maydena indicates that these
beds were in front of the glaciers. Raggatt and
David also proposed that the glendonite beds on the
Australian mainland were deposited under cold con-
ditions and possibly on mud flats or in shallow water.

Combining all the evidence from these deposits
it appears that siltstone beds were deposited in very
shallow seas, maybe on tidal flats in front of the
glaciers so that melt water and some icebergs were
coming in to the area, or the melt water was bringing
some pebbles with it. The land was gradually sinking
but deposition was able to keep pace with the sinking
so that sandy beds are interbedded with siltstone
beds. The breaks in sedimentation are represented by
the "Tillite" bands which:-

1. may be shore line conglomerate bands, with
rounded pebbles and poor sorting,

2. may be due to temporary advance of the ice
when the quantity of outwash to the sea might have in-

creased, but they were not deposited in contact with the ice.

Although the Woody Island and the Quamby Mudstone beds were deposited about the same time, the differences in the lithology of the two deposits, as discussed above, is possibly due to the environments in which they were deposited. Quamby beds indicate the presence of floating icebergs whereas the small size of the pebbles and the presence of current bedding and calcareous nodules with pyrite and carbonaceous matter in the deposits at Woody Island indicate shallow water conditions possibly with limited circulation. Perhaps the major factor responsible for the two types of deposits was the depth of the water and the distance from the ice front. Taylor (1922, p. 53) has reported deposits very similar to the Woody Island Siltstone, but on a smaller scale, being laid down in front of the Davis glacier. This suggests that under favourable conditions deposits like the Woody Island Siltstone can be laid down in front of glaciers whereas in other areas, where the sea is deeper, deposits like the Quamby Mudstone are laid down.

Darlington Limestone.

Near the base of the Golden Valley Group and near the top of the Woody Island section, a thin bed of limestone occurs. This limestone varies in thickness from a few inches to many feet, and the thickness decreases from east to west, the thickest part, about 50 ft., being at Maria Island. It contains Eurydesma cordatum, Stenopora tasmaniensis, S.johnstoni, with other fossils and erratics. It is reported from

western, northern and southern Tasmania but in the N.E.
it is only reported from Lilydale and Beaconsfield.

The similarity of the fauna, the lithology, and
other characters of this limestone suggest that during
this age the sea was open in the larger part of the
state, there was free migration of the fauna in the
sea and possibly there was not much variation in the
depths of the sea at different places. The fossil
beds contain erratics and show an increase in number
and size which had decreased near the top of the
Quamby Formation. The limestone band in the beds
which show abundant evidence of iceberg activity
appears out of place but it may be due to:

1. the warming up of the sea, so that the high
percentage of $CaCO_3$ could not be held in the water
and was precipitated.

2. the retreat of the glaciers so that less
clastic material was brought down to this part.

Both explanations are consequent upon the re-
treat of the glaciers. Bradely and Bramlette (1940,
p. 3) reported an increase in the percentage of $CaCO_3$
in the non-glacial marine deposits from the Atlantic
Ocean core, which they suggest may be due to warming
of the sea. The increase in the number and the genera
of the fossils also suggests that the conditions were
more favourable, and maybe the sea was warmer.

Golden Valley Group and Bundella Mudstone.

In northern Tasmania the Darlington Limestone
beds are near the base of the Golden Valley Group and
in the southern area of Tasmania they are below the
Bundella Mudstone.

The lower beds of the Golden Valley Group have
shales below and above the Darlington Limestone,
and there are abundant marine fossils. These shale
beds gradually pass into coarsely sorted shales and
sandstones. McKellar reported that the top part of the
formation has micaceous sandstone and quartz sandstone
which grade into the uppermost beds with thicker
bands of micaceous mudstone. In this formation the
number and size of the erratics and the number of
fossils decrease from the bottom to the top.

On the whole the area shows two conditions:
1. Possible shallowing of the basin of deposit-
ion as indicated by an increase in mica.
2. Warming up of the country as indicated by
the decrease in the number of the erratics. Had this
been due to the shallowing of the sea, possibly the
size of the erratics would not have decreased. The
top beds of the Golden Valley group are overlain by
freshwater beds which also could have been due to
warming of the land, and these beds also indicate the
rise of the land so the connection with the ice front could
have decreased.

The Bundella Mudstone beds of the Hobart area
(Banks, 1957) have siltstone bands with four bands
of sandstone. The fossils include <u>Stenopora johnstoni</u>,
<u>Eurydesma cordatum</u>.

The conditions of the deposition both in the
north and the south had probably stayed the same except
that in the south sandstone bands are thin and are
possibly connected with the waxing and waning of the
glaciers.

Deep sea cores and dredgings from the Antarctic
and Arctic regions show that, below the shelf ice and
in front of it, the deposits have a very high percent-
age of clay, but outside the shelf zone, in the outer
limit of the pack ice zone and in the iceberg zone, the
sediments are generally sandy. Menard (1953, p.1292)
suggested that this sand is possibly due to submarine
sorting of the deep sea sediments. The Tasmanian sedi-
ments of the Bundella formation and the Golden Valley
group can possibly be placed in this zone.

In the north the McRae sandstone is followed
by non- fossiliferous mudstone with carbonaceous
lenses followed by other sandstone and shales.

In the Hobart area Bundella Mudstone is follow-
ed by a thin bed of conglomerate (Fig 10) which contains
Gangamopteris. This conglomerate is overlain by sand-
stone beds with current bedding and other characters
which suggest lacustrine or paludine origin (Banks,1957).
The next is a bed of conglomerate overlain by a sand-
stone bed with worm casts and erratics. This marine
bed is overlain by a freshwater sequence with a con-
glomerate at the top. The next bed is again marine
with some marine fossils and is overlain by a fossil-
iferous deposit with Martiniopsis, Aviculopecten, and
others with some bands of Glossopteris.

The whole sequence suggests that in the Hobart
area the sea floor was oscillating so that the marine
sequence is mixed with a fresh water sequence. The
marine and freshwater beds are separated by the con-
glomerate beds, which suggest a break in the sediment-

ation.

After the deposition of the Darlington Limestone
the deposits everywhere in Tasmania indicate a gradual
shallowing of the sea. Life which was abundant in all
parts of the sea gradually decreased, and the number of
icebergs coming in also decreased, so that the number
of erratics declined. In the north all through the
Liffey group no erratics are found, while in the Hob-
art area, erratics are present but only in the marine
beds. The presence of erratics in the south indicates
that while freshwater beds were being laid down in the
north, glaciation continued but the erratics had stopped
reaching this area as there was no sea connection with
the glacial front, or the land in front of the glaciers
had also risen so that no icebergs were coming out, but
the decrease in the size of the erratics suggests that
possibly the distance of the ice front from the area of
deposition had increased.

The rise of the land during this period may have
been due to isostatic uplift; the land may have risen
after the glaciation due to the effect of the Hunter
Bowen Orogeny. Or the rise may have been due to both
reasons.

Cascade Group.

In the Hobart area, the Cascades Group was
deposited on top of the freshwater and marine sequence
but in the north these rocks are not represented. Banks
(1957) suggested that at first glance it appears simple
to explain the absence of the Cascades Group from the
north, but possibly a detailed study of the northern
area will throw some light on this problem, and some
beds equivalent to the Cascades Group may be present.

Although it is very difficult to forecast whether any such beds will be found in the Deloraine and Quamby area, the isopachs drawn by Brill (1956, p.133) show a shallowing towards the west, which suggests that perhaps in the west the Cascades Group is represented by a continuation of the freshwater beds.

Brill (1956) studied the Berriedale Limestone in detail and observed the cyclic nature of these deposits. He reported that beds of impure limestone alternate with beds of calcareous mudstone and the limestone beds tend to be thicker than the beds of mudstone. Erratics are common in the whole sequence but are more concentrated near the top. Brill concluded that these cycles were due to waxing and waning of the ice sheet (p.136). Possibly his explanation is correct as the overlying beds have a large number of erratics and also indicate an increase in the intensity of glaciation. The decrease in the number and size of the erratics in the Bundella beds was possibly due to decrease in the intensity of glaciation, because during deposition of the Berriedale Limestone the conditions had not changed, but the number and the size of the erratics had altered. The gradual increase in the number and size of the erratics from the bottom of the Cascades Group upwards indicates that possibly the Faulkner and Liffey Groups were laid down during an interglacial period, and from the beginning of the Cascades Group the glaciers were advancing again. The sediments from the bottom to the top of the Berriedale beds show an increase in coarseness. The mudstone and limestone in the lower part of the Berriedale beds give place to sandstone near the top.

On top of the Cascades Group in the Hobart
area and the Liffey Group in the north, the Woodbridge
Glacial Formation was deposited. In the northern
area McKellar divided the Woodbridge Group into three
formations, the basal mudstone with limestone bands
and layers of erratics followed by the middle sand-
stone with layers of erratics and then micaceous
mudstone at the top.

The lower beds indicate gradual deepening of the
sea, the lower 12 ft. having alternating layers of
sandstone and mudstone, followed by one foot of con-
glomerate with well rounded pebbles. The middle part
of the lower formation has bands of limestone and mud-
stone with erratics. In the Hobart area Banks (1957)
described the formation as sub-greywacke sandstone
and siltstone with a thin band of limestone 10 ft. from
the top. Erratics are common in the whole sequence
but are most abundant in the middle, and show a decrease
in numbers towards the top of the formation. Some of
the erratics are more than two feet in length.

The presence of sandy beds, marine fossils and
erratics in the Woodbridge Formation suggests that the
environment in which they were deposited was possibly
very similar to that found today in front of the Ross
ice shelf, in the outer part of the pack ice zone. It
appears that the effect of distant glaciers on the
coarseness of the sediments is in direct ratio, so
with increase in glaciation the coarseness of the sedi-
ments also increases. In the shelf ice zone the sedi-
ments are fine since they are added directly from the
melting shelf ice, and due to lack of currents they

are not well sorted so they have a large percentage
of fine grade particles. In the pack ice zone, away
from the ice shelf, the coarseness gradually increases
outwards and is possibly at a maximum near the middle
of the iceberg zone. In this zone and in the pack ice
zone currents are present which possibly help in
sorting the sediments as they settle in the sea.
There may also be some submarine sorting.

The Woodbridge Formation was deposited perhaps
in the outer part of the pack ice zone, and may have
extended up to the middle of the iceberg zone. It
shows abundant evidence of marine life, the sediments
are fairly coarse but the particles are fresh and
erratics are common and have striations. The mud
bands in these sediments possibly represent periods
of retreat when the deposits of the outer iceberg
zone were laid down on the coarser sediments.

Ferntree Group.

The Ferntree Group overlies the Woodbridge
Group and has the following succession in the north.

Eden Mudstone.	Micaceous mudstone, no erratics.
Blackwood Conglomerate.	Well rounded pebbles.
Dry's Formation.	Mudstone with mica and erratics up to 2".
Palmer Formation.	Sandstone, poorly sorted, erratics present.
Springmount Formation.	Mudstone, few erratics, coarser and fine bands near the base.
Gracia Formation	Erratics common, poorly sorted sandstone.

The sequence shows that the retreat of the glaciers which had started in the upper part of the Woodbridge Formation still continued and possibly the land was emerging from the sea since the deposits show an increase in mica content. The bands in the Springmount Formation may be due to fluctuation of the sea level or advance and retreat of the distant glaciers. Possibly the increase in coarseness in these beds is not so much the effect of marine sorting but is due rather to gradual shallowing of the basin of deposition. The aspect of the upper beds further confirm this shallowing of the basin since the Palmer Conglomerate is possibly a shoreline deposit of the retreating ice and marine fossils are absent in the upper beds.

The uppermost deposits of the Permian System are freshwater and lacustrine beds of the Cygnet Coal Measures, which are present in southern and western Tasmania, and which indicate that by the end of the Permian Period, glaciation had decreased and possibly ceased and a warmer climate existed all over the State.

BASAL PERMIAN PALEOGEOGRAPHY

In the early Permian epoch, Tasmania was experiencing a frigid climate and ice-borne deposits were being laid down on the land and in the sea. David (1950) proposed that during this time an ice sheet covered large parts of Australia and extended to within 20° of the present position of the Equator. In Tasmania in some places the glaciers may have been coming down from mountains of high relief but in most of the area they were flowing on low-level coastal plains with the relief varying no more than a few hundred feet.

The direction of ice movement, as indicated by the striations at Mt. Sedgwick and King William Saddle, was from west to east, and from the pebble orientation at Wynyard was from S.W. to N.E. The direction of the ice movement is further confirmed by the erratics in eastern Tasmania which came from western Tasmania. Along the west coast, near Malanna, and Point Hibbs, erratics of rocks belonging to these areas are absent. Another factor which could be interpreted as confirmation that the ice moved from the west is that at Maria Island and in the St. Mary's area the basal tillite is absent, but in the west wherever undisturbed Permian deposits occur basal beds are definitely present.

If the conditions of the Pleistocene glaciation can be used as a guide to the Permian glaciation of Tasmania, it can be suggested that during this time Tasmania was within 40° of the Pole and it was along the eastern side of a land mass, which was being intensively glaciated. The radial pattern of the ice movement in different parts of Australia indicated by the striations at Inman Valley (South Australia), Bacchus Marsh (Vic.), Wynyard, Mt. Sedgwick, and King William Saddle, although it cannot

be taken as a definite evidence, does not disagree with
the assumption that the glaciers which reached the sea
during the Lower Permian epoch came from a land mass
west of Tasmania and possibly the centre of glaciation
was not very far from Tasmania.

The occurrences of the tillite in the state are not
continuous but the easternmost known exposure can be taken
as the minimum limit to which the glaciers reached (Map 2).
East of these deposits the extension of the glaciers can only
be suggested on the basis of the northern extension of the
glaciers in New South Wales and Victoria, and on the pres-
ence of the deposits along the east coast and on Maria
Island which have been derived from older deposits, and
which have striated and facetted pebbles.

In the north the deposits of the tillite are limited
west of a line running from east of Latrobe through Del-
oraine to Palmerston; the deposits at Avoca, Fingal and
St Mary's indicate that the glaciers extended beyond the
present east coast of Tasmania.

Between the Great Lakes and Maydena, except King
William Saddle, no exposures of tillite are known. The
glaciers probably extended farther and they extended
beyond the east coast of Maria Island.

Twelvetrees (1915, p.13) has reported the tillites
from Rocky Plains Bay and Shoemakers Point in the south,
which indicates that glaciers at least reached to this
point and probably extended farther east and south but
there are no Permian deposits reported from southern
islands so it cannot be confirmed. If the deposits re-
ported by Twelvetrees is truly tillite it would be likely

that Port Davey was also covered by the glaciers.

In the Lower Permian Epoch Tasmania was covered by a large ice sheet or a combination of several piedmont glaciers, but the evidence as discussed earlier points in favour of a true ice sheet, which moved from west to east. The glaciers not only flowed on the land but extended into the sea. In the north they reached the sea level beyond the east coast.

At the end of the glaciation the retreat was followed by an encroachment of the sea and the tillite which had been deposited below the sea level was covered by water. The transgression was gradual, starting in the south east, then advancing towards the north west (Map 3).

The sea entered between Maria Island and Mt. Picton and it extended through the Derwent Valley in the south, to Latrobe and Beaconsfield. In the west it extended beyond the west coast of Tasmania. The Upper Mersey Valley and the Cradle Mountain area may have formed a peninsula or a chain of islands extending south from the Mt. Roland Lower Paleozoic land mass, while another peninsula extended from the north east coast lower Paleozoic land mass towards Maria Island and perhaps further south. The Inglis Valley and the Wynyard area were connected by sea from the southwest end of Cradle Mountain or another sea extended from north to south in this part. The absence of Permian beds between Cradle Mt. and the Inglis Valley may be due to post-Permian erosion, non-deposition of the Permian beds, or simply to lack of exposure.

Barn Bluff, Mt. Sedgwick, and Mt. Read possibly formed the southern extension of the North West Coast Lower Paleozoic land mass, because the considerable decrease in the thickness of the Permian deposits in this region and also the overlapping of the beds within short distances suggest nearness of the land. The sea was possibly deep towards Malanna and Point Hibbs because the thickness of the Permian deposits increases gradually to the south. The absence of the upper Paleozoic and younger rocks from the Port Davey area makes it very difficult to determine the palaeogeography but following Twelvetrees' report of the occurrence of tillites in the area it is suggested that it was glaciated but the deposits have since been removed by erosion. It is also possible that the Port Davey area may have formed a separate island in the south or may have been an extension of the land mass in the west.

Palmerston Marine Beds.

Before the ice had retreated from the State, marine tillites were laid down on top of the tillites. During this period the ice was still protecting the higher parts of the land, so the erosion had not started and the tillites were covering them, so the distribution of the land and the sea was more or less the same at the time when the retreat started (Map 4).

Quamby Beds.

The glaciers had retreated farther west and whereever the tillite was not under water the erosion had started, so that the covering tillites were removed, and the rock ridges and islands had appeared (Map 5).

The land was still sinking so that the sea had advanced farther inland. In the Mersey district, the sea had advanced in the north and the west, and it reached upto Latrobe in the north, and between the Mersey and Don rivers in the west. In the upper Mersey Valley and Cradle Mountain area, the land had gone under water, but some islands were left on which Quamby beds are overlapped by younger beds.

By the end of this time Maria Island had gone down below the sea level and the sea covered some land in the north of it.

Darlington Limestone and Golden Valley beds.

The sea had advanced farther inland (Map 6) and all the islands in the Mersey Valley and in the Cradle Mountain area had gone down under water. In the Mersey District the sea had reached up to Devonport in the north, and the Don River in the west.

In the Maria Island area the sea had probably encroached farther north, but it is not traceable.

By the end of the Golden Valley period the land had started to rise and small swamps and inland coastal lakes had developed, in which coal was deposited.

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W Y N Y A R D.

No.	Size in m.m.	Round-ness.	Spher-icity.	Soling.
1	-	0.44	-	
2	174	0.16	0.59	
3	122	0.59	0.63	
4	161	0.10	0.58	
5	113	0.57	0.67	0.10
6	99	0.29	0.69	
7	93	0.61	0.64	.02
8	92	0.15	0.67	
9	91	0.45	0.64	
10	139	0.18	0.52	
11	131	0.16	0.52	
12	83	-	0.75	.20
13	68	0.24	0.64	
14	70	0.40	0.69	
15	58	0.45	0.66	.10
16	56	0.35	0.61	
17	64	0.23	0.54	.20
18	42	0.30	0.71	
19	56	0.34	0.53	.20
20	60	0.38	0.60	.09
21	30	0.70	0.8	.28
22	32	0.50	0.75	.26
23	48	0.58	0.54	
24	44	0.42	0.55	.27
25	36	0.77	0.68	.30
26	36	0.53	0.66	.40
27	34	0.37	0.63	
28	44	0.41	0.68	.23
29	38	0.39	0.69	
30	40	0.55	0.56	.25
31	30	0.38	0.81	.25
32	40	0.36	0.53	.18
33	42	0.50	0.58	.21
34	36	0.50	0.65	.2
35	32	0.55	0.68	.16
36	50	0.39	0.53	
37	28	0.80	0.66	.28
38	58	0.36	0.59	
39	34	0.34	0.66	
40	44	0.28	0.53	
41	34	0.41	0.62	
42	32	0.44	0.66	.16
43	28	0.45	0.62	
44	44	0.30	0.48	
45	40	0.35	-	
46	40	0.27	0.66	
47	34	0.25	0.57	

No.	Size in m.m.	Round-ness.	Spher-icity.	Soling.
48	30	0.61	0.62	
49	22	0.54	0.72	.52
50	24	0.43	0.70	
51	24	0.41	0.56	
52	28	0.33	0.67	
53	34	0.48	0.65	
54	28	0.42	0.70	
55	28	0.38	0.73	
56	28	0.39	0.76	
57	26	0.58	0.54	
58	26	0.40	0.61	
59	30	0.53	0.52	
60	26	0.45	0.60	
61	20	0.50	0.79	
62	20	0.31	0.62	.43
63	18	0.49	0.76	
64	32	0.33	-	
65	24	0.58	0.75	
66	34	0.41	0.46	
67	16	0.66	0.76	
68	28	0.38	0.45	
69	24	0.53	0.65	
70	24	0.52	0.65	
71	24	0.42	0.58	.36
72	20	0.46	0.78	
73	26	0.41	0.60	
74	26	0.44	0.65	
75	26	0.44	0.65	
76	24	0.41	0.51	
77	18	0.65	0.64	
78	34	0.38	0.60	
79	24	0.50	0.52	
80	22	0.40	0.61	
81	20	0.38	0.62	
82	22	0.37	0.30	
83	22	0.53	0.61	
84	22	0.67	0.61	
85	22	0.44	0.61	
86	24	0.33	0.61	
87	28	0.31	0.49	
88	20	0.42	0.72	
89	18	0.38	0.80	
90	22	0.35	0.72	
91	22	0.43	-	
92	20	0.66	0.61	

No.	Size in m.m.	Round-ness.	Spher-icity.	Soling.
93	20	0.47	0.61	
94	16	0.70	0.76	
95	16	0.52	0.76	
96	20	0.52	0.62	
97	18	0.50	0.76	
98	20	0.44	0.62	
99	24	0.40	0.53	
100	22	0.52	0.56	
101	20	0.34	0.61	
102	16	0.37	0.76	
103	16	0.33	0.76	
104	20	0.50	0.61	
105	18	0.54	0.78	
106	18	0.38	0.37	
107	20	0.72	0.61	
108	24	0.37	0.52	
109	20	0.40	0.62	
110	18	0.55	0.78	
111	24	0.40	0.53	
112	14	0.75	0.45	

P A L M E R S T O N

No	Size in m.m.	Round-ness.	Spher-icity.	Soling.
1	70	0.27	0.67	0.07
2	44	.43	.81	-
3	52	.41	.61	0.02
4	58	.44	.60	0.12
5	54	.44	.50	-
6	42	.58	.66	0.21
7	40	.43	.67	
8	24	.55	.89	
9	36	.62	.70	0.14
10	34	.34	.79	
11	36	.29	.55	
12	36	.44	.59	
13	36	.45	.55	
14	36	.37	.59	
15	30	.42	.71	
16	32	.31	.66	
17	26	.42	.74	
18	26	.42	.78	
19	32	.43	.66	
20	32	.45	.66	0.20
21	32	.23	.64	
22	36	.35	.64	0.13
23	30	.54	.71	
24	28	.59	.72	
25	26	.54	.26	
26	36	.260	.60	
27	28	.55	.70	0.25
28	30	.39	.60	
29	24	.35	.65	0.64
30	28	.36	.75	
31	28	.57	.45	
32	34	.76	.63	
33	22	.52	.68	
34	22	.43	.73	
35	24	.54	.59	
36	22	.42	.67	0.40
37	28	.17	.57	0.25
38	22	.41	.65	
39	28	.52	.71	0.22
40	26	.55	.78	
41	26	.57	.61	
42	48	.50	.70	
43	30	.46	.66	
44	22	.57	.90	

No.	Size in m.m.	Round- ness	Spher- icity.	Soling
45	20	0.57	0.72	
46	20	.47	.79	
47	16	.66	.78	0.37
48	24	.25	.59	
49	22	.54	.63	
50	20	.53	.63	.40
51	20	.43	.70	
52	22	.31	.63	
53	26	.51	.61	
54	24	.44	.59	
55	34	.44	.59	
56	20	.54	.70	
57	20	.50	.70	
58	78	.39	.45	
59	24	.41	.65	
60	18	.40	.35	
61	20	.65	.63	
62	36	.36	.50	
63	22	.64	.57	
64	18	.38	.78	
65	20	.52	.32	
66	16	.60	.45	
67	18	.55	.70	
68	18	.58	.45	
69	24	.36	.71	
70	18	.31	.70	
71	14	.42	.48	
72	16	.60	.88	
73	22	.63	.46	.50
74	20	.41	.59	
75	18	.70	.41	.42
76	18	.60	.35	
77	20	.71	.70	
78	20	.33	.63	
79	22	.46	.63	
80	24	.52	.63	
81	16	.55	.90	
82	18	.37	.35	
83	20	.63	.40	
84	18	.50	.70	
85	22	.57	.35	
86	24	.36	.39	
87	20	.44	.75	
88	20	.61	.63	
89	22	.41	.57	
90	18	.60	.70	
91	20	.56	.41	

No.	Size in m.m.	Round-ness.	Spher-icity.	Soling.
92	20	.60	.63	
93	18	.46	.70	
94	16	.50	.77	
95	16	.57	.77	
96	22	.52	.57	
97	20	.33	.75	
98	20	.43	.63	
99	16	.50	.77	
100	16	.42	.77	

M A Y D E N A

No.	Size in m.m.	Round- ness.	Spher- icity.	Soling.
1	70	0.27	0.67	0.07
2	44	.43	.81	
3	52	.41	.61	.02
4	58	.44	.60	.12
5	54	.44	.50	
6	42	.58	.66	.21
7	40	.43	.67	
8	24	.55	.89	
9	36	.62	.70	.14
10	34	.33	.79	
11	36	.29	.55	
12	36	.44	.59	
13	36	.45	.55	
14	36	.37	.59	
15	30	.42	.71	
16	32	.31	.66	
17	26	.42	.74	
18	26	.42	.78	
19	32	.43	.66	
20	32	.45	.66	.20
21	32	.23	.64	
22	36	.35	.64	.13
23	30	.54	.71	
24	28	.54	.72	
25	26	.43	.54	
26	36	.26	.60	
27	28	.44	.70	.25
28	30	.39	.60	
29	24	.35	.65	
30	28	.36	.75	
31	28	.57	.45	
32	34	.76	.63	
33	22	.52	.68	
34	22	.43	.73	
35	24	.54	.59	
36	22	.42	.67	.40
37	28	.17	.53	.25
38	22	.41	.65	
39	28	.52	.71	.22
40	26	.55	.78	
41	26	.57	.61	
42	48	.50	.70	
43	30	.46	.66	
44	22	.57	.90	
45	20	.57	.72	
46	20	.47	.79	
47	16	.66	.78	
48	24	.25	.59	
49	22	.54	.63	
50	20	.53	.63	.40

No.	Size in m. m.	Round-ness.	Spher-icity.	Soling.
51	20	0.43	0.70	
52	22	.31	.63	
53	26	.51	.61	
54	24	.53	.59	
55	34	.41	.59	
56	20	.54	.70	
57	20	.50	.70	
58	18	.39	.45	
59	24	.41	.65	
60	18	.40	.35	
61	20	.65	.63	
62	36	.36	.50	
63	22	.64	.57	
64	18	.38	.78	
65	20	.52	.32	
66	16	.60	.45	
67	18	.55	.70	
68	18	.50	.45	
69	24	.36	.71	
70	18	.31	.70	
71	14	.42	.48	
72	16	.60	.88	
73	22	.46	.63	0.50
74	20	.59	.41	
75	18	.41	.70	.42
76	18	.60	.35	
77	20	.71	.70	
78	20	.33	.63	
79	22	.46	.63	
80	24	.52	.63	
81	16	.55	.90	
82	18	.37	.35	
83	20	.40	.63	
84	18	.50	.70	
85	22	.35	.57	
86	24	.36	.59	
87	20	.44	.75	
88	20	.61	.63	
89	22	.41	.57	
90	18	.60	.70	
91	20	.56	.41	
92	20	.60	.63	
93	18	.46	.70	
94	16	.50	.77	
95	16	.57	.77	
96	22	.52	.57	
97	20	.33	.75	
98	20	.43	.63	
99	16	.77	.50	
100	16	.77	.42	

KING WILLIAM SADDLE

No.	Size in m.m.	Round-ness.	Spher-icity.	Soling.
1	244	0.22	–	0.08
2	164	.23	0.57	0.03
3	112	.42	.65	
4	172	.13	.55	
5	178	.21	.52	
6	122	.08	.64	
7	120	.16	.65	
8	144	.24	.59	
9	130	.16	.51	
10	128	.11	.56	
11	110	.13	.46	
12	130	.15	.49	
13	190	.19	.35	
14	138	.21	.51	
15	116	.34	.55	
16	88	.20	.65	
17	108	.37	.58	
18	106	.16	.58	
19	104	.16	.51	
20	106	.16	.59	
21	94	.24	.61	.05
22	110	.17	.54	
23	114	.29	.51	.08
24	102	.23	.57	
25	104	.44	.67	
26	110	.28	.52	.06
27	81	.40	.68	
28	66	.17	.72	
29	102	.19	.61	
30	98	.14	.61	
31	96	.20	.61	
32	106	.21	.56	
33	88	.23	.63	
34	100	.31	.35	
35	94	.15	.95	
36	86	.25	.61	.07
37	100	.14	.52	
38	92	.31	.52	
39	94	.28	.57	.11
40	70	.30	.65	
41	88	.36	.50	.02
42	50	.35	.57	
43	52	.53	.56	.08

No	Size in m.m.	Round-ness.	Spher-icity.	Soling.
44	46	0.28	0.70	
45	44	.55	.59	
46	42	.33	.69	
47	42	.37	.61	.20
48	42	.30	.65	
49	-	.32	-	
50	32	.42	.70	
51	34	.90	.65	.24
52	40	.44	.61	
53	36	.60	.67	.51
54	42	.50	.45	
55	32	.60	.56	.16
56	34	.52	.71	.15
57	30	.60	.71	.22
58	42	.62	.53	
59	32	.50	.66	.21
60	30	.28	.75	
61	34	.55	.70	.17
62	40	.37	.53	
63	30	.35	-	
64	34	.36	.62	
65	36	.31	.64	
66	30	.28	.75	.31
67	32	.42	.62	
68	32	.35	.61	.24
69	32	.28	.67	
70	30	.42	.67	
71	30	.37	.52	
72	30	.37	.65	
73	28	.42	.64	.38
74	30	.31	.60	
75	30	.42	.60	
76	38	.44	.53	
77	28	.42	.57	
78	22	.40	.65	
79	22	.66	.81	.58
80	24	.50	.80	.22
81	30	.50	.60	.19
82	26	.44	.60	
83	24	.53	.66	.15
84	30	.58	.46	.33
85	32	.66	.56	.17
86	28	.33	.64	.23
87	20	.55	.70	.41
88	24	.33	.58	.32
89	22	.45	.71	
90	26	.41	.61	.38

No	Size in m.m	Round- ness.	Spher- icity.	Soling
91	28	0.50	0.57	0.30
92	22	.50	.70	.21
93	22	.46	.70	.25
94	22	.50	.65	.20
95	20	.64	.80	.26
96	28	.33	.50	
97	28	.36	.54	
98	26	.53	.75	.18
99	22	.33	.61	
100	20	.14	.65	
101	22	.60	.63	
102	26	.40	.61	
103	22	.40	.63	
104	25	.40	.55	
105	18	.40	.66	.25
106	30	.34	.66	
107	26	.60	.60	.56
108	28	.45	.50	.25
109	26	.50	.54	
110	24	.41	.54	
111	20	.50	.50	
112	22	.50	.63	
113	20	.56	.60	
114	20	.48	.66	
115	22	.33	.72	
116	18	.40	.86	.27
117	20	.50	.60	.50
118	22	.48	.74	
119	22	.50	.62	.25
120	28	.52	.50	.20
121	20	.50	.66	
122	25	.55	.65	
123	22	.75	.72	
124	20	.50	.60	.70
125	22	.50	.74	.24
126	20	.58	.60	
127	20	.50	.32	.59
128	20	.45	.66	
129	28	.40	.50	.27
130	24	.33	.54	.40
131	18	.50	.67	
132	18	.66	.67	.25
133	22	.36	.72	
134	26	.37	.49	.36
135	20	.38	.60	
136	18	.40	.67	
137	24	.53	.67	.24

LOCALITY INDEX

NAME	LAT.	LONG.	COUNTRY
Alaska Basin	50°0'N.	165°0'W.	U.S.A.
Anson Bay	41°2'S.	148°15'E.	Tasmania
Avoca	41°46'S.	147°40'E.	
Baja California	40°0'N.	120°0'W.	N. Am.
Barn Bluff	41°44'S.	145°56'E.	Tasmania
Beaconsfield	41°11'S.	146°45'E.	
Big Horn Mountain	40°30'N.	107°40'W.	U.S.A.
Cape Cod	42°0'N.	70°0'W	U.S.A.
Clumner Bluff	41°43'S.	145°15'E.	Tasmania
Cradle Mountain	40°41'S.	145°58'E.	
Dairy Plains	41°35'S.	146°31'E.	
Dalmayne	41°41'S.	148°15'E.	
Darlington	42°35'S.	148°4'E.	
Deloraine	41°31'S.	146°40'E.	
D'Entrecasteaux Channel	43°13'S.	147°18'E.	
Derwent Valley	42°45'S.	147°0'E.	
Doctors Rock	41°0'S.	145°46'E.	
Don River	41°13'S.	146°19'E.	
Dreamy Bay	Woody Island		
Eden	41°59'S.	145°18'E.	
Elephant Pass	41°36'S.	148°10'E.	
Fingal	41°36'S.	147°55'E.	
Florentine Valley	42°35'S.	146°30'E.	
Hellyer Gorge	41°16'S.	145°36'E.	
Holiks Glacier	77°55'S.	164°0'E.	Antarctica
Ida Bay	43°27'S.	146°54'E.	Tasmania
Inglis River	41°0'S.	145°38'E.	
Johnston Point	42°37'S.	147°57'E.	
King William Saddle	42°12'S.	146°6'E.	

NAME	LAT	LONG	COUNTRY
Lake River	41° 50'S.	147° 3'E.	
Lewis Point	Woody Island.		
Llandaff	41° 57'S.	148° 10'E.	
Long Island	40° 50'N.	73° 0'W.	U. S. A.
Malana	42° 1'S.	145° 13'E.	Tasmania.
Maria.Island	42° 37'S.	148° 0'E.	
Maydena	42° 45'S.	145° 5'E.	
Mersey District	41° 17'S.	146° 25'E.	
Mersey River	41° 19'S.	146° 28'E.	
Mole Creek	41° 35'S.	146° 25'E.	
Mt Dundas	41° 50'S.	145° 27'E.	
Mt.Erebus	77° 31'S.	166° 12'E.	Antarctica.
Mt. King William I	42° 13'S.	146° 7'E.	Tasmania.
Mt. Paul	42° 3'S.	146° 15'E.	
Mt.Picton	43° 7'S.	146° 35'E.	
Mt.Pelion	41° 53'S.	146° 5'E.	
Mt. Read	41° 53'S.	145° 33'E.	
Mt. Roland	41° 26'S.	146° 18'E.	
Mt. Sedgwick	42° 0'S.	145° 35'E.	
Oonah	41° 14'S.	145° 43'E.	
Palmerston	41° 47'S.	146° 59'E.	
Park Nook	41° 50'S.	147° 4'E.	
Piccaninny Creek	41° 42'S.	148° 15'E.	
Pieman River	41° 40'S.	144° 55'E.	
Point Hibbs	42° 43'S.	145° 15'E.	
Port Davey	43° 20'S.	145° 55'E.	
Preolenna	41° 6'S.	145° 33'E.	
Quamby Bluff	41° 39'S.	146° 41'E.	

NAME	LAT.	LONG.	COUNTRY
Quamby Brook	41°35'S.	146°42'E.	
Reidle Bay	42°40'S.	148°0'E.	
Rocky Bay Plain	43°35'S.	146°39'E.	
Ross Island	77°30'S.	168°0'E.	Antarctica.
Satellite	Woody Island		Tasmania.
St. Mary's	41°35'S.	148°10'E.	
Seymour	41°45'S.	148°16'E.	
Shoemakers Point	43°36'S.	146°40'E.	
Snake Creek	41°40'S.	146°15'E.	
South Cape	43°37'S.	146°43'E.	
Sunset Bay	Woody Island		
Styx River	42°50'S.	146°35'E.	
Table Cape	40°58'S.	145°44'E.	
Tamar Head	41°50'S.	146°50'E.	
Taylor Glacier	77°45'S.	162°30'E.	Antarctica.
Wasaka Bay	36°0'N.	136°0' E.	Japan.
Waratah			Tasmania.
Weld River	43°0'S.	146°42'E.	
Western Tiers	41°50'S.	147°0'E.	
Woodbridge	43°10'S.	147°14'E.	
Woody Island	43°19'S.	147°13'E.	
Wynyard	41°1'S.	145°42'E.	
Zeehan	41°53'S.	145°20'E.	

Plate 1.

Gallenkamps Sedimentation Balance

Plate 2.

Varves At Doctor's Rock.

Plate 3 A.

Pebbly Mudstone at Wynyard.

Plate 3 B.

Pebbly Mudstone Beds.

Same Area.

Plate 4.

Outwash Beds At Doctors Rock.

Plate 5 A.

Out...sh Grit Beds At Wynyard.

Plate 5 B.

Outwash Grit Beds At Wynyard.

Plate 6 A.

Outwash Beds At Doctors Rock.

Plate 6 B.

Outwash Beds At Doctors Rock.

Plate 7.

Sandstone concretion, At Wynyard

Plate 8.

Dropped Pebbles In Palmerston Beds.

Plate 9.

King William Saddle Tillite.

Plate 10.
Eldon Group
Fossiliferous ~~Ordovician~~ Pebbles.

Plate 11.

Fossil Piecs in Maydena Tillite.
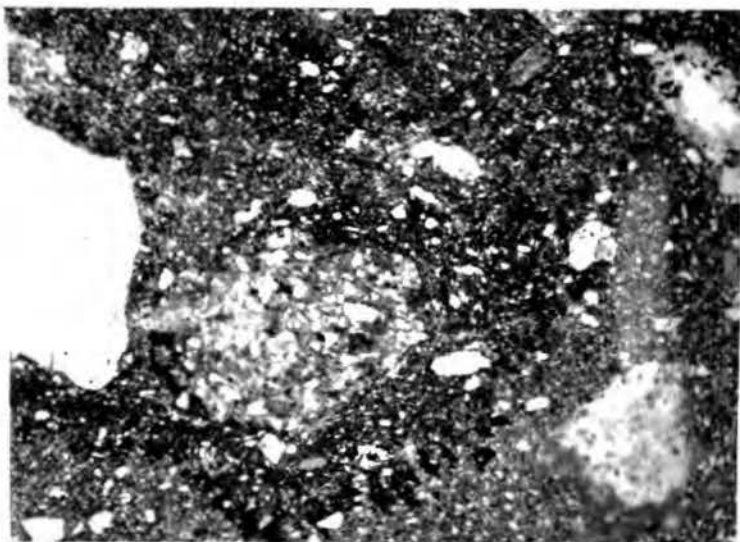
Plate 12

Photomicrograph of King William Tillite.

Plate 13.

Glacial Pebbles.
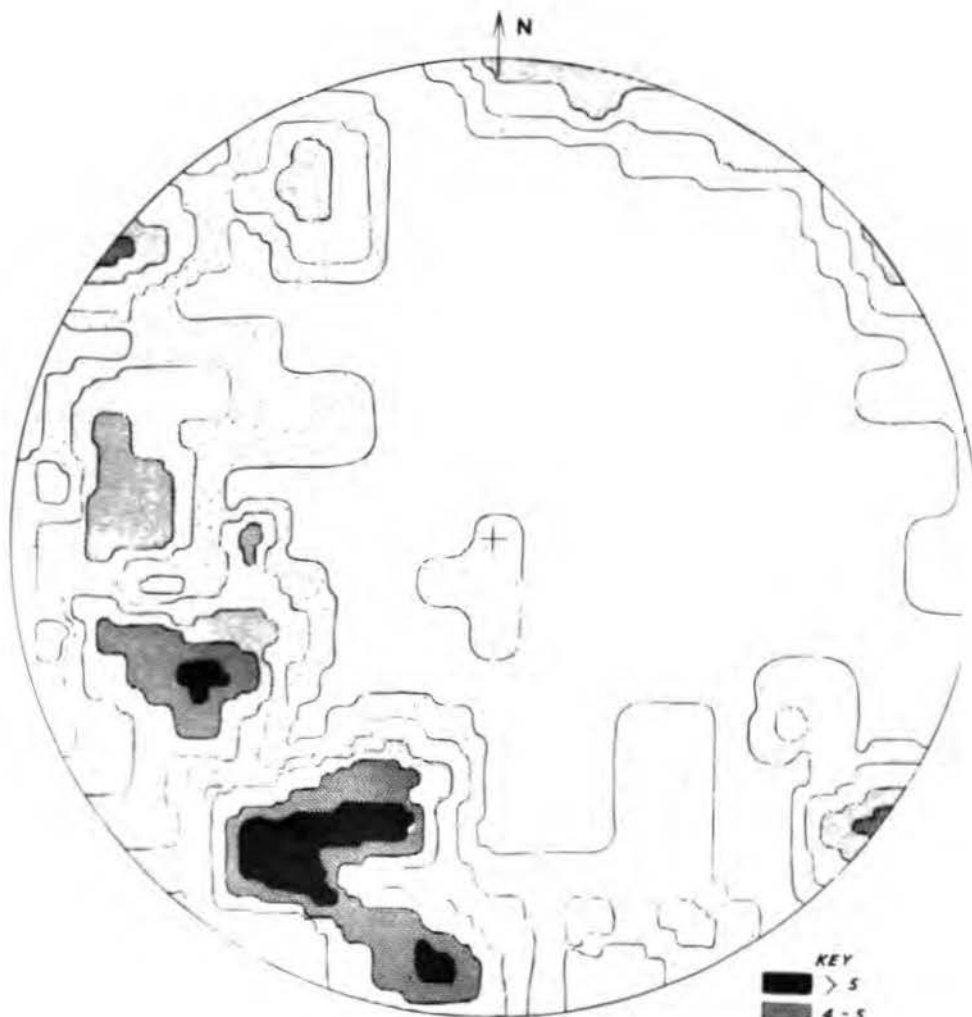
INGLIS RIVER

PLATE 14A

KEY
> 5
4 - 5
3 - 4
2 - 3
1 - 2
< 1
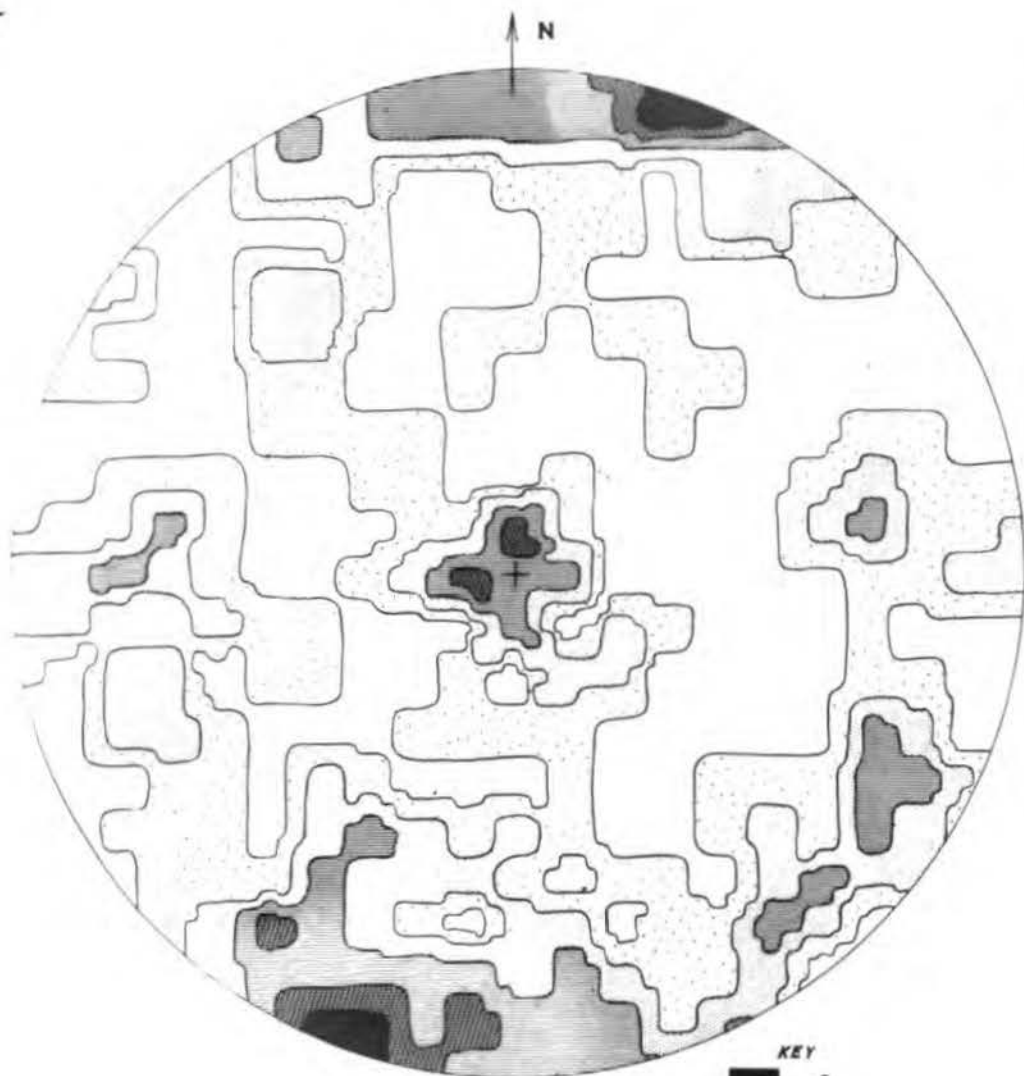
DOCTOR'S ROCK

PLATE 14B

KEY

5
4 - 5
3 - 4
2 - 3
1 - 2
1
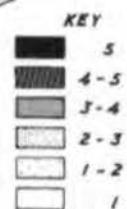
Plate 15.

Fractured Mudstone Erratic.

Plate 16.

Same Pebble

Looking at the Base.

Plate 17.
Pebble from Plate 15.

Second Side.

Plate 18.

Fracutred Sandstone Erratic.

Plate 19.

Erratic from Plate 18.
Separated.

Plate 20.

Fractured and Rounded Sandstone Erratics.

Plate 21.

Fractured Erratic.

Plate 22.

Woodbridge Glacial Beds.

Cape Paul Lamanon.

Plate 23.

Striated Pebbles.

Palmerston Beds..

Plate 24.

Basal Conglomerate Mersey Valley.

Plate 25.

Fossil and Erratic Band in
Yersby Moraine

Plat 26.

Sandstone Beds at Elephant Pass

Plate 25 A.

A Closeup Of Plate 25.

Plate 25 B.

Fossils and Dropped Pebbles In Quamy beds.

Plate 27.

Unconformable Permian and Precambrian Contact
At Wynyard.

Plate 28.

Outwash Beds At Maydena.

Plate 29.

Striated Erratics Insitu.

MAP 1

PRESENT DISTRIBUTION OF LOWER PERMIAN DEPOSITS IN TASMANIA

LEGEND

● TILL
○ QUAMBY MUDSTONE AND
   WOODY ISLAND SILTSTONE
+ DARLINGTON LIMESTONE
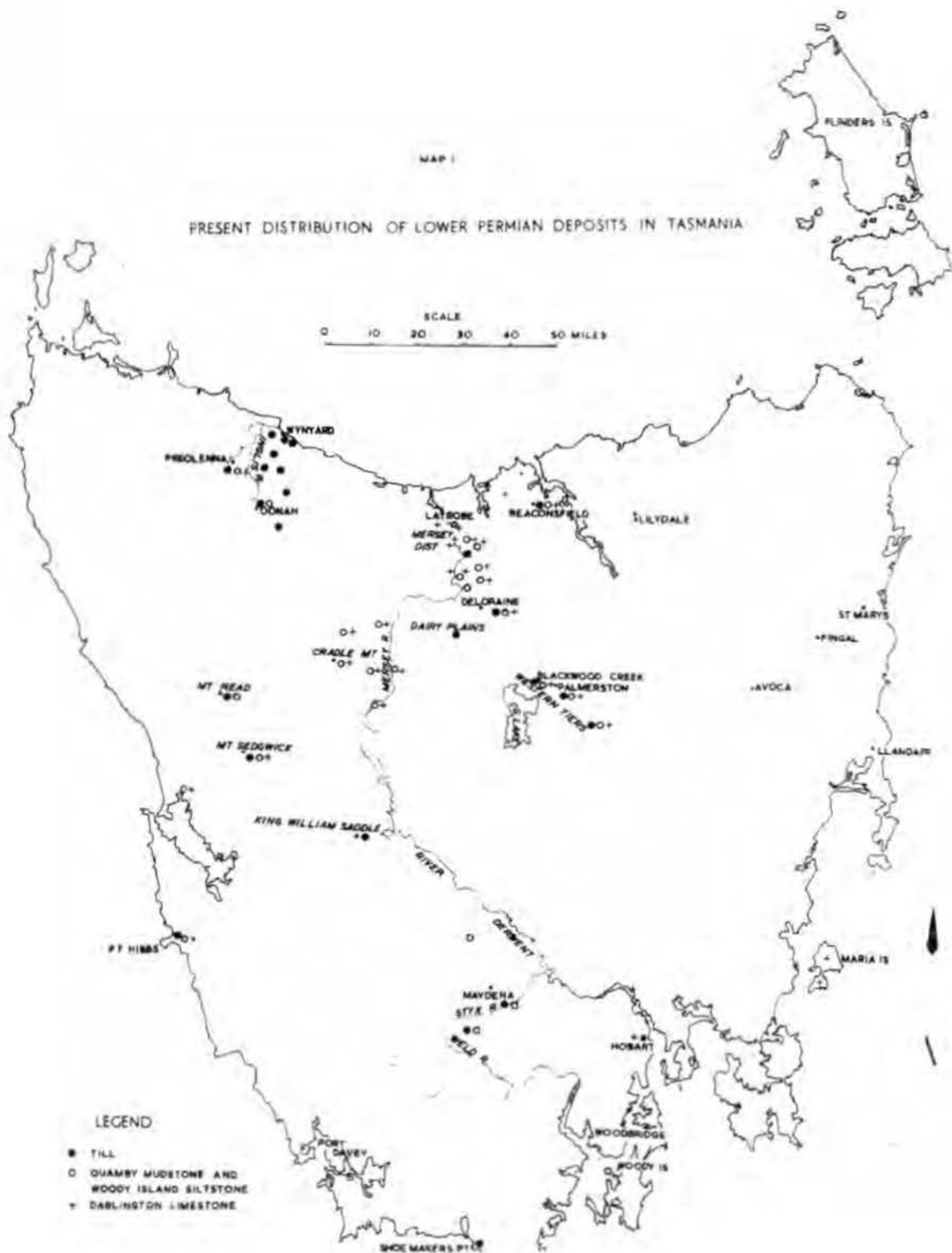
MAP 2

EASTERN LIMIT OF BASAL PERMIAN GLACIATION IN TASMANIA

BASED ON THE PRESENCE OF TILLITE

SCALE

0   10   20   30   40   50 MILES
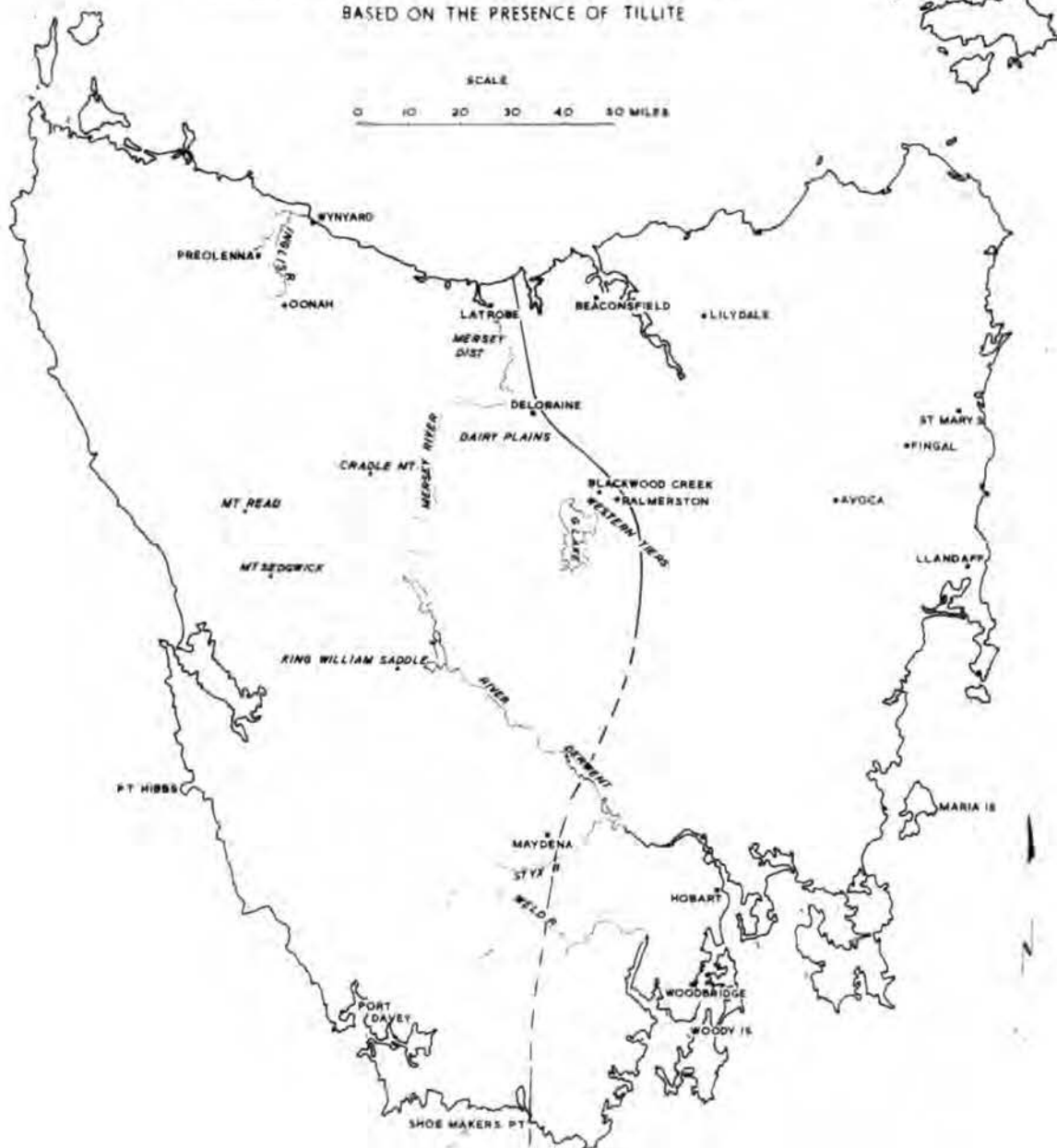
MAP 3

LAND AND SEA DISTRIBUTION AT THE END OF
BASAL PERMIAN GLACIATION IN TASMANIAN

SCALE

0   10   20   30   40   50 MILES

FLINDERS IS

WYNYARD

PREOLENNA*

*OONAH

LATROBE   BEACONSFIELD   *LILYDALE

MERSEY
DIST.

DELORAINE

DAIRY PLAINS

CRADLE MT

ST MARY S

*FINGAL

BLACKWOOD CREEK
*PALMERSTON

WESTERN TIERS

*AVOCA

MT. READ

LLANDAFF

MT. SEDGWICK

KING WILLIAM SADDLE

RIVER DERWENT

PT HIBBS

MARIA IS

MAYDENA

STYX R

WELD R

HOBART

WOODBRIDGE

WOODY IS

LAND

PROBABLE LAND

PORT
DAVEY

SHOE MAKERS PT

Correlation diagram showing stratigraphic columns for:

DELORAINE — CRESSY
AFTER
WELLS (1957) AND McKELLAR (1957)

HOBART — WOODY ISLAND
AFTER
BANKS AND HALE (1957)

ELEPHANT PASS, ST. MARYS
AFTER
K.G BRILL

Correlation after Banks

SECTION

PLAN

ROTATION OF ERRATICS IN GLACIER

FIG 1.

WOODBRIDGE GLACIAL FORMATION

(at Woodbridge) after D.Woolley

WYNYARD

PALMERSTON

MAYDENA

WEIGHT PERCENTAGE FREQUENCY

DIAMETER IN MM.
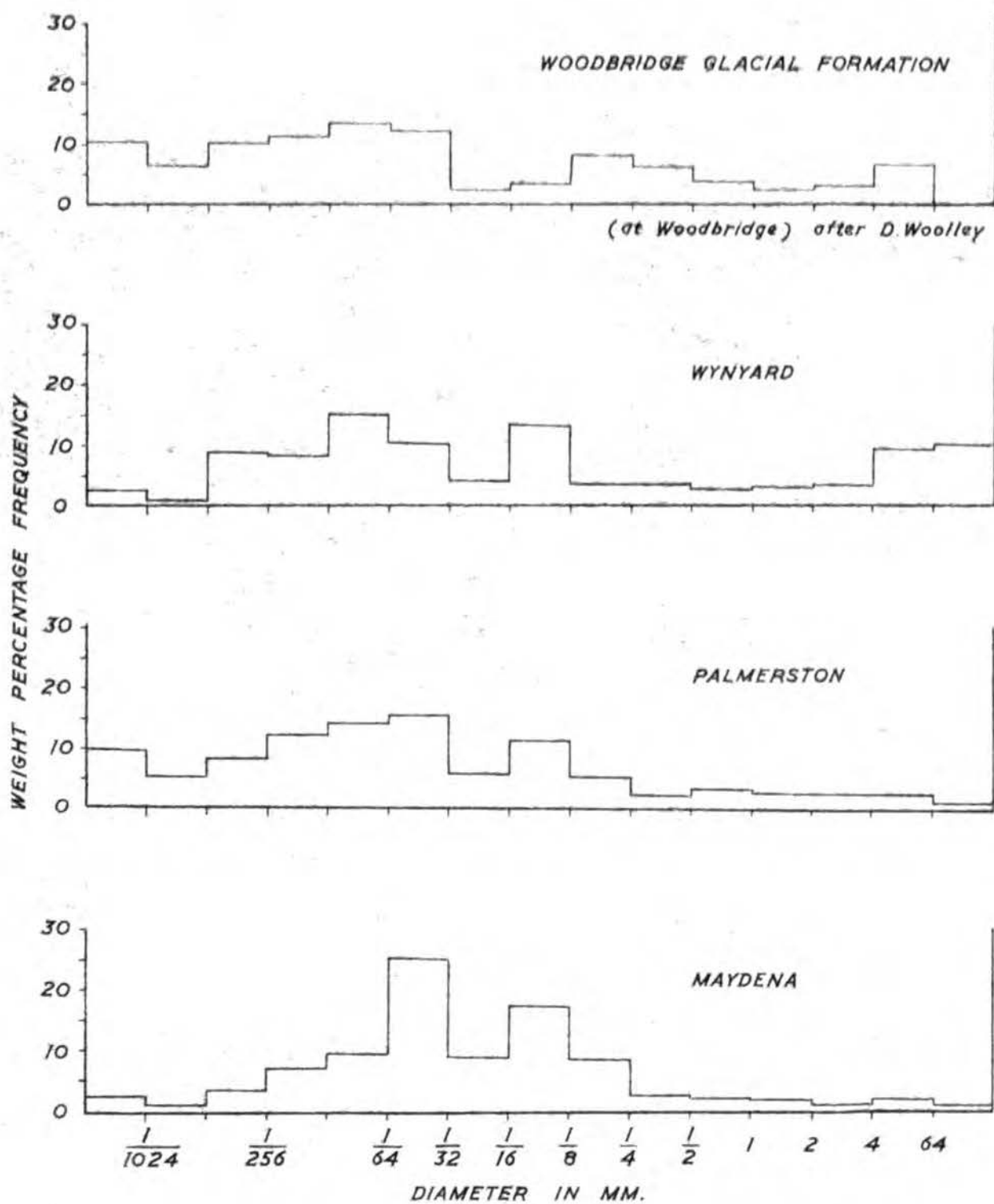
SIZE AND FREQUENCY HISTOGRAM
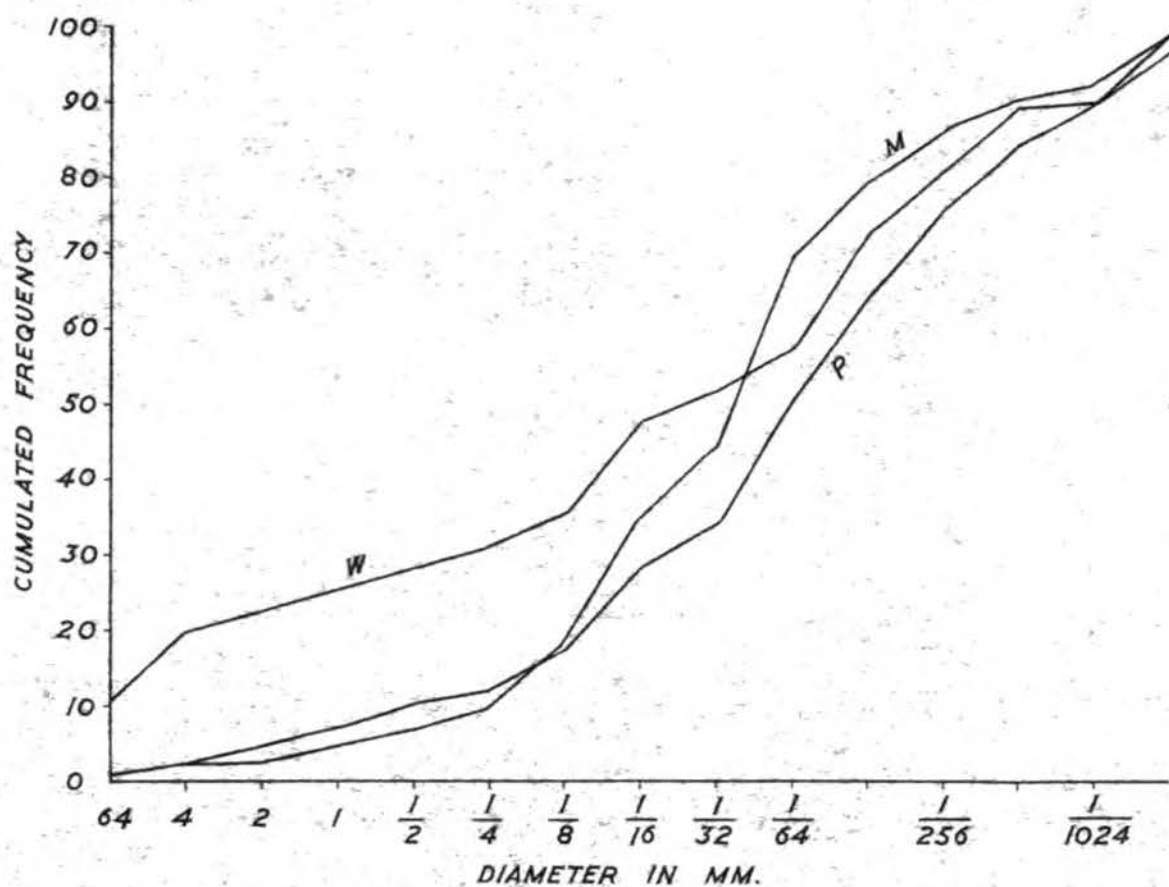
Figure 2

FREQUENCY CUMULATIVE CURVE

W   WYNYARD

M   MAYDENA
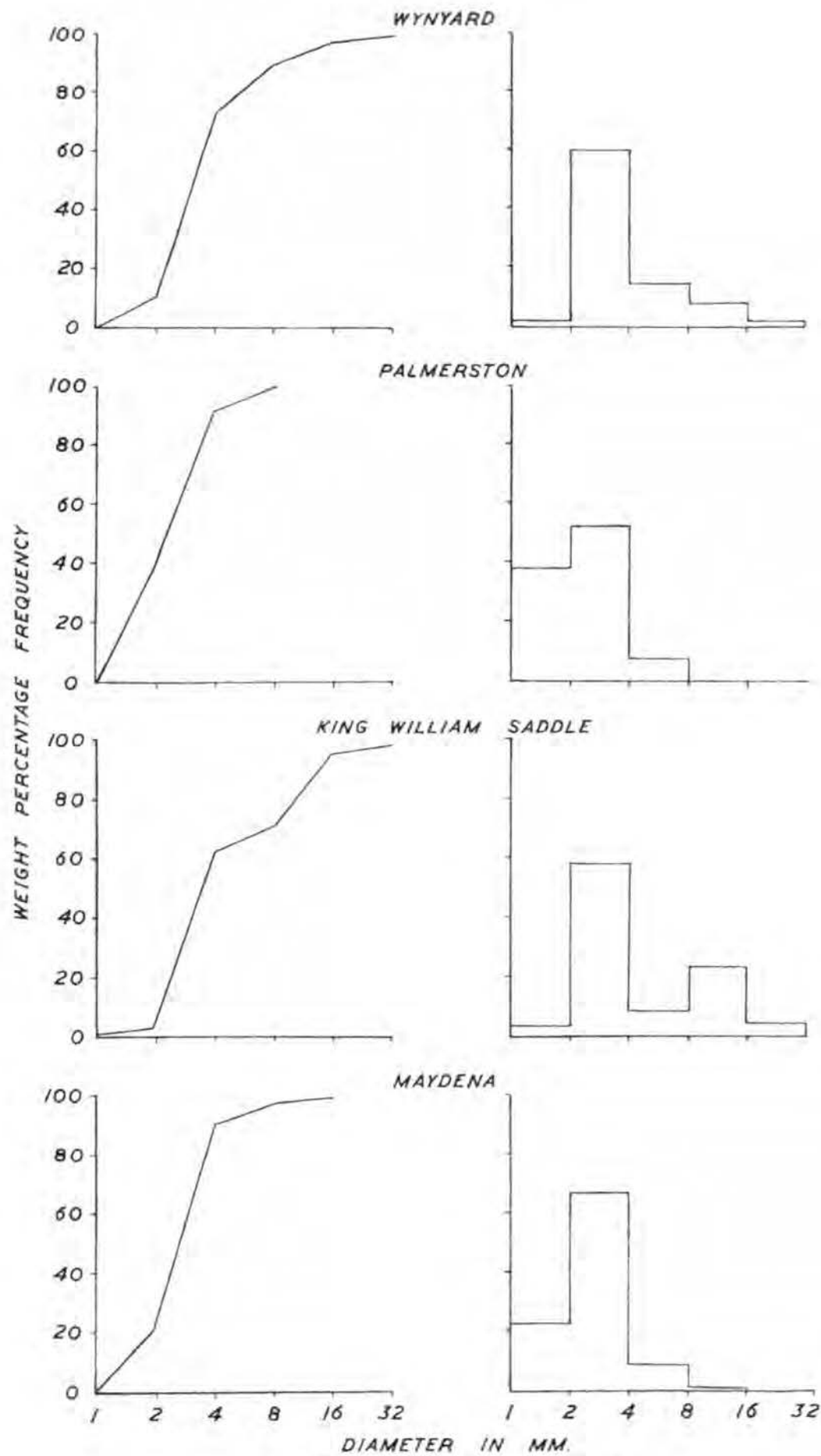
P   PALMERSTON

Figure 3
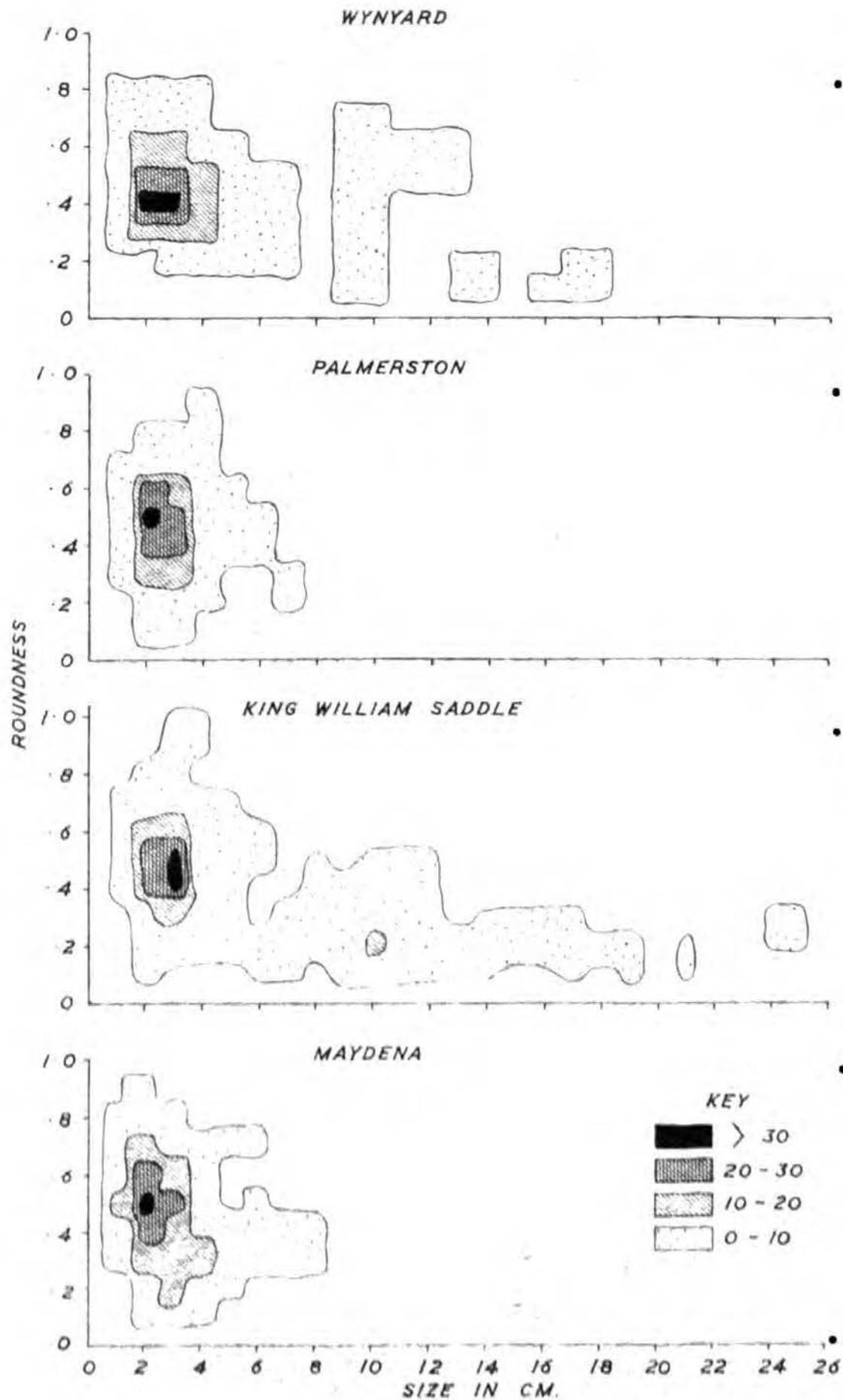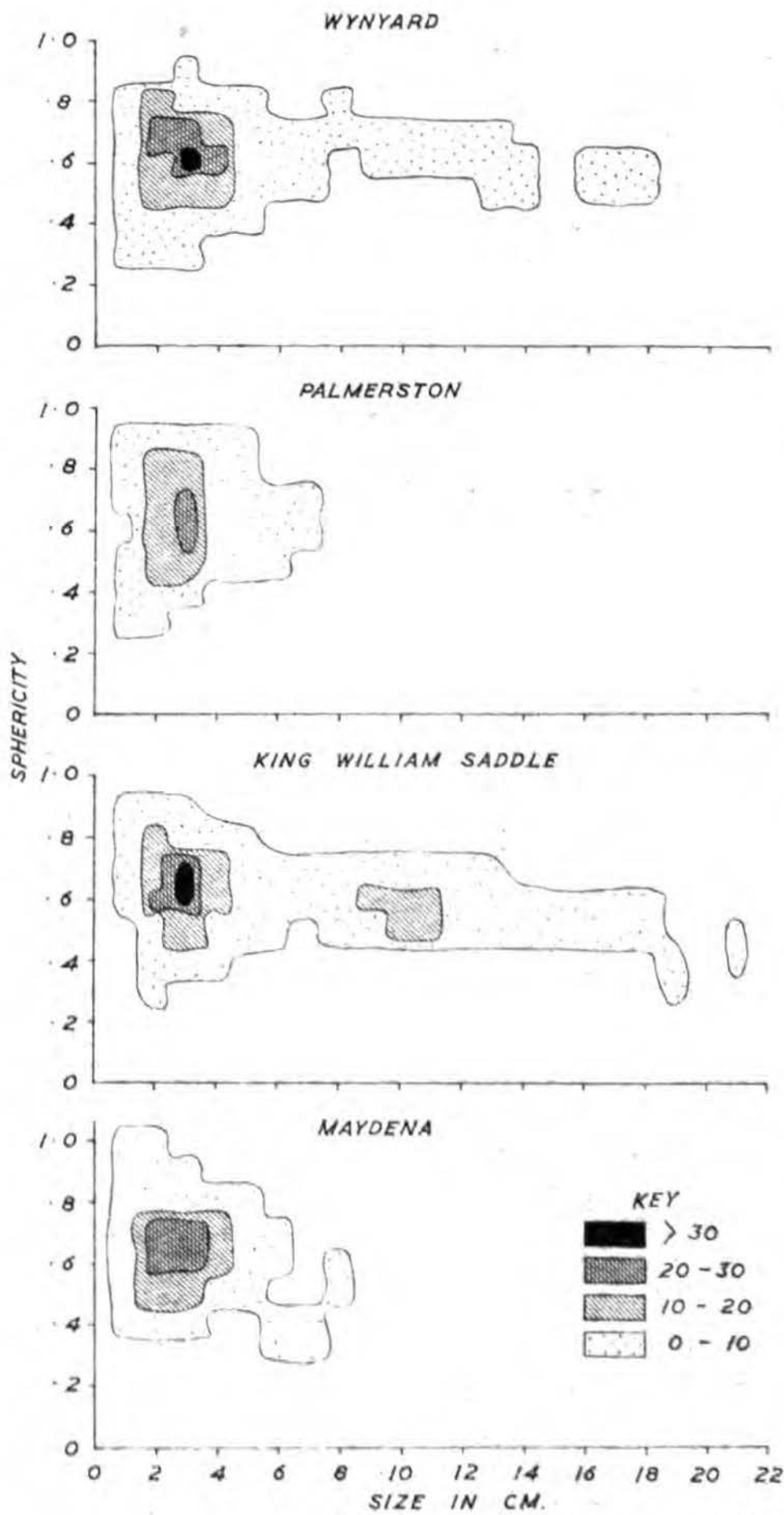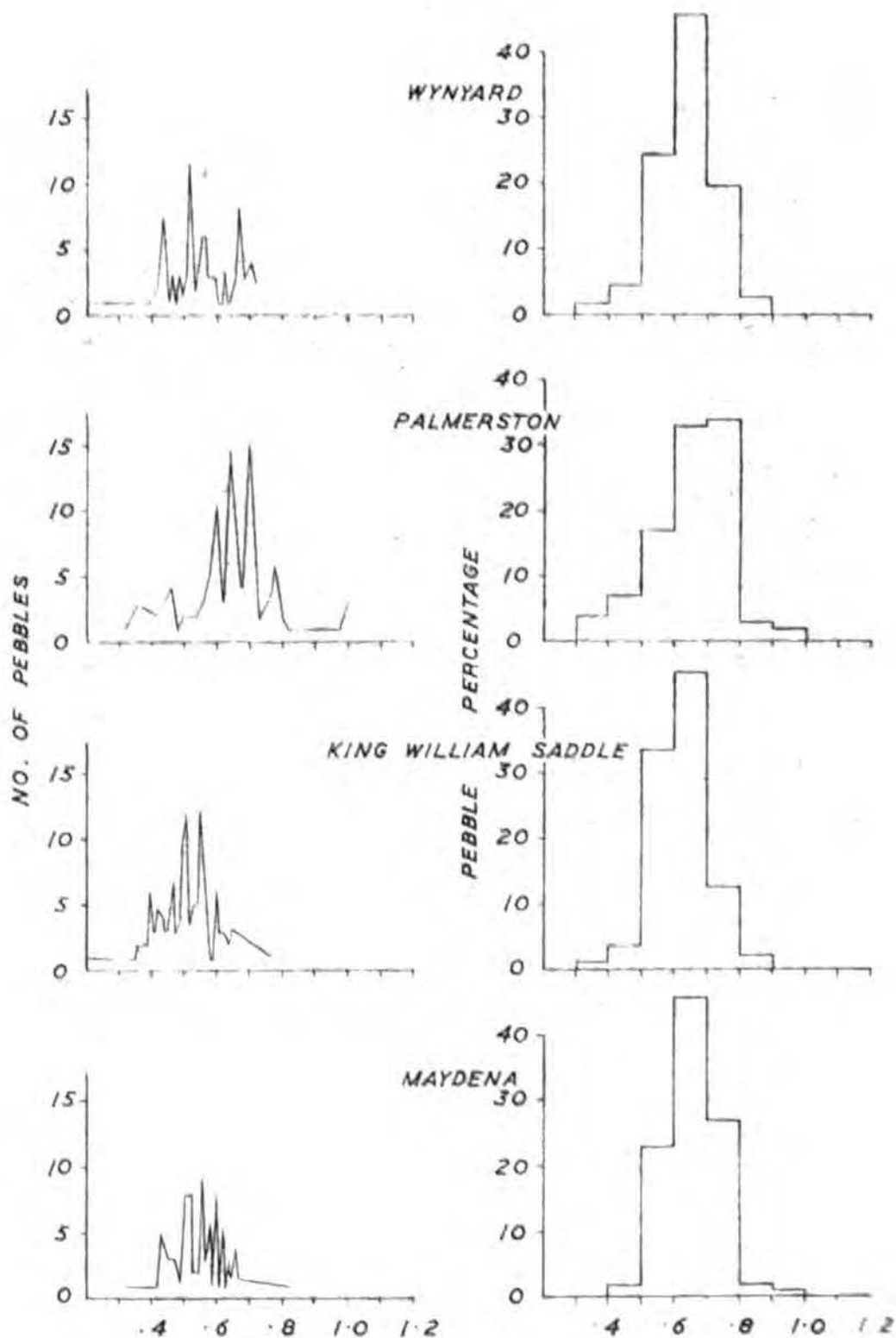
WYNYARD

PALMERSTON

KING WILLIAM SADDLE

MAYDENA

WEIGHT PERCENTAGE FREQUENCY

DIAMETER IN MM.

ERRATICS SIZE FREQUENCY

Figure 4

WYNYARD

PALMERSTON

KING WILLIAM SADDLE

MAYDENA

ROUNDNESS

SIZE IN CM.

KEY

■ > 30
▨ 20 - 30
▧ 10 - 20
□ 0 - 10

SIZE AND ROUNDNESS

Figure 5

WYNYARD

PALMERSTON

KING WILLIAM SADDLE

MAYDENA

KEY
>30
20 - 30
10 - 20
0 - 10

SPHERICITY

SIZE IN CM.

SIZE AND SPHERICITY

Figure 6

WYNYARD

PALMERSTON

KING WILLIAM SADDLE

MAYDENA

NO. OF PEBBLES

PERCENTAGE

PEBBLE

SPHERICITY AND FREQUENCY

Figure 7

WYNYARD

PALMERSTON

ROUNDNESS

KING WILLIAM SADDLE

MAYDENA

KEY

> 30
20 - 30
10 - 20
0 - 10

SPHERICITY

ROUNDNESS AND SPHERICITY   Figure 8

WYNYARD

PALMERSTON

KING WILLIAM SADDLE

MAYDENA

NO. OF PEBBLES

PERCENTAGE

PEBBLE
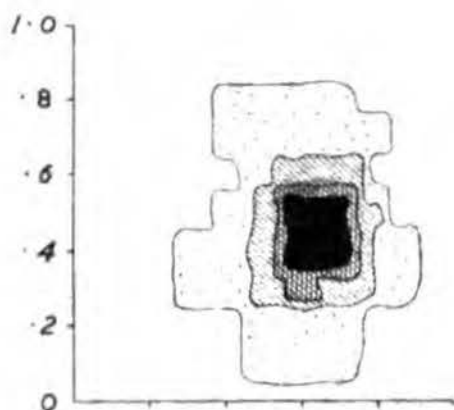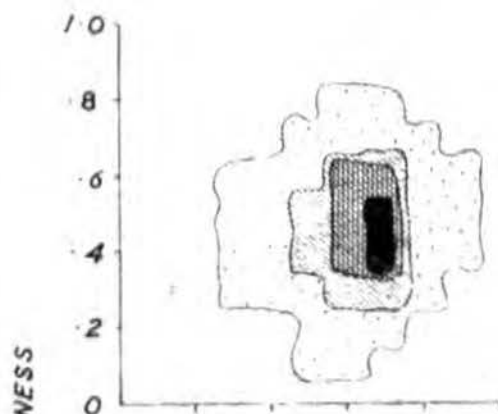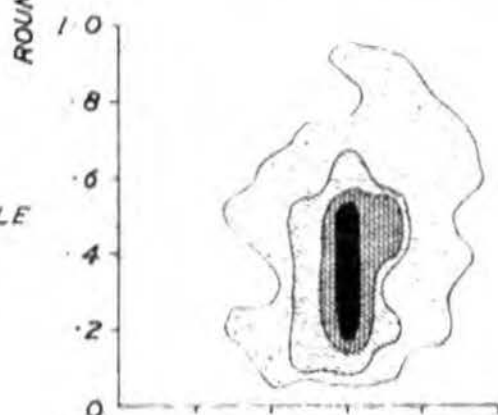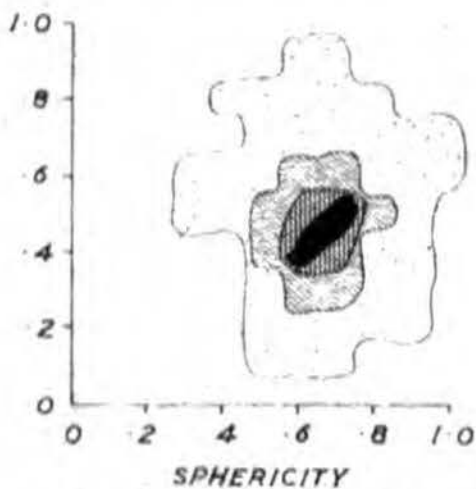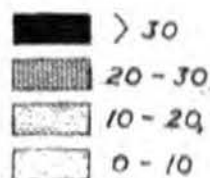
ROUNDNESS AND FREQUENCY

Figure 9
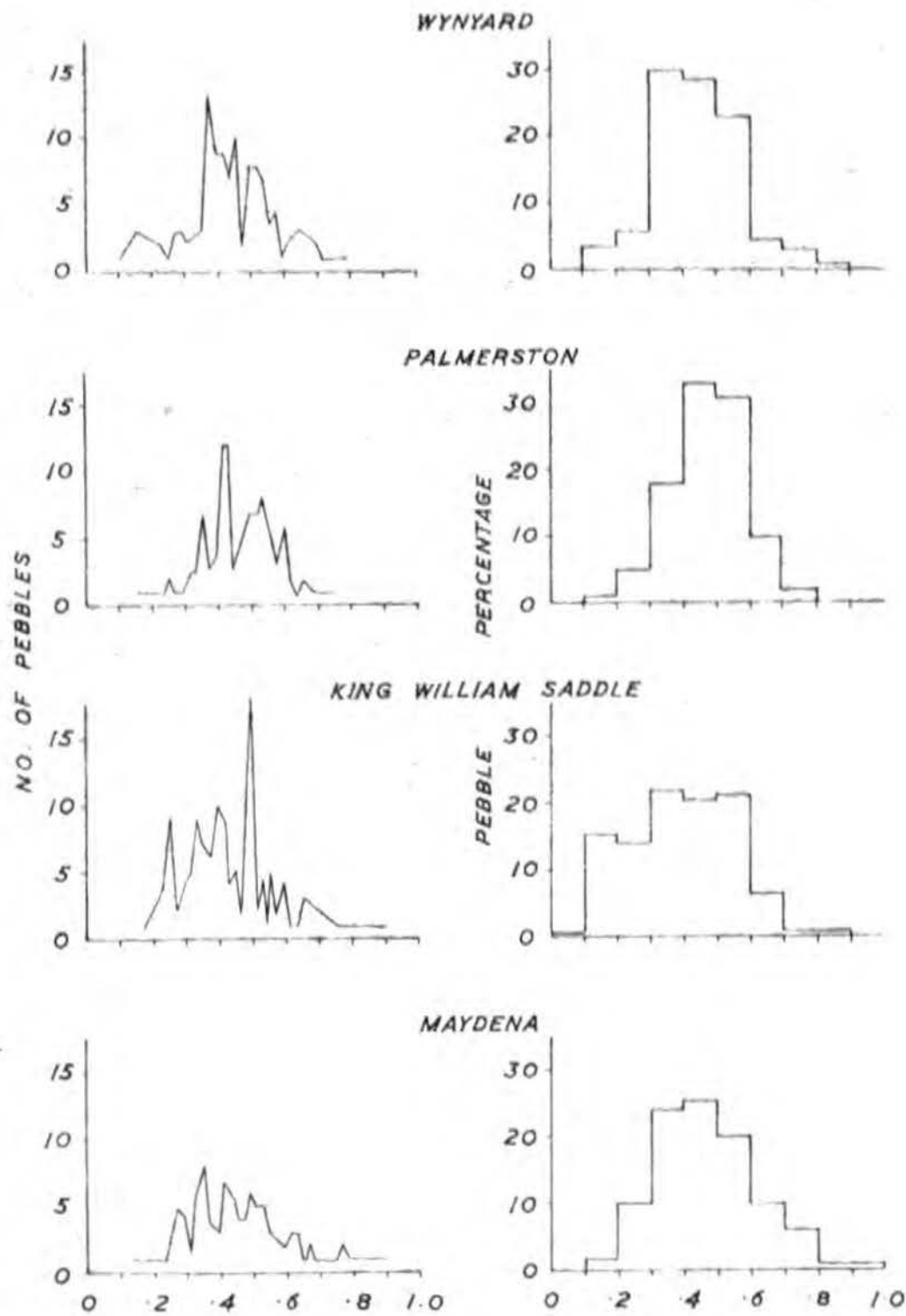
GLACIAL MARINE SEDIMENTS - THEIR ENVIRONMENT

AND NOMENCLATURE

by

S. Warren Carey and Naseeruddin Ahmad.

INTRODUCTION

Selectivity of the Geological Record

The study of glacial sediments has grown out of the study of
modern glaciers and of the terrestrial deposits left by the
Pleistocene glaciation.   Until recently comparatively little
attention had been paid to glacial sedimentation under marine
conditions.

From the point of view of the palaeogeographer however, marine
glacial sediments may be more important, in that they are much more
likely to survive in the geological record.   To-day the most
obvious and widespread glacial sediments are the terrestrial deposits
moraines, tills, and outwash.   However on the geological time scale
all these deposits are likely to be almost completely destroyed.
But in areas where the glacial sediments are included in a con-
tinuing cycle of subsidence and sedimentation, the glacial record
may be preserved, perhaps for several geological periods.   Such
areas occur in the Arctic, Alaska, and off the Ross Shelf.   The
most permanent records will be those where the glacial sediments have
been enclosed in a currently active geosyncline, where strong folds
will, in due course, take the glacial formations deep into the crust,

where several cycles of peneplanation will be necessary to erase the
last vestiges of the record.

The remains of the "Eo-Cambrian" glaciation are largely of this
type - in folded geosynclinal piles, usually resting on other marine
strata, with rarely any sign of a glaciated bedrock pavement.   There
were of course contemporaneous moraines and tills, and outwash, but.
they were the first to be wiped out.   There were also, no doubt,
shelf deposits, but they too have largely disappeared.

The record of the late Palaeozoic glaciation is partly of the
geosynclinal type but there are still very many shelf deposits, often
unfolded, often associated with glaciated bedrock pavements, but
usually leading up into a following succession of marine or lacustrine
sediments which served as a protecting shield during the erosion cycle
of the Mesozoic and Tertiary.   Terrestrial moraines and tills,
depending for their preservation on pene-contemporaneous burial
through tectonic subsidence of the shelf on which they lay, are
relatively rare.

In spite of this difference in the state of the record in
Pleistocene, Palaeozoic and Precambrian glaciations, there is still
a strong tendency to look to terrestrial glaciation for analogies for
the ancient glacial sediments, rather than to the marine environments.
This straight-jacket shows up sharply in the nomenclature.   Rock
names are carried over from terrestrial environments.   There is a
dearth of names to designate marine glacial sediments in spite of the
frequency of their occurrence, and cumbersome circumlocutions
are often used (e.g. Miller, 1953, p. 26).

Scope of Present Contribution

This paper was inspired by difficulties which arose in iden-
tifying the conditions of deposition of the marine glacial sediments
the Permian System of Tasmania.   It seemed to us that an understandi
of these sediments, in addition to its contribution to palaeogeography
could prove to be a useful complement to glaciological study, since
current marine glacial sediments are not so accessible as are those
of former glaciations.   Hence such a study, having benefited from th
stimulating advances made by the glaciologists during the last decade,
might well return new suggestions along lines not conspicuously
revealed by the glaciological techniques.

The overall design of the investigation is (1) a theoretical
enquiry as to what kinds of sediments we might expect glaciers to
develop in different marine environments, comparing this where
possible with reported occurrences (this paper).   (2) An objective
study of the properties/Tasmanian Permian glacial sediments (by K.
Ahmad).   (3) An objective study of the properties of Quaternary
glacial marine sediments off Antarctica in collaboration with the
Australian National Antarctic Research Expedition.   (This is in
the prospectus stage only).   In addition one of us (Ahmad) has
examined several Tasmanian Pleistocene glaciated areas (Ahmad,
Green and Bartlett, 1957; Jennings and Ahmad, 1957; Baker and
Ahmad, 1957).

This paper therefore attempts to review the physical behaviour
of ice sheets in so far as such processes are relevant to the
associated marine sedimentation, to classify the environments

under which marine glacial sediments may be deposited, to determine criteria whereby ancient sediments belonging to these environments may be recognised, and to point to gaps in our nomenclature which need to be filled.

Conditions for occurrence of marine ice sheet

Any glacier may protrude to sea and become buoyant, provided its rate of alimentation is sufficient to maintain it against the processes of melting, spreading and frontal calving. Because the Ross Barrier is in a bight some 400 miles wide and the same order of length, it has been suggested that a somewhat restricted re-entrant of this type is a necessary condition for an ice barrier. The Gulf of Alaska is much more open yet a floating ice sheet developed there during the Pleistocene as shown by the glacial marine sediments dredged by Menard (1953, fig. 1) throughout the gulf to more than 500 miles from land. Pliocene or Pleistocene marine glacial beds have been folded up and exposed in Middleton Island (Miller, 1953). Other factors being equal, the wider the angle of the coast, the more rapidly the floating ice will spread, but we do not agree that a restricted bight is in any way an essential condition. A number of glaciers around the Antarctic coast protrude to sea and become buoyant The Dronning Maud Land coast which is broadly convex has a fjord-like ice sheet flowing out to sea forming a floating ice shelf for many hundreds of miles of the coast (Robin, 1954). Entrenched within this general ice sheet there is a fjord-filling glacier which drains part of the interior ice plateau of the continent (figs. 1 and 2).

The Filschner floating ice shelf has a front some 500 miles long.
The West Ice Shelf off the Leopold and Astrid Coast has a 200 mile
front and projects seawards far beyond the 100 fathom line along a
convex coast.   There are very many others (see the National Geographi
Society map of Antarctica of September, 1957).

Swithinbank (1955, p. 65) after describing the form of the
Dronning Maud Land shelves suggests that "it may be true to say that
an ice shelf, unless flanked by land or by inland ice sheets, will
never extend to sea beyond the outermost shoals which could ground it"
"Without these the ice shelves might well break loose and drift
bodily to sea".

For practical purposes this induction may well be true.
However even in the area studied by Swithinbank the top of the
grounding "shoals" may be in excess of 400 m. below present sea-level.
Hence even accepting Swithinbank's suggested limitation, floating
ice shelves could extend well beyond the limits of any continental
shelf.

The most important condition would seem to be an abundant flow
of terrestrial ice.   Without this contribution a substantial thick-
ness could never be built up, because the low conductivity of snow
insulates sea ice against atmospheric cold, and bottom melting
balances surface accretion after only a small thickness is attained.

The thickness of the floating ice sheet depends on a number of
variables; the rate of outflow of ice from the feeding terrestrial
glaciers, the mean temperature of this terrestrial ice, the rate of

snow accretion on the surface, the rate of basal melting,, and the
rate of spreading, which is strongly influenced by any lateral
restrictions of the coast.   Robin (1955) has suggested a stable
equilibrium thickness for any given conditions.

Environmental zones

We may recognise the following environments of sedimentation in
relation to a glacier which extends to sea (see fig. 1).

A. Terrestrial - where the base of the glacier is above sea level,

B. Grounded shelf - where the base of the glacier is below sea-level,
but not floating,

C. Floating shelf - where the glacier is floating,

D. Inner iceberg zone - from the ice barrier to the limit of winter
pack ice,

E. Outer iceberg zone - beyond the limit of pack ice but within limit
of icebergs.

We propose to seek criteria for recognising the sediments of each
of these zones.   Before attempting this however, it is necessary to
make another distinction which we think to be of profound importance
in every aspect of glacial sedimentation - namely wet-base and dry-
base glaciers.

Wet and Dry Glaciers

Glaciologists have long discriminated between "Arctic" glaciers
and "temperate" glaciers.   A wet-base glacier is defined as one whose
base is at melting temperature, with basal meltwater.   A dry-base

glacier is defined as one whose base is below melting temperature.
A wet-base glacier may of course be a dry-base glacier further inland.
Seismic measurements by Poulter on the Ross Shelf, and by Robin on
the Dronning Maud Shelf have given conflicting pictures of the
regime of floating shelves and their thickness changes across the
line of grounding.   There has even been an implied suggestion that
one or other had misinterpreted their results.   However the respec-
tive characters as reported by the two investigations fit precisely
what our analysis leads us to expect in wet- and dry-base glaciers
respectively.   We therefore suggest that the Pencksokka Glacier of
Dronning Maud Land and its floating ice shelf is an example of a dry-
base glacier, and the Ross Barrier with its feeding glaciers is an
example of a wet-base glacier, at least in its seaward end (figs. 1-4)
These figures are based on seismic data from Robin (1954) and Poulter
(1947) respectively.   Superficially the Ross Shelf differs from the
Dronning Maud Shelf in that there is little or no change of surface
elevation where the floating shelf touches down and becomes grounded,
whereas the latter has a sharp break of surface slope at this line,
and is found to increase in thickness by four or five times.   Its
floating shelf is very much wider.   In place of the sudden increase
in ice thickness, the Ross glacier has a sudden increase of till
thickness under it at the point of grounding.   We suggest that these
contrasts are merely symptoms of very different behaviour in respect
to a wide range of properties.

In the ensuing discussion we will attempt to show that dry-base
glaciers produce marine sediments with many dropped erratics, whereas

wet-base glaciers produce well bedded marine sediments with much
rarer erratics, but often containing interbedded layers from a few
inches to several feet thick of till-like material, without drag on
the underlying sediments; that the dry-base glaciers are associated
with abundant thick-shelled fossils and limestone containing many
erratics, whereas the wet-base glacier produced more silty, non-
calcareous sediments with thin-shelled fossils; that the dry-glacier
may produce glendonites and large calcareous concretions and saline
specialised faunas; that the wet glacier produces below sea-level
thick non-fossiliferous tills, which are rather abruptly replaced
laterally by very different stratified rocks; and that in a partic-
ular facies the wet-base glacier results in strongly dragged and
rolled structures; that dry-based glaciers are more likely than the
wet to produce favourable environments for petroleum accumulation.
Moreover we have seen phenomena matching all these characteristics in
the marine glacial sedimentation of the Permian System of Tasmania.

Let us then examine systematically the physical behaviour of
these two glacier types and their geological consequences.

HEAT FLOW THROUGH GLACIERS

The variables in the thermal regions of a dry glacier are shown
qualitatively in fig. 5.

The surface temperature fluctuates between the extremes AA',
but melting prevents the glacier surface ever rising above 0°C. The
temperature fluctuation at the top of the ice is therefore between A'
and 0°C (J). The temperatures are ephemeral and no equilibrium

gradient can be established with them.   The range of seasonal
temperature variations must therefore become rapidly less with depth
owing to the very low conductivity of loose snow, and the annual
fluctuation vanishes at a shallow depth (B).   This has been con-
firmed by field measurements e.g. Wade (1945) found no temperature
variation below 45 feet.   The firn is a very poor conductor indeed,
and a good deal of the heat flow in this zone is convective through
air circulation.   Thus Poulter (1947, p. 372) reports feeling an air
draught two feet above the end of a bore ending 16 feet below the sno
surface.   The gradient at any instant may be anywhere in the figure
JBA'.   Below B the temperature gradient shows no seasonal variation.

A static ice sheet in thermal equilibrium would have to con-
duct to the surface the normal terrestrial heat flux, and at the
same time its surface would be subjected to the annual range of
seasonal temperature fluctuation.   The slope ED is determined by
the conductivity of the rocks and the regional heat flux.   The
conductivity of ice is of the same order as that of rocks, but
the conductivity of compact snow is an order of magnitude less.
Hence, DC the lower part of gradient in the ice has much the same
slope as in the underlying bedrock (a degree or so every 100 feet)
but changes rapidly to about a degree every 10 feet as the firn
becomes less compact.

There are however two further variables, the flow of the
glacier has a convective effect equivalent to bringing the very cold
surface temperatures nearer to the basement thus depressing the

basal temperature below D, and so on for all points between D and B.
Hence the gradient curve would be displaced to the position FB.
The shape of this curve is a function of the rate of glacier flow,
the rate of change of this rate with depth and the rate of change
of conductivity of the ice with depth.

The bedrock segment of the gradient EF is not in equilibrium.
It tends towards a uniform gradient (for uniform rocks) but the
longer the glaciation continues the further F moves to the right
towards lower temperatures (but always of course above the tem-
perature of B).   Heat is flowing upward through the rocks near F
at a faster rate than near E, and this leads to a progressive down-
ward migration of the point L as the chilling effect of the glacier
extends more deeply into the bedrock.   Many thousands of years are
required to approach an equilibrium gradient.

There is a complicating feedback relation between rate of flow
of glacier and the thermal gradient.   The flow rate of ice is
sharply affected by temperature,,increasing very rapidly as the
temperature rises towards melting point, other conditions being
equal.   Hence a very cold slow glacier receives more terrestrial
heat than a warmer faster one because it receives the heat flux for
a longer time.   A portion of the gradient below B may become very
steep at nearly constant temperature, receiving virtually no summer
heat by conduction from the surface, and virtually no heat from the
earth's flux owing to its convective removal by ice flow further down.

The ice also receives heat from conversion of mechanical heat by viscous flow and bottom friction slippage. Most of the potential energy lost by the ice commencing at say 3000 metres altitude and ending at sea-level is converted to heat. This is enough to raise the ice temperature by about 12^{o}C but on its own could achieve little melting. For comparison, the heat received by terrestrial heat flux during the slow journey might be an order of magnitude greater.

The three contributors of heat - friction, terrestrial heat flux and heat exchange from basement to ice with depression of temperature of the former, all lead to a progressive rise in temperature towards the base. Further, the same three contributors mean that the temperature at the base rises progressively down-flow. Hence the thicker the glacier and the greater the distance down-flow, the more the probability of a melting zone appearing at the base. This introduces a further complicating factor of meltwater.

In the zone of surface seasonal fluctuation (ABA', fig. 5), meltwater will always be present where the surface temperature is above 0^{o}C. This conducts heat as latent heat downwards giving a vertical gradient down to the limit of melting (JH, fig. 6). Likewise the temperature at the base cannot rise above zero, and if the initial rock temperature (N) before onset of glaciation was above zero, the gradient MN would be rapidly offset to K at 0^{o}C. (There is of course a small depression of the melting point by the ice load, amounting to about 1^{o}C for 5000 feet, which is neglected in this discussion.) Assuming equal conductivity of rock and ice,

the temperature of P would be below that of the prolongation of MN because of the removal of heat by glacier flow. The constant temperature segments KP and JH do not mean absence of heat flow but that the heat flow is absorbed in melting ice or freezing water until ice becomes dry at H and P. Interstitial meltwater is therefore confined to the zones of the glacier above H and below P, respectively.

Glaciers normally have a surface meltwater zone at some time of the year. The base of many Arctic and Antarctic glaciers are however well below freezing point so that there is no basal melt-water zone. In some "temperate" glaciers B and P meet and J, P, B, K are all at $0^{\circ}C$ so that there is interstitial meltwater through-out. Such glaciers do not differ hydraulically from sedimentary rocks saturated with groundwater, and interstitial flow and the seepage pressures behave as for groundwater flow.

VERTICAL SUPPORT OF THE GLACIER

While the sole of the glacier is above sea level, the whole
weight of the glacier is borne by the underlying rocks and sediments.
This is also true for dry glaciers when the sole of the glacier is
below sea level, but not yet buoyant. On the other hand in a wet
glacier below sea level the sub-glacial sediments are saturated with
groundwater which necessarily exerts a powerful uplift pressure on the
base of the glacier, and partly supports its weight, just as do uplift
pressures on the base of a concrete gravity dam. A longitudinal
section along the glacier changes little in respect to load distri-
bution from year to year, hence an equilibrium pressure gradient is
attained in the interstitial meltwater and groundwater below the
glacier. The equilibrium gradient relates the load of the ice, the
head of water on the submarine seepage outlet, the permeability of the
saturated beds and the rate of flow. Because the landward saturated
sediments are confined under the load of the ice, a strong hydraulic
gradient exists in a seaward direction. Hence the uplift pressure
in the grounded shelf zone is always greater than the hydraulic head
at an equivalent depth below sea level, and the proportion of the
glacier's weight supported by the groundwater is therefore greater
than would be the case with an equivalent column of ice freely sub-
merged to the same depth below sea level. The subglacial sediments
therefore do not bear the full load of the ice, and the ice load on
the sediments diminishes progressively to zero at the point of
flotation. This position is not materially altered if meltwater
stream channels develop in the subglacial meltwater zone, with free

flow to the sea. If there is no flow in such channels, the uplift
pressure at a given depth below sea level is the same as in the open
sea at that depth. If however there is a seaward flow in the melt-
water channel, then the uplift pressure at any point is greater by an
amount equal to the product of the rate of flow times the total
viscous resistance to flow between that point and the open sea. Low
permeability in the subglacial sediments only alters the position to
the extent that a longer time is required for the uplift pressure to
build up to its full head. Load variation at the base depends not
on the rate of flow of the glacier but on the rate at which its thick-
ness at a given point changes. This is not normally rapid, hence
the uplift pressure at the base will normally be close to equilibrium
even where the subglacial sediments have relatively low permeability.

A wet-base glacier thus has a gradual load transition which is
distributed all the way from where the sole is a little above sea
level, to where there is sufficient depth of water to float the
glacier. A dry glacier on the other hand suffers an abrupt change in
the nature of its support when it becomes buoyant. Inside this
transition its full weight is borne by stresses in the frozen sedi-
ments. There is no hydraulic contribution. Across the line
of transition its load is borne entirely by flotation. The water
below the floating section is subject to tidal rise and fall. The
abrupt transition means that there must be a hinge zone at the trans-
ition to buoyancy, even though this be hundreds of miles from the ice
barrier. Any lifting of the margin of the grounded zone along the
main shear plane during an exceptionally high tide would result

immediately in a wedge of sea ice frozen in the gap, so that the
falling tide would leave the whole of the buoyant glacier cantilevered
from the edge of the wedge.   Since the semi-diurnal tidal cycle is
too rapid for the ice to flow to equilibrium, folds and fractures
appear at the surface along the hinge zone.   This phenomenon is
well known in Antarctica (e.g. Robin, 1954, p. 199).

The inner edge of the water under a dry glacier should normally
terminate abruptly at an ice wedge connecting the sole of the glacier
to the sea floor.   This abrupt contact is greatly increased by the
rapid spreading of the floating segment due to the absence of basal
friction.

There are two kinds of flow in glaciers - down grade flow, and
spreading.   The former is the principal process when the ice is on a
sloping floor.   The latter is the principal process when the floor
is flat, or dished, or when the glacier is floating.   Both processes
may occur together and both are limited by bottom friction.   The
mechanics of downgrade flow is now well understood and good agreement
is achieved between theoretical prediction and observed surface
gradients (e.g. Nye, Swithinbank, Weetman et al).   The total shear
involved in this flow cannot exceed the friction on the base, because
as soon as this value is reached slippage occurs.   Similarly the
basal shear involved in spreading cannot exceed the friction on the
base without slippage occurring.   Where there is no friction on the
base (e.g. with floating shelf) the rate of spreading is very rapid.
Here the driving energy comes from the fact that the thinner the sheet
the nearer its centre of gravity approaches that of the displaced water

Now let us compare the behaviour of wet and dry base glaciers in this respect. In the case of the dry glacier the full load of the glacier bears on the bottom until the actual line of buoyancy, and the coefficient of friction is that between dry ice welded to dry cold rock. The friction (weight times coefficient of friction) is very high. Spreading is strongly inhibited. The cross section of ice needed to maintain a given flow is high - perhaps ten times that necessary to maintain flow in the buoyant part. Hence there is a sharp change of thickness at the buoyancy line and an abrupt rise of the surface above the level of the ice shelf to form ice hills (fig. 1A, 4B).

The wet glacier is very different. In the first place the presence of a meltwater phase at the base means the rate of flow there is greatly increased, partly because flow rates increase rapidly with rising temperature and partly because any crystalline substance flow very much more rapidly in the presence of its own liquid phase or its own saturated solution. This low viscosity zone at the base speeds up both down grade flow and spreading, so that a reduction of thickness occurs with the appearance of basal meltwater. Next as the base passes below sea level an increasing proportion of the weight is borne by the uplift pressure of the meltwater, so that the effective weight involved in friction steadily falls towards zero. The coefficient of friction between wet ice and wet till also is much lower than in the dry glacier. Hence instead of a sudden transition to spreading at the buoyancy line, the transition is spread right back perhaps for hundreds of miles. Hence there is no marked change of

surface elevation of the grounded shelf (figs. 1B, 4A).

These expectations are completely vindicated in the field.
Robin (1954) found the Dronning Maud shelf to be remarkably level for
hundreds of kilometres then/rise abruptly into ice mountains.   His
seismic reflection measurements revealed that the flat portions were
floating over deep waters, and that the ice mountains were grounded
ice still in relatively deep water (fig. 1A, 4B).   The floating
section originally about 1000m. thick and becoming buoyant in water
800m. deep thins rapidly by spreading to less than 200m. leaving 760m.
of open water beneath it, while the grounded portion still at 1000m.
thickness rests firmly on an only slightly shallower bottom.   The top
surface of the grounded portion is at more than 200m. above sea level,
whereas the top surface of the floating section is at 30m.   An
equilibrium gradient between these two sections is determined by the
glacial flow rate (elgl see Nye, 1951, 1952; Robin, 1953).   This may
result in still more of the sheet becoming buoyant until a steady
gradient of equilibrium is reached, the flow of which agrees with
theoretical predictions.

Where the ice sheet has over-deepened valleys under the more
active flow channels, it will be here that the ice will first become
buoyant.   The rapid spreading of the buoyant zone and the consequent
fall in surface level to that of the ice shelf while the ice on the
interfluves still maintains nearly the original level, leaves the
latter standing as ice ranges standing well above the re-entrant
embayments of the shelf ice, even though the rock interfluves are
far below sea level (see section 5 of figure 2).   This has been

demonstrated by seismic work (Robin, 1953) and is beautifully revealed
by the glaciological map of Western Dronning Maud Land prepared by
Swithinbank (fig. 8 of Robin, 1954).

By contrast Poulter (1947) found the surface of the Ross Shelf
to be remarkably uniform for hundreds of kilometres even though seismic
reflections proved that the shelf is grounded to within 13 km. of the
barrier edge (fig. 4A).   Also his seismic measurements show that the
thickness changes little.   The changes which occur are due to two
other causes - (1) there is a steady increase in thickness from 500 ft.
23 miles from the edge to 735 ft. 3 miles from the edge, due to the
increased precipiatation gradient towards the edge;   (2) progressive
thinning from 755 feet to 525 feet in the outermost three miles due to
rapidly increasing bottom melting by inflowing sea water under the
shelf, and a similar thinning by bottom melting 13 miles in, where a
3 knot "warm" current under the shelf follows the outer edge of the
subglacial sediments.

The only surface indication of the buoyancy line in a wet glacier
is the tidal cracking.   The semi-diurnal tidal movements are too
rapid a cycle for the subglacial seepage pressures to remain in
equilibrium with them.   Hence long straight fractures develop along
the buoyancy line, which may eventually lead to the calving of
tabular bergs many tens and even hundreds of miles in length (Poulter,
1947).

Shelf Tongues:

Even far out on a self-spreading floating shelf the flow pattern

may be dominated by the active trunks of the feeding glacier. For example Robin (1954, p. 198 and fig. 8) describes a tongue projecting from the free margin of the Dronning Maud Land ice shelf, which is apparently the terminal outflow of a main trunk glacier which has maintained its identity right across the shelf. For several hundred kilometres in either direction, the shore is a fjord coast with a mountain ice sheet extending to below sea level and continuing out over the continental shelf as a floating ice shelf (see map fig. 2). On the Greenwich meridian however there is the outlet of the Peneksokka Glacier flowing from the high continental ice sheet 1000 Km. inland. The floating ice shelf at the mouth of this fjord extends about 200 Km. seawards and has about the same width. In the line of flow of the trunk fjord glacier however the floating ice shelf protrudes a further 50 Km. over a width of 50 Kms. (fig. 3). Such phenomena may substantially affect the pattern of contemporaneous sedimentation.

A number of other tongues protruding many miles beyond the rest of the floating shelves are shown on the most recent map of Antarctica ((National Geographic Society, 1957).

Subglacial Relief:

Reports of the Maudheim expedition have emphasised the extreme relief of the subglacial topography in Antarctica (e.g. Robin, 1953, fig. 1). However Robin's published sections were drawn simply along the field traverse route, which crossed and recrossed the main lines of glacier flow. Critical examination of the data makes it clear tha

the sharp relief is in directions transverse to the glacier flow.

The contrast of longitudinal and transverse bottom profiles shown in figure 1A and figure 2, in the same area as Robin's section makes this relationship clear.

SEDIMENTATION BELOW GLACIERS

Terrestrial Zone:

The factors which cause deposition of sediment by a glacier are necessarily very different from those causing deposition by a stream. The principal cause of deposition in streams, reduction of velocity, has no effect in a glacier, which can continue to transport its full load no matter how slowly it moves. The second critical threshold to water transport, size of body, is also irrelevant in a glacier, which can carry huge blocks as readily as clay - hence the characteristic lack of size sorting in till. Conversely, bottom melting, the most significant factor in glacial deposition, has no counterpart in streams. Because the principles of fluvial and glacial deposition are dramatically different, geologists need a conscious effort to free their minds of fluvial models when contemplating glacial deposition.

There are two critical surfaces in a terrestrial glacier, the upper limit of meltwater (fig. 6), and the lower limit of interstitial ice, which may be taken as the boundary between glacier and deposited sediment (fig. 8). Traced downflow, both surfaces should rise with respect to the basement since the extent of offset of K with respect to N (fig. 6) should diminish in that direction. Traced upflow first one then the other surface (if the glacier is cold enough) should reach the basement. These surfaces are almost parallel to the flow lines but cross them at very gentle angles. (The angle depends on the balance of heat convection and heat conduction in the ice, hence the faster the glacier the flatter the angle).

Since the till is deposited under the full load of the glacier it is born as a consolidated sediment, which is another point of sharp contrast with water-laid sediments which are "normally loaded" (in the soil mechanics sense). Since the till develops in situ by the replacement of interstitial ice by meltwater, through a transitional zone of considerable width, the englacial flow fabric is preserved. (Compare Harisson, 1957, p. 296).

The upstream limit of melting (locus of Y, fig. 8) would not be regular. Glacier ice is far from isotropic. It consists of many ice-streams convergent from different nevée fields at different elevations, conveyed thence by feeder routes of different thickness an different rates of flow. Different streams may converge with quite different basal temperatures. These differences gradually vanish, but only by conduction, which is slow. Again the detritus carried by the different streams differs widely in amount, composition and grainsize. This may greatly affect the rate of melting, because in the zone of melting the ascending terrestrial heat is absorbed only as latent heat (gradient KP, fig. 6), hence the rate of upward migration of the surface of complete melting is directly proportional to the ice-sediment ratio. The sediment variation of the different streams also affects the glacial flow parameters both in the sub-zero and water-saturated states. Again the streams with the higher sediment content have the slowest velocity and hence receive more terrestrial heat per mile of flow and melt first. In addition there are irregularities in the basement. Quartzites are better heat conductor

than basalts.   Basement projections cause upward and lateral
deflection of the flow lines with reduced rate of flow in their lee,
and increased rate of flow elsewhere.

All these variations cause the upper limit of melting to be a
wavy surface very dentate in transverse section although smooth in
longitudinal section in the direction of flow, and the line in which
this surface meets the bedrock basement is extremely dentate with
very long upstream projections and outliers.

Long before permanent sedimentation becomes general it might
well occur in the lee of basement projections, forming lee drumlins
(rock drumlins or "crag and tail").   In the threshold region where
sedimentation is about to become general melting would/occur on the
                                                             first
most favourable/lines.   Deposition here would inevitably cause
           flow
deflection of the flow lines upwards and laterally, so that a drumlin
is born.   Since the favourable ice stream would continue to come to
this point, the drumlin would grow in size, with successive additions
in the streamline shape.   As the melting became more general, con-
tinuous ground moraine would develop, but its upper surface would
still be drumlinoid, such as is beautifully illustrated in the
U.S.A.A.F. air photograph   of the Carp Lake area of British
Columbia (published   as fig. 12-12, p. 196 of Longwell
and Flint, 1955).   The deflection of the flow lines by the drumlin
deposit increases as it grows.   This crowds the flow lines at the
front which implies greater rate of shear, and widens them behind
which implies less shear.   Material is therefore dragged from the

upstream end and deposited at the rear slope. This is accentuated by the fact that there is probably bottom slippage between glacier and basement in this zone.

One of us (Carey, 1953, p. 76-78) has drawn a dimensional analogy between dunes and drumlins and of lee shadow sand drifts with lee drumlins. In each case we are dealing with a fluid (wind and ice respectively) flowing over an inert basement on which there is loose material being shaped and transported by the moving fluid. The viscosity of ice is some 10^{18} times that of air, but wind velocities are 10^9 times as great as glacier velocities. A more viscous wind would need to move more slowly to produce a similar drag and if the viscosity of the wind were progressively increased until it reached the viscosity of a glacier, the velocity required to mould the "dune" into comparable shapes might be of the order of that of glacier flow.

Although the most important cause of deposition is basal melting, the question arises as to whether deposition can occur upstream from where wet ice is in contact with the rock floor (Y, fig. 8), and again whether disposition can occur above where PP' rests on the basement, i.e. when dry sub-zero ice rests on the basement. Let us first be clear on what we mean by "deposition" under these circumstances. Can subglacial sediments be said to be deposited when its voids are occupied by ice continuous with the ice of the glacier above? The whole of the shear involved in the flow of the ice above must be transmitted through this material, and it must surely flow even if comparatively slowly. If so then deposition begins at the line of basal melting

(Y, fig. 5).   The most that could occur would be stagnation of rock loaded ice to await the ultimate melting of this part of the glacier.

Let us now examine the question of bottom slippage of the glacier in the different cases of a wet glacier on its already deposited till, a wet glacier on bedrock, and a dry glacier on bedrock.   In all three cases the total shear involved in the flow of any column cannot exceed the cohesion between the glacier and its bedrock, for if it reaches this level slippage occurs.   In a dry glacier the cohesion between dry ice and sub-zero rock is high.   However the shear needed to cause flow in a sub-zero glacier is also very high.   Hence the thickness of ice must accumulate until the total flow through the cross section balances the total precipitation.   Other things being equal the thicker the ice the greater the shear at the base, but as this cannot exceed the cohesion, the latter determines the maximum permissible thickness.   If the precipitation rate is such that a greater thickness is required to remove it by flow alone then slippage must occur at the base.

An equivalent wet glacier will flow at a very much faster rate under the same load, therefore a very much thinner section will transmit the same precipitation with a correspondingly lower shear load on the base.   However the shear cohesion of wet ice on wet rock is also very much lower, so again slippage could occur under appropriate conditions.   In the case of the wet glacier on its own till, again the flow rate in the ice will be high and the corresponding cross section necessary to transmit a given flow will be low.   However we

have to consider the relative rates of shear under given loads in the wet ice, in the wet till, and the limiting friction between the wet ice and the top of the wet till. It seems that in all cases the deposited till will be much more resistant to shear flow than the wet ice. The lower part of the ice contains the same constituents as the till but they are separated by ice wet with its own liquid phase. The ice has mechanisms of flow not possessed by the till for example crystal glide plane slippage, two phase deformation, energy transfer from crystal to crystal. The till on the other hand has a wide range of grainsize coupled with load precompression so that it will contain little interstitial water and should have relatively high shear strength. Although it contains "clay" grade material in the sense of grainsize it contains little clay in the sense of phyllosilicate minerals. There should therefore be little if any deformation of the till as the ice flows over it. The most rapid rate of shear might be in the lower part of the ice, and there might also be slippage between the ice and deposited till perhaps rolling some of the surface layer of the till. The till should preserve the flow fabric of the basal portion of the glacier, that is with long axes of pebbles in the transverse direction or parallel direction or both (see Glen, Donner and West, 1957).

The empirical geological evidence seems to indicate definitely that slippage at the base of the glacier is the general rule. Where-ever glaciated surfaces are revealed either from the Pleistocene or Palaeozoic glaciations, striations, grooving, roches moutonnées, and plucking of the basement seems to be universal. From this evidence

we would judge that both wet and dry glaciers normally slide on their floors as well as flow within their mass. The empirical evidence however is more open on the question of wet glaciers sliding on their own tills.

TRANSITION TO BUOYANT STATE

A glacier with interstitial meltwater at its base resting on water-saturated sediment suffers no abrupt changes during the transition to buoyancy. There is no change of basal temperature, there is a gradual change of vertical support, and there is a transition from distributed flow throughout the ice to nil drag on the base of the glacier.

By contrast a dry base glacier suffers abrupt changes in the heat flow, in the nature of support from solid to hydraulic, and in the conditions of horizontal shear. Each of these changes has marked effects both on the behaviour of the glacier and the nature of the sediments deposited by it.

When a dry base glacier, whose temperature is below the freezing temperature of sea water (about $-1.9^{\circ}C$), flows out to sea and becomes buoyant, sea water rapidly freezes and welds on to the bottom of the glacier and forms an insulating layer. The temperature gradient now has the form shown in fig. 7. The temperature at the base of the new ice is at freezing point. There is a steep gradient to S where the original gradient/inherited from the non-buoyant stages. Heat flows rapidly across the steep gradient RS so that S moves up towards B and eventually after a long time equilibrium is reached in a form

like RTB, the curvature of which is due to the decreasing conductivity
of the upper levels of the ice.   The heat required for this warming
of the glacier implied by the change from gradient RSB to RTB is
derived from the underlying sea water with progressive freezing of
more sea water which is added to the bottom of the glacier.   Before
this equilibrium curve is reached, melting will have commenced at the
base (since the sea water is normally above melting temperature).
Even a sub-zero glacier which adds sea ice to its base at its land-
ward end begins melting from the base well inside the terminal barrier.
In such a sub-zero glacier, the underlying rock floor is as we have
seen also in a frozen permafrost state (fig. 5) and will freeze sea
water if brought in contact with it.

Sea water may be brought in contact with new surfaces of sub-
zero ice by one of three processes - the outward movement of the
glacier, tidal movements, and on a much larger scale by landward
extensions of the buoyant area, such as must occur during a time of
climatic amelioration because of reduced alimentation, thinning
through surface melting, and eustatic rise of sea level.

During the retreat stages of a large marine ice sheet, when many
hundreds of square miles of previously grounded ice becomes buoyant,
considerable volumes of dense brines must be produced, because, not
only are new sub-zero surfaces of glacier bottom and bedrock bottom
brought into contact with sea water, but the buoyant ice sheet spreads
very rapidly to a fraction of its thickness, bringing very large areas
of sub-zero ice, much colder than the original base, very close to the
sea water.   Much sea ice must be frozen on to the bottom forming an

insulating blanket or filling joint cracks.   The resultant cold
brines would sink to the bottom then form outward gravity currents
out to the deep ocean floor beyond the shelf and be lost.   If however
there are barred basins either under the floating ice barrier or
beyond the ice barrier, the brines would flow into such depressions
and lie there perhaps to crystallize slowly with rising temperature.
This will be discussed further later.

SEDIMENTS PRODUCED BY SUBGLACIAL MELTWATER

Subglacial meltwater sediments are conspicuous features of
terrestrial glaciation and as such have been studied and described at
length, but they are not directly part of this study.   However
subglacial meltwater drainage and sedimentation may produce sediments
in association with submarine products which if not properly under-
stood and interpreted could lead to a quite erroneous concept of the
climatic sequence recorded by the sediments.   It is therefore
necessary to examine critically the terrestrial subglacial drainage
processes and sedimentation, particularly the long sinuous ridges
called eskers (in the sense used by Flint, 1947) or osar (in the
sense used by Charlesworth, 1957, not eskers of Charlesworth).   To
avoid confusion we will use the term os, plural osar, following
Charlesworth.

The origin of osar has produced long controversy which is
summarised comprehensively by Charlesworth (loc. cit.) with extensive
bibliography.   This will not be repeated here.   Suffice it to say
that for the following reasons we are satisfied that osar (or perhaps

only some of them, those being the ones which will concern the rest
of the discussion) are formed in sub-glacial streams flowing under the
ice with or without connection with surface meltwater of the glacier.

No adequate explanation seems to have been published concerning
the mechanism of development of open subglacial channels.   Mere
opening of a crevasse right to the sole of the glacier cannot produce
such a channel.   If the glacier is sub-zero in part, the crevasse
will freeze.   If it is water saturated throughout or at least not
sub-zero, the water head at the bottom of the crevasse, even when
filled to overflowing, is little more than the hydraulic head pre-
viously existing in the meltwater below the glacier, and may even be
less because artesian pressures may exist in sub-glacial meltwater.
Sub-glacial channels can only develop from the seepage outlet of the
meltwater at the terminus of the glacier.   The subglacial till and
the meltwater-saturated part of the glacier are simply water saturated
rocks and behave so hydraulically.   The production by seepage of open
channels through water saturated sediment well below the surface is a
process well known in soil mechanics under the name of "piping" or
"underground erosion" (see for example, Terzaghi and Peck, 1948,
pp. $^{502-12}$) and has led to the failure of many earth dams.   We suggest
that os channels are merely examples of this process.   Such channels
cannot develop in a sub-zero glacier or in permafrost sediments.
They commence at the seepage outlet of the meltwater, under the
terminus of the glacier on land, or where the meltwater seepage
enters the sea below the end of a grounded glacier, or where a
floating glacier becomes buoyant.   The seepage pressure of the melt-

water at the outlet is sufficient to wash out the finest particles, thus enlarging the water-bearing spaces and hence increasing the rate of flow.   Silt particles can now be washed out, then, as the process develops, sand and gravel, so that an open channel appears. This takes the steep hydraulic gradient further back into the water-bearing material, so that the process of underground erosion works continuously back further from the terminus of the glacier.   As the groundwater bears the full static load of the overlying ice, there is always an extremely steep hydraulic gradient at the head of the piping channel.   In fact the thicker the glacier, above them the more rapidly may the channels work back.   Underground channels may therefore develop as far as interstitial water is present in the subglacial till irrespective of any contribution of meltwater from the surface.   They may still develop at the base even though a zone of sub-zero ice intervenes between the basal meltwater and any surface meltwater. Osar are known to lengths of as much as 150 miles.   The process will be stopped where the freezing isotherm passes below the rock floor, or by working back to the plucked crag of a large roche moutonnée or where sediments are met which are of such a grading that they form a natural filter (see Terzaghi and Peck, loc. cit., p. 51), or which are too impermeable to seep at a sufficient rate.   Herein perhaps lies the reason why osar are not universally developed (see Charles-worth, 1957, p.423).   They require a suitable grainsize distribution in the till.   It would be useful to compare the grainsize histograms of tills associated with osar with those that are not.   The grain-size of the os sediment is not relevant since all the fines are

washed out and the coarse material regraded.

Osar develop at the top of the already deposited till. The till below is already compressed and though water saturated has low permeability. The basal part of the saturated glacier with a considerable proportion of its ice already melted is most vulnerable. The interstitial meltwater is milky with fine sediment. If it is near the head of an os this water will seep rapidly towards the os. As the weight of the glacier is bridged across the developing channel at least temporarily the seepage pressure removes more clay than sand as fast as it is freed by melting, and so the os extends. Water pressure in the tunnel is of course high enough to keep the tunnel full, - even contribute substantially to the support of the back. The pressure is sufficient to drive the water up grades "over hill and dale". Erosion into the underlying till occurs to some extent. The channels widen downstream as melting progresses and rise upwards with the melting. Tributaries may develop. For comparison one of us (S.W.C.), in the engineering investigation for foundations of a heavy structure 50' below the surface found an unsuspected complex dendritic drainage pattern of underground channels developed in silty sand below cemented conglomerate which had worked back from an original groundwater seepage outlet (or spring) below low water level.

It is abundantly clear that the terrestrial osar as we know them have suffered little deformation by glacier flow. Any channels oblique to glacier flow are rapidly closed by any flow. Moreover osar in their lower reaches are often associated with undisturbed

crevasse fillings.   Hence the osar finally left must be the product
mainly of the moribund melting stage of the glaciation.   But os-like
water channels are not incompatible with glacier movement, which is
necessarily slow, too slow to forbid piping processes.   The glacier
may flow and alter the form of the channels to streamline shapes.
But all voids of both the till and glacier are saturated, and water
does not compress much even under the total weight of the ice above.
Hence the water filled openings, though deformed, remain, and a hydrauli
gradient must build up quickly across any constricted links, sufficier
to widen them $r_a$pidly or to erode the underlying till sufficiently to
pass the water flow.   The ispatinows of Saskatchewan (Tyrrel, 1895,
and fig. A, plate XIII of Charlesworth, 1957), ridges of drumlin form
but of os material, and regarded by Charlesworth (p. 423) as related
to osar, probably belong here.

The conclusion that is important to the present study, is that
there seems no reason why osar should not form in the transition zone
below sea level, extending right out to where the glacier becomes
buoyant, which might be at a considerable water depth.   Osar so
formed could be preserved during ice retreat since any warming should
lead both to thinning of the glacier and rise of sea level, each of
which tends to lift the glacier off the bottom, leaving the os to be
covered by marine glacial sediments.

Such deposits should be looked for in ancient glacio-marine sedi-
ments.   Since the ice sediment contact in the zone is the locus of
unusually strong shear, the osar would probably be remoulded into the
ispatinow form.   Osar are unmistakable on recently glaciated surfaces

because of their form in plan. But they might easily pass unrecog-
nised in a Permian cliff section.

SEDIMENTS OF THE GROUNDED SHELF ZONE

In a dry glacier the grounded shelf zone does not differ in any
significant way from the terrestrial zone. The full load of dry ice
loaded with debris grinds the pavement. There can be little if any
sedimentation.

A wet glacier is however profoundly different. We have already
shown that the weight of the glacier borne by the rock or sediment
beneath diminishes progressively to zero, and that this is accompanied
by progressive thinning of the glacier as the frictional inhibition to
spreading declines. Further that the till is deposited at the base of
the glacier as the level of complete melting rises. The conditions
in this till are those of low vertical load but very great horizontal
shear. Hence a great deal of shear will occur at the top of the till
The lighter the load, the shallower is the effective depth of pene-
tration of this shear into the till and the greater its horizontal
component. The till which still bears part of the weight of the
glacier will be dragged or rolled forward and left where the water
depth is just sufficient to give complete buoyancy. Here a submarine
slope will develop at the angle of rest for the material. Under
steady conditions material will be constantly dragged or rolled for-
ward to the edge of this slope and pushed over the edge as a foreset
deposit. This slope will move forward steadily. Warming climate
will cause a thinning of the ice and rise of sea level so that the

buoyancy line will retreat landwards.   A new foreset slope will develop here and advances over the till already deposited.   Such conditions may produce meltwater sediments interbedded with till. Meltwater outwash silts deposited ahead of the foreset slope may develop strong drag, slump or roll structures.   One of us (Ahmad) has studied balled and rolled structures at Wynyard at the top of the till just at the transition into submarine drift sediments.   These might have developed as the ice sheet, previously partially resting on the till, gradually lifted off it as the buoyancy line moved back.

Climatic cooling would produce thicker ice, and more weight on the base and great dozing of the upper sediments which would be pushed forward to a new foreset slope front.

Unlike truly terrestrial tills, the tills and meltwater sediments deposited below sea level would suffer only partial or even no load compression.   Hence during increasing glaciation when a thickening ice sheet emerges to a fallen sea level the extra ice load would consolidate the sediments to lower void level.

Whether the ice is stationary, advancing, or retreating these processes have the effect of determining a base-level for till accumulation analogous to wave base for marine settling.   Till builds rapidly up to the base of the ice or is cut down to that level then stays at that level while its foreset slope is advanced.   The top of the till and the thickness of the ice sheet may remain uniform for hundreds of miles.   Such conditions were found by Poulter in the Ross sheet.   In this case the slope of the foreset slope was

surprisingly steep for marine conditions. Seismic reflections indicate that the foreset slope opposite Lindbergh Inlet drops from 800 feet to 1700 feet depth in about 2 miles, with a maximum slope approaching one in five.

Submarine slopes of such steepness must be very unstable, and as fresh loads of till are dozed over must from time to time inevitabl[y] produce mud slides down the slope. Studies of submarine slump phenomena have now established that such mud flows may roll out for many miles from their starting point. Since the slumping material is till the resulting deposit would be difficult to distinguish from till. It might occur in a thin or thick bed in stratified marine drift. They might be completely unstratified with abundant erratics in an unsorted matrix. The erratics would show no signs of dropping through water. The underlying sediments might be undisturbed. Such flows might well pick up marine fossils by eroding channels in the sediments over which they move. They might over-run and incorporate living shell beds, crinoid beds or the like. They would however lack the characteristic fabric of true till.

This is a very important conclusion, for such submarine flow "tills" could easily be mistaken for true tills leading to a very different interpretation of the palaeoclimatic sequence. Several "tillite" bands interbedded in well-bedded siltstones have been described from the Marine Permian of Tasmania. Thus Banks, Hale and Yaxley (1955, p. 224) have described on Woody Island, the D'Entrecasteaux Tillite, six inches thick, interbedded between stratified siltstones which contain occasional erratics. The

"tillite" is rich in erratics and rarely contains spiriferid shells, and the formations above and below it contain marine fossils. Interpreted as a tillite (sensu stricto) we can offer no feasible explanation for its occurrence as such a thin bed in undisturbed marine strata, and for the presence of fossils in it. Interpreted as a submarine flow from a till foreset deposit, its occurrence and characteristics are completely satisfactory and logical.

Quite as important as the till flows is the role of the melt-water which appears on the foreset slope at the embouchures of the os streams. The total flow of subglacial water laden with clay and silt must be very great and turbidity currents flowing down the foreset slope and thence far out to sea must surely be a dominating phenomenon Hence the sediments below the floating ice shelf and in the iceberg zones should be dominantly of this type. Such sediments should consist originally of unweathered finely ground rocks and mineral, of sand and clay grade. There are many processes whereby these sediment should develop laminar or rhythmic or thick bedding. Annual changes in the current circulation, or annual or longer range patterns of glacier thickening or thinning altering the bottom load in the sedi-ments, could easily produce bedding.

None of these till-flow and turbidity current phenomena should be expected with dry base glaciers. Nor does the contrast end there. For the pattern of iceberg transport is quite different. The wet base glacier suffers basal melting perhaps for hundreds of miles back from the buoyancy line. The lower sediment-rich layers of the glacie

drop their sediment by melting miles inland and if this sediment ever reaches the open sea it does so either by the dozing action of the glacier to the foreset slope, or by the meltwater transport of fines, or by turbidity currents. By the time the ice reaches the ice barrier where icebergs are calved, there may be little sediment left in the ice. Many observers have reported the lack of sediment in the Ross barrier bergs even where differential melting tips them to expose a complete section. This has been regarded as a puzzling feature, but in fact is what might have been expected. As a result the 'berg zone sediments off wet glaciers are siltstones and sand-stones and clay stones with only occasional dropped erratics. The dry base glacier by contrast carries all its sediment right to the buoyancy line. There it lifts and spreads rapidly, freezing sea water to its base because of its sub-zero temperature. Thus the erratics and clay alike are trapped in the floating ice shelf, which is for a comparable ice flow many times wider than the buoyant shelf zone of wet glaciers (fig. 1). The dry glacier, lacking meltwater turbidity currents and foreset till flows, produces instead mud-stones with abundant erratics all of which have clearly dropped through water. These two contrasting types of sediment association are both characteristic of different parts of our Permian succession.

Still another sharp contrast exists in respect to the salinity of the waters. The wet base glacier discharges large volumes of meltwater into the sea at its buoyancy line, quite comparable to the discharge of a major river. No sea water is frozen on to the glacier base when it becomes buoyant. The net effect therefore is to reduce

the salinity of the sea water in its velocity.

When a dry-base glacier becomes buoyant however both the ice and the exposed bedrock are below freezing temperature, and sea water is frozen on to both forming an insulating layer at freezing temperature.   However as there is lower temperature within heat continues to flow from the sea water with slow addition to the frozen layer.   Since the newly buoyant glacier rapidly spreads to a fraction of its original thickness (as has been previously explained) the cold heart of the glacier is continually brought very close to the sea water so that a large surface area is exposed to a steep freezing gradient.   These processes result in production of cold and brines in the sea, which sink because of their density and flow along the bottom as salinity currents, which might flow out and over the continental shelf and be lost, or on the other hand might be trapped in any barred basins which might exist.   This process would be especially characteristic of times of general glacial retreat when previously grounded dry glaciers find their thickness reduced and the depth of water increasing, with the result that very many square miles of sub-zero ice and bedrock are exposed to the sea in a short time.   During advancing glaciation the phenomenon would be much reduced but still present because wherever the dry glacier eventually becomes buoyant it quickly spreads exposing its cold heart to the sea.

There are at least two important consequences to this process, the precipitation of calcite and the precipitation of sulphates. Most sea water is nearly saturated in $CaCO_3$, but the solubility of

$CaCO_3$ increases as the temperature of the water drops towards zero.
The effect of the freezing of some sea water is to increase the titre
of $CaCO_3$ but it may not precipitate immediately from the very cold
water.   As it flows outward however into a warmer environment, $CaCO_3$
is inevitably precipitated, either directly or via organisms building
it into their shells.   Hence the first consequence is that dry base
glaciers have an off shore calcareous facies, or a calcareous facies
in their glacial retreat sequence whereas wet base glaciers produce
environments unsaturated with lime.   The limestones associated with
the dry base glaciers will contain abundant erratics since as we have
shown 'bergs from this kind of glacier still carry their erratics and
sediment when they become buoyant whereas the wet base do not.   In the
Tasmanian Permian the Darlington Limestone and the Berriedale Limestone
fit these environments closely and we shall show presently that the
fossil ecology also fits this environment.

The next possible consequence of the salinity currents associated
with dry base glaciers, is the precipitation of sulphates.   We offer
the suggestion that the large glendonites characteristic of certain
horizons of the Permian sediments of New South Wales and Tasmania
probably formed in this way.

Glendonites, calcite pseudomorphs after glauberite ($Na_2SO_4.CaSO_4$),
were first described by David, Taylor, Woolnough and Foxall, (1905)
from New South Wales, and have subsequently been reported from many
localities in the two States.   Raggatt (1937) has summarised the New
South Wales occurrences, and Banks and Hale (Ms.) have recently
collected the information on Tasmanian occurrences including several

new localities discovered by them.   Conditions common to these
many occurrences are:

(1) Host rock usually dark siltstone or sandstone
(2) Glacial erratics present in all cases
(3) Large calcareous concretions commonly present.   In some cases
     the concretions have grown around the glendonites.
(4) Pyrite nodules often present
(5) Marine fossils commonly present, usually sparse but rarely
     abundant.   In some cases the glauberite seems to have grown on
     fossils.
(6) Characteristic of particular horizons.

   These conditions all fit the environment we now suggest.   The
characteristic presence of dropped erratics implies floating icebergs
or a floating ice shelf.   The fossils imply marine conditions.   The
dark colour, and the pyrite suggest the enclosed basin environment
necessary to trap the salinity currents.   The calcareous concretions
suggest high salinity water with rising temperature because solu-
bility of $CaCO_3$ decreases with temperature.   The crystallisation of
sodium sulphate suggests abnormally high salinity.

   However there has been much discussion about the enigmatic
presence of glauberite instead of mirabilite ($Na_2SO_4.10H_2O$).   Direct
evaporation of sea water produces mirabilite, and no glauberite has
been crystallized below 25°C.   Freezing of sea water likewise forms
mirabilite.   This has been shown experimentally by Nelson and
Thompson (1954) and has been observed in nature by Debenham when

pools of sea water were frozen on the surface of the Antarctic ice. However the experimental evidence so far as it goes is in agreement with the process herein suggested by us, although further work is necessary to establish the process beyond doubt.

Thompson and Nelson (1956) have investigated the partial freezing of sea water to various salinities with withdrawal of the partially concentrated brine. They report that at moderate sub-zero temperatures the first precipitate is a very small quantity of calcium carbonate. This is followed on further freezing by a crop of mirabilite. The precipitate of the $CaCO_3$ occurs with rising temperature after the brine has been separated from the ice. The less the degree of refrigeration and the lower the salinity the longer the time delay. "Thus for brines with temperatures somewhat below the freezing point of the original sea water, a period of several days at room temperature was required before precipitation became noticeable" (loc. cit., p. 232). This time interval was decreased to several hours for refrigeration to $-22.9^{\circ}C$ and to less than an hour for refrigeration to $-36^{\circ}C$. Under the conditions of ice formation at the base of a sub-zero glacier where the sea water, having yielded some ice, and actually crystallizing or about to crystallize mirabilite, would sink to the bottom and flow down the ocean floor at very slowly rising temperature towards a maximum well below "room temperature", many days or weeks would be required for the $CaCO_3$ precipitation. The conditions then existing would be appropriate for the further reaction described by Thompson and Nelson (loc. cit., p. 237) in

which saturated Na_2SO_4 solutions react very slowly with calcium carbonate to form sodium carbonate and anhydrate, gypsum or glauberite according to the conditions. The slowness of this reaction is indicated by the observed rate of the rise of pH (due to sodium carbonate replacing sodium sulphate in solution) - 7.31 initially, 8.99 after several days, 9.31 after three months (loc. cit., footnote, p. 237).

Thus glauberite is a most unlikely product from a surface closed system freezing of sea-water - where mirabilite is formed. However glauberite is a likely product of slow crystallization in sea floor basins receiving brines derived from partial freezing at the base of a sub-zero glacier. It is only during fairly rapid retreat of a sea-going sub-zero ice sheet that these conditions are likely to be fulfilled. We suggest that glendonite horizons in dark marine sediments containing erratics might be so interpreted.

Finally we ask whether the pyritic nodules characteristic of these glendonite horizons are not themselves a consequence of the high sulphate content in a reducing environment.

FOSSILS ASSOCIATED WITH MARINE GLACIAL SEDIMENTS

No life of geological consequence can exist below the either dry or wet base glaciers before they become buoyant. Even after an ice sheet becomes buoyant however there are severe limiting factors such as light, salinity and oxygen. Complete and utter darkness must prevail at relatively short distances in from the ice barrier. An ice barrier only 70 feet high implies that its

base is some 500 feet below sea level.   Light is already severely
filtered at this depth, and transmission under the ice depends only
on light scattering.   Even if this were not so and full daylight
were projected underneath from the base of glacier, the absorption
of the water itself would bring on total darkness before a mile
had been penetrated.   There can be no question then that Stygian
darkness prevails in the water beneath buoyant glaciers of any
areal extent.   Since all life depends directly or indirectly on
photosynthesis for food, absence of light is an important restriction

It has already been established that animals exist on deep
ocean floors well below the limit of penetration of daylight.
However, here there are lighted seas overhead teeming with life
of various kinds.   An ultimate source of food is therefore
available.   Moreover turbidity currents bring sediments to the
ocean depths, containing no doubt some organic matter, which can
start another food chain for the bottom dwellers.

Beneath a floating ice sheet however there are no lighted
populated seas above to shower crumbs, and the sediments are of
glacial origin, which have been brought thither at the sole of
the glacier which is also sterile.

The only source of food would seem to be currents under the
ice.   Such currents have been shown to exist under the Ross Shelf
right in to the buoyancy line (Poulter, 1947), and it is possible
that favourable bottom configuration may result in the tidal flow
taking the form of a wide circuit under the ice.   There might

also be convective circulation - cold briny waters sinking and
flowing outwards along the bottom and warmer water flowing in just
under the ice.   There could therefore be a sparse food supply for
those who can do without light and can remain in the path of current
circulation.

There are however other restrictions.   Any barred basin
areas associated with dry base glaciers would be highly saline and
in any case they would have no food supply.   It is likely too
that oxygen would be deficient, since there would be no plants to
generate it, the minerals of the sediments would for the most part
be fresher and unoxidized and hence reducing, and any animal life
would reduce the oxygen concentration and hence check its own
multiplication.   Currents could however bring in some oxygen.

To sum up, whereas we may not be able to maintain that life
is impossible far beneath/floating ice sheet, it is true that there
are stringent limitations.   We would expect sediments of this
zone, although marine, to be largely if not entirely unfossili-
ferous.   Abundance of life however appears in the iceberg drift
zones.

Thick-shelled faunas and thin-shelled faunas:

Many visitors to the Tasmanian Permian sections have been
puzzled by the abundance of glacial erratics in limestone, full of
very large, thick-shelled fossils.   The common view has been that
thick-shelled fossils suggest tropics.   Thus Murphy (1928) and
Kirk (1928) state that organisms secreting large amounts of lime
are most abundant in the tropical seas and at a minimum in cold
waters.

What then is the explanation of the Tasmanian Permian where fossils with calcite shells, quite exceptionally thick by present day standards or in comparison with fossil faunas throughout time are associated with abundant glacial erratics? We suggest that this may be another consequence of the sub-zero, dry base glacier.

A wet base glacier results in dilution of the sea with large volumes of fresh water. A dry base glacier by contrast causes saturation or actual precipitation of lime along the bottom zones in the path of the outward salinity currents. This may affect shell thickness in three ways: (a) the physiology of some organisms may be such that they precipitate more $CaCO_3$ into their shells in waters in which these ions are most concentrated; (b) any organisms selectively adapted to high $CaCO_3$ may multiply at the expense of others which can dominate them in less saline waters; (c) the lime saturation of the bottom water inhibits re-solution of shells and produces a limestone.

We should not therefore be surprised to find erratic-bearing coquina limestones with abundant thick shells off the embouchure of a dry base glacier, during a time of glacial retreat. The Darlington and Berriedale Limestones of the Tasmanian Permian are examples. Here we find riotous multiplication of large _Eurydesma_ with shells approaching an inch in thickness, large pectens, gastropods and thick spirifers and martiniopsis. The bryozoans such as _Stenopora_ and the Fenestellidae show prodigious development with very calcareous colonies. It is perhaps significant that these are all benthonic forms, for the high salinity should be essentially confined to salinity currents flowing outwards along

the sea floor. This being so we should not be surprised if the calcareous facies changed rapidly into a siltstone facies along the boundary of the salinity current. Brill's mapping (1953) of the isopachs of the Berriedale Limestone suggests the embourchure of a dry glacier to the south-west of Hobart and a salinity flow north-eastward. These limestones show a conspicuous limestone to calcareous siltstone rhythm (Brill, loc. cit.). This is not unexpected if our interpretation is correct, because the amount of brine produced depends on the area of the ice sheet which becomes buoyant, with the result that the bottom salinity current would necessarily be very sensitive to any climatic pulse super-imposed on a general retreat.

There has also been a tendency to interpret the thick shells as indicative of shallow water and strong waves. But we question the validity of this. The characteristic environment off a sub-zero glacier implies water several hundred feet deep if not deeper. The rather rare association of high $CaCO_3$ content with fairly deep water may be the reason for the special development of large calcareous bryozoans characteristic of these beds.

The dry base glacier also produces an environment which might well be favourable for the petroleum source beds - abundant life, high salinity and a likelihood of barred basins. It may not be mere coincidence that the Berriedale Limestone has a marked foetidity and a bituminous odour.

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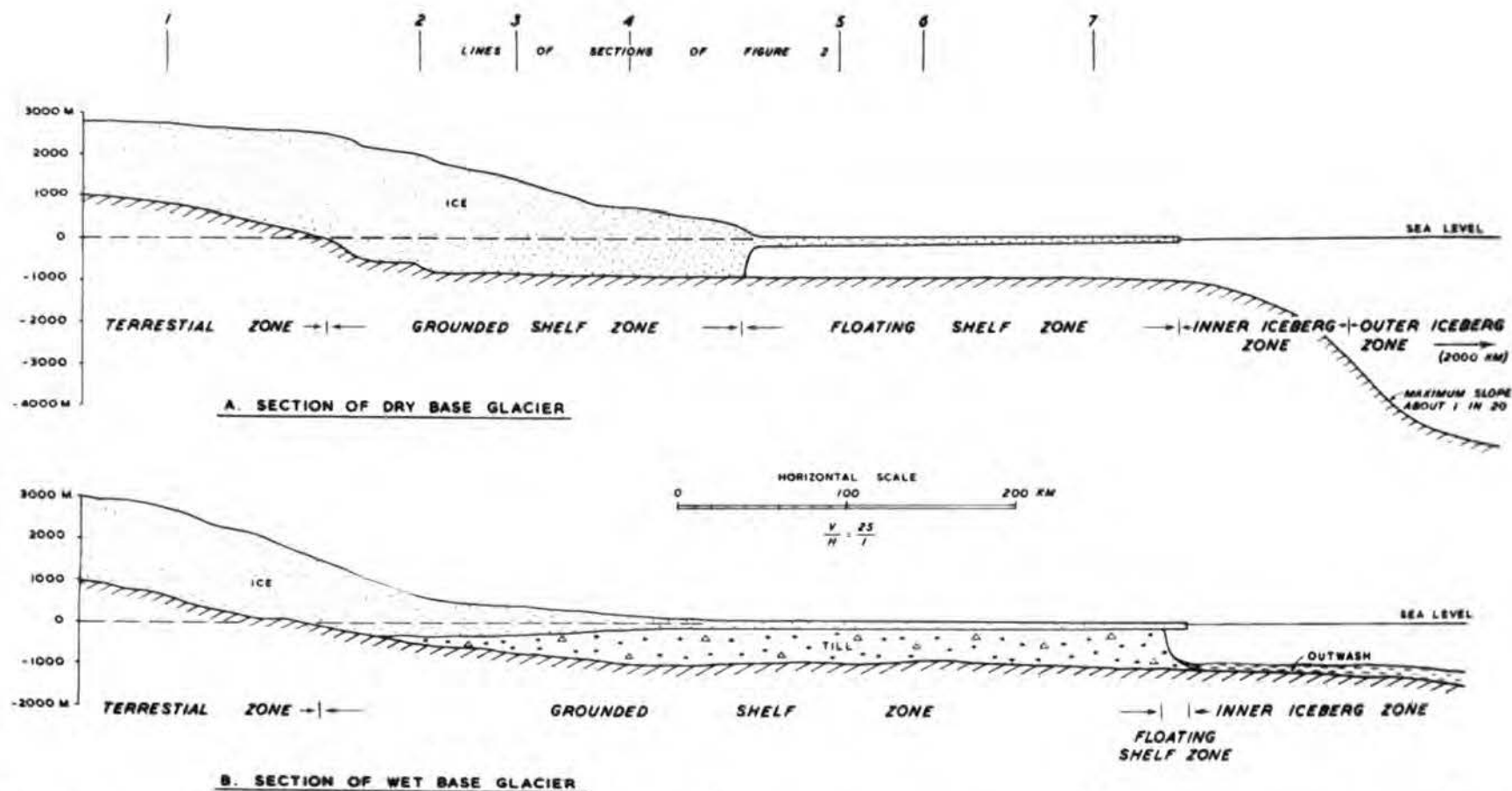
Figure 1. Comparative profile sections of wet and dry base glaciers showing environments of
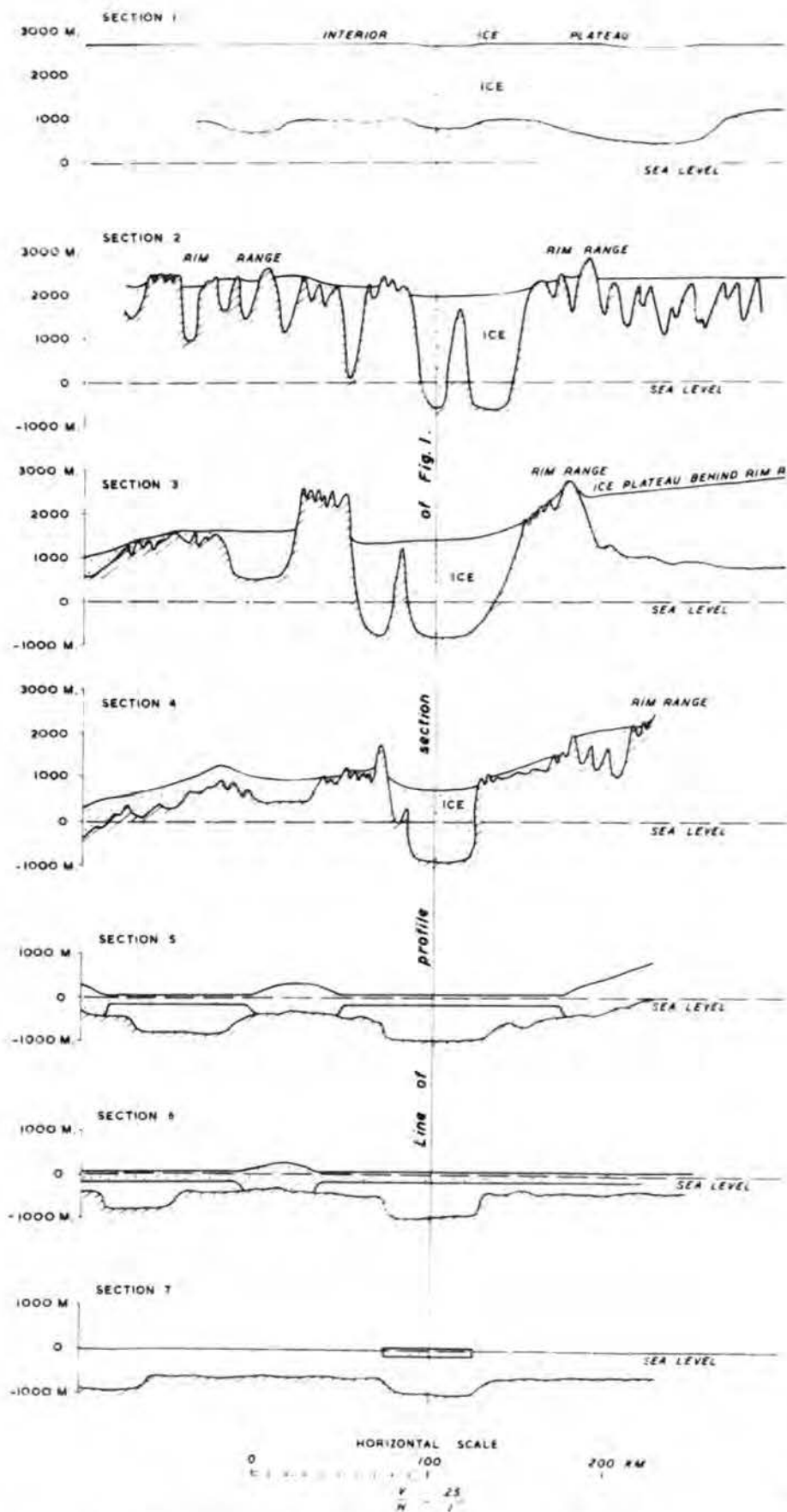
Figure 2. Cross-sections of Pencksokka Glacier.
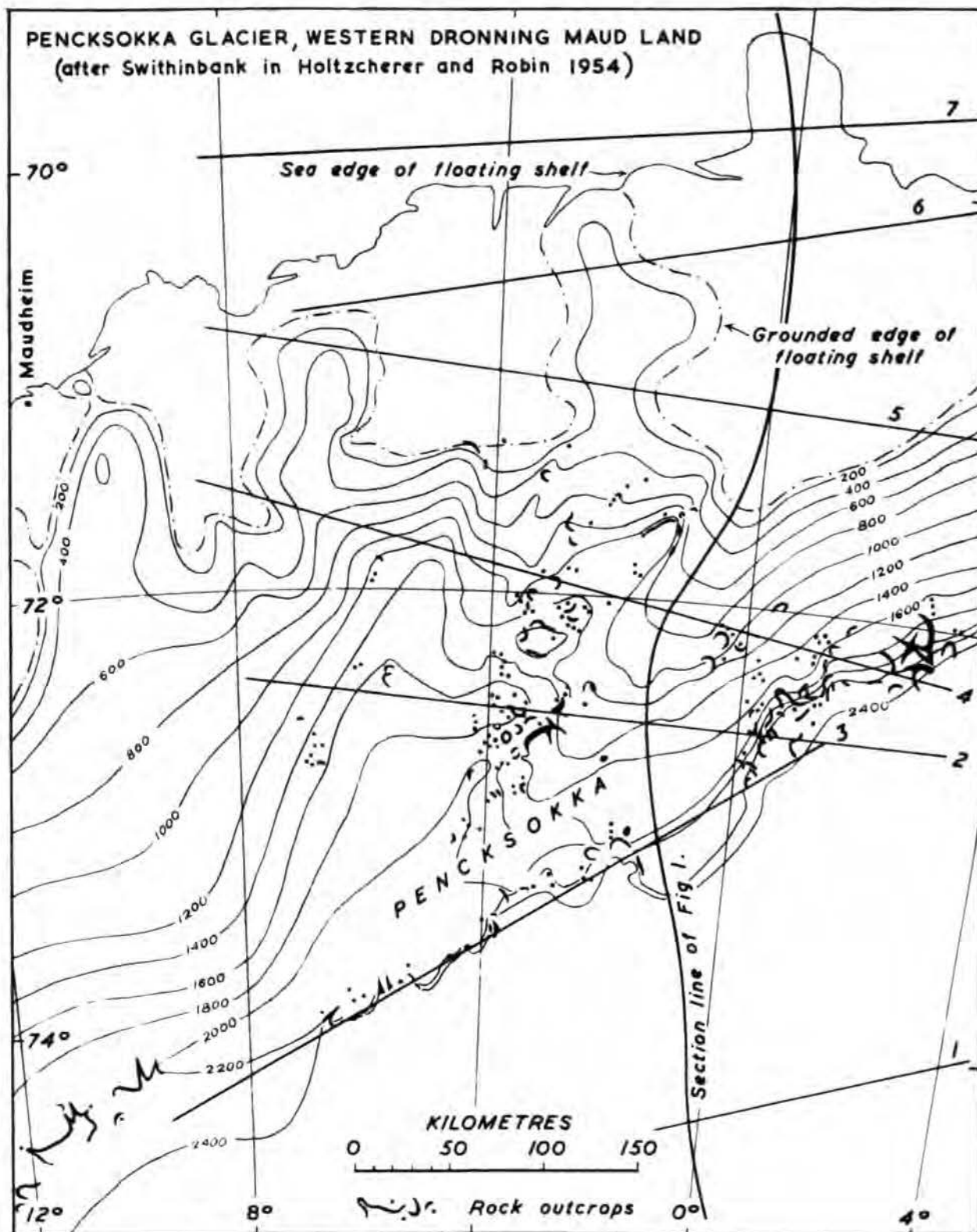
Figure 3. Map of Dronning Maud Land showing Pencksokka Glacier.

500 M.

SEA LEVEL

0

ICE

TILL

Slope up to
1 in 5

OUTWASH

-500

-1000 M.

A. PROFILE ROSS ICE SHELF (ADAPTED FROM POULTER'S DATA)

1000 M.

500

0

SEA LEVEL

200 km. to barrier

ICE

-500

-1000 M.

B. PROFILE PENCKSOKKA GLACIER OF DRONNING MAUD LAND ICE SHELF
(ADAPTED FROM ROBIN'S DATA)
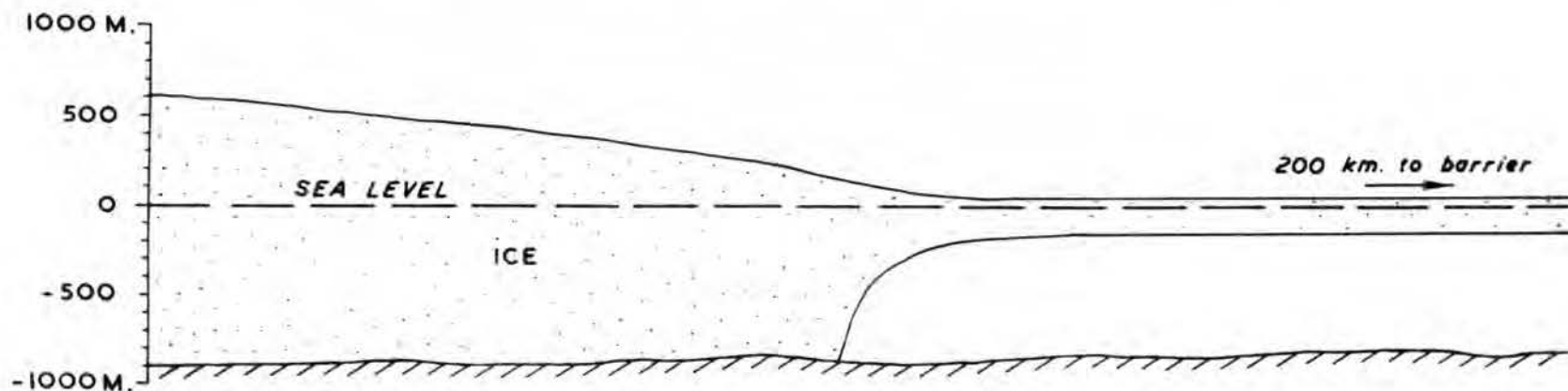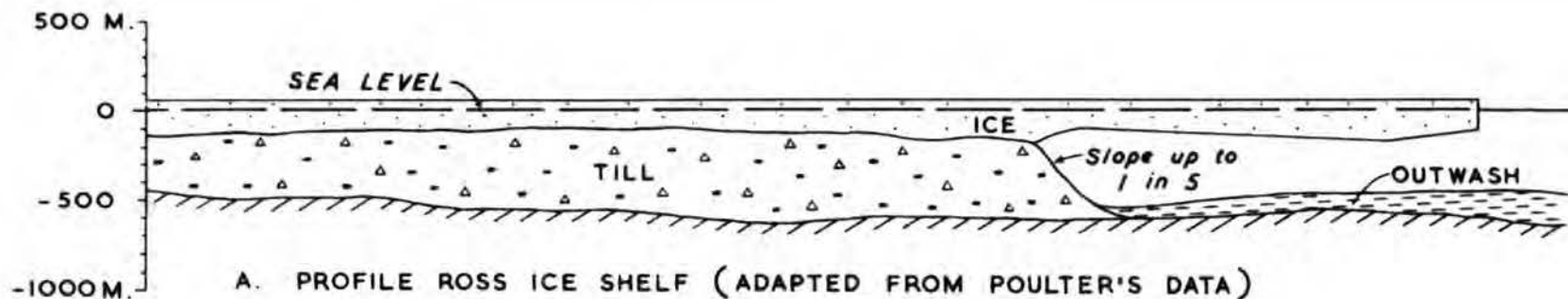
SCALES :—

0   1   2 KM. VERTICAL
0   10   20 KM. HORIZONTAL $\frac{V}{H} = \frac{10}{1}$

Figure 4. Comparative profile sections of transition from grounded to floating shelf
in wet and dry base glaciers.
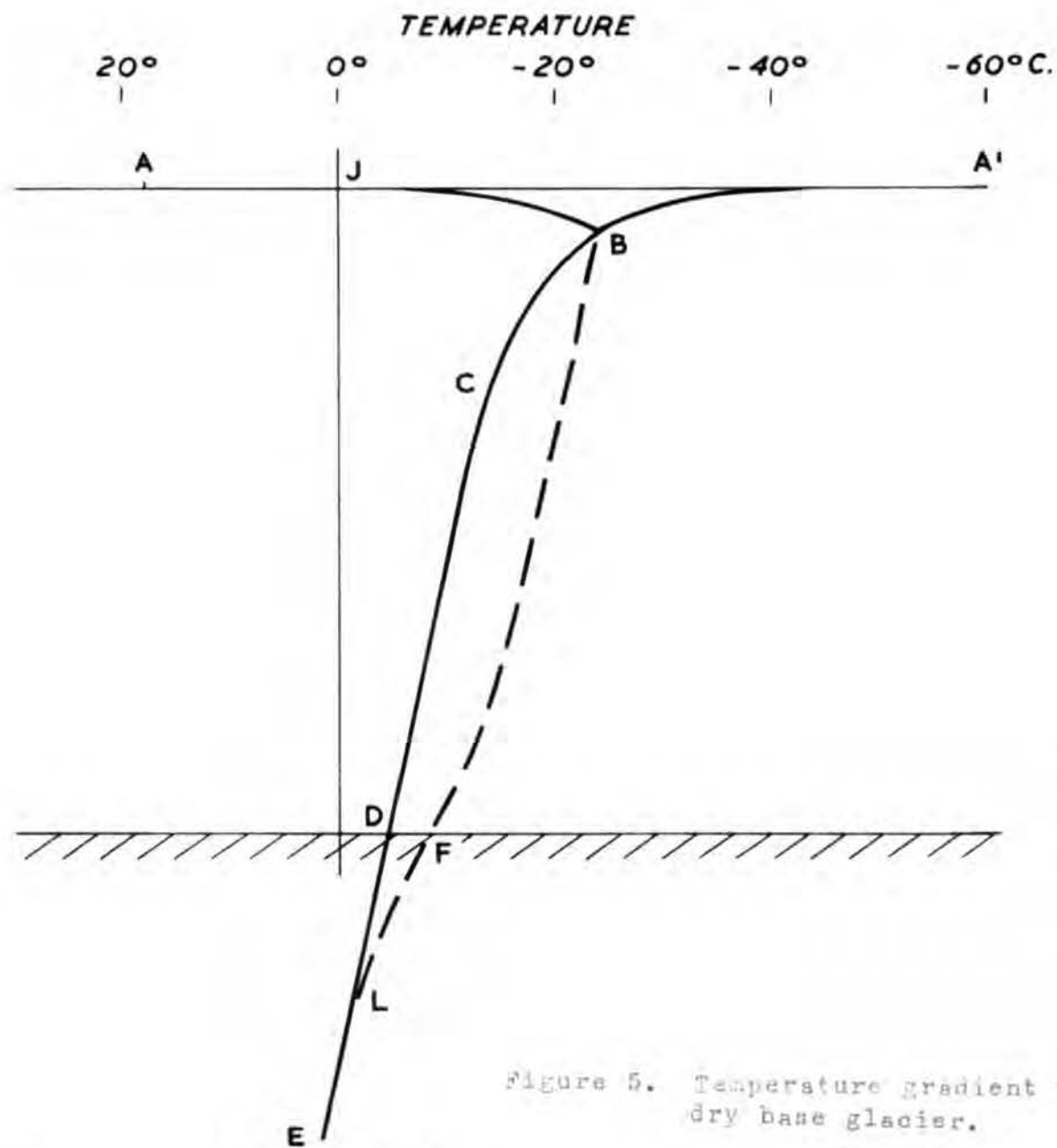
Figure 5. Temperature gradient through a dry base glacier.
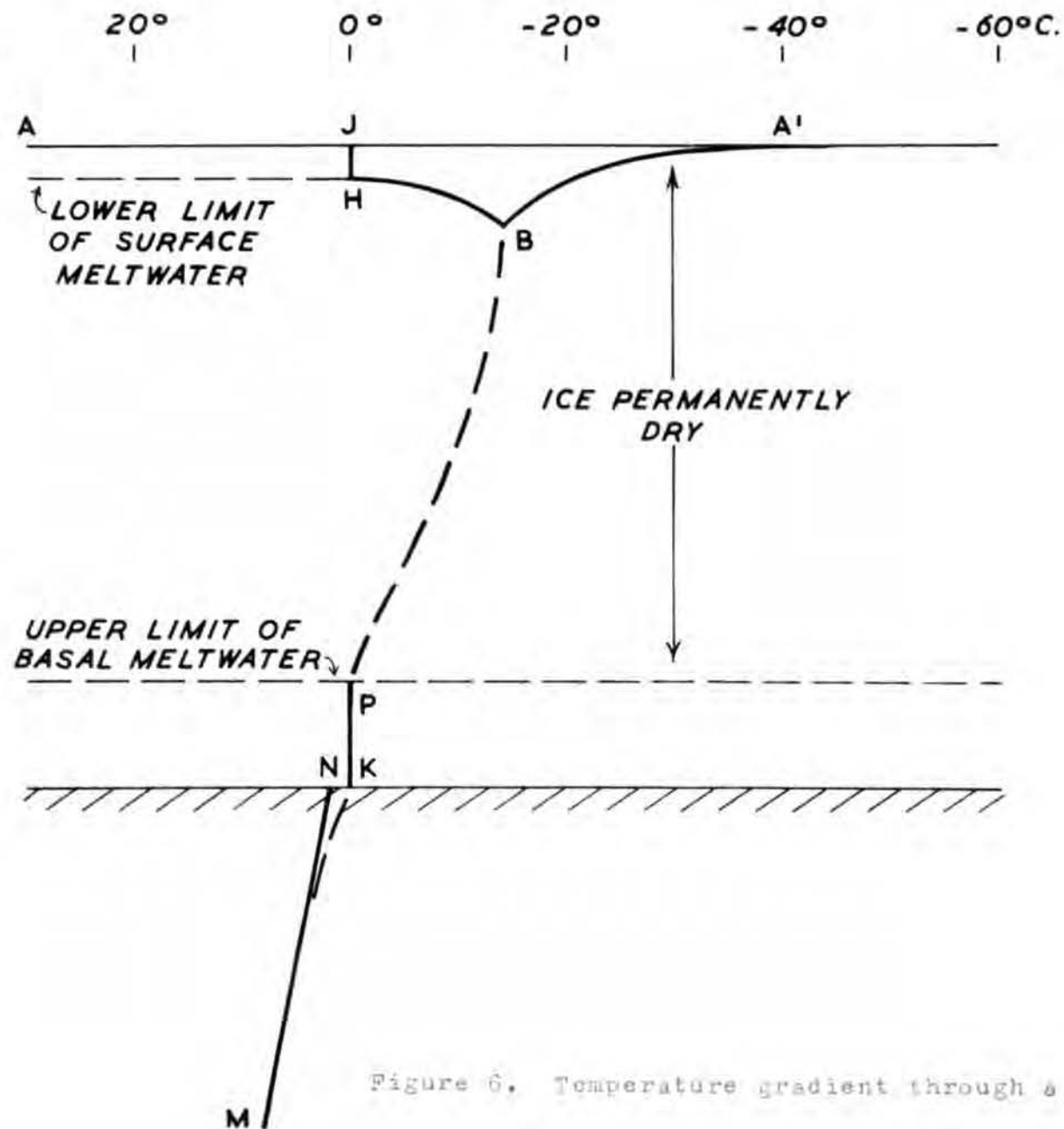
Figure 6. Temperature gradient through a wet glacier.
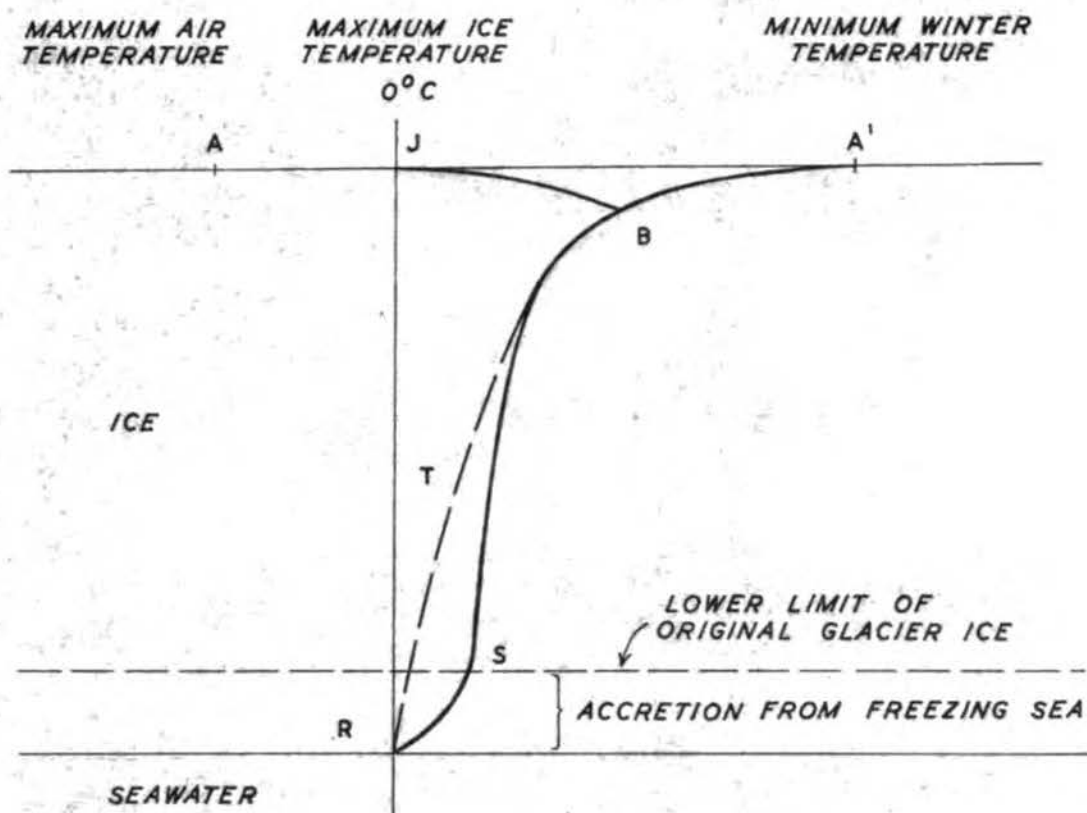
Fig. 7.

Temperature gradient of the floating ice shelf
derived from a dry-base glacier (assuming
absence of surface meltwater zone).

J

H

WET ICE

J'

DRY ICE

H'

P'

P

WET ICE

Y'

WET GROUND MORAINE

Y
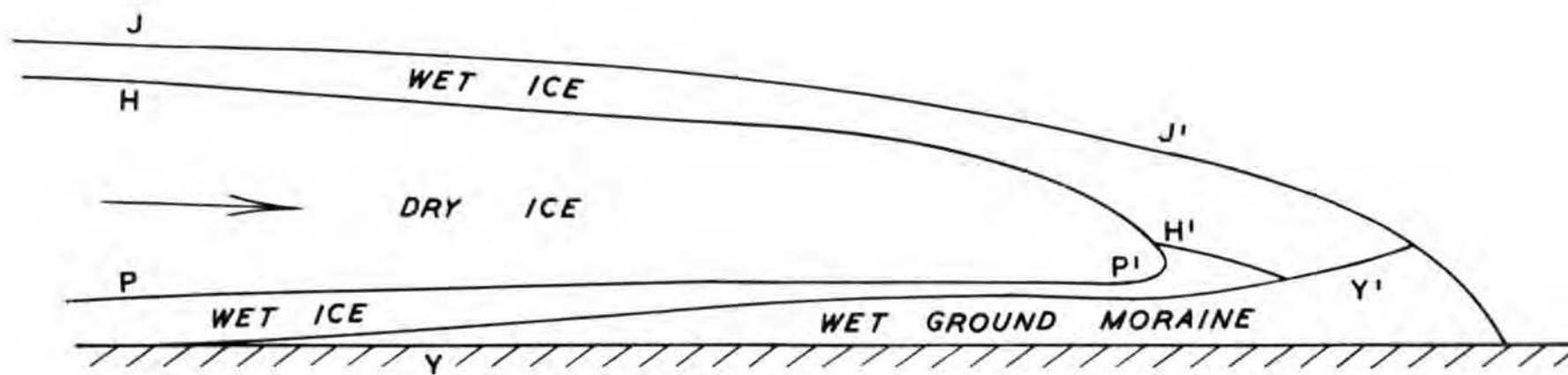
Melting zone at base of glacier.

THE LEGACY OF AN ICE CAP

The Lakes of the Western Part of the Central Plateau of Tasmania

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Introduction

Late on a clear afternoon the traveller flying between Hobart and Launceston and looking west will be impressed by the glaring patchwork of evening sun reflected from the intricate patterns of standing water of the western part of the Central Plateau. Here are to be found the "Nineteen Lagoons", the "Forty Lagoons" and the "Ninety-nine Lagoons" of local parlance. In fact, the water surfaces are numbered in hundreds (Fig. 1, folding map [page 64]). On that portion only of the Central Plateau which is included in the Du Cane 1/63360 sheet, over 4,800 lakes and ponds are figured. Admittedly most of these are but ponds in sphagnum bogs or rock pools, but there are over 500 lakes of 1 hectare (2.5 acres) or larger. The pattern of the flowing water channels is equally involved, indeed labyrinthine in parts; moreover many small areas including substantial lakes (e.g., south-east of Pillans Lake) are shown with no outlet at all. In this climate there is of course outflow; but here the drainage is so indeterminate that it defies recognition on the air photos.

This western part of Tasmania's "Lake Country" is geographically unique in Australia. It has long been recognised that it is the result of Pleistocene ice action (A. N. Lewis 1923, 1932; W. H. Clemes, 1924) and that to Canada's Barren Grounds or to Finland, the here we have this continent's nearest approach "Land of Ten Thousand Lakes". There has, however, been no very close consideration of the origin of the lakes, though air photographs are now available and good contoured maps have been prepared for most of the area[1]. It is the purpose of this paper, based on air photo study, a special flight over the area and four weeks' field work in it, to go some way to fill this gap. It can be a reconnaissance study only, because it was not possible either to take a portable boat for soundings or to make borings in the glacial drift. In fact, the paper is offered more as providing regional instances for teaching purposes in Australia than as contributing to the systematics of geomorphology, though there is perhaps some general interest in the method employed to determine the pattern of ice movement.

Extent, Structure and Relief

The Central Plateau is a well recognised unit with a particularly clear boundary on the north and east in the high scarp of the Great Western Tiers, which rise to well over 4,000 feet at many points, though declining as a whole to the south-east. This scarp, 2-3,000 feet in height, is regarded as directly due to Tertiary faulting—a true fault scarp (Carey, 1947).

On the west there is a clear limit also, even though it has not always been used as the limit of this physical region; this runs along the deep glaciated troughs of the upper Mersey[2], the Narcissus, and Lake St. Clair and along the upper Derwent. East of this line the plateau has suffered little dissection. Again along this western margin altitudes are generally above 4,000 feet.

Southeastwards and southwards the plateau declines fairly gradually in height, and it is much harder to discern a boundary. It can

1. Reference should be made to the excellent Middlesex (No. 45), Du Cane (No. 52) and Great Lake (No. 53) sheets of the 1/63360 State Aerial Survey which cover a good deal of the area involved. The Lands Department, Hobart, kindly made available unpublished 2 inches to 1 mile surveys with 25 ft. contours of these three sheets. There remain topographically unmapped small parts of the area of study which will be covered eventually by the Quamby (No. 46) and St. Clair (No. 59) sheets. The drainage of Fig. 1 is based with permission on the accurate Lands Department maps to the extent they were available; the southernmost part and a small area in the north-east corner rely on uncontrolled air photos and here the map is much less reliable. Since the map was printed Lake Latona has been officially named Lake Sappho, its original name, by the Nomenclature Board of Tasmania.

2. Recognition that the Mersey valley, at least above the Arm Junction, has been heavily glaciated is recent and is due in the first place to J. L. Davies. This has been confirmed in field work by parties led by A. Spry and E. Ford and by the present writers.

be taken to run from the Wentworth Hills to Table Mountain, more or less following the 2,000 feet contour. There are scarps here, some associated with faulting but nowhere are they so well-defined, high and continuous as the Great Western Tiers.

Geologically the Central Plateau is simple, being largely developed on dolerite (Fairbridge, 1948; Prider, 1947; Voisey, 1948, a, b). With accompanying faulting this tough rock was intruded in Jurassic times into Permian and Triassic sediments as sheets or sills (possibly of lopolithic nature). Then in late-Mesozoic and early Tertiary times the country was peneplaned and the covering rocks removed to expose the dolerite over most of the plateau; smaller areas of sandstone, mudstones and shales of Permian and Triassic age do appear on its surface, more particularly in the south central parts. The peneplain was block faulted in mid-Tertiary times, when the plateau was uplifted and defined. Valleys were cut into its southern slopes, and in the late Tertiary (probably in the Pliocene) basalts were poured out to fill several of these valleys. Faulting accompanied and may have followed the vulcanicity. Despite the latter, dolerite remains by far the most widespread rock and, though a tough rock resistant to many forms of denudation, if not to chemical weathering, it is everywhere thoroughly shattered by numerous planes of weakness—joints, shear planes, and faults of varying magnitudes—resulting from the tectonic history outlined.

This paper is directly concerned only with that part of the Central Plateau lying northwest of the Lyell, Marlborough and Lakes Highways. There is a big contrast between the west, where there are innumerable moderate-sized and small lakes, and the centre and the east, where there are only few (but large) lakes and where the drainage pattern is much simpler. The relief of this western part is characterised essentially by a gentle decline in height southeastwards from a high rim on the northern and the western margins. But a few other major characteristics must be noted. Starting from the knot of hills which comprise the Walls of Jerusalem, a substantial scarp strikes southeastwards to divide the area roughly into two halves, the northern half standing some 400 feet above the southern half along this line. This Great Pine Tier very probably follows a fault. On the west also the plateau is broken by features which must be at least fault-guided; thus south of the Fish River, the plateau drops down to the Mersey valley in two irregular steps. The edge of the main plateau southwards from the

Walls of Jerusalem follows a curving line east of Lakes Adelaide and Meston round to the Du Cane Gap. These two lakes lie on the first step down, which includes also the broad ridge to the west from Howells Bluff to Mt. Rogoona. Lakes Myrtle, Louisa and Bill belong to the next step. Both steps are cut off on the S.W. from the Cathedral Mountain plateau by a remarkable linear cleft, in which Chapter Lake and Cloister Lagoon lie. On the main high plateau, a significant variation in the amount of local relief is to be seen. Apart from the neighbourhood of the Walls of Jerusalem, there is much more hill and valley at the northern end (north of the Julian Lakes) and in the southwest (near the Mountains of Jupiter and Travellers Range) than in between.

The Nature of the Glaciation and the Pattern of Ice Movement

Although A. N. Lewis (1932) was doubtful about the extent and degree of glaciation in the centre and east of the Central Plateau, he had no doubt about the fact of strong glacial action to the west of Great Lake. He stressed in particular the prevalence of end moraines, whereas Clemes (1924) was impressed more by the erosional effects of ice, in particular on the Travellers Range in the south-west. Both were right; it was in fact possible to recognise in the air photos and later to confirm in the field a zone of predominant erosion and one where deposition held sway.

From the plains N.W. and S.E. of Wild Dog Tier, this latter belt stretches south-west past Lakes Augusta and Ada to the Great Pine Tier: beyond, in the more wooded country, it has only been possible to sample, but numerous moraines were found near Blue Lagoon, Lake Ina, Travellers Rest Lake and Clarence Lagoon. Here the moraines lie near to and invite correlation with the great series of festooned moraines reaching south and east from Lake St. Clair, the work of a great piedmont glacier rising in the Du Cane Range but partly fed also from the Central Plateau.

North-west of this belt moraines are by no means absent, but are not prominent except in certain places. Instead there is a tremendous amount of rock exposed and a corresponding absence of soil. Almost everywhere evidence of ice abrasion and plucking is to be seen and it is possible to walk for mile after mile with scarcely a break over roches moutonnées. The depth of ice must have been considerable, since nearly every eminence was overridden, including even Mt. Jerusalem (4,788 ft.) which rises between 400 and 800 ft. above its imme-

LAKES
OF WESTERN PART OF
CENTRAL PLATEAU
TASMANIA

LOCATION

TASMANIA

MILES

Structural Lineaments (Faults, Joints, Etc.)

Ice Movement

Moraine

Till Swell

Outwash Plain

Corque-like Feature

Ice Eroded Plateau Margin

Frost Shattered Plateau Margin

Plateau Margin of Uncertain Origin

Possible Nunatak
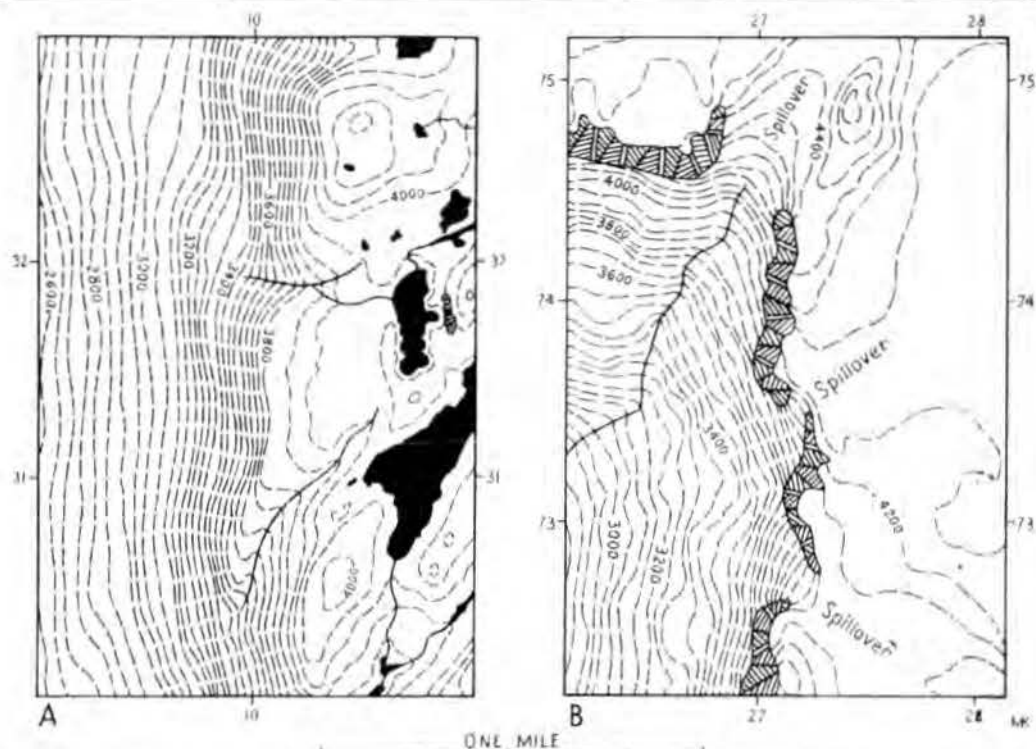
Figure 2.—A: *Plateau margin along Narcissus valley, highly abraded by ice from plateau.* B: *Plateau margin south of Western Bluff; frost-shattered "windows" alternate with ice "spillovers."*
—Drawn with permission from Lands Department maps.

diate vicinity; its high steep western face (the East Wall of Jerusalem) has been vigorously eroded by ice and has only a little scree below it, the product of post-glacial frost weathering. Much of the facing Wailing Wall and West Wall was abraded too, but the highest points here may have escaped burial under ice to stand out as nunataks; certainly the cliff above the Damascus Gate is deeply gullied, shows much more sign of frost shattering, and has bigger screes below.

Signs of ice erosion generally reach right to the edge of the plateau on its northern and western sides. In the south-west, for instance, overlooking the Narcissus valley, the edge is vigorously rounded by ice abrasion; here considerable thicknesses of ice must have flowed over into the valley (Fig. 2). But as the margin is followed northwards down the Mersey valley, such abraded margins become more frequently interrupted by craggy cliffs where frost shattering has long had free action, e.g., Howells Bluff. These cliffs must have constituted windows in the ice cover at least in a late phase of the glaciation; indeed in some cases (e.g., Western Bluff) it seems likely that there were in fact small areas right on top of the plateau projecting as nunataks (Fig. 2B;

plate 1). In the north, where these nunataks and windows are more frequent, the ice which overflowed between them, eroded shallow scoops on the plateau edge. Below these "spillovers" true glacial till can be seen in some instances, e.g., where the Mole Creek Track rises to the top of the Tiers. More often great block moraines are found running down from the spillovers with no rock cliff above: these can be much more massive than the screes below the neighboring frost-shattered cliffs. Distinguishing these block moraines from periglacial blockstreams, which also occur on the tiers, can be a difficult matter. On the western side much of the ice flowed down to join great valley glaciers, but on the north the ice tongues cannot have reached far down the Tiers; at any rate no very obvious traces were left below them.

For the analysis of the lake pattern, it is important to know the system of ice movement in the area. Unfortunately the dolerite plateau is in many respects unfavourable to the determination of this system.

(a) *Striations and Associated Features.* In accordance with the experience of previous workers, the present writers observed not one occurrence of striations, groovings, or friction

Plate 1: In foreground glacial catenary where ice from Central Plateau overflowed into Mersey valley. In middle ground is Cathedral Mountain, a possible nunatak with a glacial "spillover" on its right. In background Mt. Pelion East, another possible nunatak.

cracks on the plateau. Dolerite does not preserve such minor erosional features, presumably through a susceptibility to chemical weathering (Walker & Poldervaart, 1949). In fact the only *striae* observed by the authors during the field work were in quartz-schists at the low elevation of 1,700 ft. in the Mersey valley at the northern end of Howells Plains.

(*b*) *Erratics.* The almost completely doleritic nature of the plateau means that little help can be derived from erratics. A train of vesicular basalt fragments from the western side of Lake Augusta south-eastwards to Lake Kay was at first puzzling in that all known nearby outcrops of basalt lay east of this train whereas other evidence of ice motion indicated a contrary movement; this difficulty was largely removed by the later discovery of a small outcrop of such basalt at 457441. The only basalt erratics north-west of this were along the shore of the lake and might have been transported there by other means. Slightly metamorphosed banded siltstone erratics were also found between Pillans Lake and Lake Kay; the provenance of these rocks was not established. The two localities of Permo-Trias encountered within the area of study were unlikely to have yielded them. One of these

is a small raft of grey-black carbonaceous shales and grey felspathic sandstones exposed in a rotational landslip on the flanks of The Temple (279499) : this outcrop did not yield suitable rock material for the erratics. The other locality was in the Mersey valley in the Lake Meston-Junction Lake neighbourhood, where in heavily drift-covered country a number of exposures of Permo-Triassic sediments were seen. It is not impossible that suitable erratics could be derived from this area; on the other hand other evidence suggests that flow was not such as to carry material from this area to the places where the siltstone erratics occurred.

(*c*) *Moraine Morphology.* The most characteristic moraine yielded by the dolerite in the western part of the Central Plateau is an irregular spread of very bouldery drift, which is not readily related in its morphology to the direction of ice movement. No true drumlins were seen, possibly because of the high proportion of boulders in the drift;[3] the nearest approximation was found in several very low oval mounds which formed N.W.-S.E. promon-

3. Till fabric analysis was not undertaken. The large numbers of big spheroidal boulders in the till make it a difficult and unpromising line of attack.

tories along the S.E. shore of Lake Augusta.[4] These mounds were so flat that no clear distinction in slope between the ice approach and the opposite side could be made. But the axial direction did correspond with the direction of ice movement established by other means for this vicinity. A number of the end moraines found in the area of study were arcuate in plan (e.g., near Travellers Rest Lake, Lake Mackenzie and Fishers Lagoons) and from these a rough estimate of flow direction could be obtained. Some of these end moraines were recessional and lay within valleys on the plateau and it might be expected that the movement which produced them would be more subject to the control of local relief and different from the general flow at maximum glacial development. There were in fact a few cases of discrepancy (e.g., at Stumps Lake) between the direction of ice flow indicated by small crescentic valley moraines and the regional ice flow established by other evidence.

(d) *Erosional Morphology.* This other evidence consists chiefly of ice-erosional features on a larger scale than striations, and in the production of these one property of the dolerite, namely its excessively jointed nature, was decisively favourable. The joints promoted ice plucking and so strong contrasts between onset and lee slopes are apparent in the zone of ice erosion—smoothed, gradual slopes face the direction of ice approach whereas steep, quarried faces were produced where the ice moved away from the rock. Large hills as well as roches moutonnées show this contrast, which is not restricted to isolated projections, but applies also to long ridges and valleys. Where either of these had lain across the former ice movement, asymmetry could be produced with a more gradual abraded slope on one side and a steeper plucked slope on the other. Furthermore, although the broad character of a U-shaped valley cannot reveal whether ice moved up or down its length, the detail of its steps and sidewalls can (W. V. Lewis, 1947). Any slight projection in the valley sidewall will show plucking on the lee side in terms of ice movement. There are some cirque-like features on the plateau, which will be referred to later; they also afford a rough determination of ice flow direction.

Even without plucking on one flank, a roche moutonnée can often give some indication of direction of ice flow since the onset side is often worn to a gentler gradient than the lee side. But plucking was very prevalent on this dolerite plateau: indeed a common form of roche moutonnée was plucked along the two sides parallel to ice movement as well as on the lee side. An interesting case in the James river valley at 426493 was furnished by a roche moutonnée which was plucked on the lee side and otherwise normally abraded except for a neat excision, looking almost as if artificially quarried, on one flank of the onset end. The *general* tendency of quarried slopes to face a particular direction was the criterion to be used to determine the movement, though it was recognised that odd exceptions would occur.

Chiefly then by means of the distribution of abraded and plucked slopes a map of ice movement was prepared (Fig. 1).[5] First of all this was done from the air photographs and then checked in the field as widely as possible. The two methods agreed well, the field work usually providing additional instances on a smaller scale which confirmed the indication of the larger features. In view of subsequent argument it is important to stress that the orientation of lake basins themselves was not used in determining the ice movement.

The pattern of ice movement shown on the map largely explains itself, but some comment is necessary. Above all it reveals the existence of one large elongated icecap, covering with one system of outward flow nearly all of the western plateau. There was a major ice divide running roughly S.W.-N.E. In the S.W. this divide lay very close to the Mersey valley at a distance of one to two miles from the margin of the high plateau; nevertheless the amount of ice overflowing north-westwards was considerable, since it overswept the two steps and eroded them vigorously before falling into the Mersey valley. Between Junction Lake and the Du Cane Gap the ice fell directly into the uppermost part of the Mersey valley, the divide here in fact running W.N.W.-E.S.E. On the opposite flank of the ice divide, movement was outwards in roughly radial plan. W.N.W., S.W., S.S.W. and S. into the Narcissus-Lake St. Clair valley, S.S.E. into the Travellers Rest Lake area, S.E. into the Nive, Little and Pine river valleys.

Across the Great Pine Tier the ice divide retained much the same trend as before, but here of course the western edge of the high

4. As it existed before the building of the dam; the natural shore is shown on Fig. 1 and can still be examined at low reservoir levels. On Fig. 1 only the major ice depositional features which were examined in the field are shown. Much other moraine in the wooded country is not included because air photo identification is not sure enough on its own.

5. On the scale if Fig. 1 it is not possible to show the evidence by which the ice movement was determined, but only the directions.

plateau lay farther away and there was a closer balance between the areas on either side of the divide. The ice moved W. through the Walls of Jerusalem neighbourhood to the Mersey and Fish valleys, N.W. into the Little Fisher and Fisher valleys, N. and N.E. to topple over the Great Western Tiers at many points, S.E. to cover the plains around Lake Augusta and Lake Ada.

The movements described so far formed one system of common outflow, the expression of an icecap some 30 miles long by 20 miles across. It is possible that there were two domes, north and south of the Great Pine Tier respectively; there is indication of flow, for instance, down this Tier. But otherwise within this area there was no marked convergence of ice-streams.

On the western and northern periphery, however, the topography favoured the development of small independently functioning centres, though the evidence for them is in some cases distinctly less clear. The Cathedral Mt. plateau acted as one such centre, the bulk of its ice moving N.E. and converging around Lake Myrtle with the ice of the main plateau. There was some flow southwards into the Mersey valley and some north-westwards down the spillovers between Cathedral Mt., Bishop's Peak and Dean Bluff. Behind Clumner Bluff on the plateau tongue between the Fish and Little Fisher valleys, flow is not easy to determine; south of the Bluff itself there was a nunatak but some overflow went southwards as well as northwards. It is possible that there was a small separate cap here. Slightly larger than the Cathedral Mt. centre was another north of the Devil's Gullet; here the main flow was southwards from the high tiers east of Western Bluff. There was convergence with the main ice north-west of Lake Mackenzie. It seems likely also that the high tier of Bastion Bluff acted as a small independent distributive centre, but the outflow was closely confined on the west and south by strong currents from the main icecap.

Subsequent ice movements are of course likely to remove or at least confuse the effects of previous movements of different pattern; the pattern reconstructed above must in fact represent the conditions at the time of the latest substantial glaciation of the western part of the Central Plateau.[6]

8. A comparison of the map of ice movement presented here and the much smaller scale map by A. N. Lewis (1932) would reveal many similarities but not a few contradictions. It is not proposed to analyse these differences and defend them since Lewis' map seems to be mainly based on general considerations of topography (possibly also of precipitation) and not on more direct evidence of movement such as has been used here.

The Lakes

Amongst more than 60 lakes examined in the field the range of geomorphological type is quite wide. The object now will be to illustrate this variety of type and then to discuss certain more general questions relating to the lakes.
(a) *Simple Origin.* At one extreme are lakes entirely due to an irregular surface of glacial deposition; these are best exemplified on the till plain S.E. of Lakes Augusta and Ada (Fig. 1). Here the bouldery drift has a gentle sag-and-swell topography superimposed on the general S.E. slope; the swells rise no more than 10-30 ft. in height and show no marked

Figure 3: Clarence Lagoon, a moraine barrage lake. —Drawn from air photographs; approximate form-lines only.

elongation and orientation. It is ground moraine rather than end moraine as A N. Lewis (1932) described it. Occupying a number of sags are shallow lakes, rounded in plan with simple curving shore lines; typical are Rocky Lagoon, Lake Paget and Carter Lakes. Lakes Chipman and Botsford, though compact, have a slightly more complex shape, each showing a projecting bulge on the eastern shore which corresponds with the form of a till swell. The shores of these lakes are usually lined by dolerite boulders, which often indeed break the surfaces also, testifying to their shallowness. Certain of these lakes, notably Double Lagoon, First and Second Bar Lakes, have a further characteristic feature in the form of ice-pushed ramparts which divide or nearly divide them into twin lakes.[7] The ramparts vary from vegetated sand and boulder

7. S. W. Carey drew our attention to these features and to their mode of origin. Griggs, R. F. (1909) was apparently the first to discuss such dividing ramparts, though the effect of ice-push on lake shores was understood earlier (Gilbert, 1890, Bulkley, 1900).

ridges tens of feet across to single lines of large boulders, their materials demonstrating conclusively that they are not wave constructional features. Such dividing ramparts can only form in shallow lakes and the excessively blocky nature of the drift on the plateau would also be favourable.

Lakes dammed by a moraine barrage are more frequent and more widespread in their occurrence. Clarence Lagoon (Fig. 3) furnishes a very simple case. This is situated near the margin of a tier overlooking the Lyell Highway and in a broad valley between two rock ridges rising about 300 feet higher. Running transversely across the mouth of the valley right on the lip of the tier is an end moraine. The stream outlet of the lake is at the eastern end and the moraine ridge rises to a maximum height of about 50 ft. above the lake, declining to some 30 ft. above it at its western end. The lake looks very shallow and the shore shows no solid outcrop at all, consisting variously of reedswamp, moss bog and moraine. There is nothing to suggest that any factor other than the end moraine is responsible for the lake. Blue Lagoon (3330), draining into the Nive River above Gowan Brae, is very similar in general nature, though here the impounding moraine rises only several feet above the lake.

Lake basins due to the melting of ice blocks are particularly associated with fluvioglacial deposits, e.g., kettleholes in pitted outwash plains and esker troughs. However, no fluvioglacial landforms were recognised in the area of study other than a small, practically featureless outwash plain south of Travellers Rest Lake. Ice-block basins do occur in till as well, and these are represented here. Towards the north-eastern end of the main ice divide the plateau possesses a minor knobbly relief which gives it a distinctive air photo pattern somewhat resembling beaten copper. Where examined on the ground this relief was found to be due chiefly to rather vigorous small moraine hummocks, though roches moutonnées did occur as well. Occupying some of the rather steep-sided depressions amongst these hummocks were small lakes and ponds, often with no surface outflow; it is very likely these are due to the melting of ice-blocks (Fig. 4a). The water bodies which could safely be attributed to this origin are of small size, up to about 500 ft. in maximum dimension. In fact no large-scale features of dead-ice topography were recognised on the plateau.

Also on a similar small scale are most of the lakes entirely surrounded by solid rock, due solely and obviously to glacial overdeepening. The hollows in amongst roches moutonnées provided very many instances of this type, where one could walk on the bare rock round the whole shore. Often these also had no regular streams flowing from them, but points of exit after heavy rains could usually be discerned. There are very many of these tiny rock-basin tarns and it will suffice to give the map references of a few examples from the Du Cane sheet—246435, 428491, 137357. Generally they seem quite unrelated to any conceivable preglacial relief and are very completely the product of the icecap. With increasing size, the rock-basin lake rapidly increases in depth, in contrast to the depositional types, and the shape becomes intricate and often rectilinear. A case in point is furnished by the unnamed lake lying between L. Myrtle and L. Louisa (Fig. 4B, 4C). The planes of
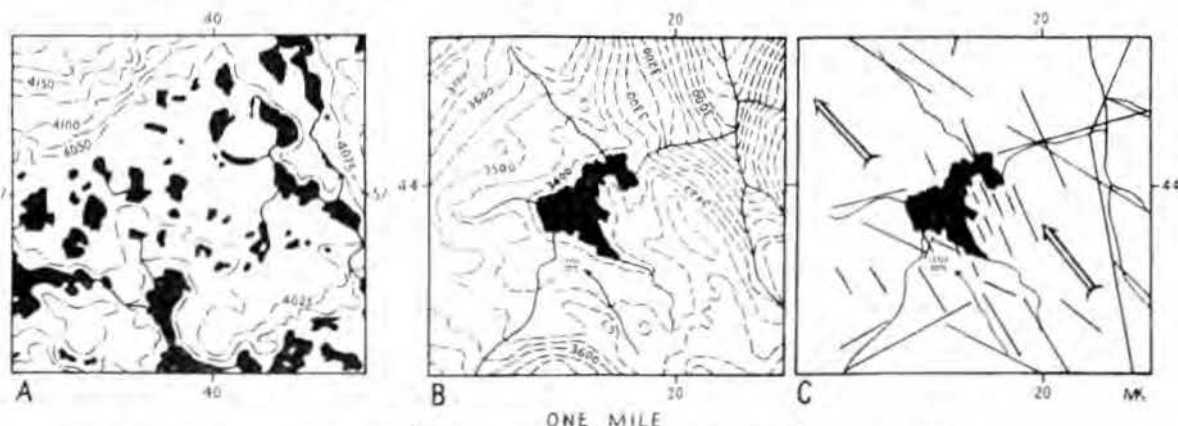
Figure 4.—A: Small lakes south of Lake Nameless formed by melting of ice blocks in moraine. B: Small rock basin lake between Lake Myrtle and Lake Louisa. C: Ice flow direction and structural lineaments in this vicinity.
—Drawn with permission from Lands Department maps.

weakness in the dolerite are clearly responsible for the angular shape.

(b) *Compound Origin.* With the small lakes and ponds an origin by one process is common enough. This becomes rarer amongst the larger lakes, where frequently both erosional and depositional work are involved, though often the relative importance of the two (or even the certainty of the presence of the former) is not determinable without sounding of the lake and boring along the moraine barrier.[8]

Thus Lakes Ada and Augusta (Fig. 1) are held up on their S. and E. shores by the till plain. Extremely flat drumlin forms complicate the former shore-line of Augusta as mentioned earlier. But on their N.W. and W. flanks, the lakes are encompassed by rock ridges and indeed low plucked cliffs lie close to parts of their N.W. shores. S.W. of L. Augusta is a rock ridge exposing dolerite and basalt at several points but largely plastered with a veneer of drift; the outlet of L. Ada runs over drift between two swells of drift, each however revealing a rock core. There is then the possibility that the two lakes have basins which are partly due to ice erosion. However both lakes appear to be comparatively shallow and borings in connection with the dam and embankment construction of the L. Augusta reservoir showed that considerable thicknesses of drift were quite common, depths reaching to more than 80 ft.[9] The probability is thus more in favour of the lakes being entirely impounded by the glacial drift.

Also in the belt of predominant deposition, Lake Ina (Fig. 1) offers much the same degree of probability of solely depositional origin but with erosion not completely eliminated. End moraine ridges enclose its broad southern end, but rocky knolls surround its narrow northern extension and reach more than halfway down its western side so that a rock basin is not impossible at the N. end. However, the presence of an ice-pushed rampart almost isolating the narrow northern arm suggests shallowness and the presence of a submerged moraine across the mouth of this arm; this part also may be solely moraine dammed.

Farther S.W. the Travellers Rest Lake presents a picture which leads to a contrary inference. The southern half of this elongated lake is shallow and drift-lined; moraines project into it and scattered boulders point to their continuation beneath the water. The river runs out through a very low ridge of moraine, which is followed by a small gravel outwash plain. This leads to an extremely shallow lake, 2-3 ft. deep, which is dammed by a well-defined moraine ridge. Below is another arcuate moraine and not far beyond this the river topples over the edge of the dolerite in a waterfall to enter a V-shaped fluvial valley course. The river fall from the foot of Travellers Rest Lake to the top of the waterfall is quite moderate, of the order of 20-30 ft. In contrast the northern half of the lake lies in a quite well-defined glacial U-shaped valley, with solid rock outcropping frequently on the eastern shore, less so on the western. High plucked cliffs surround the head of the lake, except where the valley continues northwards, narrowing and with a rock step in its profile just above the head of the lake. From above, the lake looks very much deeper in this northern half and it appears very likely that this end is over-deepened by ice erosion.

Throughout the area where solid rock outcrops extensively and where glacial erosion was master, there are very many lakes which in the air photographs appear to be simple rock basin lakes in origin. Characteristically they have extremely intricate shapes with many rectilinear shores and sharp angles; the water laps up against roches moutonnées along the margins and roches moutonnées project as small islands within them. But field examination in many cases shows that though for the most part predominantly hewn in the country rock, the last few feet in depth are frequently held up by moraine; this is usually in the form of a thin spread rather than a positive ridge. Lake Explorer (Fig. 1) is typical. The broader southern end lies in fairly open plateau, with roches moutonnées lining much of the southern shore while moraine runs from the other shore up the slopes of the solid ridge to the north. North-westwards the lake narrows, and is gradually hemmed in between the walls of a more confined glaciated valley, to become eventually just a broad river channel. Where this channel breaks into rapids to mark the real end of the lake, there is but a thin skin of boulders over the solid. The general ice movement here was S.-N., and so the rock bar and moraine veneer are broadly at the downstream end in regard to that flow. In the case of the nearby Lake Nameless, the ice drained northwards right along its length and over the rock col on the north to L. Lucy Long, but the river outlet is at right angles to the west through a thin spread of

8. Marr (1895-6), analysed classic instances of the fallacies into which the investigator of glacial lakes may be led.

9. We are grateful to Mr. I. Jennings and Mr. R. Preston of the Hydro-electric Commission for making these records available to us.

drift between two large roches moutonnées into Snake L., which itself overflows in a similar context into L. Explorer.

The present-day major water divide between drainage westwards to the Mersey and south-eastwards to the Derwent lies in places west of the former ice-divide, and so lakes occur which drain to-day in the opposite direction to the iceflow which fashioned their basins. This is true of Lakes Ball and Toorah. Toorah (Fig. 1) is a very fine rock-basin lake, with its origins complicated by a little drift on its south-eastern shore at the stream outlet; but below this the solid soon outcrops. The ice flowed N.W., gouging out the rock basin on its upward flow over the main crest of the plateau; two cols in the abraded crest rise only about 20 and 40 ft. above the lake level. L. Ball (Fig. 1) is in many ways similar; it is H-shaped and beyond three of the extremities lie cols in rock between 20-50 ft. above the water. The Pine River drains out at the fourth extremity through boulder moraine. About a furlong from the outflow and a few feet below, a rock spur reaches the river from the Great Pine Tier; less than fifty yards of moraine separate this from the gently sloping rock slope of the other side of the valley. It seems that only a few feet of the lake's depth can be attributed to the moraine dam.

The examples quoted so far all give surface indications from which a reasonable inference can be made as to the nature of the lake basin concerned. Many are more cryptic and the possibility of a drift-filled valley away from the present line of outflow must always be borne in mind (J. E. Marr, 1895-6). Pillans L. is a case in point (Fig. 1). It has so many ramifying arms, in a part of the plateau where local relief is slight, that without a good deal of boring certainty as to its nature would be impossible. At the present outlet, bedrock is found in the James R. bed less than 5 ft. below the lake level, the water flowing through boulders above this. Between the stream and rock ridges on either side there is not much room for or indication of a drift-filled former valley course to one side. On this evidence alone the lake would appear to be primarily of erosional basin formation. But beyond the rock ridge on the north there is an extensive area of drift with minor hummocky relief leading over to the Ouse valley. A valley formerly leading the drainage this way cannot be excluded on present evidence, and the drift there could be completely responsible for Pillans L., much of which is seen in the air photos to be quite shallow. The long southern arm of the lake is however deeper and fairly closely hem-

med in by rock ridges; it remains very likely that this at least is a rock basin.[10]

(c) *Lake Orientation.* In certain parts the lakes show clear indications of preferred orientation, reinforced by sympathetic alignment of the streams; this is for example very evident throughout the lake district south of the Great Pine Tier. Orientation of glacial lakes is attributed elsewhere to the direction of ice movement in till areas, to differential erosion of rocks of varying resistance, and to erosion along faults and other planes of weakness. Here it is the last process which is operative, for there is little tendency to orientation in the Lake Augusta-Ada till plain, while in the erosional zone variations in lithology can only be involved in one or two places.

All structural lineaments discernible in the air photos were traced, though many had to be omitted in transference to Fig. 1.[11] As with plotting ice movement, the lakes themselves were ignored in this process. The lineaments range from shear zones of considerable magnitude, such as the one limning the Cathedral Mt. plateau on the N.E., down to simple joints. In the area as a whole, N.W.-S.E. lineaments preponderate heavily (cf. Fairbridge, 1948) and their influence on ice erosion in the form of elongation of lake basins and in rectilinear shorelines is most obvious where the iceflow coincided in direction with this structural trend, e.g., in the country between the Great Pine Tier and the Ling Roth Lakes.

But within this set of lineaments directions can range between N.N.W.-S.S.E. and W.N.W.-E.S.E. Moreover there is definitely at least one other major set of lineaments between N.N.E.-S.S.W. and E.N.E.-W.S.W. in trend. There are also a few N.-S. and W.-E. features which it would be difficult to attribute to either of these sets. Consequently there may be planes of weakness in almost any direction in a given area. One result of this may be that where ice attack is strong, the structural

10. Late- and post-glacial processes of non-glacial nature have contributed on a small scale to the lake complex of the area. Dunes, possibly formed in late-glacial conditions when much more bare soil favoured sand-drifting (but now under active blowout destruction, consequent on burning and grazing) have been constructed along the eastern shore of several lakes. In the case of L. Ada, these have isolated part of the main stretch of water to form two separate, small, and very shallow lakes. Post-glacial alluviation of glacial lakes to isolate smaller lakes of fluvial origin has also occurred. Thus L. Kay is entirely surrounded by alluvium. Near here and also N.E. of Lake Mackenzie, streams have built levees above the level of alluvial plains and isolated small lakes in backswamp depressions (cf. Jennings, 1953).

11. Also no interpretative joining of aligned minor lineaments was attempted. As a result some of the greatest features, e.g., the Great Pine Tier fault, do not appear on Fig. 1.

trends which most nearly coincide with the direction of ice movement will be selected and eroded. Thus in the extreme S.W. of the plateau around L. Payanna the radial movements of the ice are roughly reflected in a radial arrangement of the lakes. N.E.-S.W. and N.N.E.-S.S.W. elements preponderate in the structural lineaments revealed in the air photos where the ice moved these ways. Since the normally dominant N.W.-S.E. trends are present, this local preponderance of other trends is thought to be a matter of surface expression only and not a true indication of structural conditions.

This selection of lineaments parallel to the direction of ice flow can only operate within certain limits. Thus in the Pillans L. area, one of comparatively weak glacial erosion, certain trends of the N.N.E. set are strongly reflected in the lake pattern, notably in the long southern arm of Pillans L. itself, although these lineaments are disposed transverse to the ice flow hereabouts. In the north, L. Lucy Long and Westons Lagoon are similarly transverse and this is in an area of more vigorous ice-action. If the structural weakness is pronounced enough, it can find surface expression whatever its relationship to ice movement.

The best examples of transverse basins of glacial erosion in this area are undoubtedly to be found in Lakes Adelaide (Plate 2) and Meston. That the ice moved directly across their lengths is fully evidenced by the morphology of the high plateau immediately east of them and of the ridges to the west. It can be exemplified by one instance: half a mile from the N. end of L. Adelaide there is a high plucked cliff in the dolerite below the strongly abraded crest of the main plateau. Directly opposite this on the other side of the lake a train of great boulders comes out from deep water and runs straight up the hillslope to thin out on the flat ridge-top into scattered perched boulders on well preserved roches moutonnées. It is in fact a magnificent "glacier garden" made all the more attractive by its present setting of gum woodland. This ridge is only 200-300 ft. above the lake, but Mt. Rogoona which was also over-ridden by ice from the main plateau, rises more than 1,000 ft. above L. Meston.

Lake Adelaide (Plate 2) is a deep rock basin lake, overflowing laterally through clefts in vigorously eroded roches moutonnées in a gap in the rock ridge which bounds the lake on the west. The valley which continues northwards in the line of the lake is rock-floored only a quarter of a mile from it. Southwards

the line of the lake is continued by a through valley to L. Meston. A low ridge of moraine alone constitutes the divide; nevertheless it is rock which outcrops round the southern shore of L. Adelaide. L. Meston is not quite so simple a case. A rock ridge and roches moutonnées line three-quarters of the shore along its southern foot; their detail testifies to ice movement from S.E. to N.W. Very deep water comes right up to these roches moutonnées and the outlet flows over a rock bar through them. But at the S.E. corner of the lake the shore is of till and derived sand; not far from this shore, boulder moraine stretches from the plateau wall to the river. However rock in situ occurs at so many points on the stream course in the 100 ft. of descent in the first half mile that it is almost certain that the lake is mainly a rock basin lake; the possibility that there is a deep drift-filled valley very close to the southern wall is slight. Dolerite constitutes the rock barrier, but within 50ft. of the lake level Permo-Triassic sediments outcrop in the stream bed, though there is a further dolerite dyke just over 100 ft. below the lake giving rise to rapids. The plains southwestward to Junction L. are thickly covered in hummocky ridges of ground moraine, but outcrops in stream beds reveal different rock types of the Permo-Triassic sequence, which undoubtedly underlies most of the plain. Junction L. is dammed by a further dolerite dyke which is responsible for the 40 ft. waterfall a furlong below the lake and for the very clearcut step in the Mersey valley here. Thus, though surrounded by moraine on the east and the north, Junction L. may well be of erosional origin.

Lakes Adelaide and Meston are due then to glacial over-deepening by powerful ice streams transverse to their length. In the case of L. Meston there is a real possibility that differential erosion may be involved; it is quite likely from the observations recorded above that its basin reaches down into the weak Permo-Triassic sediments.[12] Zumberge (1955) relied on obstructed extrusion flow of ice to explain rather similar lake basins, but many authorities doubt whether such flow does occur. Certainly in the case of L. Adelaide there is no need to call in such flows. Here the pre-glacial step in the plateau margin, itself probably due to faulting, would give rise to an icefall; this would promote rotational glacier movement of the type which has been investigated by W. V. Lewis and his Cam-

12. Closer search N.W of L. Meston may reveal Permo-Triassic erratics on the dolerite there and so prove this hypothesis.

bridge co-workers and which provides the most convincing mechanism for glacial rock-basin formation yet recognised (Lewis, W. V. (1949); Clark, J. M. & Lewis, W. V. (1951); McCall (1952); Glen (1956)). L. Meston is more problematic since Mt. Rogoona rises higher than the crest of main plateau east of the lake.

The Degree of Glacial Modification of Relief and Drainage

The preceding discussion of lake orientation has begged the rather intractable question as to how much the pre-glacial relief has survived through the Pleistocene. Some attention must now be given to this question.

A. N. Lewis (1932) attributed the deep valleys of the Mersey and its eastern tributaries, and of the Ouse within our area, to post-glacial river erosion. It is now known that the Mersey system was fully glaciated to a point north of the Arm junction. Glacial spillovers into the Devil's Gullet show that this feature also existed prior to ice invasion though it may have been extended at its head postglacially in some degree. The Ouse valley at the L. Augusta reservoir dam is deeply filled with drift. Moreover, although fairly strong

arguments have been given for post-Newer Basalt faulting (inter alia, Fairbridge, 1948), there does not seem to be any solid grounds for the claim (A. N. Lewis, 1944a) that there is Pleistocene faulting subsequent to an early glaciation of the plateau. In fact the most reasonable assumption is that before glaciation the major elements in the relief of the area of study were fundamentally as they are to-day.

If this assumption is correct, the present-day drainage divides remain much where they were preglacially, the vast bulk of the country draining south-eastwards. There is much complication in detail through glacial erosion and deposition, but broadly the system is a simple one closely related to the structural pattern. It is true that divides have sometimes been eroded almost to the point of disappearance, e.g., between the S.E. drainage of the Ling Roth Lakes and the Mersey drainage. But on the whole there seem few cases of breached divides and drainage diversion (cf. Linton, 1949). It is possible that the Fisher Lagoons drainage preglacially was directed S.E. into the Ouse system and that the ice draining N.W. destroyed a low divide between the Fisher Lagoons and the present Fisher valley running N.N.W. into the Devil's Gullet so

Plate II: Lake Adelaide, a rock basin lake due to transverse glacial erosion. Ice fell from main plateau on left, gouged out the lake basin and overrode lower ridge on right.

that the waters were deflected first E. and then N.N.W. to feed into the Mersey.[13]

A lesser but more certain drainage change is to be found in the L. Adelaide-Meston circumstances described above. The Mersey seems to have risen north of L. Adelaide preglacially, flowed southwards along the present length of that lake, and thence across to the site of L. Meston and so to the present Mersey course. Erosion of the Adelaide basin and of the outlet gap through the ridge to the west has diverted the upper part of the former Mersey headwaters. North of L. Adelaide there seems to have been further diversion of this former drainage line by ice movement and erosion across it, so that the preglacial Mersey has been twice dismembered.

In general, then, the ice conformed to, rather than remodelled completely, the larger relief elements; but some contrasts have to be noted. As a rule, the greater the local relief, the fewer the lakes, but on the average they are larger and deeper, e.g., in the north around L. Nameless and near the Mountains of Jupiter as compared with the more level plateau between. Superimposed on this pattern is a gradient of increasing glacial denudation south-westwards; thus the amount of glacial erosion in the L. Nameless area seems to be much less than in the Mountains of Jupiter. But the biggest contrast is between the two flanks of the ice divide. South-east from the divide, glacial erosion though definite enough was not often great; here were very gradual slopes and slighter precipitation, a region of passive glaciation. The comparison is with the lobate northern margin of the Vatnajökull in Iceland. But on the north-western flank, gradients down into the Mersey valley were steeper and precipitation greater. The glaciation would be more active here in terms of a dynamic classification of glaciers (Ahlmann, 1948) and the steep southern outfalls of the Vatnajökull are more comparable. Here, in a locus of ice congestion, the four lakes—Adelaide, Meston, Myrtle and Louisa—may well prove to be the deepest lakes of the Central Plateau. In the introduction this lake country as a whole was likened to the Finnish Lake Plateau, but this western margin is in fact more like the Norwegian fjell.

These differences reflect once more the old antithesis in the effects of glaciation—here weak, there strong—which was in part the basis of the two schools of thought—glacialist and protectionist—with regard to ice action. Where there are marked topographical features and dynamic glaciers, there is a great deal of erosion; where relief is more uniform and where the supply and wastage of ice is less rapid, the effects may be slight (cf. Tanner, 1944).

The Glacial History

A. N. Lewis (1923, 1932, 1933, 1938, 1944) considered three separate glaciations were represented in the Central Plateau. The earliest or Malanna extended over virtually the whole of the plateau and was an ice-cap glaciation. Soils have had time to reform where not subject to later glaciation and forest cover to re-establish. The Clarence Lagoon area was ascribed to this glaciation. Next followed Yolande glacial period, which was restricted "to the western edge of the plateau" and characterised by a general absence of soil. He does not specify the nature of the glaciation here, though in other parts of Tasmania dendritic glaciers are the most extensive type claimed for it. The third or Margaret glaciation is definitely regarded as a cirque glaciation only, and is thought to be represented on the western edge of the Central Plateau and also on the north at Pine Lake.

The present authors can only recognise one glaciation in the western part of the plateau, and that is an ice-cap glaciation. It is true that there are certain cirque-like landforms (Fig. 1) to be seen: the lake immediately west of L. Malbeena (2334) is hemmed in on three sides by a cirque; the Little Fisher river drops over the edge of the plateau into a good cirque below Mersey Crag; the lake (4559) south of L. Meander also lies in a cirque. But in all the cases examined (with the possible exception of one small and poorly developed example (3373) practically at the top of the Great Western Tiers) these cirques had clearly been completely overrun by the general ice-cover and could well have formed during the advancing hemicycle of the icecap glaciation (cf. Tanner, 1944). Moreover, provided there are nunataks, cirques can form contemporaneously with an icecap. Nor were any of the cirques marked by very recent moraines which suggest reoccupation by cirque glaciers after the wastage of the icecap.

End moraines were seen both in the zone of deposition and within the zone of erosion, but there was no such marked difference in degree of morphological preservation as would prevent them all from being regarded as retreat phenomena of a single glaciation. Moreover the freshness of form and absence of dissection of the moraines, together with the

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13. Capture here is unlikely to be pre-glacial in the light of the absence of dissection up to the supposed elbow of capture.

absence of soil from the eroded areas, make this icecap glaciation very young in the Pleistocene. If it is ascribed to Margaret, it is completely at variance in its intensity with Lewis' views for this area and Tasmania at large. Alternatively if it is equated with Yolande (which because of its extension and intensity would seem more reasonable) Lewis' correlation of the latter with the Riss must be abandoned. The numerous lakes confirm this view. In the N. European Plain, the distribution of lakes provides a first-class indication of the distribution of the Weichsel (Würm) glaciation (Majdanowoski, 1956). Though the synchroneity of Pleistocene climatic history in the two hemispheres must still be regarded as not proven, the geomorphological evidence indicates fairly reliably that the icecap glaciation of this area must have occurred within the last thirty or forty thousand years.

The Contrast Between West and East on the Central Plateau

The question of the contrast between the lakes of the west and east of the Central Plateau can now be discussed.[14]

In his great work on the landforms of Finland, Tanner (1938) found a close correspondence between the numbers and areas of the lakes and the topography. The flatter the topography the greater is the development of lakes. This however does not appear to be the solution here; some areas of many small lakes in the west are fully matched in lack of relief by areas in the east which have one or two large lakes or alternatively no lakes at all.

Is the contrast to be sought in structure? It is true that Tertiary basalts and Mesozoic sediments are more extensive in the centre and east than in the west, with greater possibilities of differential erosion. Voisey (1948a, b) has attributed Great Lake and Arthur's Lakes to the glacial removal of basalt from valleys in the dolerite. However some of the large lakes such as L. Echo lie entirely in dolerite country. And even if differential erosion could explain the presence of very large lakes, it could not explain the absence of small ones.

Implicit earlier in this paper is the idea that the west and east have had different glacial histories and this might be the answer to the problem. No attempt was made to map the limits of the icecap glaciation of the west, though the marginal belt is thought to lie

between the neighbourhoods of Lake St. Clair and Liaweenie. No clearcut outermost line of moraine ridges has been left and tracing the limits of the moraine spreads in this wooded country is not easy. It is clear however that glaciation of the same age does not reach into the country east of the Lake and Marlborough Highways (except possibly in the highest tiers north of Great Lake). If these parts were glaciated, it must have been in an earlier glaciation as Lewis maintained. From such an earlier glaciation the small lakes might have been effaced by sedimentation and only the large lakes survived; there are many marshes in this eastern area which might be the sites of infilled lakes.

This is an attractive solution but it faces difficulties. The fact that subsequently the west alone was extensively glaciated suggests that in the earlier glaciation it would have been the locus of greatest ice accumulation. Present-day climatic gradients support this contention. If so, how was it that the greatest lakes were produced in the centre and east? We have seen that the structural argument of differential erosion does not suffice to explain this. Moreover if the smaller lakes of the east have been entirely lost through sedimentation, the large lakes could be expected to show greater diminution by these processes than they do. In addition, glacial erosion of lakes as large as these would surely result in deep basins at least here and there. The only soundings available are those of Legge (1904) in Great Lake. These do in fact suggest rock bars isolating separate basins, but the soundings are scarcely close enough to eliminate the possibility of valleys crossing them. More indicative, however, are the uniformly flat floors at depths of only 15-20 feet of the intervening basins. This characteristic is strange for glacial rock basins and if it is due to post-Malanna sedimentation, where are the complementary encroaching deltas? A last difficulty lies in the fate of the vast amounts of moraine which must have been excavated from these large lakes if of erosional origin; no investigator of the eastern part of the plateau has been able to point to much undoubted moraine. Lewis (1932) himself had doubts about the glacial origin of the large eastern lakes, though in his case they arose from his idea that the Malanna icecap here was only about 100 ft. thick, the dolerite tiers projecting as nunataks.[15] Therefore Lewis was inclined to call in tectonic forces as at least a supple-

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14. The writers have made no proper field examination of the eastern lake area, and the following comments rely mainly on previous literature with regard to them, fortunately a little more voluminous than on the western lakes.

15. In the light of present knowledge it seems unlikely that there could be a true icecap of that thickness; it would not move (R. F. Flint, 1947).

mentary cause of the large lakes. Fairbridge (1948) goes farther and assigns L. Echo primarily to tectonic subsidence across the fall of the drainage. The arguments presented here support the probability of this alternative mode of origin. The lake legacy of the icecap glaciation of the western part of the Central Plateau thus has not only its intrinsic geomorphological interest but a distinct bearing on the interpretation of the larger lakes of the east.

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RE-EXAMINATION OF THE FJORD THEORY OF

PORT DAVEY, TASMANIA

by

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Port Davey, a large drowned inlet in the extreme south-west of Tasmania, has been considered by Lewis (1924, 1928 and 1945) to be a fjord, the result of Pleistocene glaciation to below sea-level followed by post-glacial drowning. As such it has been cited as the only example of a fjord in Australia, and has been widely quoted not only in Australia but in international reviews (e.g. Flint, 1947; David, 1950; Valentin, 1954; Charlesworth, 1957). Nye, on the other hand, (1930) had described Port Davey as a drowned river estuary.

In the course of recent work for the Mines Department of Tasmania during which we spent six weeks together in the area, the authors were afforded the opportunity of examining the region around Port Davey in some detail. One of us (W.E.B.) had previously spent several weeks in the area. As a result of these studies we support the view of Nye (1930).

The statement by Lewis (1945) is as follows: "The most impressive feature of the Malanna glaciation is Port Davey, a glacial fjord, the only one in the Australian region - not in the same category as the fiords of Norway or New Zealand, but, nevertheless, a true example, with white cliffs rising to sharp tinds some 5200 feet above the deep sinuous channels".

The authors have found no evidence of glaciation at sea level around the shoreline of Port Davey. Not one example of a polished, faceted or striated pebble was found amongst the large number of superficial boulder deposits exposed around the shoreline, nor have

any striated or plucked rock surfaces been found, nor any other
forms characteristic of ice sculpture. Stereoscopic examination of
aerial photographs of the region reveals no sign of glaciation near
Port Davey. Clear examples of glacial erosion and deposition are
evident in the Norold Mountains although the position of the terminal
moraines shows that the limit of glaciation was some 15 miles to the
north-east of Bathurst Harbour.

Bathurst Channel, the alleged fjord, has none of the
attributes of a glacial valley. There are no truncated spurs. There
are no parallel walls dropping steeply to a flat floor, no valley
shoulders. The plan has numerous jagged promontories and inlets,
quite typical of that of a drowned river valley, and unlike that of
a fjord. Whilst a glacial valley may retain a sinuous course during
its development, the highly sinuous nature of Bathurst Channel,
particularly through the relatively soft rocks of Long Bay (Figure 1),
would not have persisted under the conditions of glacial erosion.
Angular projections such as Eve Point and Joan Point would have been
truncated. The cross section of the valley, whether taken between
opposing mountains (Figure 2, Section A-A) or opposed inlets
(Figure 2, Section C-C) reveals the normal "V" shape of the river
valley. Tributary valleys join at normal valley level (Figure 2,
Section B-B) and there is no suggestion of the hanging valleys to be
expected at the junction of minor tributary glaciers with the trunk
glacier.

Figure 3 is an echo-sounding run up Port Davey. This has a gross exaggeration of the vertical scale with respect to the horizontal, and in addition the line of steaming does not follow faithfully the sinuous course of the deepest part of the channel, so care is necessary in interpreting the echogram. Nevertheless it does show that the lateral spurs continue down normally (for river erosion) below the water and show no signs of truncation as would surely have occurred in a fjord.

Characteristic of fjords is the threshold, and "skerryguard" islands are common features (Charlesworth, 1957, p. 343). Breaksea Islands could fit the latter role admirably and there are bar-like shoalings of each channel at the mouth which could be a low threshold. The water shoals to 5 fathoms in the north channel and to 7 fathoms in the south channel (the maximum interior depth is 22 fathoms). But bars of these dimensions are equally characteristic of estuary mouths. There is a substantial catchment of high rainfall country draining through Bathurst Channel, and although the coarse sediments are largely dropped in Bathurst Harbour suspended material tends to be flocculated at the esturary mouth. The mouth then is compatible with the fjord theory but is equally compatible with drowning.

Finally the fjord-making glacier if it existed, can only have come from Bathurst Harbour. This is a broad flat-floored pan several miles across at sea-level (see fig. 4). To feed the fjord glacier, this would have to have been filled by an ice-cap. Now this ice-cap would have had equally easy outlet via Melaleuca Inlet to

Cox's Bight. This too would have been occupied by a distributary glacier of at least comparable dimensions to that in Bathurst Channel. There is no evidence to suggest a glacial valley or fjord here, and mu to the contrary.

CONCLUSIONS

Since Lewis did not, to our knowledge, publish any evidence to support his view, there is no existing case to be answered. As a result of the total absence of positive evidence of glaciation, and the strength of the evidence presented above inconsistent with glaciation the authors do not agree with Lewis's view, nor generally accepted, that Port Davey is a fjord. We conclude that Port Davey, Bathurst Channel and Bathurst Harbour are a drowned river estuary, the drowning having caused dismemberment of a fairly large river system.

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BATHURST HARBOUR AND BATHURST CHANNEL
showing section lines

5 FATHOM LINE
10 FATHOM LINE

SCALE OF MILES

0   1   2   3   4   5
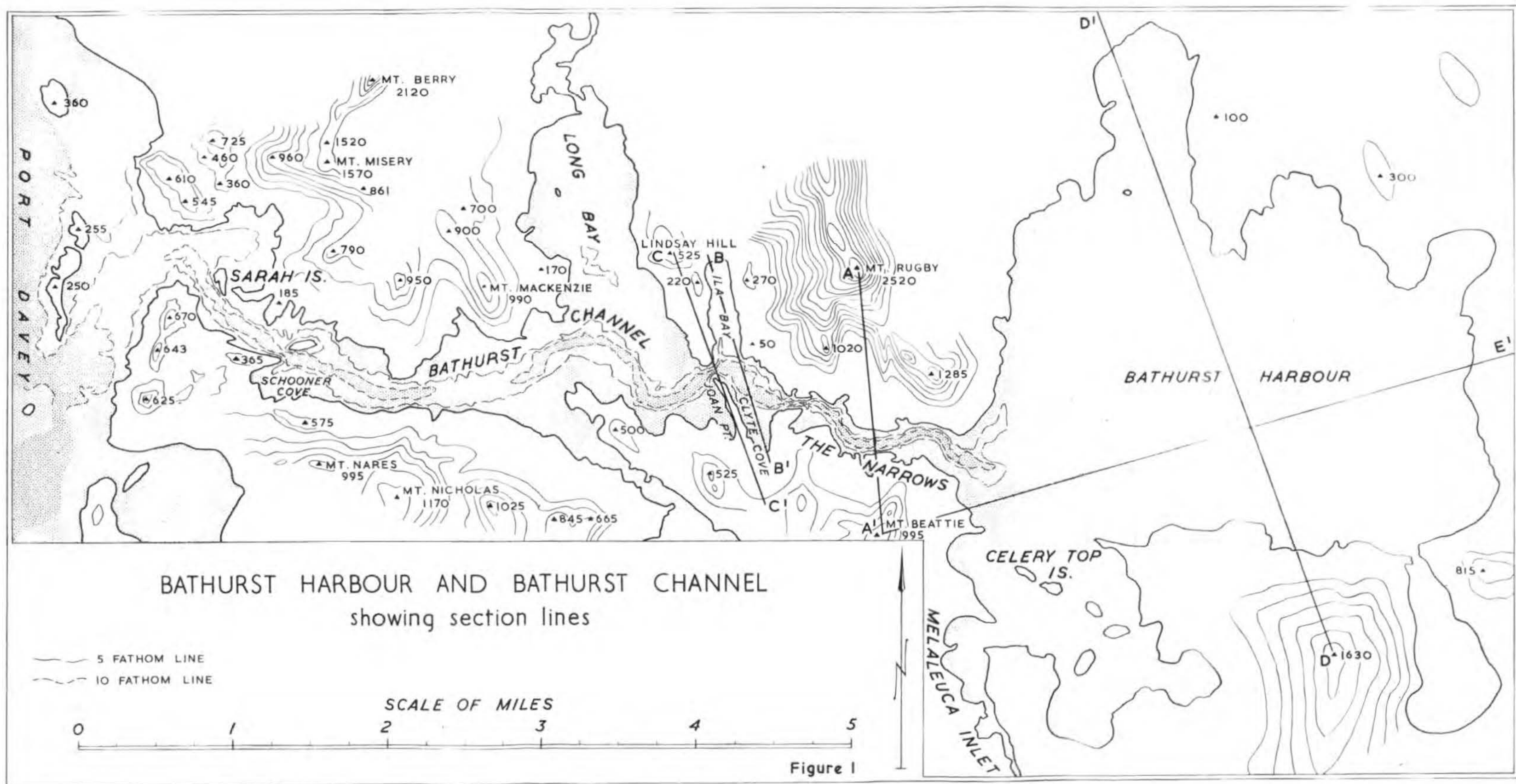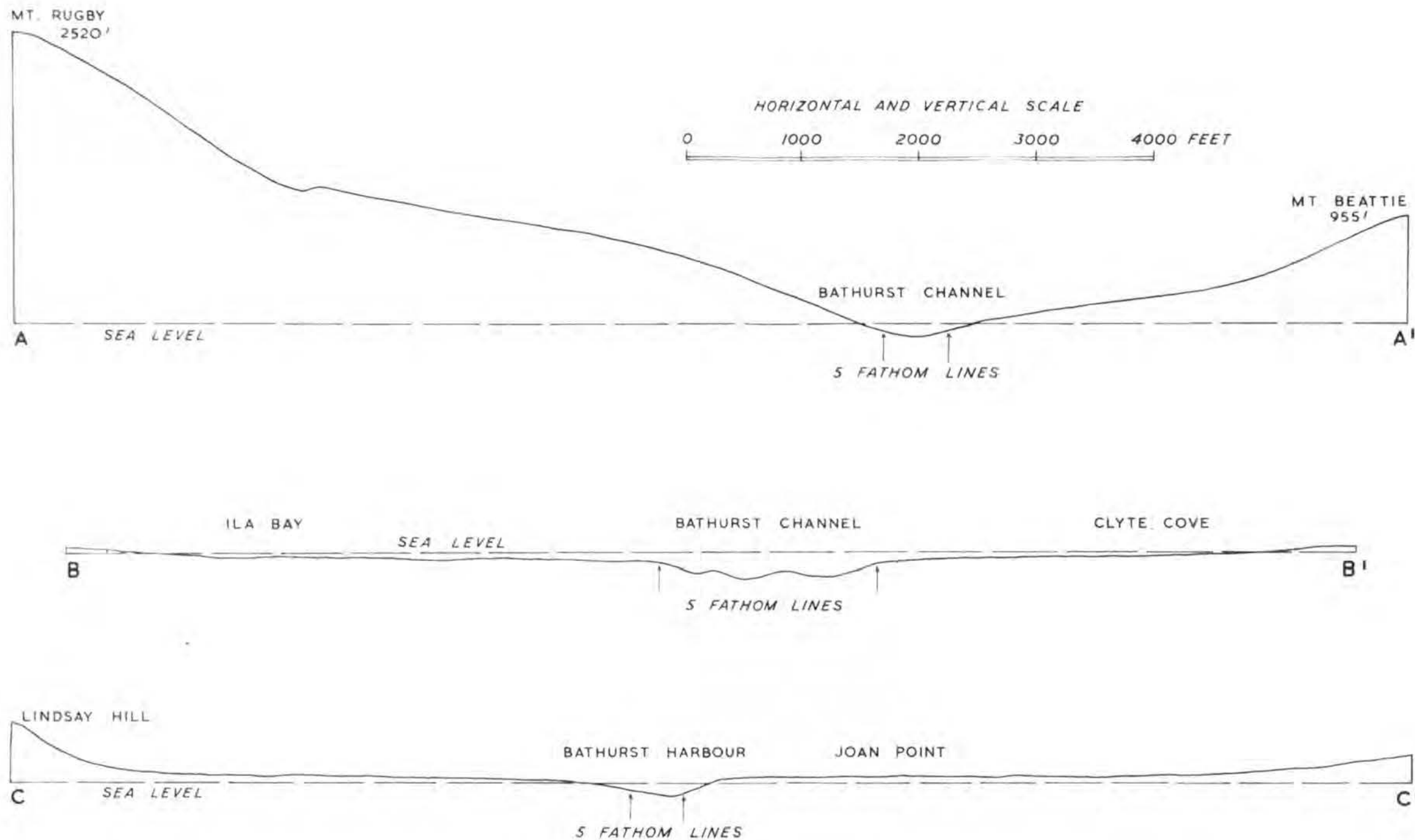
Figure 1

Figure 2

CROSS SECTIONS OF BATHURST CHANNEL AND ADJACENT TOPOGRAPHY

D   SEA LEVEL   BATHURST   HARBOUR   D'

MT. BEATTIE   SEA LEVEL   BATHURST   HARBOUR   SEA LEVEL

A'   E'

VERTICAL AND HORIZONTAL SCALE

O   I   2 MILES   Figure 4

FATHOMS

SARAH IS.

SCHOONER COVE

Entrance LONG BAY

ILA BAY

THE NARROWS

BATHURST HARBOUR

CELERY TOP IS.

MELALEUCA INLET

APPROXIMATE HORIZONAL SCALE

MILES
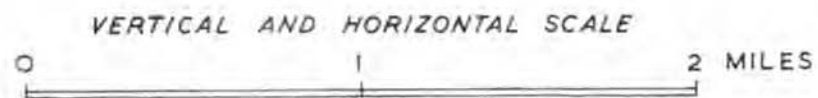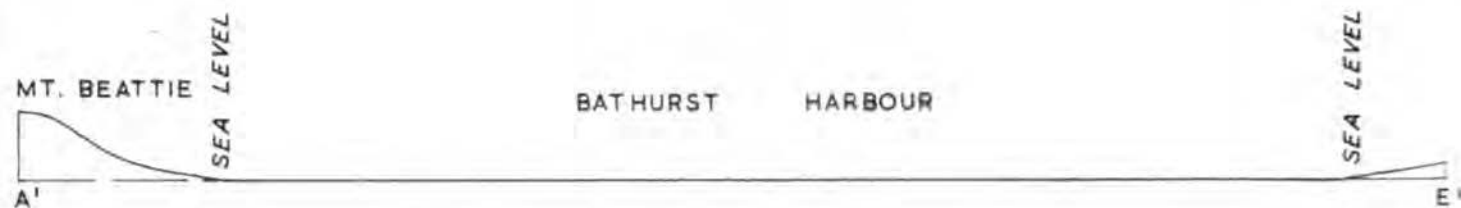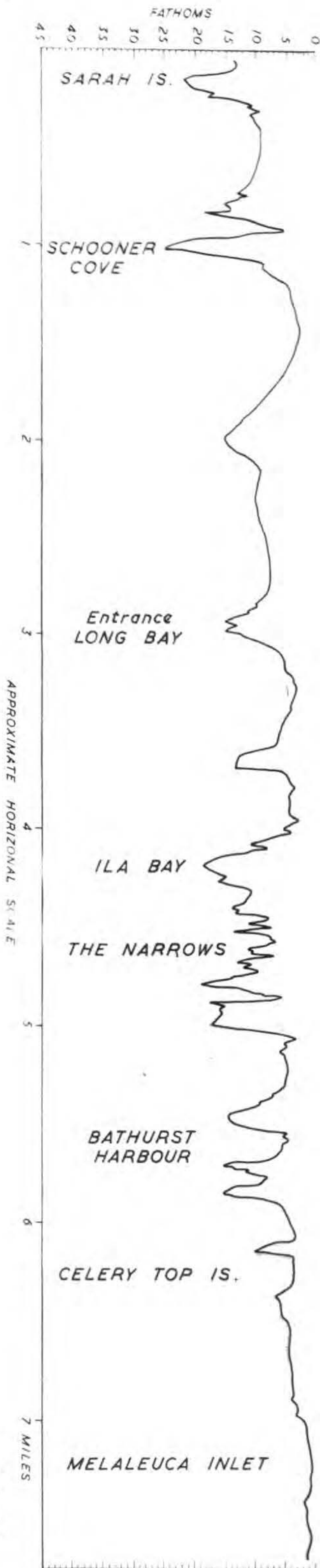
Figure 3

THE GLACIATION OF THE KING VALLEY

WESTERN TASMANIA.

by

N. AHMAD,

H.A. BARTLETT,   and

D.H. GREEN.

ABSTRACT.

The King Glacier resulted from the join-
ing of two major glaciers flowing in the North Eldon
and South Eldon Valleys. These were fed respectively
by ice from the east side of the Tyndall Range and
from the Eldon Range.

At its maximum the King Glacier was over
1000' thick at the eastern end of Mt Lyell but from
there decreased rapidly in thickness so that 4 - 5
miles south its thickness was of the order of 100'.
The rapid change in thickness was probably mainly due
to loss of ice to the Linda Valley and Nelson Valley
distributaries and to the change in valley width from
about 1 mile to almost 3 miles.

The glacier reached the northern part of
the Crotty Plain at some stages but for a longer period
remained barred behind a quartzite ridge crossing the
valley about 3½ - 4 miles south of Linda Valley entrance.
No large terminal moraine was formed but there are very
extensive outwash deposits both north and south of this
ridge. The distinct moraines of the Linda Valley and
Nelson Valley were deposited during early retreat stages.

It is considered that in this area there is
no evidence of three glacial stages but rather the evid-
ence suggests that the Margaret Stage and Malanna Stage
(as represented by the King Glacier (Lewis 1936)) are
contemporaneous. Accepting Gill's (1956) radio-carbon
dating, then the single glacial advance and retreat is
roughly equivalent in time to the Wisconsin Glaciation
of North America.

THE GLACIATION OF THE KING VALLEY - WESTERN TASMANIA

Introduction:-

The existence of glaciation in the West
Coast Range area has long been known. (Sprent (1886),
Montgomery (1891), Power (1892), Dunn (1894), Johnston
(1894), Moore (1894), Officer, Balfour and Hogg (1895),
Moore (1896), and Gregory (1904)). Of these many
reports those of Dunn (1894), Moore (1896), and Greg-
ory (1904) are of the most value. The first two of
these discuss the glaciation of the Tyndall-Sedgwick-
Dora region and Gregory, in reporting on the Linda
Valley area, concluded that the glacial deposits of
the Linda Valley came from a distributary of a larger
glacier in the King Valley. This view has been supported
by the work of Bradley (1954) and Carey (1955) but no
actual study of the King Valley glaciation and espec-
ially of the limits of the glacier has been previously
reported.

The present study consisted of a week's
field work in the southern part of the King Valley and
in the Linda and Nelson Valleys and a careful study of
aerial photographs of the West Coast Range and Eldon
Range area. Two of the authors (Ahmad and Bartlett)
have also visited the areas around Mt. Sedgwick and
Lake Dora.

PHYSIOGRAPHY: -

The King Valley runs almost north-south on the eastern side of the West Coast Range and about 20 miles from the western coastline. The valley is crossed by the Lyell Highway about seven miles from Queenstown and there is a vehicle track running south to Crotty Plains where the King River turns west to flow in a 1000' gorge between Mts. Huxley and Jukes.

The area which must be considered in a discussion of the King Glacier can be conveniently divided into 4 physiographic units:-

(a)   The western margin of the Central Plateau - a strongly dissected but topographically youthful area extending west and southwest from near Castle Mt. and Rocky Hill.

(b)   The Eldon Range - an east-west range which changes from plateau-like near Eldon Bluff and Castle Mt. to the sharp Eldon Peak at its western end.

(c)   The West Coast Range - a series of rugged peaks with Mts. Tyndall (3800'), Geikie and Sedgwick (4000') in the north forming the western and southern margins of the Dora-Spicer plateau. South of Mt. Sedgwick the range is broken by the broad Comstock Valley (about 1000' above sea level) and again on the south of Mt. Lyell by the Linda Valley (1000'). South of this valley are the several peaks of Mt. Owen (3600') and further south-east the Thureau Hills form an east-ward extension of the range.

(d)   The King Valley - the North Eldon Valley between Eldon Peak and the West Coast Range, and the South Eldon Valley on the southern side of the Eldon

Range, join southwest of Eldon Peak and from there the King Valley extends south for a further 8 miles as a broad, (1 - 3 miles), flat-floored valley in which the King River pursues a meandering course.

East of the Thureau Hills the river passes through a short but narrow gorge before emerging into the Crotty Plains.

The Princess River and Bull Creek are small streams in wide valleys separated from the King Valley north of the Lyell Highway by Princess Ridge (1435') and its northerly continuation. The Nelson River flows west parallel to the Lyell Highway until the latter crosses a 1030' divide to the wide Princess Valley. The Nelson river, however, turns sharply south and finally joins the King River east of the Thureau Hills. In its lower part it is separated from the King by a range of hills about 1500' high.

It may be stressed here that the main control of the physiographic features is the geological structure. The King, Linda, Comstock and Nelson Valleys all owe their existence primarily to the presence of relatively soft Gordon Limestone or Eldon Group sediments. Modifications of the physiography which are due to glacial action are present to a varying extent.

THE SOURCE OF THE KING GLACIER.

The King Glacier resulted from the junction of glaciers flowing in the North Eldon and South Eldon Valleys. The main collecting ground for the ice in the North Eldon Valley was in the Tyndall Range to the west. A substantial ice field on the east of the Tyndall Range between Mts. Tyndall and Geikie gave rise to glaciers which dominantly moved eastward across Lake Dora and Lake Spicer to spill over the plateau edge into the North Eldon Valley. The western wall of this valley is an abrupt but moulded and smoothed slope capped by morainal material. A tributary of the North Eldon River, rising just north-east of Lake Dora flows eastward through a distinctly U-shaped valley and this was probably one of the principal feeders of the North Eldon Glacier.

From the north side of Mt. Tyndall ice probably also moved northward through Lake Rollesten. The Hamilton Moraine resulted from ice originating probably on the east slopes of Mt. Geikie and flowing southwest through Lake Margaret. There was probably no great quantity of ice moving off the plateau in this direction, but, coming from the highest parts of the plateau, and falling sharply on to the Henty Peneplain, the terminal position of the glacier probably stayed essentially constant over a long period of minor fluctuations recorded in the King Glacier and even perhaps until after the retreat of the latter to the plateau edge. This would account for the magnitude of the Hamilton Moraine as compared to the magnitude of the moraines in the King Valley.

The area around Mt. Sedgwick is problem-
atical as it is not known whether this acted as a sep-
arate source area or whether the main ice flow from the
north flowed past Sedgwick and spilt south into the
Comstock Valley. Ice also moved southeast to join the
King Valley by way of Lake Beatrice.

There are no cirques of any size on the
western side of Eldon Peak but two glaciers from the
north side of the peak moved north-east and have left
very large lateral moraines. These glaciers meet the
valley of a tributary of the North Eldon (flowing north
west here) at right angles but do not appear to have
flowed down this valley but rather to have melted at
its edge.

From the southeast of Eldon Peak several
glaciers fed into the South Eldon Valley. From the El-
don Range some ice moved south into the South Eldon
Valley and some north into the Canning River. From
Eldon Bluff a rather larger glacier moved sout-east
and then south-west along Danube Creek to join the South
Eldon Valley. There is an ice divide running approx-
imately from Castle Mt. to Rocky Hill. East of this,
ice from High Dome moved into the Canning River and
thence to the Murchison River. From the southern side
of the South Eldon Valley several glaciers came from
the vicinity of Last Hill to flow into the South
Eldon Valley.

The South Eldon and North Eldon glaciers
seem to have been of similar size by the time they
joined to form the King Glacier but the glacier flowing
through Lake Beatrice was probably somewhat smaller.

FEATURES DUE TO GLACIAL EROSION.

The walls of the King Valley show evidence of glacial erosion only in the presence of truncated spurs and occasional striated pavement (as at King Bridge (Carey, 1955)). On the floor of the valley there are a number of roches moutonnées, the most prominent of these being some 2 miles south of the Lyell Highway and east of the Crotty Track. A further 1½ miles south a quartzite ridge (1030' max. height) stands sharply from the flat depositional plain and trends across the valley. The ridge is breached at the eastern end by the King River, here flowing on basement rock and also, near the centre of the ridge, by a dry, U-shaped valley.

About ¾ mile further south the 70' gorge above the Crotty Plain has a V-section but the eastern and lower wall is of glacial deposits. A section showing the distribution of bedrock could well be U-shaped and the V-shaped valley cut in a depositional flat to the east, be due to later river action.

The Gormanston Gap at the western end of Linda Valley is U-shaped for a distance of several hundred yards on the western side of the crest but beyond this it exhibits a sharp V-section.

GLACIAL DEPOSITS:-

Glacial deposits north of Mt.Lyell were not examined in the field. An examination of the aerial photographs show that there are probably small terminal moraines marking retreat stages of the South Eldon glacier located in the river valley south and south-east of Eldon Peak. On the plateau edge east of Lake Dora and particularly Lake Spicer there also appears to be an irregular terminal moraine. This would mark a late retreat stage of the North Eldon glacier when the ice melted just prior to the sharp descent into the North Eldon Valley. There are a number of other moraines on the Eldon range on the area around and east of the Tyndall Range, and south of Mt. Sedgwick. These are of somewhat lesser importance in this study although it should be noted that the Hamilton Moraine, west of Lake Margaret, is the most impressive in the area. At the western end of the Comstock Valley are morainal deposits somewhat similar to those near Gormanston in the Linda Valley.

In the Linda Valley till occurs to a height of 1400' near Gormanston and varves and till lie on basement rock at 900' near Linda township. Between these two there is continuous till, outwash and varves. A typical part of the section exposed in a gully near the Linda-Gormanston road is as follows:-

1090' - 1080'	Till.
1080' - 1050'	Bedded Till - Outwash.
1050' - 1020'	Varves.
1020' - 1015'	Till.
1015' - 960'	Varves.
960' - ?	Till.

The interbedding of varves and till seems to represent successive advances and retreats of the ice. The deposition of till and varves from a height of 1400' down to 900' and at a maximum thickness of 100' - 200' indicates an overall retreat downhill of the glacier so that successively lower and northeast-ward (downhill) deposits are progressively younger.

Till, varves and outwash deposits at the mouth of the Linda Valley indicate another retreat pos-ition of the ice. These are actually lateral moraines of the King Glacier rather than terminal moraines of the Linda distributary. North-west of this erratics and till occur to a height of 1700' on the eastern slopes of Mt. Lyell. This is 1000' above the present valley floor at the King Bridge. Benches occur on this slope at 1700', 1660', 1560', 1220', 1120'. These benches are apparently of morainal material and probably are small lateral moraines. Many have depressions, now filled with gravel behind them.

East of this, erratics were found at a height of 1440' on the summit of Princess Ridge. Fur-ther east still, dolerite erratics occur to a height of 1160' north of the mouth of the Nelson Valley. At the entrance to the Nelson Valley the Lyell Highway crosses two end moraines. The most easterly of these is at 1050' and is semicircular and concave towards the Princess flats. Till does occur a little further up the Nelson Valley (height 1015') but no distinct end moraine exists. The Nelson River prior to the glac-iation probably flowed westward, approximately following the present Lyell Highway, to join the Princess River before the latter joined the King River. The end

position of this easterly distributary of the King
Glacier and the moraine it built up, caused the div-
erting of the Nelson River to the south so that it is
separated from the King River by a range of hills for
a distance of about 4 miles and finally joins the King
River just before this enters the Crotty Plain.

A second and later end moraine occurs
parallel to and slightly west of the above mentioned
moraine. It is at a height of 965'. The moraines
consist of till (dominantly Jurassic dolerite) inter-
bedded with current bedded sand, gravel and some clay
beds.

Two to three miles south of the mouth of
Linda Valley the 1040' roche moutonnée mentioned before
is flanked at its northern end by a very low morainal
ridge rising about 5' above the flat plain and apparent-
ly extending right across the valley. This is most
probably a small terminal moraine of the King Glacier.
The plain extending from here to the quartzite ridge
crossing the valley (3½ miles south of the Linda Valley
entrance) is probably an outwash plain formed between
the retreating glacier and the quartzite barrier ridge.
The roche moutonnée has erratics to at least 1030' though
none could be found between 1030' and 1040'.

The quartzite ridge crossing the valley
about 3½ miles from the Linda Valley entrance has a
surface veneer of till to a height of 940' at the
western end of the ridge. There is no evidence of an
extensive terminal moraine in this area and a moraine-
like ridge immediately south of the point at which the
track cuts through a gap in the ridge is probably

solid with only a thin veneer of moraine. Lateral
moraine occurs on the lower slopes of the Thureau Hills.
There are several terraces to the north-east of the
ridges but these were probably cut by earlier courses
of the King River and its tributaries. The dry U-
shaped valley cutting through the centre of the ridge
has a fan-like deposit at its southern end - this seems
to be outwash material.

The small plain between the quartzite bar-
rier ridge and the gorge north of the Crotty Plain has
several terraces cut in it, presumably by old courses
of the King River and tributary creeks. The plain is
dominantly of outwash material. One section showed sand
and clayey sand with some boulders and pebbles, rest-
ing on basement rock. A section by the vehicle track
showed sandy varves. Generally the character of the
plain material was that of glacial outwash with little
actual till. It is considered that these deposits
formed in a shallow lake between the quartzite barrier
ridge to the north behind which the glacier lay, and
the moraine dammed gorge north of the Crotty Plain.
The till and outwash of the Crotty Plain show that the
glacier at one time reached there, and during its re-
treat it probably stayed just north of the gorge long
enough to deposit an end moraine in that area.

THE RECONSTRUCTION OF THE KING GLACIER.

From the junction of the North Eldon and
South Eldon Valleys, the King Glacier travelled a
further 8 - 10 miles south at its maximum extent as
far as the gorge north of the Crotty Plain and at times
even on to the plain itself. In its northern part the
King Glacier was very thick and the presence of low,
wide valleys to the west and east resulted in several
distributary lobes moving into and melting within these
valleys. The Comstock valley was probably filled with
ice in this manner and morainal material deposited at its
western end. A complicating feature however was the
ice moving south into the valley from near Mt Sedgwick
and at times the ice could have moved out of, rather than
into, the Comstock Valley.

The presence of erratics at 1700' on the
eastern end of Mt. Lyell, a height of 1000' above the
valley floor, indicates that the glacier in this area
was at least 1000' thick. This is supported by the
fact that ice over-rode Princess Ridge (1440') east of
Mt Lyell. The floor of Linda Valley rises from 800'
at the mouth of the valley to 900' near the Linda town-
ship and 1360' at Gormanston Gap. Erratics occur to
1700' at the valley entrance (north side) and to 1400'
on Owen Spur south of Gormanston. The thickness of ice
would have been over 1000' at the valley entrance, over
500' above Linda township and over 40' at the Gormanston
Gap. The fact that, at a short distance east of the
Gap, erratics occur to a height of 40' above the level
of the Gap, suggests that it is likely that some ice
spilt over the Gap and rapidly melted on the steep slope

of Conglomerate Creek. The U-shape of the Gap supports
this but the sharp V-section of the Conglomerate Creek
favours the idea that only meltwater and not ice flowed
down this creek. During the retreat of the ice from
this maximum extent the large Gormanston Moraine and
the lesser moraines at the mouth of the Linda Valley
were deposited.

Since the ice thickness in this vicinity
has been shown to be of the order of 1000' the Princess
Ridge (1440') would be crossed by about 260' of ice
moving towards the Nelson Valley. Ice would possibly
also cross the divide between the Princess watershed
and the South Eldon - King watershed at points higher
up the valley. Also the lower levels of ice would no
longer be barred from the Princess Valley south of the
end of Princess Ridge so that ice may have eddied back
up the Princess valley from near the King-Princess
junction. This is unlikely during maximum glaciation
as the southward movement of ice having over-ridden the
Princess - King divide would probably be sufficient to
prevent such a back eddy.

The net result was that ice moved towards
and into the Nelson Valley. Since erratics on the
north side of the valley entrance only occur to 1160'
and the valley floor is about 900' the thickness of
ice at the valley mouth would be something over 260'.
The ice did not move any great distance up the valley,
but two relatively long-lived retreat stages caused
deposition of two terminal moraines at the entrance to
the valley. Since the valley floor at the Princess
Ridge is 705', the ice thickness there during maximum

glaciation would be over 465'. This is compatible with
the ice thickness in the King Valley of 1000' if it
represents the thickness of ice crossing the slightly
lower northerly extension of Princess Ridge.

South of a line running through the Linda
Valley, Princess Ridge and the Nelson Valley, the King
Glacier rapidly decreased in thickness. This is to
be expected as at this point the glacier emerged from
a valley about 1 mile wide into a valley almost 3 miles
wide. Another important factor was the loss of ice into
both the Nelson and Linda distributeries.

At a point 2 - 3 miles south of the en-
trance to Linda Valley erratics occurring to a height
of 1030' on a roche moutonnée indicate that the ice
was at least 230' thick - this figure is based on
the height of the plain on the western side of the
hill (900') and estimating about 100' as the thickness
of glacial deposits on the valley floor. Similarly
the erratics at 940' on the ridge 3½ miles south of
the Linda Valley entrance (valley floor at 830') indic-
ate a thickness of ice of not less than 170' - 220'
allowing 50' - 100' as the thickness of till.

It is improbable that ice ever passed
between Thureau Hills and Mt. Owen, as the northern
entrance to this valley is barred by a solid rock
ridge (960') which has till on its north face but
none on the crest or southern face.

Further south where the King River passes
through the short gorge above the Crotty Plain, the
upper limit of till at 740' above a rock basement at
650' indicates an ice thickness of about 100'. The
presence of till and outwash on the northern part of

the Crotty Plain shows that at the maximum glaciation, and probably only for a very short time, the ice passed beyond this gorge. However for the greater part of its history it is thought that the ice remained behind the quartzite barrier ridge about 3½ miles south of Linda Valley entrance, sending small tongues of ice through the lower parts and depositing outwash material south of the ridge.

From this position - probably corresponding with the period of deposition of the Gormanston Moraine and the higher Nelson Moraine - the ice retreated, irregularly depositing a small terminal moraine near the 1040' roche moutonnée and forming a broad, flat outwash plain north of the quartzite barrier ridge. The lower Nelson moraine, the lower levels of the Gormanston moraine, the moraine at the entrance to Linda Valley and the several lateral moraines on Mt. Lyell all reflect retreat stages of the glacier, but it is not possible to correlate these stages in different areas.

AGE OF THE GLACIATION.

The authors have found no evidence to suggest more than one glacial phase in the whole of the King, Eldon, Lake Dora and Lake Margaret areas. Rather they consider that the whole observed features can be explained as a single, though fluctuating, advance and retreat of the ice with the various depositional features recording stages of that retreat.

It has been established that the Gormanston Moraine was formed as a terminal moraine of a distributary of the King Glacier (Bradley 1954, Carey 1955 and this paper). Lewis (1936) considered this glaciation to belong to the Malannan Stage (the oldest of his three stages), whereas the Hamilton Moraine belonged to the Margaret Stage (the youngest of the stages). The authors contend that the Margaret Glacier (which deposited the Hamilton Moraine) and the King Glacier were both formed during a single glacial stage.

Gill (1956) has reported the result of the radio-carbon dating of fossil wood from varves in the Gormanston Moraine. Unfortunately the exact locality from which this wood was obtained cannot be ascertained. If, as seems likely, it came from a locality rediscovered by one of the authors (H.A.B.) several hundred yards upstream from the Linda Hotel, then the deposit would seem to have been formed during the initial advance of ice up the valley. The sequence of beds suggests that a forest became a swamp, then a shallow lake and then covered by a layer of till.

Presumably ice then moved over this till to reach the higher land near Gormanston and later till and varves were deposited during the retreat of the ice.

Gill (1956) reports the result of the radio-carbon dating as 26,480 ± 800 years which corresponds to the beginning of the Wisconsin glaciation of North America. If the fossil wood is actually from the locality described above, then it appears that the dating of the King Valley glaciation is 26,480 ± 800 years for the initial, advancing phase.

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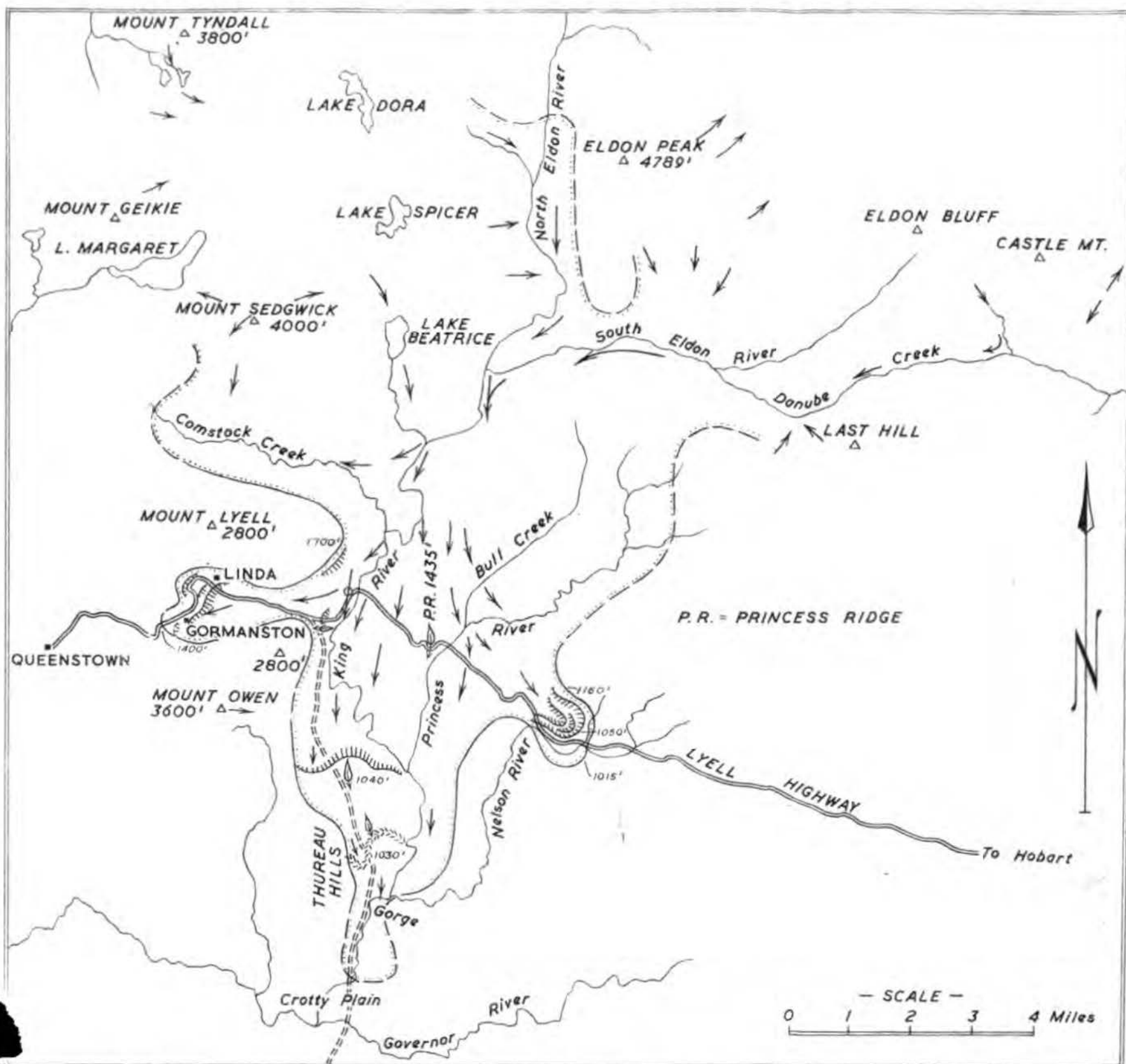
GLACIATION OF KING RIVER VALLEY

King Valley Looking North .

Eldon Peak In The Centre And N. and S.

Eldon Valley On The Two Sides Of It.

Princess Ridge.

Eldon Peak and the Two Eldon Valleys

In the Background.

King Valley Looking South, and Linda
Valley Mouth In the Foreground and
Roche Moutonnees and Quartzite Ridge
In the Distance

Looking North From The King Gorge
The Outwash Plain.

Princess Ridge.
Looking From South-West.

Looking Along Conglomerate Creek
Mt. Owen In The Background.

MERSEY VALLEY GLACIATION.

BY

N. AHMAD

The Mersey River and its tributaries, Arm, Fisher, Little Fisher and Fish rivers, originate along the western margin of the central plateau. The Fish joins the Mersey on Howells Plain, about one mile south of the rapids in the Mersey River north of Walter's Marsh, the Little Fisher joins the Fisher north of Dublin Plain, and the Fisher joins the Mersey near Parangana. The Arm originates on the February Plains and joins the Mersey north of Maggs Mountain.

Physiography.

The Mersey River flows out of Lake Meston on the Central Plateau; for about two miles it flows southwest but near the Upper Mersey Falls it takes a U turn and starts on its northerly course to enter Howells Plain. In the short distance of six miles the Mersey loses more than 2000 ft., and on Howells Plain it meanders until it reaches the lower Mersey Falls where its valley is restricted in width to about 500 ft.

The Fisher River rises from Lake MacKenzie on the plateau and flows through a narrow valley where it has cut a deep gorge. At its source it has an east to west course, but after it is joined by the Little Fisher it takes a north westerly course.

The Little Fisher flows down from the plateau as a small stream. At the beginning it flows through a small semicircular plain which is bounded by high plateau walls on all sides, except for the gap through which it runs out to join the Fisher.

The Fish drains a small glaciated valley south

of Clumner Bluff and flows down from this hanging valley as a waterfall to join the Mersey on the west.

The Arm river originates on the wide, glaciated February Plain and flows northwards through a narrow valley to join the Mersey north of Maggs Mountain.

Evidence of Glaciation

The Central Plateau has been widely glaciated, and much of it was probably covered by an ice cap during part of the Pleistocene. The general slope of the plateau is from north-west to south-east, and along the western margin it drops steeply down towards the west and north. During the last stages of the glaciation the ice divide was probably somewhere along a line running through points a mile or two south of Lake Meston and east of the Walls of Jerusalem. In this region there is a large number of rock basin lakes. The extent to which the relief of the plateau is inherited from pre-glacial times or the work of the ice is difficult to determine, but it can be safely said that at least along its western margin ice followed older river valleys and moulded their floor into wide glaciated valleys. All the rivers of the area except the Fisher, have well pronounced broad U-shaped valleys on the top as well as at the foot of the plateau. The Mersey, like others, has a wide U-shaped valley up to the Maggs Mountain, but north of it the valley changes into a V-shape. This change is so gradual that a precise line cannot be given to separate the two sections. South of Maggs Mountain there are no spurs in the valleys, but north of it, truncated spurs appear, and beyond Parangana spurs are quite common. Flowing through the Devil's Gullet the Fisher River appears to have a V-

shaped valley but the profile section of the valley
shows a crude U shape.

Along the eastern side of the Mersey, between
Howells Plain and Dublin Plain (42050 E, 86150 N),
roches moutonnées trending from south-east to north-
west were observed, on Precambrian quartzite and schists.
These features are elliptical in shape; where-ever they
have developed near the top of the hills on flat surfaces
they are smooth at both ends, but whenever they have
developed near the edge of the hills the western end
is plucked and they have the shape of roches moutonnées.
These roches moutonnées are on a surface which approx-
imates to the ripped pre-Permian surface but they are
probably Pleistocene in age, as the direction of the
ice movement indicated by them is very nearly opposite
to the inferred direction of the Permian ice sheet,
from the west and north of this region, and where thin
layers of Permian rocks still cover the Precambrian beds,
these features do not show up either on the surface or
in the section.

Broad U-shaped hanging valleys come down from
the plateau all along its western edge and along the
northern edge towards Mole Creek. One of these runs
along the southern side of the Devil's gullet over-
flowing where the Little Fisher joins the Fish and
another is at right angles to the Gullet, running from
the western Bluff and overflowing on the northern side
of the same junction.

About two miles from the H.E.C. Hydrology hut
along the road to Howells Plain Hut, (855 YN, 418 YE)

a good striated pavement in schist was found with north-south striae. This is the only instance where striations were found in the area.

Depositional evidence.

All the valley floors of the region are covered with till. The thickness of the till could not be determined as the base is not exposed, and possibly a large quantity of the till has been removed by the rivers since the end of the glaciation. In the Fish, the first exposure of the till is seen at a height of 2210 ft., and above 2260 ft. varves were found interbedded with till.

In the Little Fisher valley till extends down to the junction with the Fisher, and at several places the exposed thickness is over 30 ft. The absence of the till below the junction is possibly due to post-glacial erosion, by the Fisher.

In the Arm valley till is exposed at several places in the river bed near the huts (863 YN, 415 YE), and it is overlain by varves, and end moraines.

In the Little Fisher plain a set of three end moraines was examined. These moraines are 25 ft. high and they run perpendicular to the course of the river. The river has cut along the east of the first moraine, but the other two have been eroded along their western ends. These moraines contain sand, dolerite boulders and they also include varves. These are the best example of end moraines seen in the area.

In the Arm valley the moraines are not well developed, but on the leeward side of the moraines a 5 - 10 foot deposit of varves is to be seen in the river bed west of the southern hut.

On Howells Plain east of the hut (415YE, 855 YN) and on Dublin Plain a mile and a half south of the hut, (423 YE, 8645YN), some deposits were seen which could be the renmants of an end moraine. A ridge of dolerite boulders runs across Dublin Plain, and on Howells Plain two mounds of boulder, clay and sand occur. These heaps are more than 20 ft. high and may have been part of an end moraine.

In general the till is mostly composed of dolerite boulders, and erratics of dolerite can be seen on the surface of Berriedale, February and Dublin Plains and along the banks of the Mersey, on bedrock of basalt, dolerite and schist respectively.

On the Borradaile Plain dolerite erratics are on basalt bedrock, and along the banks of the Mersey on quartzite and schists. Some of these erratics have been transported for several miles and are more than 25 ft. across.

Direction of Ice movements.

With all this evidence, there can be no doubt about the glaciation of the area but at the same time one is faced with the fact that it indicates two diff- erent directions of ice movements. Roches moutonnées on the east side of the Mersey, the evidence from the western margin of the Plateau (Jennings and Ahmad, 1957), the glacial features north of the February and Borra- daile Plains and the occurrence of the dolerite erratics indicate south east to north west direction of the ice movement, but the striations in the Mersey valley, and the east-west trend of the moraines in the Little Fisher and February Plain, indicate south to north direction of

the ice movement.

The two different directions of the ice movements may be explained in different ways:

1. When the whole area including the Mersey Valley was covered by the ice cap the upper layers moved from S.E. to N.W. and the lower layers moved from south to north.

2. The area was first covered by an ice cap which moved from S.W. to N.E., which later retreated or it melted away, during an interglacial period, and was followed by glacial advance in which the ice moved from south to north.

3. On the plateau south of Howells Plain and on higher level in the Mersey Valley, the direction of the ice movement was from south-east to north-west but in the lower parts of the valley the direction of the flow was governed by the valley walls, so it was from south to north.

The general distribution of the erratics on Borradaile and February Plains, and the direction of the ice flow as indicated by the structures north-west of Lake Myrtle and Lake Louisa, indicate that in this part the ice was not only entering the Mersey Valley but that it was also extending towards Mt. Pillinger. In Howells Plain, Walters Marsh, and Dublin Plain, the ice was flowing down from the plateau and its movement for some distance on the plain was from south-east to north-west. In the plains the ice probably melted more quickly, and the valley was deeper than farther south, so it could not extend towards north-west, and the western walls of the Mersey Valley diverted it towards the north. The small valley (86550 N, 42230 E.) north

west of Dublin Plains hut and the roches moutonnées
along the east side of the river indicate that up to
this part of valley the ice had maintained the north-
west direction of the flow, and the striations along
the west side of the river indicate that near the
western wall of the valley the direction had changed,
so that it flowed towards the north.

North of Howells Plain the thickness of the ice
on the plateau was probably less than in the south. In
this part separate ice centres had developed (Jennings
and Ahmad 1957) and the direction of the ice movement
was also changed, so that it flowed from south to north
(along the northern margin of the plateau), north-
east to south-west (from Western Bluff) and from west
to east (on the two sides of the Devil's Gullet and
along the east side of the Little Fisher Valley). If
in this area the ice had been thick enough to fill the
Mersey Valley, it would certainly have gone across
the Devil's Gullet towards Western Bluff, and would
not have moved parallel to it.

It appears that the ice from the plateau was
able to cross the Mersey Valley in the south but in the
north the ice was possibly not thick enough to cross
the valley, so it moved northwards. To the east of the
Mersey river the ice was moving down from the plat-
eau and carved the roches moutonnées with south-east
to north-west axes, but west of the river the ice,
guided by the valley, gradually changed the direction
of movement and moved from south to north.

In the upper parts of the Little Fisher Valley
some of the hanging valleys face towards the east, and

the end moraines run from the east to west which sug-
gests that, although from the west side of Clumner
Bluff ice was moving towards north-west, in the east
it flowed down on the Little Fisher Plain and followed
the older valley.

All the above evidence suggests that possibly
while in some parts of the Mersey Valley the ice was
moving from south-east to north-west, in other parts
it was moving from south to north, and possibly these
were not due to two separate ice advances or glaciation.

Lewis (1944) proposed that during the second
glacial stage Yolande, the end moraines were deposited
above 2200 ft. In this area all the Arm Valley end
moraines were deposited at 2000 ft. height, and the
Little Fisher Valley end moraines were deposited at
2600 ft. above sea level. If Lewis' basis of classif-
ication is strictly followed, the Arm Valley end mor-
aines were deposited during first glaciation and the
Little Fisher Valley moraines were deposited during
second glaciation, but in the locality there is no
evidence to support this twofold classification of the
glaciation. All the evidence in this area indicates
that it was glaciated at the same time as the central
plateau and if there were any earlier glaciation, its
traces have been obliterated by this glaciation.

The relevant field work was carried out in the
company of Messrs R.J.Ford and A. Spry, in December 1955
and with Mr J.N.Jennings, in March 1956. The author
wishes to thank these gentlemen for observations and
discussions which have contributed to this paper. The
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during the field work.

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Striated Pavement.

Mersey Valley.

Dolerite Erratic.

South Of Dublin Plain.

Davel's Gullet and The Hanging Valley.

Looking East From Dublin Plain.

Little Fisher Moraines.

Looking From the East Side.

THE PERMIAN SYSTEM IN WESTERN TASMANIA

BY

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ABSTRACT

Permian rocks occur at Mount Read, Mount Dundas, Mount Sedgwick, Zeehan, Firewood Siding, Strahan and Point Hibbs in Western Tasmania. The basal formation is exposed at Mount Read, Mount Sedgwick, Zeehan and Point Hibbs and consists of tillite. Striations on the basement at Mount Sedgwick indicate ice moving from the west. Sections up to and including the Golden Valley Group (Lower Artinskian) occur at Mount Sedgwick and Point Hibbs and a section from the Mersey Group up to and including the Cygnet Formation is found in the Firewood Siding area near the mouth of the Henty River. The sections at Mount Sedgwick and Firewood Siding are much thinner than corresponding ones in north-western and south-eastern Tasmania but that at Point Hibbs is as thick as or thicker than corresponding sections.

INTRODUCTION

The first record of Permian rocks in the area studied seems to be that of Montgomery (1894) who made brief reference to coal on the

Henty River, (see map, figure 1).   Johnston (1892) recorded some
fossil plants from the Henty River area and correlated the coal
measures there with the Mersey Coal Measures.   In 1894 Dunn noted
the tillite near Mount Read and commented on its similarity to the
Dwyka of South Africa and to the conglomerates at Wild Duck Creek
(Derrinal), Victoria.   In the same year Moore noted the Permian
fossiliferous and glacial beds on Mount Sedgwick and those at Zeehan
(Zeehan Tillite) which he also described as Permian.   The fluvio-
glacial beds near Strahan were first described by Officer, Balfour
and Hogg (1895).   Several later workers dealt with the deposits
mentioned above but no new work was added until Hills (1914) dealt
with the Point Hibbs section.   In 1925 Reid noted the probable
presence of Permian rocks on Mount Dundas.   Voisey (1938) included
references to this area, particularly to the Point Hibbs and
Malanna sections in his work on the Permian of Tasmania.   Edwards
(1941) noted the exhumed Permian surface on Mount Sedgwick and the
Permian of Mount Sedgwick was mentioned by Bradley (1954).   Some of
the Permian rocks at Firewood Siding, near Malanna, were described
by Gill and Banks (1950).   Others have also commented on the Permian
rocks considered in this paper but only as repetitions of earlier
work.

OCCURRENCES of PERMIAN ROCKS
in
WESTERN TASMANIA
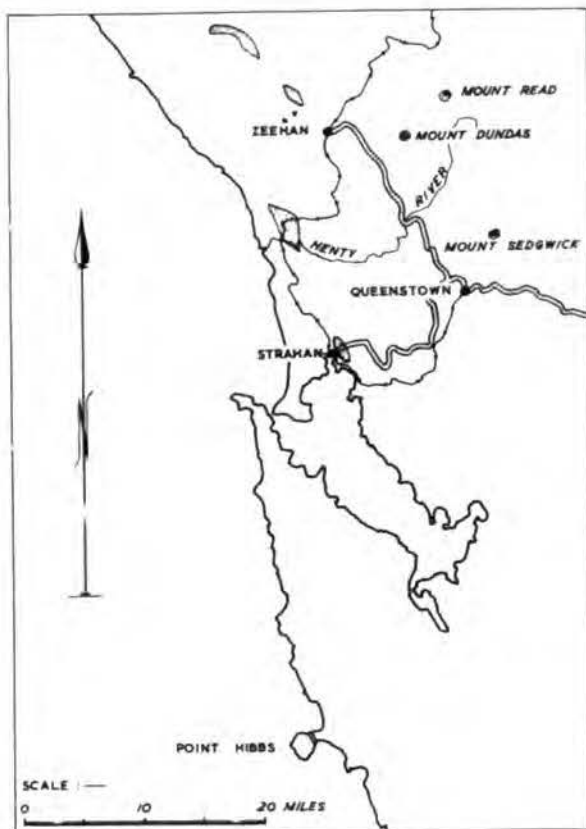
FIGURE I

Serious investigation of the Permian sections in this area began in November, 1953, when Professor K. G. Brill, Visiting Professor at the University of Tasmania, G. E. Hale and M.R. Banks spent a week measuring sections in the Malanna area. In January, 1957, the authors measured sections in the Malanna area, at Point Hibbs and on Mount Sedgwick and made observations on the Permian rocks on Mount Read and at Strahan. During the 1953 trip to Malanna sections were measured in railway cuttings and creek beds

using a steel tape and abney level.   The Mount Sedgwick section
was measured by using a Brunton compass as a level and measuring
cliff sections.   The Point Hibbs section was measured by laying
a steel tape along the dip of the vertical beds and reading off
thicknesses directly until the fault zone was met and then by
using the abney level.   Thicknesses in the section in Geologists
Creek, near Malanna, studied in 1957, were only estimates due to
thick undergrowth.

During the work the authors were aided by the explicit
directions on the route to Mount Sedgwick given by geologists of
the Mount Lyell Company.   This company also made available the
services of Mr. Jock Gilphillan who was of considerable assistance.
The Lyell-E.Z. Exploration Company made the work at Point Hibbs
possible by making available to the authors one trip each way in
a helicopter and later made aerial photographs available.   The
Electrolytic Zinc Company kindly allowed us to use their facilities
and made Mr. John Druett available as a guide.   The authors
acknowledge with gratitude the assistance of these companies and
their officers.   The authors are also indebted to Mr. M. Longman,
Geologist at the Tasmanian Museum for access to plant bearing shales
from which Johnston had described fossil plants from the Henty River.

All bearings are related to true north.

MOUNT READ

The earliest mention of Permian rocks near Mount Read is that,
of Dunn (1894) who mentioned a conglomerate with a great variety of

pebbles on the south side of the track half way from Mount Read to Moore's Pimple. He remarked on its similarity to the Dwyka of South Africa and the conglomerate on Wild Duck Creek (Derrinal) Victoria, both considered as Permian. Mills (1915) gave further details. He considered that it was Permian as it contained fragment of undoubted "Silurian" rocks (now known to be Ordovician). Bradley (1954, p. 199) also mentioned this occurrence as showing that the Carboniferous peneplain in this area has a undulating surface with variations up to 80 feet in height.

A complete survey was not made by the present authors but two areas of Permian rocks were examined. On the track from Mount Read to Zeehan about half a mile south of the "L" Lode open cut on the south-west side of Mount Hamilton, tillite was found in a small depression (co-ordinates 8.5 cms. S.S.W. by S. of C.P., Zeehan 8, 23627) (see map, figure 2).

SOME PERMIAN DEPOSITS
ON MOUNT READ

Based on air photo Zeehan Run 6 no. 23627

FIGURE 2

The Permian rests on "sheared pyroclastics" of probable Cambrian age which have a steep easterly dip and the Permian has a horizontal fissility although no bedding could be seen. The rock is greenish grey. It is very poorly sorted with boulders up to 18 inches long in a fine-grained (clay and silt grade) matrix. The boulders are angular and sub-angular and are grey Owen Conglomerate, pink Owen Conglomerate, Eldon Group quartzites, green sandstone, black slate, quartz and rare feldspar porphyry. The rock is fairly well lithified. The tillite is probably not more than two feet thick and is a small remnant preserved in a hollow between hills of "sheared pyroclastics" which rise locally to more than 50 feet above the level of the Permian. There is no internal evidence in this exposure of a Permian age. The lithology and degree of lithification are very similar to known Permian tillites elsewhere in the state and quite dissimilar from those of the Pleistocene till in the West Coast area. The top of the plateau at Mount Read shows no signs of Pleistocene glaciation so that all available evidence from this outcrop suggests a Permian age.

The Permian age is confirmed by the other exposure of tillite and associated rocks. In a depression between hills of "sheared pyroclastics" about three quarters of a mile south-west of "L" Lode Open Cut, tillite is found, resting on the pyroclastics (at point 10 cms. S.S.E. by S. of C.P. Zeehan 8, 23627) and in a runnel of an old track a surface of the pyroclastics shows a polished surface with striations trending 0° M approximately. This may be part of the pavement beneath the Permian but this could not be established beyond

doubt as the striations may be due to log hauling. Erratics in this vicinity reached a length of 33 inches and in addition to the types reported from the first locality include a black fine-grained quartzite. Otherwise the tillite at this second locality is very like that from the first. About 100 yards north-east along this track (at point with co-ordinated 9.7 cms. S.S.E. of C.P. Zeehan 8, 23627) olive grey siltstones are exposed dipping at a moderate angle to the west off a small hillock of "sheared pyroclastics". These are well-sorted in that erratics are rare and small. They are poorly bedded. These siltstones contain articulated crinoid columns and attached cirri of a type common in the Permian System in Tasmania. The siltstones are overlain by tillitic material and similar siltstones occur, apparently above the tillitic material, further north along the track and are again overlain by tillitic material. On topographic grounds it seems probable that the section is basal tillite, siltstone with crinoids, tillite, siltstone, tillite, but in view of the poor exposure the succession cannot be regarded as established. The presence of the crinoid of Permian type established the age of the succession.

The surface of the plateau north and east of these occurrences was not examined and there may be further outcrops. The two areas of Permian rocks found occupied small depressions in the surface of the "sheared pyroclastics" which rises perhaps as much as a hundred feet above the base of the Permian. To some extent the present topography on this part of Mount Read is a slightly subdued expression of the pre-Permian topography.

SECTION AT MOUNT DUNDAS

This section was not visited due to inaccessibility. Reid (1925) postulated that Permian sediment occur on the south-western fall of Mount Dundas on the evidence of fossiliferous boulders in some of the creek systems near Dundas. Elliston (1954, p. 172) stated that a thin layer of mudstone occurs between dolerite sills on Mount Dundas.

SECTION AT ZEEHAN

In 1894, Moore discovered the tillite north-west of Zeehan and considered it (1896a, p. 60) to be Permian in age on lithological grounds as also did Twelvetrees and Ward (1910). It has also been regarded as Precambrian (Hills and Carey, 1949; Carey, 1953), and Cambrian (Carey and Scott, 1952, p. 70; Elliston, 1954, p. 177) but Banks (1956, p. 193) regarded the age of the Zeehan Tillite as not then established. More recently Spry (1958) has suggested that the Zeehan Tillite is Permian on structural grounds and because it contains fragments of Dundas Group and Eldon Group rocks. He has also found a further occurrence of it north of the Pieman River and describes the rock from the different areas in some detail. The authors did not visit these areas. Campana and King (1958) give detailed evidence for a Permian age for the Zeehan Tillite.

SECTION AT MALANNA (MOUTH OF HENTY RIVER)

The Permian rocks here were first noted by Montgomery (1891,

p. 42) who gave the section as coal bearing beds overlain by sand-
stones and limestones with marine fossils and by white grit or sand-
stone. Johnston (1892) recorded Glossopteris browniana,
Gangamopteris spatulata, G. obliqua and Noeggerathiopsis hislopi
associated with curious botryoidal concretions from the coal bearing
beds and correlated them with the Mersey Coal Measures. Twelve-
trees (1902a) noted that the impure limestones and mudstones overlie
the coal measures. Twelvetrees (1902b, 1903) recorded details of
two bores put down near Malanna in a search for coal and suggested
(1902 b, p. lxxii) that the coal occurred on two horizons, one
exposed near Malanna and the other below the limestone. Voisey
(1938, p. 322) considered that only one formation containing coal
was present, that above the beds with marine fossils. In this he
was possibly influenced by the undoubted presence of coal bearing
or carbonaceous beds above the marine beds in the railway sections.
However, investigations by the authors suggest that there are
two coal bearing formations separated by marine beds. The authors
were unable to find any of the "curious botryoidal concretions"
in the railway cuttings nor any Noeggerathiopsis and it is clear from
Montgomery's statement (1891, p. 43) that the coal being investi-
gated was on the flats just north of the Henty River, not as far
north as the railway cutting. This has been checked in conversation
with local residents. Specimens of the shale containing the plant
fossils described by Johnston are in the Tasmanian Museum. The
rock is a micaceous siltstone containing decomposed concretions
probably of pyrite, now in the form of melanterite. It is thinly
bedded with beds of dark grey and medium grey siltstone alternating

from 0.1 inches to 0.25 inches thick. It is well lithified and shows some curved slickensided surfaces. No fossil seeds, nor sphenopsids, nor Vertebraria are present in the Museum specimens but large specimens of Glossopteris are present. One of the "botryoidal concretions" referred to by Johnston may be present but this is not certain. This material is much more lithified than any of the plant bearing siltstones seen in the railway cuttings and described later and the types of fossils are somewhat different in the two formations. The specimens in the Museum are characterised by extraordinarily large specimens of Glossopteris while specimens from the railway cutting contain numerous seeds, Vertebraria and sphenopsids. On the grounds of difference in degree of lithification and in the overall aspect of the plant assemblages it is considered that Johnston's specimens did not come from the railway cuttings, and on the grounds of Montgomery's description and Johnston's description, further supported by conversation with local residents, it is considered that they came from the flat ground on the north side of the Henty River, probably between Geologists Creek and the railway bridge. Thus, the coal measures referred to by Johnstone as equivalent to the Mersey Coal Measures are thought to be below the marine beds but there are further coal measure beds above these marine beds and the present authors suggest that these are equivalent approximately to the Cygnet Coal Measures. Gill and Banks (1950, pp. 266-7) described rocks from two formations close to Firewood Siding and suggested that the marine ones might be equivalent to the "Granton Formation" (= Cascades Group).

MAP OF HENTY RIVER (MALANNA) AREA
SHOWING PERMIAN LOCALITIES
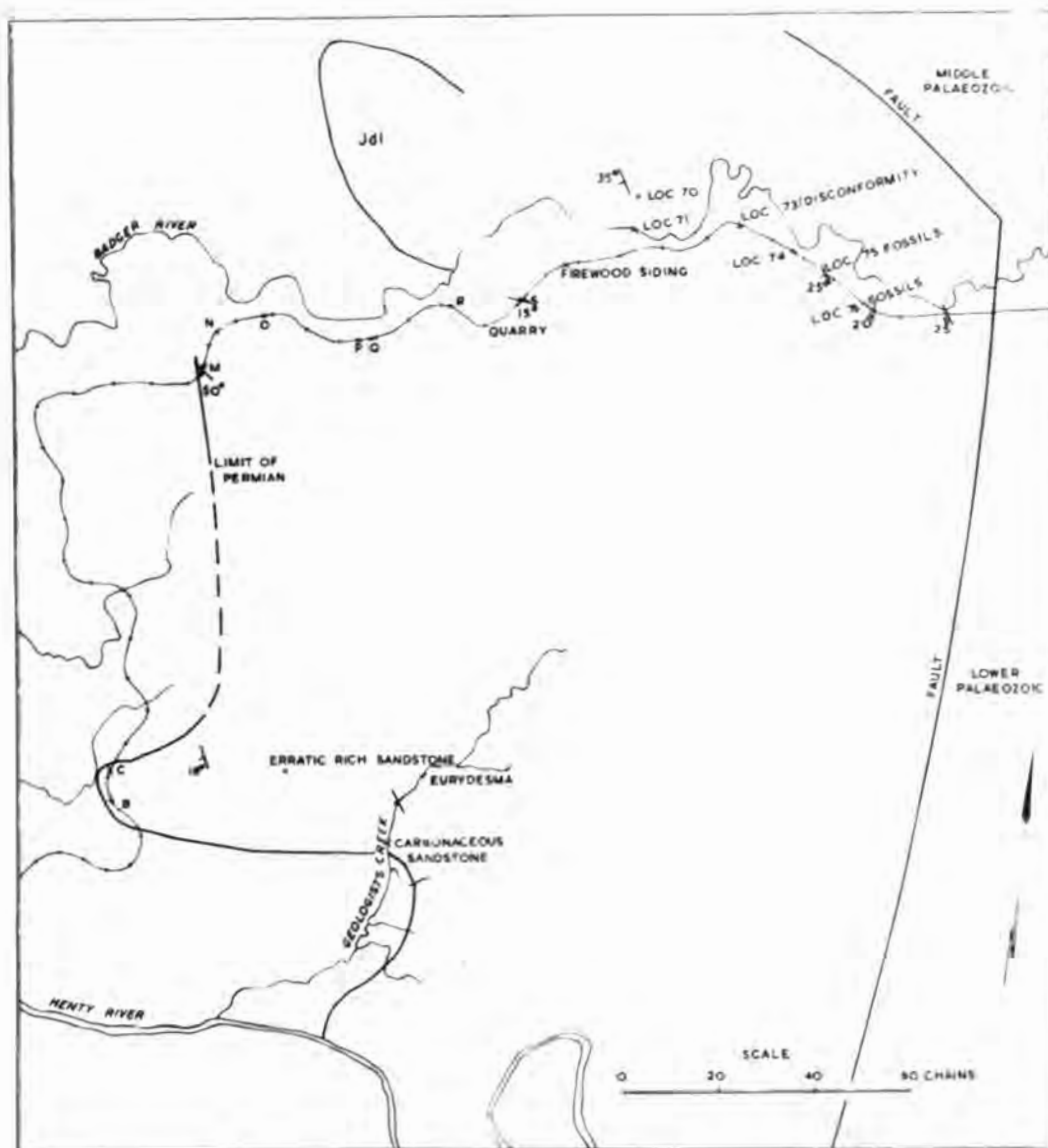
FIGURE 3

The present study was made in a traverse down Geologists
Creek, in cuttings along the railway line and in a section along a
creek flowing north into the Badger River just west of Firewood
Siding (see map, figure 3).

Due to thick vegetation and flooding of Geologists' Creek no measure-
ments of thickness were possible. Sections of the upper coal
measure beds in railway cuttings and a creek were made with a steel
tape but due to faulting and irregularities in dip correlation between
them is poor. Exposures of the marine beds in the railway cuttings
were too discontinuous for measurement and the thickness is cal-
culated trigonometrically. Localities mentioned are shown in the
map (figure 3).

The lowest formation in the section consists of carbon-aceous, micaceous, quartz-rich sandstones which are well sorted. These occur in the lower part of Geologists' Creek where they are apparently associated with black shale containing *Glossopteris browniana*, *Gangamopteris spatulata*, *G. obliqua* and *Noeggerathiopsis hislopi* (Montgomery, 1891, and Johnston, 1892).

This is followed after an interval with no obvious out-crop by alternating sandstone and fissile siltstone containing some fossils and a few erratics. In one place this shows north-west trending jointing planes less than a foot apart suggesting some faulting but there is no obvious displacement of beds and no change of dip. There follows a sandstone and siltstone alternation in which fenestellids, stenoporids, spiriferids and one specimen of *Eurydesma cordatum* were seen.

This is overlain by a fissile, calcareous siltstone containing spiriferids, stenoporids and predominant fenestellids. This is perhaps one of Montgomery's impure limestone beds.

After a further gap in the section the next unit is a buff sandstone with rare small erratics which forms small flats above creek level. This is richly fossiliferous with fenestellids, *Stenopora*, *Strophalosia*, *Terrakea* (?), *Dielasma*, spiriferids (including *Neospirifer*), *Spiriferina*, *Notospirifer* and pelecypods. Of particular interest is the occurrence in bands of numerous small, inarticulate brachiopods.

A few feet above this and also forming a small flat area is a greyish feldspathic sandstone with small erratics.

This is followed by a considerable thickness of greyish, feldspathic sandstone with numerous erratics up to 4 inches long occurring in bands. The bedding in this sandstone is thick and the unit forms several large waterfalls. The final unit, exposed in the uppermost waterfall in the gorge of Geologists' Creek, is a white, quartz-rich, well sorted sandstone with a few pebbles of white quartzite and this type of rock covers much of the surface between the Henty River and the railway line west of Firewood Siding. The basal sandstone and the plant-bearing shale formation is of the order of 40 feet thick and the marine sequence between the two formations of siliceous sandstone is of the order of 300 feet thick based on measurements of height of the top of the plateau cut in the Permian (525 feet above the level of the Henty River) at the railway bridge) and on estimated heights in the gorge of Geologists' Creek. The beds in the creek section are essentially horizontal. The thickness given is considerably in excess of that given by Montgomery which is thought to be much too small. The section in Geologists' Creek is summarized as figure 4.

Columnar Section of Permian Rocks
in Geologists Creek

QUARTZ SANDSTONE

THICK BEDDED, ERRATIC RICH SANDSTONE

FELDSPATHIC SANDSTONE
FOSSILIFEROUS SANDSTONE

FOSSILIFEROUS SANDSTONE AND SILTSTONE

SANDSTONE AND SILTSTONE

CARBONACEOUS SANDSTONE

FIGURE 4

In a railway cutting ("B") (Strahan 11, 5292, 1", E.N.E. of
G.P. and see map, fig. 5) a section of Permian rocks is exposed.
This shows:

Top 20 feet: medium to fine grained feldspathic sandstone with
                occasional erratics or lenses of erratics.

1 foot 8 inches: greenish-grey, medium-coarse grained sand-
                stone rich in erratics (up to 18 inches long) includ-
                ing quartzite, quartz schist, mica schist, chlorite
                schist and granite, very angular, un-oriented,
                unsorted; large (4" wide) "Martiniopsis" (?)
                subradiata and other spiriferids.

3 feet: dark grey, medium-fine grained sandstone with small

erratics and cylindrical bodies outlines by carbonaceous matter (? worm tubes).

4 feet: dark greenish yellow sandstone with a few erratics.

4 feet: greenish yellow, fine grained sandstone with occasional small angular and rounded erratics.

The next cutting to the north ("C") also contains erratic bearing sandstones but fossils are commoner and occur in lenses. The fossils include Stenopora (massive type like crinita, and ramose types), fenestellids, Strophalosia, spiriferids, including Notospirifer subradiata, Spiriferina Aviculopecten, other pelecypods and calcareous worm tubes.

East of Firewood Siding a series of railway cuttings expose Permian beds. The easternmost one (co-ordinates 3.5 cms. S. CP. Zeehan 1,23435) contains alternating thick-bedded impure sandstone and fissile siltstone with sandstone predominating in thickness in the alternations. Erratics up to an inch long are present but no fossils have been found. The boundaries between the members of the alternating beds are gradational. The next cutting west (Locality 76 of Gill and Banks, 1950, p. 267 and pl. III) contains fissile, grey, fossiliferous siltstone with a few small erratics interbedded with thin beds of feldspathic sandstone containing ostracodes and fenestellids. The fossils from these beds include Fenestella, Polypora, Protoreterpora, Stenopora, Notospirifer subradiata, other spiriferids, Schuchertella,

Strophalosia, Merismopteria macroptera, Eurydesma cordatum, Edmondia, Chaenomya, Aviculopecten, other pectinaceans, Platyschisma oculus, Camptocrinus and other crinoids and Conularia inornata. The erratics include quartzite, schists and hornfels.

About forty yards from the eastern end of the cutting the fissile fossiliferous siltstones are overlain with a sharp contact by a bed of sandstone about a foot thick which grades up into a non-fissile siltstone, with a few erratics and no obvious fossils. Bedding is not clear and if present is thick to very thick. Erratics are up to 8 inches long. Towards the western end of the cutting some bands of pebbles occur in the siltstone as also do some bands of very thinly bedded siltstone without erratics. These, however, are only a minor part of the sequence. In the next cutting west (co-ordinates 3.8 cms. S.W. of C.P., Zeehan, 1,23435; Locality 75 of Gill and Banks, 1950) thickly-bedded, grey siltstones with pebbles of schist which are well-rounded and ellipsoidal, are associated with greenish grey sandstone. Fenestella, Strophalosia, Dielasma, Notospirifer oviformis, Spiriferina duodecimcostata and other spiriferids are present. A fine-grained greenish brown siltstone occurs in the next cutting to the west (Loc. 74 of Gill and Banks, 1950) and contains slightly rounded pebbles of quartzite, schist and one large facetted cobble of grey granite eight inches long. Similar rock types occur west of the cutting (co-ordinates 5.2 cms. W.S.W., C.P., Zeehan 1,23435; Locality 73 of Gill and Banks, 1950)

where the erratic bearing siltstones are overlain disconformably by well-sorted, thickly bedded siliceous sandstone with carbonaceous fragments. The disconformity is marked by a narrow band of small siliceous pebbles (see fig. 5).
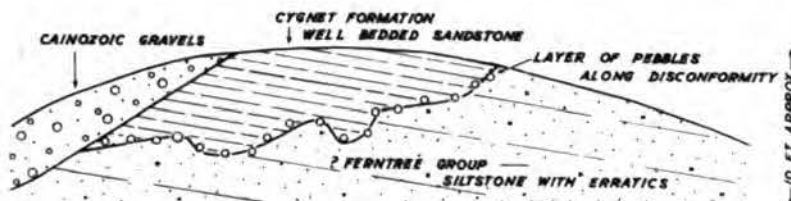
SKETCH OF DISCONFORMITY (Locality 73)

FIGURE 5

The apparent thickness of marine or erratic-bearing beds exposed is of the order of 800 feet. The section east of Firewood siding is summarised as figure 6.

FIGURE 6

The total thickness shown is thought to be excessive.
Dips are up to 25°, indicating proximity to faulting, and
possible faulting within the section. The numerous long gaps
in the section in addition render the overall thickness quite
unreliable and even the section is doubtful. Similar erratic
bearing beds occur in some of the depressions south-east of the
railway line where they outcrop beneath the quartz rich sandstones.
One such outcrop occurs at a locality 1" N.N.E. C.P. Strahan 11,
3293 and consists of grey, poorly-sorted sandstone with angular
to sub-angular boulders of schist, slate, phyllite and quartzite.

Above the disconformity the basal beds are quartz-rich
sandstones with carbonaceous fragments. These are thickly-bedded
and well-sorted but have a few sub-rounded to rounded pebbles
of quartz and quartzite. Between this cutting and Firewood
Siding sandstone and fine-grained carbonaceous sandstone occur in
the cuttings. At the eastern end of the Firewood Siding cutting
brown to grey micaceous sandstone outcrops. This is medium-
grained and thinly-bedded with some cross-bedding in the thinner
beds which have carbonaceous laminae. Plant fragments are
present but not common. The rocks outcropping in the Badger
River and on the cuesta to the north have been described earlier
(Gill and Banks, 1950, pp. 266-67).

In the first cutting ("S") west of Firewood Siding
(co-ordinates 3.4 cms. S.S.E. C.P. Zeehan Run 1, 23437) sandstone
and pebbly snadstone are the main rock types with rare beds of

carbonaceous siltstone. Cross-bedding dipping west or north-west is present.

In a small quarry south of the railway line about a hundred yards south-west of this cutting sandstone overlies a bed of grey, carbonaceous micaceous siltstone which is very rich in plant remains including Glossopteris, Gangamopteris and Vertebraria. Pyrite occurs associated with the carbonaceous siltstones.

The next cutting to the west ("R") (co-ordinates 3.5 cm. S.S.W. C.P. Zeehan Run 1, 23437) is thought to be that figured by David (1926, p. 102). At the eastern end of the figured cutting a bed of brecciated siltstone underlies a strongly jointed sandstone in a small syncline. Just to the west in what is structurally an anticline is sandstone with pebble bands and thin carbonaceous bands. Beyond a fault is a sandstone with a few thin carbonaceous siltstone bands, one of which shows circular, approximately horizontal sand-filled tubes, overlain by interbedded thin beds of sandstone and carbonaceous siltstone. To the west of another fault the basal section of the cut is in carbonaceous siltstone with thin beds of sandstone overlain by 2 feet of sandstone, carbonaceous siltstone and finally 11 feet of thinly-bedded sandstone. In the basal siltstone unit there are worm tubes, current ripple marks (currents from west) and cross-bedding occurs in the sandstone layers indicating mostly currents from the east or south-east but some from the west. Next to the west is a fault zone in sandstone and

carbonaceous siltstone. In the western section of the cut
the basal portion consists of an alternation of thin beds of
sandstone with very thinly interbedded sandstone and carbonaceous
siltstone. This is overlain by a brecciated sequence of carbonac-
eous siltstone and thin sandstone bands. This is followed by
several thick beds of sandstone which shows no brecciation or
faulting like that in the underlying bed. The top unit in this par
of the cutting is a thinly-bedded sandstone sequence.

Sandstone occurs in low cuttings further west in the axis
of a flat anticline. The sandstone is medium-grained at the
base and finer grained and thinly-bedded above. In the next
major cutting ("Q"), (co-ordinates 5.3 cm. S.W. Zeehan 1, 23437)
a fault divides the cutting. At the eastern end there is a
thickness of 9 feet of medium-grained sandstone with thin bedding
and cross-bedding mainly dipping south-west and overlain by 9
feet of thinly bedded fine grained sandstone. Some cross bedding
dipping to the west and more rarely to the east occurs in the
basal part. West of the fault the following section was measured.

Top

2 feet: Sandstone, white, micaceous, feldspathic, with
           clayey and carbonaceous partings producing flaggy
           breaks.

5 feet 6 inches: Brown and black carbonaceous and micaceous
           siltstone with minor superimposed on major rhythms,
           i.e. carbonaceous siltstone and siltstone alternate
           and each member consists of alternations of

carbonaceous siltstone and siltstone;   there are
8 cycles, the carbonaceous siltstone members being
the thinner; a prominent sandstone band from 3' 6"
to 4' above the base.

3 feet: white, micaceous, feldspathic sandstone with thin
bedding.

The basal sandstone contains bodies of concentric laminae of
carbonaceous matter, the exact nature of which is unknown.   The
siltstones of the second unit contain worm tubes and ripple marks
and the top is brecciated and crumpled.   This unit is very
reminiscent of the brecciated beds in the previous cutting ("R").

Another section is exposed in the next cutting to the west ("P
co-ordinates 5.6 cms. S.W. C.P., Zeehan 1, 23437) and is as follows

Top

1 foot 6 inches: Thickly bedded, medium-coarse grained
feldspathic sandstone.

2 feet: fine-grained, feldspathic, micaceous sandstone;
quartz grains angular; cross bedding rare; flaggy
splitting.

1 foot: medium to coarse grained quartz sandstone with
angular grains; thickens to 6 feet at east end of
cutting.

3 feet: black to brown, carbonaceous, micaceous,
feldspathic siltstone, very thinly bedded with
irregular bedding and some worm tubes.

5 feet 6 inches: white, friable, thinly bedded, flaggy-
platy feldspathic micaceous sandstone becoming
carbonaceous near the top; some bedding planes
show rippling.

20 feet: (top 4 feet in cut, bottom 16 feet in cliff to
north) white, medium grained feldspathic sand-
stone.

Cross bedding in this cutting dips south-west, west and
north-west.

A section was measured up the bed of a creek tributary to
the Badger River, with the co-ordinates 4.6 cms. S.S.E. C.P.
Zeehan 1, 23438 (on railway line). This section is given in
detail below:

Top

Interbedded carbonaceous and non-carbonaceous
siltstones with shattering, overthrusting, and
normal faulting (as in section figured by David).

30 feet: fine to medium-grained, cross-bedded sandstone
with quartz, feldspar, muscovite; thickly-bedded.

40 feet 2 inches: medium to coarse-grained sandstone, with
quartz, feldspar and a few quartz pebbles up to
50 feet above the base then some pebble bands,
thickly-bedded.

5 feet 9 inches: coarse sandstone to fine conglomerate with
large pebbles (up to 25 mms.) of quartz, red
chert and quartzite; angular to sub-rounded

pebbles; thickly-bedded.

4 feet 9 inches: fine to medium-grained quartz sandstone with a few pebbles of quartz and quartzite sub-angular to sub-rounded; thickly-bedded; flaggy to massive.

10 feet 9 inches: brown-yellow, fine grained quartz, mica, feldspar sandstone with very thin to thick bedding and some carbonaceous partings, flaggy and platy.

11 feet: interbedded black micaceous siltstones with disseminated carbonaceous matter, and brown micaceous siliceous siltstones; very thinly bedded; distinct band of quartz pebbles 8 feet above base; some distinct plant fragments.

1 foot 5 inches: white to pink, medium-grained sandstone with conglomeratic bands with pebbles of quartz, quartzite and some schist; quartz, pink feldspar and muscovite in matrix; some carbonaceous partings.

2 feet: greenish grey sandstone with muscovite and biotite; fine-grained, angular grains; erratics, present especially about 18" above base, consist of quartz, quartzite, slate, argillite and pink granite.

The first unit is lithologically very like the beds below the disconformity east of Firewood siding and distinct from those higher up. The higher units may be considered as parts of two

cycles as under

  8. siltstone

  7. fine to medium-grained sandstone

  6. medium to coarse-grained sandstone

  5. coarse sandstone to fine conglomerate

  4. fine to medium-grained sandstone

  3. fine-grained sandstone

  2. siltstone

  1. medium-grained sandstone

The grainsize decreases upwards to unit 2 then increases upwards to unit 5 with later decrease to unit 8. This latter is overlain by sandstones in the railway cuttings but the sequence is broken. From the top of unit 8 to the highest point on the hills to the south on which the quartz rich sandstones occur, is well over 50 feet so that a minimum thickness for this formation is 150 feet.

further to the west (cutting "O", co-ordinates, 4.2 cms. S. C.P. Zeehan 1, 23438) white, siliceous sandstones outcrop. Some of these are pebbly and there are beds of siliceous and carbonaceous siltstones. Small limonitic concretions are present and there are also curved cylindrical worm burrows in it. Cross-bedding dips north-west to south-west with some dipping east, and ripple marks and slump structures are also present.

In the next cutting west ("N", co-ordinates 4.7 cms. S.S.W. C.P., Zeehan 1, 23438) a fault divides the cutting. At the eastern end of the cutting a cross-bedded sandstone is overlain

by a micaceous sandstone containing plant fragments, a silt-
stone containing a sphenopsid, _Glossopteris_, both small and
large species, and _Vertebraria_, and finally by a coarse sandstone
with rare boulders up to 4 inches long. At the western end of
the cutting the following section was measured.

Top

4 feet;   white quartz sandstone, with much feldspar.

6 feet;   alternating fine micaceous sandstone and
carbonaceous siltstone; sandstone beds $\frac{1}{2}$" to 2"
thick, siltstone bands up to $\frac{1}{2}$" thick.

4 feet: white to brown carbonaceous, micaceous, feldspathi
sandstone; carbonaceous partings and bands,
thick bedding.

2 feet;   carbonaceous , micaceous, siltstone; thinly-
bedded.

4 feet 8 inches: brown-yellow sandstone; fine to medium-
grained; few quartz pebbles at 1' 6" above base
sub-angular to sub-rounded; above this bed
becomes more carbonaceous and micaceous with
some plant remains.

6 inches: carbonaceous, micaceous brown-black siltstone
with _Glossopteris Gangamopteris_, _Vertebraria_,
Phyllotheca, _Schizoneura_, and seeds.

5 feet: white, thick bedded sandstone, with fine
conglomeratic bands of angular to sub-rounded
quartz pebbles; cross bedding dipping south-
west.

This section shows an alternation of sandstone and silt-
stone with some of the siltstone units themselves composed of
alternating sandstone and siltstone beds. The sandstone units
are consistently thicker than those of siltstone.

The final cutting in Permian rocks ("M", co-ordinates
5.5 cms. S.S.W. C.P. Zeehan 1, 23438) consists mainly of sandstone
as shown in the following section.

Top

1 foot: Cross-bedded, white, medium grained sandstone

2 ins.: sandstone

6 ins.: clayey sandstone

9 ins.: sandstone

1½ ins.: clayey sandstone

6 ins.: sandstone, top surface ripple marked

1 in.: clay

2 ins.: sandstone

4 ins.: yellow clay

9 ins.: sandstone

9 ins.: micaceous, feldspathic sandstone with 1 in. clay seam

1 foot 8 ins.: yellow brown limonitic, clayey sandstone

6 feet 8 ins.: yellow quartz sandstone with cross-bedding;
          muscovite and feldspar present, grains angular;
          bedding planes about 1 inch apart.

5 feet 1 inch: conglomeratic sandstone; pebbles of quartz,
          quartz schist up to 1½ inches long, sub-rounded
          to sub-angular; matrix coarse sandstone to fine

conglomerate, mainly quartz with some feldspar, very angular; lower surface uneven.

1 in.: white sandy micaceous clay

2 feet 6 ins.: medium-grained, micaceous, (muscovite), feldspathic sandstone with occasional pebbles of quartz; grains angular; thickly bedded.

The prevalence of faulting, lack of distinctive marker beds, and common occurrence of cyclic sedimentation makes correlation between all these sections virtually impossible without very detailed work. On dip the last section ("M") should overlie the second last ("N") and this latter should overlie the creek section and due to lack of any possible correlation between them may well do so. However, the presence of faults is such that this superposition cannot yet be proved.

This, the highest formation in the Permian section in this area, consists of siliceous sandstones dominantly but with some minor conglomerates mainly quartz-rich with some muscovite and feldspars, well-sorted, with some rounded pebbles of resistant types in a matrix of angular grains. Bedding varies from thin to thick and cross-bedding on a fairly fine scale is common. No consistent current direction is shown but currents from the eastern quarter seem to have been somewhat commoner than those from the west with very few from north or south. As exposed the sandstones are mainly white. Cyclic sedimentation is well shown in several sections with a major sandstone siltstone alternation on which is

superimposed finer alternations of fine grained white siliceous sandstones with carbonaceous siltstones. These cycles represent changes in competency of the streams in the depositional area with perhaps the development of peaty swamps during times of low competence. The causes of the variations in competence may have been climatic or tectonic but more regional work is needed to establish the cause of the variation. There were at least three major cycles and many minor ones. The siltstones commonly show slumping, cross-bedding and ripple marking as well as the presence of worm tubes of several types. On at least one horizon plant remains are common and include Glossopteris spp., Gangamopteris, Vertebraria, Phyllotheca, Schizoneura and seeds.

The bores described by Twelvetrees (1902, 1905) are of some interest although neither can now be located accurately. Eden Bore No. 1 was placed north of the railway line (and probably south of the Badger River) fifteen and a half miles from Strahan probably somewhere near cuttings M, N, O as Eden Bore No. 2 was stated to be 1½ miles further north east on the Eden Coal Companies section (probably 4210,43 M on the West Coast Mineral Chart) and must certainly have been south-west of the fault north of Firewood Siding. This means that bore No. 1 was almost certainly not east of Cutting Q nor west of Cutting M. In Bore No. 1 it seems likely that the sandstones, shales and coal markings down to 115 feet belong to the topmost formation of the present authors. The beds from 115 feet down to 291 feet consisting of pebbly sandstone and mudstone may

be the equivalent of the Ferntree Formation in the Geologists
Creek section, the conglomerate between 291' and 309' to the
Risdon Sandstone and the underlying pebbly and calcareous mud-
stones to the marine beds low in the Geologists Creek section. The
order of thickness of the marine beds under this correlation is the
same as that found in Geologists Creek. The correlations must,
however, be regarded as very tentative only in view of the lack of
detailed information about the rocks in the bore. Eden Bore No. 2
passed through 108 feet of sandstone before entering a hard,
indurated, broken up slate which may well belong to the Eldon Group.

The Permian section in this area could be considered as
consisting of five formations, a basal carbonaceous, micaceous well-
sorted sandstone associated with plant-bearing shales, followed by
poorly-sorted sandstones and siltstones with marine fossils, then an
erratic-rich sandstone, a thick bedded impure sandstone and finally
a well-sorted siliceous sandstone with plant-bearing siltstones.
The sequence is summarized as figure 7.

Composite section of Permian System in
Malanna Area

FIGURE 7
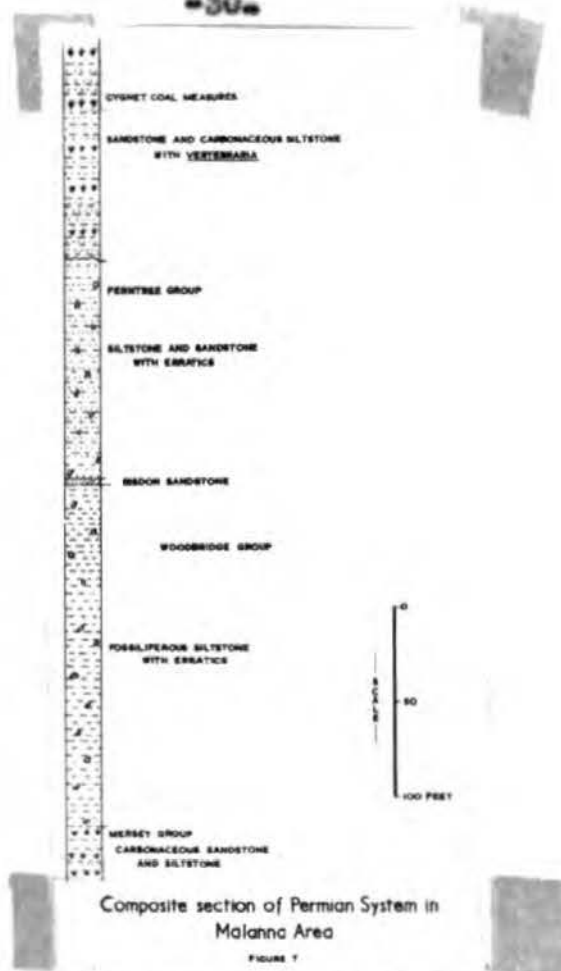
The lowest formation, in addition to Glossopteria and
Gangamopteria, contains Noeggerathiopsis hislopi which is known
elsewhere in Tasmania from the lower or Mersey Coal Measures, now
roughly correlated with the Liffey Group of McKellar (1957) and the
Faulkner Group of Banks and Hale (1957). The succeeding formation
contains Stenopora crinita (in typical "Woodbridge Glacial Format-
ion" type of preservation), and martiniopsids like those in the
"Woodbridge Glacial Formation". On palaeontological, lithological
and stratigraphical evidence this formation is equated to the
"Woodbridge Glacial Formation". As in the Deloraine area and other

parts of the north-west there is no clear indication of the
Cascades Group as developed in the south-eastern part of the state.
The absence of this group from the area near the mouth of the Henty
River cannot be definitely established until more detailed palaeo-
ontological work is done. It present it is either in a different
facies or extremely thin. The thin, erratic-rich sandstone may be
correlated on lithological and stratigraphical grounds with the
Risdon Sandstone but this is not at all certain. The thickly-bedded
impure sandstones are in the stratigraphical position of the Fern-
tree Mudstone but are coarser and contain more erratics. The
uppermost formation has in addition to sphenopsids, Glossopteris
and Gangamopteris, many specimens of Vertebraria, which is known
elsewhere in Tasmania from terrestrial sediments correlated with
the Cygnet Coal Measures. Thus on the occurrence of Vertebraria
this formation is correlated with the Cygnet Coal Measures in the
south-east.

In this area the marine beds and the correlates of the
Cygnet Coal Measures all show signs suggesting closer proximity to
the shoreline or source than those near Hobart. The "Woodbridge"
correlate is rather coarser than its Hobart equivalent and in
addition contains inarticulate brachiopods indicative of shallow
water, perhaps shoreline conditions. The correlates of the
Ferntree Mudstone are much coarser and contain more erratics than

that formation. The equivalent of the Cygnet Coal Measures has generally much coarser sand grains and a greater preponderance of pebbles and boulders than in the south east. Of interest also is the fact that the marine beds estimated at 300+ feet thick, are thinner than at Hobart where corresponding beds total at least 885 feet thick.

SECTION ON MT. SEDGWICK.

This seems to have been first noticed by Moore (1894, p. 148) who assigned a coal measures age (=Permian) to it. He became involved in a controversy with Montgomery (1894, p. 161) who suggested that the beds observed by Moore were due to Pleistocene redistribution of Permian material. Edwards (1941, pp. 19-22) also dealt with the area as part of an exhumed Permian or pre-Permian surface and correlated the beds on Mt. Sedgwick with Voisey's (1938) Achilles Stage on the grounds of the presence of Spirifer and Aviculopecten. Bradley (1954, p. 199) mentioned the Permian rocks briefly.

On the southern side of Mount Sedgwick (fig. 8) a section of Permian rocks is exposed in creek and cliff sections overlying quartz feldspar porphyries with slate fragments and overlain by dolerite which has baked the Permian. The dolerite contact is irregular and where best exposed is dipping west at a steep angle and cutting across the bedding. The Permian beds have a very low dip to the west. The basement on which the Permian rests varies

considerably in height. Just east of Mount Sedgwick Owen Con-
glomerate occurs at least 200 feet above the base of the Permian
and has no Permian on it. Just to the west dolerite rests on
porphyry about 200 feet above the base of the Permian and the Perm-
ian is absent. Thus it seems that there is either a valley in the
pre-Permian surface at least 200 feet deep or a post-Permian graben.
No evidence can be advanced as yet to favour either hypothesis.
The depression occupied by the Permian is no more than 1,500 feet
across (i.e. in an east-west direction).

MAP SHOWING OCCURRENCE OF PERMIAN
SOUTH OF MOUNT SEDGWICK (Based on air-photo,
Lyell, Run 2, no.774)

MOUNT SEDGWICK

Jd1

COVERED BY TALUS

P

𝕮 - CAMBRIAN
P - PERMIAN

— SCALE —
0   20   40 CHAINS

W.P.

—O→ DIRECTION OF ICE SCRATCHES ON PRE-PERMIAN PAVEMENT
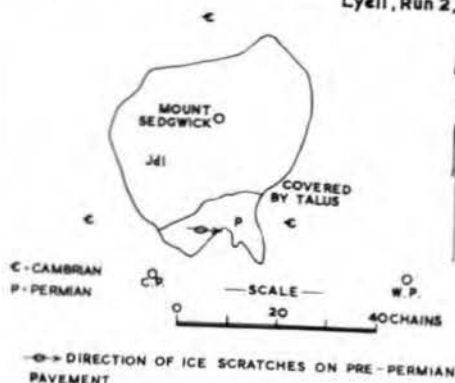
FIGURE 8

On the southern side of Mount Sedgwick two streams flow
south in the Comstock Valley. The section measured began on base-
ment about 100 feet west of the eastern-most and bigger stream.
The section was carried to the north along cliff sections on the
western side of the creek until a narrow shelf was reached which
ran back to the foot of the dolerite cliffs. The section was
carried north-westerly across this shelf and to the foot of a cliff
in Permian then up this cliff to the dolerite contact. The total
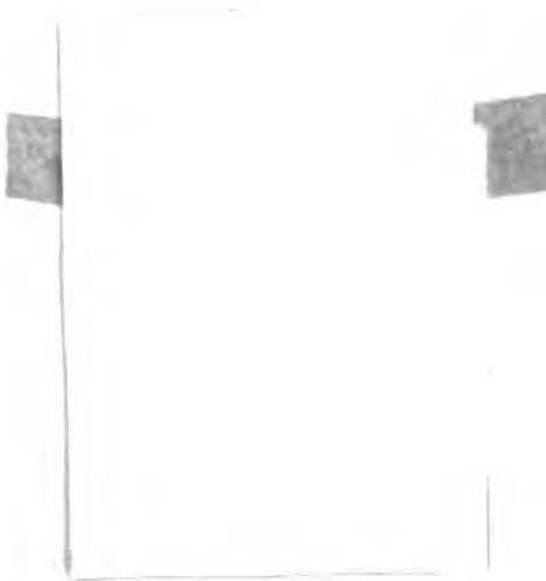
thickness measured between basement and dolerite was 210 feet approximately.

The contact between the Permian and the underlying rocks is exposed in a cliff section west of the creek mentioned above (1.8 cm., E.N.E., C.P., Lyell Run 2, no. 774). The contact is smooth and striated in some places but is jagged and irregular in other places within ten feet of the smoothed areas. Striations occur on both the slate fragments in the porphyry and in the porphyry itself but are clearer in the slate. The striations trend 106° and are deepest on the west side and shallowest towards the east thus indicating movement of the ice from west to east approximately. The contact is somewhat curved in section suggesting a roche moutonnée and it is perhaps significant that the surface is smoothed to the west and jagged to the east. Thus if the contact locally is the surface of a roche moutonnée, which cannot be positively established, the smooth upstream surface is to the west, the jagged plucked downstream surface is to the east, this agreeing with the evidence of the striations.

The basal Permian unit which is 39 feet thick (see text figure 9)

Columnar Section of Permian Rocks
on Mount Sedgwick

DOLERITE

SILTSTONE WITH FOSSILS

SILTSTONE WITH ERRATICS
AND FOSSILS

LIMESTONE

FOSSILIFEROUS SILTSTONE

CONGLOMERATE

SILTSTONE WITH ERRATICS

CONGLOMERATE

CONGLOMERATE

PEBBLY TILLITE

TILLITE

CAMBRIAN PORPHYRIES

SCALE

0

20

40 FEET

consists of well-striated boulders up to 4 feet in length consisting of porphyry, Owen Conglomerate, slates, quartzites and other rock types, in a poorly-sorted, dark grey clayey matrix with poorly marked horizontal fissility. No bedding was apparent. About six inches above the base at one place where the basement is jagged, there is a lense-like body of varved siltstone which is about 6 inches thick and about 30 inches wide. This lenticular body is slumped along the north-south axis. There is a slight disconformity just above the varved siltstone body. The basal unit is a tillite.

The second unit is a well-sorted pebbly tillite which is 6 feet thick. Much of the smaller material is lacking and the pebbles are well-rounded. Facetting and striations are rare. No bedding is present. It is possibly an outwash deposit of supra-glacial material. This is followed by 14 feet of silicified conglomerate, rather resembling Owen Conglomerate, which differs from the underlying unit only in degree of silicification. The next unit consists of 6 feet of compact conglomerate with small, rounded, mainly siliceous pebbles. This is well-sorted. It tends to form

the lip of a cliff. It is probably an outwash deposit. A
thickness of 50 feet of unfossiliferous thickly-bedded, erratic-
rich siltstone follows the conglomerate. The pebbles and boulders
show some facetting and striations. This rock was possibly depos-
ited by a floating ice sheet in very shallow water. There follows
a thickness of 11 feet of conglomeratic material with somewhat
rounded pebbles and few large boulders. This is possibly an
outwash deposit.

The next unit represents a marked change in lithology.
It is eleven feet thick and consists of dark grey, fissile silt-
stone with a few small pebbles. Near the base this is unfossil-
iferous but higher up becomes fossiliferous, the main fossils being
ramose stenoporids (S. tasmaniensis) but with some spiriferids and
crinoids.

A limestone, 21 inches thick, succeeds the siltstone. It
is medium to dark grey in colour and fine-grained with a few small,
rather rounded erratics. Fossils are common and are dominantly
stenoporids (S. tasmaniensis) although Eurydesma cordatum is common
and crinoid plates are present. The Eurydesma shells are disart-
iculated and usually convex side up. The stenoporid colonies are
broken and the crinoids fragmented.

The succeeding unit is about 63 feet thick and consists
of a thickly-bedded, grey, erratic-rich siltstone with numerous
fossils of which Eurydesma cordatum is prominent. However,
spiriferids and stenoporids also occur in considerable numbers.

The last unit is at least 11 feet thick and is a richly fossil-
iferous siltstone containing some erratics. Close to the dolerite
contact it is baked and assumes a light grey to creamy colour. It
is thinly-bedded and rests on the underlying unit with a sharp
contact. The fossils are dominantly ramose stenoporids (<u>S.</u>
<u>tasmaniensis</u>) some of which show marked current orientation. Other
fossils however are common and include fenestellids, spiriferids
(including <u>Neospirifer</u>), <u>Eurydesma cordatum</u>, aviculopectinids and
other pelecypods, <u>Mourlonia</u> and other gasteropods, and <u>Camptocrinus</u>
and other crinoids (as disarticulated plates).

The succession consists of a basal formation of tillite
and outwash conglomerate, then another similar formation differing
from the lower one particularly in possessing bedding. The lower
one is correlated on lithological and stratigraphical grounds with
the Wynyard (=Stockers) Tillite and the higher one with the
erratic-rich beds at the base of the Quamby Mudstone at Deloraine.
The fissile siltstone which follows this is lithologically like
the Quamby Mudstone. The succeeding limestone unit can be
correlated with the basal limestone of the Golden Valley Limestone
and Shale and with the Darlington Limestone as it is lithologically
similar, occupies a similar stratigraphical position and contains
<u>Eurydesma cordatum</u> and <u>Stenopora tasmaniensis</u>. The Darlington
Limestone correlate is followed by an erratic-rich fossiliferous
siltstone and this by a fossiliferous siltstone. These are
correlated on stratigraphical and palaeontological grounds with
the higher parts of the Golden Valley Group near Deloraine and the

formations above the Darlington Limestone at Darlington (Banks, 1957). The section does not go high enough to include the Liffey Sandstone and its correlates.

The basal formation indicates initial terrestrial glaciation followed by an ice retreat (units 2,3,4). The next formation marks a second advance of the ice sheet which was possibly not grounded and was floating in a shallow water, possibly a marine basin. This second advance was less intense than the earlier one and the sheet thinner. This was followed by further retreat leading to the development of outwash. Shallow marine conditions succeeded the formation of outwash, in which the icebergs deposited small erratics in the silt. Then the limestone unit represents reduction in the amount of clastic material supplied to the site of deposition and rich marine life. Some icebergs were still present. The next formation indicated a third ice advance but in a deeper sea with icebergs depositing erratics. This third phase of the glaciation was less intense than either of the earlier phases. The topmost formation shows a decreased intensity of glaciation, possibly marking the retreat from the third phase maximum. Lack of outwash deposits also supports the idea of floating ice.

SECTION AT STRAHAN

This deposit was probably first noted by Moore (1896b, p. 74) in a paper read before the Royal Society of Tasmania in August, 1895 and by Officer, Balfour and Hogg (1895) who noted

a deposit of unstratified or faintly stratified clay of great
hardness........." which contained some boulders which bore
striae. They remarked that similar deposits occurred on Mount
Sedgwick and Mount Tyndall. The Strahan deposits were considered
to be Pleistocene by Moore because they contained no erratics
"foreign to the country". Moore (1896b, p. 75) also noted
another outcrop of a similar sort of rock two miles north-east
of the outcrop being discussed on "Gould's old track". Lewis
(1939, p. 165) commented on these deposits which he considered to
be a Pleistocene moraine, associated with the "fluvio-glacial"
terraces at Strahan.

Road cuttings on the Queenstown-Strahan Road between
3 miles and 1½ miles from Strahan expose a tillitic conglomerate.
It shows some fissility and rough bedding which dips south-east at
a low angle. This rock contains boulders up to 2 feet in length
of greenish quartzite, quartz, very fine-grained black quartzite,
slate, greywacke conglomerate like those of the Dundas Group
and a biotite rich granite. Despite careful search no boulders
of Owen Conglomerate or dolerite were seen. Some of the boulders
are facetted and striated. The rock shows marked fracture planes
and in places both matrix and boulders are badly shattered. There
are a few beds of sandstone with rare rounded boulders and good
bedding. On the whole sorting is very poor but there are lenses of
comparatively well-sorted material which may be buried ice deposits
or deposits of melt water streams. Boulders in the main body of

the rock are more rounded than might be expected in a sub-glacial terrestrial till, and this rounding may have been produced by water transport.

On the first cutting south of the S 2 milepost steep faces trending 76° show slickensides which are either horizontal or dip slightly south and indicate north side west movement.

The present authors regard these beds as Permian because of their degree of lithification reflected in the fracturing of matrix and boulders, the jointing and the slickensiding. They have the same degree of lithification as is common in the known Permian beds in this region (e.g. the Firewood Siding and Geologists Creek area and on Mount Sedgwick) and considerably greater than that of the alleged Pleistocene beds at Strahan and Malanna. Lack of dolerite boulders and of Permian boulders also points to a Permian age but this lack can also be explained on the assumption of a Pleistocene age with a source in the West Coast Range just east of Strahan where neither Permian sediments nor dolerite are known. However, if the source was in this area the lack of Owen Conglomerate boulders is very difficult to understand. The slickensiding and movement implied by it are in keeping with slickensides and movements in the Permian near Firewood Siding. Thus the bulk of evidence suggests a Permian rather than Pleistocene age but no Permian fossils were found in it. This however, is common in the glacial beds near the base of the Permian System in Tasmania.

The poor sorting of the bulk of the material together

with facetting and striation of some boulders indicate some
glacial transport. Increase in percentage of smaller boulders and
clay and silt grades indicates some sorting, although of a low
order, which is confirmed by crude bedding and sandstone layers.
On the whole the deposit indicates the possibility of deposition
in aqueous conditions of some of the sediment. Possibly these
deposits were laid down by an ice sheet which was not thick enough
all the time to remain grounded but when the ice sheet was grounded
the meltwater lenses may have been deposited whereas when floating,
the tillitic conglomerate was formed.

Lithologically, these beds resemble those between 65 and
115 feet above the base of the section on Mount Sedgwick (see
figure 9) and beds near the top of the Stockers Tillite and the
base of the Quamby Mudstone in the Deloraine area.

SECTION AT POINT HIBBS

The only previous work on Permian rocks at Point Hibbs is
that of Hills (1914) who noted the presence of fossiliferous
mudstone conglomerate which he regarded as equivalent to the basal
beds of the Permian System in other parts of Tasmania. His work
was later quoted by Voisey (1938). The authors have examined
the Permian section both north and south of Point Hibbs (see map,
figure 10).
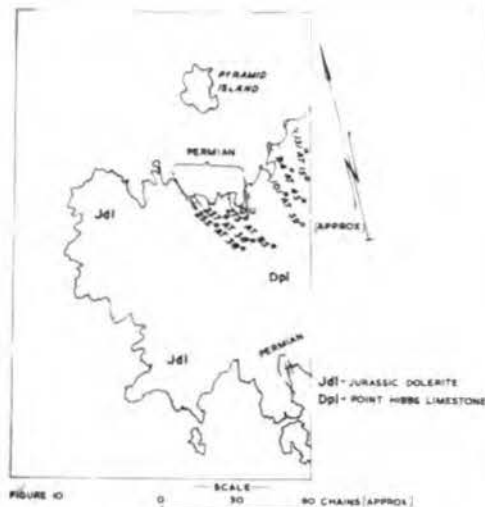
On the northern shore the Permian is faulted against a (?) Devonian limestone at the eastern end. Near the contact the limestone is sheared but the Permian is affected by the fault only to the extent of a shearing trending north-westerly in some of the finer beds near the contact. The Permian at the contact dips 264^{o} at 85^{o} and maintains this steep westerly dip for nearly a thousand feet. At this point it is affected by another fault and after a belt of variable strikes and dips some 30 yards wide maintains a dip of 38^{o} to the south-west (226^{o} to 244^{o}) for about a hundred yards before dipping under dolerite. Almost at the dolerite contact the dip is 244^{o} at 38^{o} and the contact although irregular in detail, trends 307^{o} over the length of its exposure on the shoreline. This trend would carry the contact west of Pyramid Island, which is reported to be dolerite, so that there is probably either a marked swing in the trend or a fault. The dolerite has produced intense contact metamorphism in the Permian sediments for a hundred yards from the contact.

Because of the faulted contact with the Devonian limestone, the lower part of the Permian section may be missing. The lowest bed exposed is a conglomerate about two feet thick composed of

numerous small angular rock fragments in a matrix of angular, coarse sand. The grains in the matrix are equant. There is no bedding within the unit but there is a north-west trending fissility, probably due to shearing. The rock is medium grey in colour. Some of the boulders in it are striated and reach a length of one foot. One remarkable feature noticed was the lack of boulders of the adjacent limestone. This is followed by 11 feet of conglomerate, interpreted as an outwash, which is relatively well sorted and contains somewhat rounded pebbles and boulders up to a foot long and lenses of tillitic material. The boulders include many of granite, some of a feldspar porphyry and quartzite including a green quartzite.

The outwash conglomerate is followed by 93 feet of tillite containing several thin beds of outwash conglomerate near the base. After a gap of 30 feet, siltstone outcrops for a thickness of 85 feet. This contains a lense of pyritic limestone twelve feet above the base and pyrite nodules at higher levels. It is a dark grey, somewhat fissile rock with wavy laminations in places. A few pebbles which tend to be rounded occur near the base. A thinly-bedded siltstone unit 84 feet thick follows. The bedding is usually less than an inch thick but may reach 2 inches in thickness. Some of the siltstone is calcareous and there are a few cross-bedded sandy bands. Erratics are rare but pyrite nodules are common. The only fossils present are some worm burrows.

The next unit is 205 feet thick and consists of twenty alternations of fine-grained sandstone (or coarse siltstone) and

erratic-rich sandstone in which the fine-grained sandstone is dom-
inant as far as thickness is concerned. The fine-grained sandstone
consists of angular, equant fragments of quartz, feldspar, rare
white mica and carbonaceous material with a few erratics. It is
medium grey in colour and bedding planes are 4 inches to 8 feet apart.
It is brittle rather than fissile. Fossils are absent or rare in
this rock type in the lower part of the unit but a little more
common higher up. They include Stenopora, Fenestella, spiriferids
Strophalosia, Eurydesma cordatum, ariculopectinids, a euomphalid
gastropod and crinoid columnals. Cross-bedded sets up to eight
inches thick are present and the currents came from a southerly
direction. One sandstone band about 10 feet above the base
showed ripple marks with a trend of 340° when restored and cross-
bedding formed by a current flowing from 225°. This sandstone
contains plant fragments. Associated with the fine sandstone are at
least twenty-one beds of erratic-rich sandstone which stand up
several inches to two feet above the platform cut in the fine
sandstone. These bands vary from six inches to nine feet thick.
They are composed of the same minerals as the fine sandstone but
they are perhaps a little coarser. In places the cement is
calcareous. Bedding varies in thickness from six inches up to
several feet. Their characteristic feature is that they contain
numerous erratics up to four feet long which are angular to sub-
angular and include granites, porphyries, quartzites, quartz,
quartz schist, gneiss, green quartzite and rarely limestone like
that immediately to the east. These erratic-rich bands are not

tillites as they appear to contain little if any clay matrix and within any one band there is some sorting, although it is only fairly good. Another feature of these bands is their richness in fossils especially Eurydesma cordatum and gastropods such as Keeneia but spiriferids, Stenopora and gastropods like Mourlonia also occur. The Eurydesma is articulated, disarticulated, or fragmented, the fragments lacking orientation. About 110 feet above the base the fine sandstone contains large limestone lenses with Stenopora, Fenestella, Strophalosia, spiriferids, Eurydesma, Aviculopecten and Calcitornella.

This unit of alternating sandstone and erratic-rich bands is followed by fifty-six feet of siltstone with rare erratics up to six inches long and numerous fossils. It is very dark grey. The siltstone contains rare small calcareous concretions which contain Calcitornella and numerous glendonites which are commonly single crystals up to six inches long and only rarely rosettes with up to three crystals. One glendonite grew around a specimen of Eurydesma. Fossils are very common and include numerous Stenopora tasmaniensis, fenestellids (very common in some bands), spiriferids including Grantonia, Neospirifer and Notospirifer, Eurydesma cordatum, Kalamya, aviculopectinids and Keeneia. The numerous extensive colonies of Stenopora tasmaniensis are preferentially oriented in many places suggesting currents from the north-west, north or north-east.

The next unit, which is 450 feet thick consists of four cycles, each cycle consisting of a basal member of banded conglomeratic siltstone and siltstone and the higher one of siltstone. The lowest cycle is 65 feet thick and of this 59 feet are conglomeratic siltstone and siltstone and 6 feet are siltstone. In the second cycle there are 9 feet of siltstone and conglomeratic siltstone and then 37 feet of siltstone. The next cycle is 125 feet thick with the basal member 103 feet thick and the upper one 22 feet thick. The final cycle is 295 feet thick with a basal member only five feet thick. There is a further major cycle in that in the first and third cycles the basal member is the thicker one while in the second and fourth the upper member is the thicker.

The lowest cycle contains numerous fossils, especially in the erratic-rich bands, and these include Eurydesma cordatum, Keeneia platyschismoides, Stenopora tasmaniensis, spiriferids, Mourlonia and fenestellids. The siltstones have wavy laminations. In the basal member of the third cycle fossils are again very abundant on some horizons and in the top member an 18 inch thick limestone bed occurs. This is very impure and contains numerous erratics. Fossils are not common in the limestone but include worm burrows and a bilaminar Stenopora very like S. johnstoni. The higher member of the fourth cycle is fossiliferous and the fossils include Stenopora and Eurydesma. Erratics up to 18 inches long occur in the siltstone member and a band of erratics occurs 146 feet above the base of the member. Erratics again become common near the top of the formation.

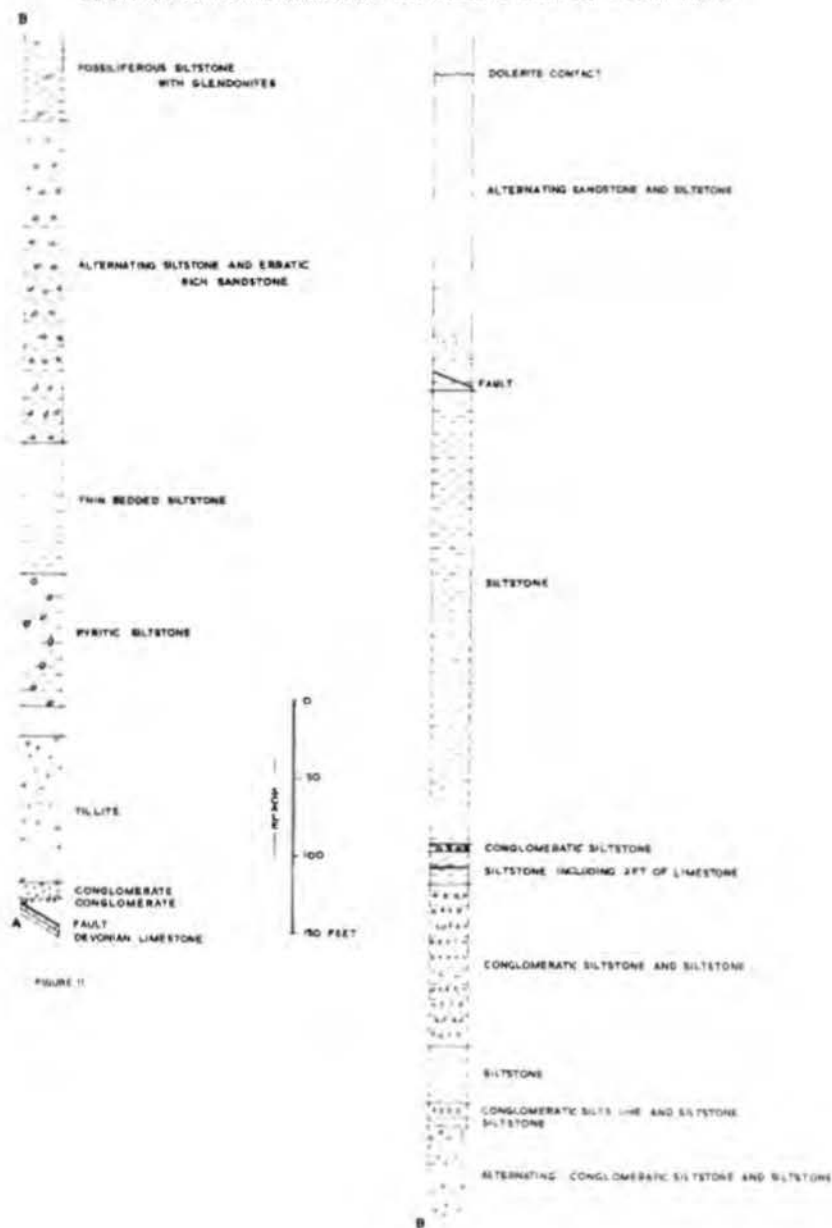# SECTION OF PERMIAN ROCKS ON NORTH SIDE OF POINT HIBBS

FOSSILIFEROUS SILTSTONE
WITH GLENDONITES

ALTERNATING SILTSTONE AND ERRATIC
RICH SANDSTONE

THIN BEDDED SILTSTONE

PYRITIC SILTSTONE

TILLITE

CONGLOMERATE
CONGLOMERATE

FAULT
DEVONIAN LIMESTONE

FIGURE 11

DOLERITE CONTACT

ALTERNATING SANDSTONE AND SILTSTONE

FAULT

SILTSTONE

CONGLOMERATIC SILTSTONE

SILTSTONE INCLUDING 2FT OF LIMESTONE

CONGLOMERATIC SILTSTONE AND SILTSTONE

SILTSTONE

CONGLOMERATIC SILTSTONE AND SILTSTONE
SILTSTONE

ALTERNATING CONGLOMERATIC SILTSTONE AND SILTSTONE

0

50

100

150 FEET

The belt of shattering and variable strike interrupts the section at this level. Beyond it at least 200 feet of alternating sandstone and siltstone occur in which the beds are usually 2 to 5 feet thick but one more than 20 feet thick is present. Bands of erratics are present in the sandstone and erratics are present in the siltstone. The erratics are dominantly quartzite and are up to 6 inches long. Fossils are abundant on some horizons. They include gastropods, fenestellids, stenopora, Aviculopecten subquinquelineatus, Eurydesma cordatum, E. cordatum var. sacculum and spiriferids including Notospirifer. The section is terminated by a dolerite intrusion.

The Permian section on the north shore of Point Hibbs is summarized here as figure 11.

It will be seen that at least 1,200 feet of clastic sediments are present.   The lowest 106 feet of dominantly glacial origin might be correlated with the Wynyard Tillite. The next major unit could be considered as composed dominantly of pyritic siltstone with some calcareous concretions and rare sandstone bands. It is at least 174 feet thick but there is a 20 feet gap between it and the underlying formation. This might be considered on lithological and stratigraph-ical grounds as equivalent to the lower part of the Quamby Group (=Quamby Mudstone of Wells, 1957) and the Woody Island Siltstone. The next major lithological break is at the top of the alternating sandstone and erratic-rich sandstone, 205 feet thick.   Fossils become abundant in the next unit, 56 feet thick, which consists of siltstone with glendonites. The fossils indicate a position low in the Permian sequence in Tasmania and the presence of glendonites strongly suggests correlation with part of the Woody Island Siltstone as these pseudomorphs are known from this formation and its correlates in eastern and northern Tasmania. The formation showing the four cycles follows this and is 430 feet thick. The thin limestone bed in the third member may be the Darlington Limestone as it is roughly in

the correct stratigraphic position and contains some of the fossils
from that formation. However, this cannot be regarded as established.
It occurs within a unit 20 feet thick of siltstone much more richly
fossiliferous than the adjacent beds and this strengthens the
correlation with the Darlington Limestone. If the richly fossiliferous
beds elsewhere are accepted as being at or near the base of the
Golden Valley Group (= Formation of Wells, 1957), the base of the
correlate of this group in the Point Hibbs area might well be con-
sidered as the base of the 20 feet of richly fossiliferous beds. The
beds above this 20 foot unit consisting of at least 449 feet of
alternating siltstone and conglomerate then sandstone and siltstone
might then be considered equivalent to the higher units of the Golden
Valley Group such as the Macrae Mudstone and Billop Sandstone of
McKellar (1957) and the Bundella Mudstone of Banks and Hale (1957).
The Mersey Group of fresh-water beds does not seem to be present but
it is not impossible that they are represented here by marine sediments.

Permian rocks also occur on the south side of Point Hibbs
(see fig. 10). They are faulted on the east against a Devonian lime-
stone which dips steeply east, and are intruded on the west by dolerite.
The total thickness would not exceed 500 feet of which 250 feet are
exposed in the shore section at the head of the bay. The Permian
itself dips steeply to the east near the limestone and is apparently
overturned and on the west side of the bay dips west at about 80°.
The lowest (i.e. easternmost) exposed bed is an erratic-rich sandstone
with fragmentary fossils. Higher up are siltstones with erratics and

and some beds of limestone. Stenopora tasmaniensis is particularly
common in these. The topmost unit consists of interbedded sandstones
with erratics and siltstones. The sandstones are richly fossiliferous
The dolerite has highly indurated the sandstone. Fossils in the
Permian rocks in this bay include Stenopora tasmaniensis, S. johnstoni
and a fine ramose stenoporid, fenestellids, Strophalosia, spiriferids
including a large Notospirifer, pelecypods including numerous large
Eurydesma cordatum and aviculopectinids and gastropods. The beds
close to the western side of the bay where limestone is commonest are
probably close to the Darlington Limestone in age judging from the
presence of S. johnstoni and the lowest beds exposed would be about
the level of the "Erratic Zone" on Maria Island.

One feature of the Point Hibbs section is the cyclic sed-
imentation. On a fine scale there are alternations of siltstone and
sandstone, siltstone and conglomerate or sandstone and sandstone with
numerous erratics. In many cases fossils are much commoner in the
beds rich in erratics but this is not invariable as some of the silt-
stones with few erratics are also fossiliferous. These alternations
are themselves grouped into larger cycles with the banded sediments
alternating with siltstone. On an even greater scale the succession
may be considered as glacial beds, followed by siltstone, banded
sandstones, siltstone, and then five cycles of alternating banded
siltstone and siltstone. Explanation of these cycles requires more
regional data than it is appropriate to present here.

SUMMARY AND CONCLUSIONS

Permian tillite and fossiliferous siltstone occurs on the western end of Mount Read occupying depressions in a partly exhumed pre-Permian surface of Dundas Group rocks. The section is not clear but the Wynyard Tillite and Quamby Group are thought to be represented At Mount Sedgwick the Permian rests on a striated surface of Dundas Group rocks. The striations indicate movement of ice from west to east. The section includes a basal tillite then rocks of the Quamby and Golden Valley Groups but these are considerably thinner than in eastern and north-western Tasmania. Near the mouth of the Henty River Permian rocks are faulted against older Palaeozoic sediments. The oldest Permian rocks recognized are correlated with the Mersey Group and include some coal. These are followed by the "Woodbridge" Group, Ferntree Group and Cygnet Coal Measures. All these groups are thinner than in south-eastern Tasmania and generally of coarser grainsize. Near Strahan a fluvio-glacial conglomerate occurs and is correlated with the top part of the Wynyard Tillite. At Point Hibbs a section from somewhere within the Wynyard Tillite up at least into the Golden Valley Group is well exposed. The Quamby and Golden Valley Groups are comparable in thickness with their equivalents near Deloraine and Cressy and perhaps thicker than the sections in south-eastern Tasmania (see figure 12)

PERMIAN SECTIONS IN WESTERN TASMANIA
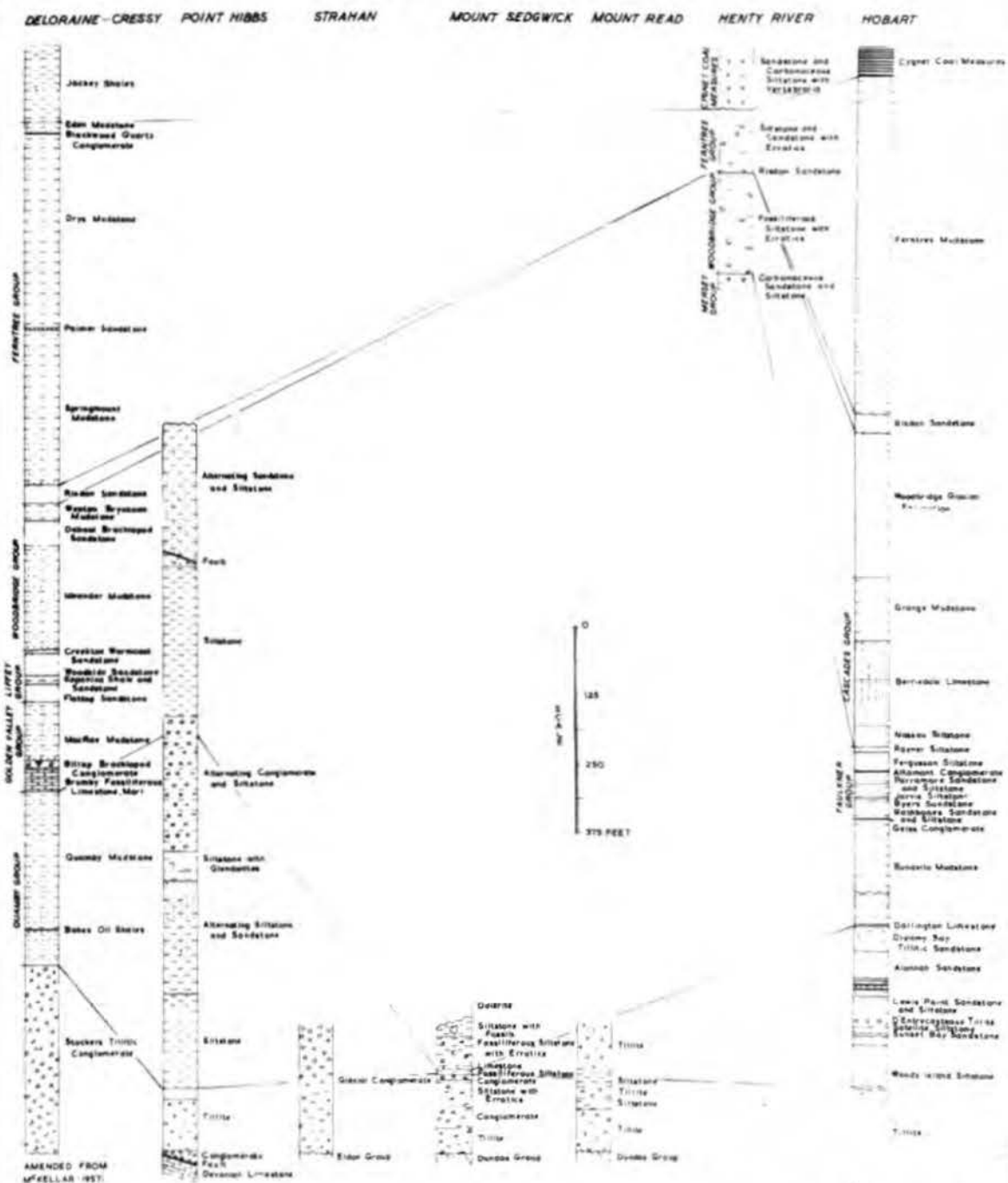
FIGURE 12

Evaluation of these variations in thickness must await more regional data.

GEOLOGICAL HISTORY

It would appear from the evidence quoted earlier that early in the Permian Period, Western Tasmania was covered by an ice sheet which in places moved from west to east over a surface with a relief of possibly a couple of hundred feet. On the evidence of the boulders present in the basal tillite and overlying fluvio-glacial or glacial beds the area overridden by the ice included conglomerates like the Owen Conglomerate, greywacke conglomerates like those in the Dundas Group, quartzites like those in the Eldon Group, some of them with fossils, black quartzites, green sandstone, slates including black slates, porphyries of several types including feldspar porphyries and a biotite rich granite. The lack of erratics similar to local Precambrian rocks is remarkable.

The ice later retreated depositing a sequence of glacial and fluvio-glacial beds. At least one readvance of the ice sheet is recorded in the section on Mount Sedgwick. Following retreat of the ice sands and erratic bearing sands were deposited in a cyclic fashion, the cycle representing perhaps minor retreats and advances of the glacial front. At Point Hibbs at least these sands were marine. Erratics of quartz schists and gneiss of a local Precambrian type occur at this level. Perhaps contemporaneously at Mount Sedgwick erratic rich silts were being deposited. Lack of the cyclic nature may be due to close proximity to the ice front.

Further retreat of the ice front led to deposition of siltstones containing few erratics, calcareous concretions and glendonites. This type of rock is much thicker at Point Hibbs than at Mount Sedgwick where it also lacks the glendonites, and concretions. These siltstones are marine and were probably formed under conditions of poor circulation. They contain no structures or fossils characteristic of deep water but at the same time contain no clear evidence of shallow water deposition. From their consistent position between tillites or fluvioglacial beds and shallow water sandstones or limestones here and in other parts of Tasmania it would appear that they were deposited as shallow water near-shore silts derived from a low-lying land surface.

There is a break in the record of Permian sedimentation in Western Tasmania the next rocks exposed being the lacustrine and paludal beds of the Mersey Coal Measures. These terrestrial conditions were followed by an inundation of the land by the sea. Erratic-bearing siltstone, sandstone and impure limestone were deposited and at times marine life was abundant. It was mainly of shallow water benthonic forms. Some evidence suggests that this area was closed to the source and that the sea was shallower than at Hobart. After retreat of the sea erosion occurred before deposition of pebbly sands, silts and carbonaceous silts in lakes and swamps. There were several variations in the competency of the currents in the area of deposition. Plant life was abundant and included pteridosperms and equisetales.

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LOCALITY INDEX.

	Quadrangle	Latitude South	Longitude East
Badger River	Strahan	41°59'	145°14'
Comstock Valley	Lyell	42°02'	145°39'
Darlington	Maria Is.	42°35'	148°04'
Deloraine	Quamby	41°31'	146°22'
Derrinal, Victoria		36°53'	144°36'
Firewood Siding	Zeehan	41°59'	145°16'
Geologists' Creek	Strahan	42°00'	145°16'
Henty River	Strahan	42°03'	145°18'
Hobart	Hobart	42°52'	147°20'
Malanna	Zeehan	41°59'	145°13'
Maria Island	Maria Is.	42°37'	148°06'
Moores Pimple	Zeehan	41°02'	145°28'
Mount Dundas	Zeehan	41°52'	145°29'
" Hamilton	Murchison	41°50'	145°30'
" Read	Zeehan	41°50'	145°30'
" Sedgwick	Lyell	42°00'	145°35'
" Tyndall	Murchison	41°57'	145°34'
Pieman River	Pieman River	41°40'	145°02'

	Quadrangle	Latitude South	Longitude East
Point Hibbs	Pt. Hibbs	$42^{\circ}43'$	$145^{\circ}15'$
Strahan	Strahan	$42^{\circ}10'$	$145^{\circ}20'$
Zeehan	Zeehan	$41^{\circ}53'$	$145^{\circ}17'$

NOTES ON THE CAINOZOIC HISTORY OF WESTERN TASMANIA - THE

"MALANNA" GLACIATION

by

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ABSTRACT

The Cainozoic history of the Malanna area included faulting,
deposition of sediments by streams or in lakes and swamps, two
epochs of planation, and stream erosion. There is no physiographic
depositional or structural evidence for glaciation in this area,
the type area of Lewis' Malanna Phase of the Pleistocene glaciation
This phase must be considered invalid on the evidence available in
the type area and the term "Malanna Phase" should be abandoned.

INTRODUCTION

Gregory (1904, p. 51) first described the sediments near
Malanna which were later considered to be morainal and adopted by
Lewis as the type feature for his Malanna Glacial Phase of the
Pleistocene glaciation. Gregory considered them to be glacial in

origin because of the presence of boulder clay containing boulders
of very decomposed dolerite. Because of the depth of weathering
of the boulders and lack of "indication of recent glaciation in
this locality" he provisionally assigned a Carboniferous (Permian
of present day nomenclature) age to the glaciation.   Later Loftus
Hills showed these deposits to David who was impressed by the
shattering of the Permian rocks at Malanna and influenced by the
flat-floored, steep walled valley of the Eden Rivulet (now Badger
River) and perhaps also by the morainal form of some of the hills
near Firewood Siding and Koyule (see fig. 2.).   David (1926) also
considered the deposits near Malanna to be glacial and because of
the impressions he had gained of the physiographic effects of glac-
iation, he considered them to be Pleistocene. On the grounds of
depth of decomposition of the dolerite boulders and amount of river
erosion since the supposed glaciation he considered that the
"Malanna Moraine is at least as old as Mindel, and possibly even
as old as Günz". The presence of dolerite boulders, the lack of
known outcrops of dolerite in the neighbourhood and the supposed
glaciation forced David to postulate transport of dolerite from
Mount Sedgwick, twenty miles away, by an ice sheet, as Mount
Sedgwick was, and still is, the nearest known dolerite capped
mountain showing glacial effects. In 1934 Lewis designated these
deposits as the type evidence for his Malanna Phase.

Gill and Banks (1950) visited the area and made observation

on the older rocks. In 1953 K. G. Brill, G. E. Hale and M. R. Banks visited the area and K. G. Brill made the important discovery that there was a large outcrop of dolerite less than one mile north of Firewood Siding, and less than half a mile from the nearest "morainal" deposit. Detailed sections were measured in some of the cuttings. In 1957 the present authors visited the area in an attempt to solve the outstanding problems connected with the glaciation but after a few days further observation concluded that there was no evidence at all for Pleistocene glaciation in this area and that all the features could be produced by Tertiary faulting and normal erosion.

Method of Measuring Heights.

Heights quoted along the railway line and south of the Henty River are those for a series of survey pegs which were established by R. Braybrook, Hydro-Electric Commission surveyor, as part of a air-photo control survey. The heights of the survey pegs are related to mean sea-level, Hobart. The four heights not on the railway line were measured in 1957 by an aircraft altimeter capable of reading to ten feet. The heights were found by reading on the survey pegs, then on the various physiographic features and then onto the same or another survey peg. Maximum time between successive readings on survey pegs was twenty minutes.

The authors wish to acknowledge helpful criticism of the manuscript by J. L. Davies, Geography Department, Professor S. W. Carey and Dr. E. Williams, Geology Department, University of Tasmania.

PHYSIOGRAPHY

The Malanna area is situated north of the mouth of the
Henty River a couple of miles inland from the western coast of
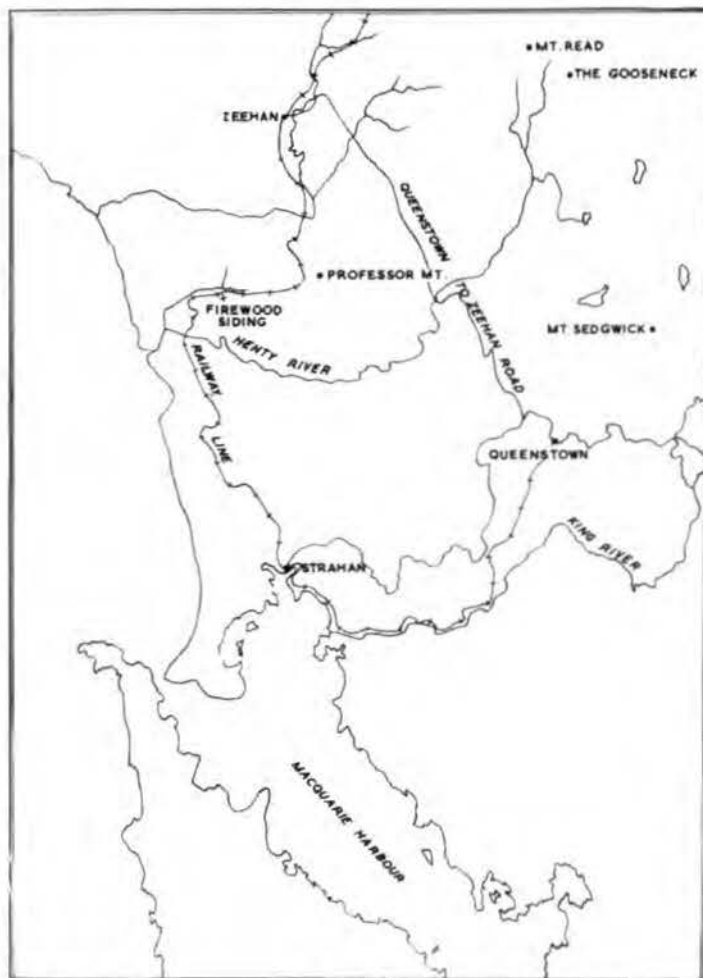Tasmania (see map fig. 1).

FIGURE 1   SCALE

The main drainage of the area is by the Badger River to
the north and the Henty River to the south. The Badger River
rises about four miles to the east on the western slope of Mount
Professor (see Gill and Banks 1950 pl. III). From its source to

about half a mile east of Firewood Siding the river flows in a
swampy plain about half a mile wide which has a height of about
360 feet above sea-level about a mile east of Firewood Siding. The
plain lies between steep scarps on both sides and the river flows
mainly near the western or northern boundary of the plain (see
Gill and Banks, 1950, pls. I and III). There is one low, sharp
ridge on the plain which parallels the western and northern wall
and is about twelve chains from it. The main tributaries above
Firewood Siding enter the Badger River from the east and south and
are approximately at right angles to the scarp bounding the plain.
In this part of its course the western and northern scarp is in
Crotty Quartzite, the river plain in Gordon Limestone with a sandy
band in it which is represented topographically by the small ridge
on the plain, and the eastern and southern scarp a dip slope in
Owen Conglomerate and Caroline Creek Sandstone. From its source to
about half a mile east of Firewood Siding, the Badger River and its
tributaries are strongly structurally controlled; the river fol-
lows the strike of the Gordon Limestone and the tributaries are
perpendicular to the strike around the noses of a north-west plung-
ing anticline and syncline. Minor streams only join the Badger
River from the scarp of Crotty quartzite. The Badger River is
separated at its source in the north by a low divide from a stream
cutting down through the Crotty quartzite and finally flowing into
the Little Henty River. The divide is within the plain and is only
a few feet high. After crossing the Eden Fault (see fig. 2) half
a mile east of Firewood Siding the Badger River enters a valley

tract (of Hills, 1946) in which it is cutting into Permian rocks. The main tributaries now enter from the north and are more or less parallel to the strike of the Permian rocks.   About thirty chains west of Firewood Siding the river enters a mountain tract (of Hills, 1946) in which it is entrenched in the sandstones of the Cygnet Coal Measures.   Further downstream the relief is lower but the stream still in mountain tract until about one and three quarters of a mile west-south-west of Firewood Siding it opens out into a valley tract before turning south and flowing between vegetated dunes to the east and intermittently moving dunes to the west.

FIGURE 2
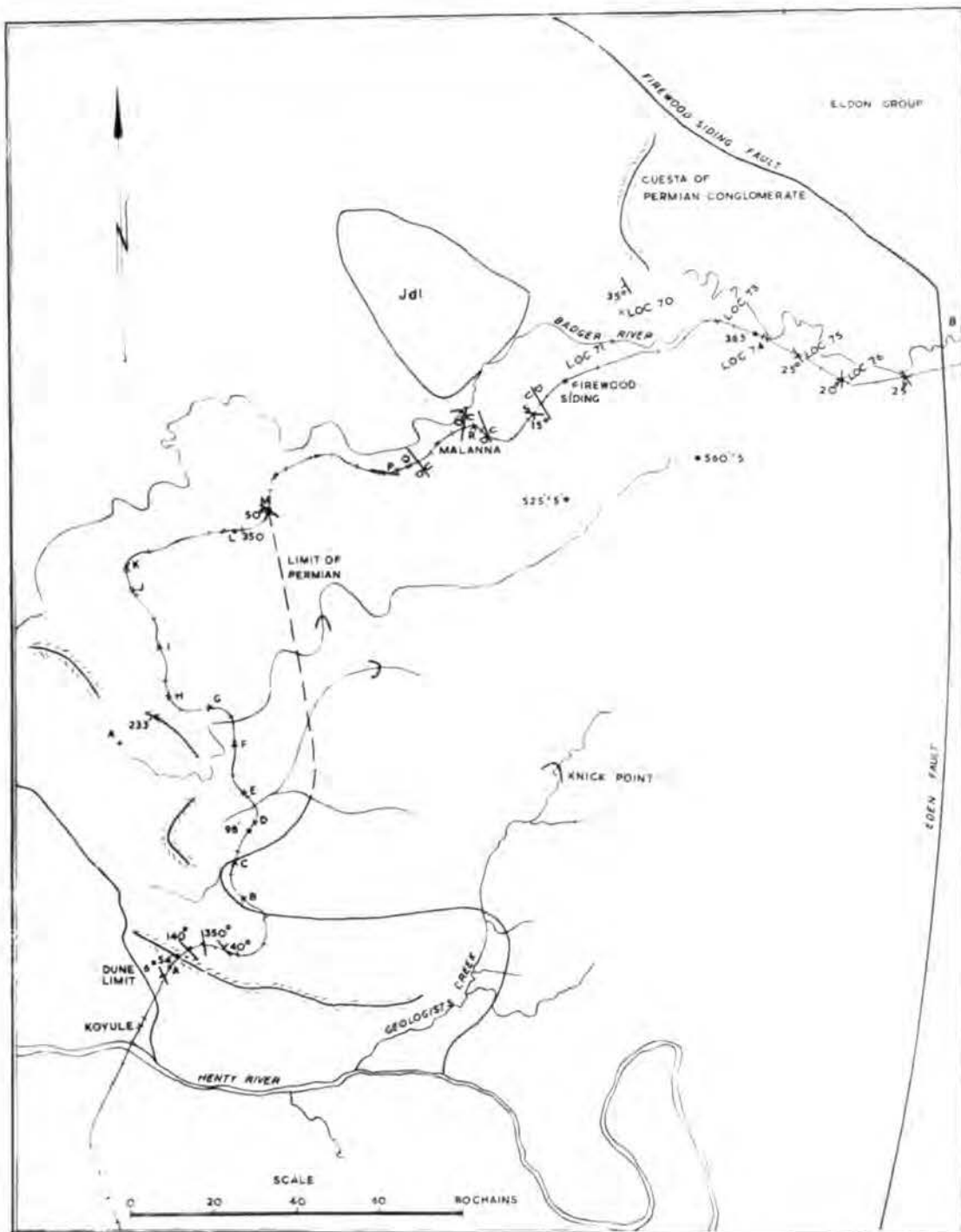
Loftus Hills is credited by Lewis (1926, p. 88) with the discovery of a small glacial valley superimposed on a broader glacial one in the Eden Valley. David (1926, p. 91) implies that the Eden Valley (now Badger Valley) had suffered "very ancient glaciation", but in neither case was any detailed physiographic evidence offered. At first sight the Badger Valley might appear glacial because of the steep side slopes, the flat floor and the gentle curve of the valley in plan. These features of the valley are the result of normal atmospheric and stream erosion in a topography of rocks of varying competence, the incompetent Gordon Limestone lying between the resistant formations, Owen Conglomerate and Crotty Quartzite. A local base level developed in Permian rocks just east of Firewood Siding has caused the erosion to that level of the Gordon Limestone in the Badger Valley. The entrenchment of the Badger River west of Firewood Siding may be due to rejuvenation, possibly associated with faulting or fall in sea-level, with headward movement of the knick point (see fig. 2). The Badger River valley shows no physiographic evidence of glaciation.

The main stream draining the area is the Henty River which rises about sixteen miles to the north-east on the slopes of Mount Read and the Gooseneck and flows through the Henty Surface as an entrenched, youthful stream to within a mile or two of the Malanna area. It is still in mountain tract to within a mile of

the railway bridge but below this shows some depositional features
near present river level.   The Henty River north of the Queenstown-
Zeehan Road is glaciated and the terminal moraines of the glacier which
occupied the Henty Valley occur a few hundred yards below the road
bridge over the Henty River (Bradley, 1954, p. 196 and Yolande River
Sheet).   The exact level of the foot of these moraines is not known
but it is probably not more than forty feet below the bridge which has
a height of 282 feet above sea-level (H.E.C. Bench Mark).   Below
these moraines the Henty River is in a typical mountain tract and
there are no further signs of glaciation.

The area between the railway line and the Henty River
is drained by west-flowing streams and by Geologists Creek.   In all
cases the headwaters are in wide, low valleys and swampy conditions
are common.   Further down their courses the streams pass into steep,
narrow, mountainous valleys before entering short valley and plains
tracts prior to joining the Badger or Henty Rivers.

The surface topography of the area might perhaps be
considered in terms of a number of erosional surfaces.   The highest
of these is the Henty Surface (of Gregory 1903, not Bradley 1954).
This extends from the foot of the West Coast Range where it has a
height of from 1100 to 1200 feet above sea-level to the Malanna area.
In the Malanna area a height was measured on a hill of Crotty quartz-
ite (Gill and Banks, 1950, Pl. I, western photo, 1.5 cms. S. W. of
centre point) which forms part of this Henty Surface.   The height
is 720 ($\pm$10) feet above sea-level giving an average slope to the sea
of about 60 feet per mile. Immediately south of Firewood Siding the

heights of two hills in Permian sandstone and conglomerate were
measured and found to be 525(±5) and 560(±5) (averaged heights) feet above sea-level.
The higher, easternmost one, is about two miles from the hill of
Crotty Quartzite measured and the seaward slope is of the same order
as that for the main part of the Henty surface. North of Firewood
Siding a sharp ridge of Permian and a hill of dolerite appear to reach
a similar height to the hills south of Firewood Siding and to be
accordant with flat-topped hills to the north forming part of the
Henty surface. Thus the hills immediately south of Firewood Siding
are considered to form part of the Henty surface.

Sections of Malanna Area Showing Physiography and Structure
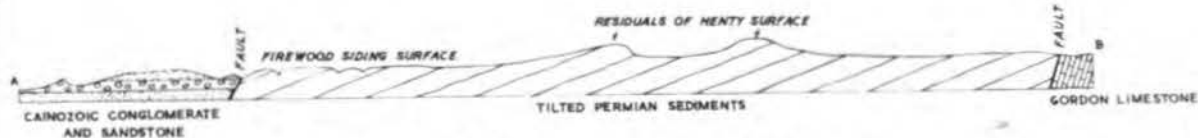
a DIAGRAMMATIC SECTION SHOWING RELATIONSHIP OF SURFACES TO STRUCTURES

b STRUCTURE OF CAINOZOIC SEDIMENTS AS SHOWN IN RAILWAY CUTTINGS PROJECTED
ON TO SECTION BETWEEN CUTTING L AND RAILWAY BRIDGE OVER HENTY RIVER

FIGURE 3

Between these hills and the Henty River is a flattish
area cut in dipping Owen Conglomerate and tilted Permian sediments.
This rises abruptly along steep scarps to the north and falls abruptly
to west and south along steep scarps. To the east it continues across

Permian rocks and the Junee Group until it reaches a more resistant
bed in the Owen Conglomerate which forms a steep dip slope about one
and three quarters of a mile south-east of Firewood Siding.   This
surface is drained by wide, flat valleys.   It is an erosional surface
truncating the tilted rocks (see figs. 2 and 3).   No heights have
been measured on this surface but from ground observations it would
appear at its north-western end to be a few feet lower than the top
of the Tertiary sediments (350 feet above sea-level).   It rises some-
what to the east where it is a little higher than the height of the
Badger River plain about a mile east of Firewood Siding i.e. about
360 feet above sea-level.   This surface at from about 350 feet to
about 400 feet might be called the Firewood Siding Surface.   Separating
this surface from the present plain of the Badger River is an east-
facing obsequent fault-line scarp of the Eden Fault about 40-50 feet
in height.   The plain east of this scarp is in Gordon Limestone that
to the west in more resistant Permian mudstone and sandstone.
Another flattish area occurs on the northern side of the Badger River
north-west of Malanna.   This also is eroded in tilted Permian sediments
and is at about the same level as the Firewood Siding Surface.

The sharp ridge north of Firewood Siding is a compound
one as a fault appears to pass across it near its southern end.   Near
the southern end it is made up of two connected cuestas but a structural
control for its northern end is not clear.   The hills south of
Firewood Siding forming part of the Henty Surface have steep southerly
facing scarps and gentler, although still steep, northern slopes.

The southern scarp is a cut-off scarp of the creek shown on the map
(fig. 2) now dissecting the Firewood Siding Surface.   The northern
slope is not a dip slope as shown by numerous dip readings in the
Permian rocks in the railway cuttings.

On the hills east of Henty Siding, a siding about a mile
south of the railway bridge over the Henty River, there are exposures
of gravels containing large blocks of Permian sandstone and dolerite.
These form part of a surface at about 306 feet above sea-level.   The
hill west of cutting H has a height of 235 ($\pm$5) feet above sea-level,
and is in more or less accordant with the hills immediately to north and
south.   The maximum height reached by the Tertiary sediments (in
Cutting L, fig. 2) is 350 ($\pm$5) feet above sea-level.

Further west is a low sand covered area, now vegetated
and beyond this an area of active dunes.

The two surfaces in the area, the Henty Surface and the
Firewood Siding Surface, show no signs of glaciation in this area.
The Firewood Siding Surface is shown on Gregory's map (1903, pl. XX)
as part of his Western Peneplain. On neither surface are there any
erratics locally, no ice polished, or ice-scoured surfaces and no
sign of ice plucking. Although David noted roches moutonnées in the
Eden Valley he gave no details and the present authors are unable to
find any.   There are no signs of the passage of ice over any of the
scarps and some of the ridges, e.g. that just south of Firewood Siding
containing the measured hills, have steep irregular faces to the east
(the alleged direction from which the ice sheet came) and a more

gentle slope to the west. The sharp crested ridge north of Fire-
wood Siding is very steep sided and, in plan, convex to the west.
It could on superficial examination be mistaken for a moraine,
especially as it has blocks of rock scattered over its surface.
However, detailed examination of this ridge shows it to be, at least
partially, a cuesta in Upper Permian quartz sandstone and conglomerate
and the scattered boulders to be entirely of these materials.

The Firewood Siding Surface is succeeded to the west
and south by an area of lower hills. (see fig 3,a)   Just south of
the Badger River is a south-east trending fairly sharp-crested ridge
cut by a south-westerly flowing stream. In many places the crest of
the ridge is demonstrably pebbly. Its height is 255 ($\pm$5) feet above
sea-level.   South of the end of this ridge and separated from it by
a deep valley is a sharp peaked hill with a slight tendency to a
south-south-westerly trend then a swing to the south-east. No out-
crop occurs but boulders and pebbles are spread over the surface.
This is separated by a creek valley and a swampy flat from a ridge
which is cut by the railway line just north of Koyule and trends
east-south-east.   The surface of this ridge is gravel covered by the
railway section reveals the presence of other types of sediments.   It
is clear that the ridge is not a depositional feat/ure and where cut by
the railway line it is anticlinal in structure.   A superficial
examination of these ridges may well suggest terminal moraines, more
or less convex to the sea, later breached by streams. However, the
section through the southern one clearly shows that it is not
morainal and sections in railway cuttings just behind the others

suggest the same. These considerations together with lack of glaciation in the hinterland indicate that these ridges are not moraines. Thus in the Malanna area there is no physiographic evidence of glaciation.

CAINOZOIC DEPOSITS

The earliest record of Cainozoic deposits in this area is that of Montgomery (1890) who noted the presence of clays in the first creek valley west of the railway bridge and near the Henty Ferry and suggested their equivalence to the Macquarie Harbour Beds. In 1892 Johnston noted the presence of lignite in the same area and recorded a _Fagus_ close to _F._ (now _Notofagus_) _cunninghami_ and an _Acacia_ close to _A. melanoxylon._ To him the close resemblance of these two forms suggested that the lignites were "of a more recent date than any other lignite formation hitherto described". Gregory (1904, p. 51) described some of the rocks in the railway cuttings. Boulder clays with boulders of Owen Conglomerate and decomposed dolerite up to 2 feet across were mentioned. Gregory described the boulders as lying at all angles and having a shape characteristic of ice-action most of them having one or more flattened surfaces. David (1926, pp. 94-95) described the blocks in the northernmost railway cutting as up to 5 feet in diameter and all rounded, although elsewhere on p. 95 he states that the shape of many is obviously glacial. In a footnote on page 102 David notes that the sequence is more complex than he had depicted. He states that the redistributed glacial beds at Henty Siding pass below sea-level, are capped by lignitic shale and sandstone and faulted.

It will be convenient to describe the deposits exposed in the railway cuttings between the railway bridge over the Henty River and Malanna in order from south to north.

Cutting A (see Map).

At the southern end of the cutting a succession dipping 240° at 6° is exposed. At the base is a bed of gravel at least twelve feet thick and consisting mainly of rounded boulders of dolerite up to 30 inches long with some boulders of Permian sandstone, siltstone and conglomerate. Some of the boulders are sub-angular to angular. There is a suggestion of an upward decrease in grainsize although this is not marked. This is overlain by 2 feet of clayey sand and then 1 foot 10 inches of clayey pebbly sand with lignitic fragments used in radiocarbon dating. This latter bed of sand is cross-bedded, and contains some conglomerate bands in which there are pebbles of siltstone. In both of these sandy beds there are branching, cylindrical ferruginous concretions which are in some cases around lignitic fragments. The next bed is a conglomerate composed mainly of fragments of Permian siltstone with some lignite fragments. This is 1' 8" thick at its northern end but thickens to the south-west and becomes more conglomeratic in that direction. Some cross-bedding is present in the sandy matrix and the currents came from the north-east or north. This conglomeratic bed has an irregular lower surface and this forms a prominent overhang in the face of the cutting. The final bed in this succession is at least 15 feet thick and consists of gravels with sandy lenses showing cross-bedding indicating currents from

north-east or north. Some clay lenses are also present. The
main boulders are composed of Permian rocks and dolerite.

This succession is affected by two normal faults forming
a small graben with a throw of a few feet. Further north in the
cutting the succession is hidden for an interval by sand and vege-
tation. Beyond this gravels again occur. They are sandier than
those in the southern end of the cut and contain rounded boulders of
dolerite near the base with boulders of Permian rocks and Owen Con-
glomerate becoming more common near the top. At the extreme north-
ern end of the cutting these are overlain by a lignitic bed, then
more gravels, sand and finally a lignitic bed. The lower of these
lignitic beds dips 50° at 40°. About halfway along the western wall
of the cut sands and interbedded carbonaceous sand or peat abut dis-
conformably against the gravels.

Cutting B.

This cutting is mainly in Permian rocks which are described
elsewhere (Banks and Ahmad, in press). The Permian rocks are over-
lain by a bed of gravel 2 feet thick which in one place near the
southern end of the eastern bank of the cutting occupies an old gully
a few feet deep. The gravel is sandy and contains rounded to sub-
angular boulders of Owen Conglomerate, quartz, quartzite and Permian
sandstone. This is overlain by about three feet of soil.

Cutting C.

In this cutting Permian rocks are overlain by a couple
of feet of gravel and then about three feet of soil.

Cutting D.

The beds in this cutting dip 85° at 28°. At the base is a cross-bedded sandstone with some gravelly layers, which becomes lignitic and clayey towards the top. This is followed by sand and then after a slight gap, sand with layers of peaty sand and boulders. Another cross-bedded sandstone follows. It is pale yellow brown in colour and has rare boulders. The cross-bedding is due mainly to currents coming from the north-east. This sandstone passes up into sandy clay, clay and then lignite. The top beds in the cutting are pebbly sands with three thin beds of boulders near the base and a bed of pebbles higher up. Cross-bedding dipping south-west is present. Boulders in these cuttings include those of Owen Conglomerate, Permian sandstone and weathered dolerite. These beds are overlain by a gravel as in cuttings B and C.

Cutting E.

At the southern end of this cutting there is a gravel with boulders up to four feet long which is rudely bedded and contains pockets of pebbles. There is a high proportion of boulders in the gravels which range in size from a quarter of an inch up. The boulders are well rounded but the sphericity is frequently low. The boulders include those of dolerite, weathered and fresh Permian rocks including conglomerates and sandstone, quartzite, clay and Tertiary sandstone. The small dolerite boulders are completely weathered, the larger ones to a lesser extent. This part of the cutting is overlain by about three feet of soil.

To the north the gravels are overlapped by white cross-bedded, sandstones with interbedded clays and lignified wood, seeds and leaves with some pyrite nodules. Some of the larger fragments of lignified wood are still standing upright (i.e. are in growth position). The cross-bedding appears to be due to currents coming from the north-east. The succession in this part of the cutting dips $230°$ at $26°$ and is overlain unconformably by a surface gravel with boulders of Owen Conglomerate.

The succession was measured in some detail and is shown below:

Top

> Gravel with boulders of Owen Conglomerate
> Unconformity

18 feet: white, fine to medium grained sandstone with thin bedding.

1 foot: coarse siliceous conglomerate with boulders of Permian
> sandstone.

4 feet: white, medium-grained, unbedded quartz sand.

1 foot: coarse siliceous conglomerate with boulders of Permian
> sandstone.

4 feet 6 ins.: grey clay with lignified plant remains near top and
> about 1 foot from top; top foot is fissile; some pyrite.

3 feet 6 ins.: white, grey, or white with red streaks, clay with
> cylindrical, spherical, branching and irregular limonitic
> nodules.

7 feet: yellow to white, fine to medium grained sandstone consisting
> of quartz with clay cement; friable; grains sub-angular to

sub-rounded with a few rounded pebbles.

5 feet; white, medium grained sandstone with occasional bands of
pebbles of Permian sandstone.

1 foot 6 ins.; conglomerate with rounded, elliptical pebbles and
cobbles up to 8 inches long, mainly of Permian sandstone.

1 foot; medium grained sandstone without pebbles.

9 feet; sandstone with coarse bands of angular grains, with 9 inch
pebble band with angular to rounded pebbles of several size
of quartz, quartzite, Owen Conglomerate and Permian sandstone

15 feet; very pale, sticky clay.

4 feet; gap in section.

6 feet 6 inches; pale brown, cross-bedded sandstone.

2 feet 6 inches; clay, grey with plant stems and some carbonaceous
bands.

6 feet 6 inches; pale brown, cross-bedded, finely bedded sandstone;
quartzose, mostly fine grained bed some beds of medium to
coarse grained sand; in the latter of which the grains are
distinctly angular.

16 feet; grey clay with plant stems and some carbonaceous bands.

26 feet; sandstone, white to yellow, medium-grained, quartzite,
thinly bedded.

5 inches; conglomeratic, yellowish sandstone.

13 feet; white to pale yellow sandstone with a few conglomeratic
bands.

9 feet; conglomeratic sandstone with cross-bedding on small scale
dipping south-west; sub-rounded, sub-angular and some

rounded pebbles and cobbles up to 6 inches long of Owen Conglomerate and Permian sandstone.

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133 feet (approx)

At the northern end of the cutting the south-westerly dipping basal beds in the above succession are overlain unconformably by gravels with boulders of Owen Conglomerate.

Cutting F.

At the southern end of this cut a fault throws cross-bedded sandstone to the south against beds of gravel and cross-bedded sand to the north. The cross bedding dips north. The beds are almost horizontal and are from one foot to two feet thick. The boulders which are up to 3 feet long and are sub-angular to sub-rounded and rounded. They consist of Permian sandstone and siltstone, quartzite, quartz and dolerite. Many of the boulders are deeply weathered. The total thickness is about 30 feet. Further north again are sands containing a few large pebbles.

Cutting G.

The main rock types is a conglomerate lacking bedding and containing sub-angular to sub-rounded boulders up to several feet in diameter with a few rounded. They consist mainly of Permian sandstone which is deeply leached and exfoliated. The matrix is sandy and there are some sandy bands, one of which at least is in a washout.

Cutting H.

Gravels are overlain by sands. The boulders are up to four

feet long and are mainly Permian sandstone near the base but a few dolerite boulders occur near the top. The sands are cross-bedded with the dip of the cross-bedding to the south.

Cutting I.

Sands at the base are overlain by lenses of gravel, followed by more sand and then gravel. The dip is at a low angle to the south. The top sands are cross-bedded with some of the cross-bedding traces in the cut dipping south-east while most of them dip north-west.

Cutting J.

Gravels with boulders more than 3 feet long are present.

Cutting K.

The gravels in this cutting contain boulders up to 5 feet long. The boulders include those of Permian sandstone, and leached siltstone, with some quartz and quartzite. The gravels are inter-bedded with a fine grained sand showing cross-bedding dipping north. A washout in the sands is filled with the conglomerate.

Cutting L.

This is the last cutting in which Cainozoic deposits occur. They are gravels with many boulders up to a few feet long of Permian sandstone with a few boulders of dolerite.

The Cainozoic deposits exposed in the railway cuttings con-sist then of more or less unconsolidated rocks, with gravels, cross-bedded sands, clays and lignites being represented. The gravels are commonly bedded and the boulders in them are mainly

sub-rounded. No striated pebbles were found although they were looked for. The rock fragments consist mainly of Permian sandstone, siltstone or granule conglomerate, dolerite, Owen Conglomerate, quartz and quartzite and more rarely fragments of clay or clayey sand or lignite. Some of these boulders are now deeply weathered. It is notable that there is a general increase in grainsize in the boulders from south to north, boulders up to 5 feet long occurring in Cuttings K and L, but all the gravels are not necessarily contemporaneous. It is also significant that the rock types present are all potentially of local derivation and could all come from within three miles to the east. The matrix of the gravels is predominantly sandy and they contain little clay.

Cross-bedding in the sands dips mainly to the south-west, south of Cutting F but north of this cutting the dip of the cross-bedding varies and tends to be northerly in several places. The presence of lignite indicates that some of the beds at least are paludal and no marine macrofossils were seen. There are numerous disconformities and some unconformities suggesting folding or at least tilting before deposition of the later beds.

The radiocarbon dating indicates an age greater than 32,000 years for some of the lignites in Cutting A. (Rubin, M, analysis W444, letter to E.D. Gill). This specimen was submitted on the mistaken idea that the conglomerates in this cutting were morainal and associated with the Malanna glaciation. A considerable age for most of the gravels is indicated also by the extent of weathering of

the dolerite boulders. The deposits are clearly post-dolerite, the dolerite probably being Lower Jurassic. Some of the material from the lignitic beds in Cutting E was submitted to Dr. I. Cookson for palynological analysis. Seeds and seed cases on cones of Banksia of marginata were reported from the sample, so that a Cainozoic age seems likely and possibly an upper Cainozoic age in view of the close resemblance of the seed cases to forms still living in the area. Final dating must await detailed palynological work but the beds might best be considered Upper Cainozoic and this would be in accord with Johnston's record (1892, pp. 12-13) of Acacia from this area.

The rocks in the cuttings thus provide no evidence of a glacial origin and there is no physiographic evidence for glaciation in the area. On these grounds alone the hypothesis of the existence of a Malanna Phase of the Pleistocene Glaciation in the Malanna area, must be considered invalid. Before providing an alternative explanation for the observed facts, a final point in David's argument must be dealt with, that of the "brecciated pavements". To put the features so interpreted by David into their correct perspective the structure of the area must be considered.

STRUCTURAL GEOLOGY.

The structure of the Lower and Middle Palaeozoic rocks to the east has been figured and briefly described by Gill and Banks (1950), and is not relevant to the advancement of the argument and will not be considered further. The Permian rocks near Firewood Siding are faulted against the older rocks. North of Firewood Siding

is a fault, here called the Firewood Siding Fault, trending west-
north-west and downthrowing to the south-south-west. This continues
to about a mile east of Firewood Siding where a fault, the Eden
Fault, trending north-north-east and downthrowing west becomes the
main structure. In general the Permian rocks in the downthrown
block dip to the south-west at angles varying from a couple of
degrees to 50^o, the latter occurring in Cut M, that nearest the
Tertiary beds, and suggesting a fault downthrowing to the south-
west. The dips are all shown on the map (fig. 2).

In cutting S there is a normal fault trending 330^o (all
bearings related to true north) and causing a distinct drag dip
showing west side up. Joints trending 10^o and 325^o are common
in this cutting. On some of the 325^o joints are slickensides
dipping west at a low angle and indicating west block north movement.
Slickensides also occur on some of the bedding planes and show that
the top moved west or north-west over the bottom.

Cutting R is that containing the rocks figured by David
(1926, p. 102). At the eastern end of the cut the beds dip 115^o
at 30^o and show joints trending 360^o and dipping steeply west.
Near the western end is a fault striking 340^o and dipping west at
about 45^o and this is associated with much brecciation. In addition
to these main faults there are many minor normal faults dipping
steeply west with a few normal faults dipping east and several
minor thrusts dipping west. The beds affected are a formation of
quartz sandstones and thinly interbedded fine quartz sandstone and

carbonaceous siltstones. The section figured by David is in the fine sandstone-siltstone alternation and these beds are consistently the ones showing the most brecciation and minor faulting. It is notable that in the main body of the cutting a bed of thickly bedded quartz sandstone at the top of the cut is not affected by brecciation nearly as much as the underlying alternation of fine sandstone and siltstone. The structure along this section is shown as figure 4.

FIGURE 4
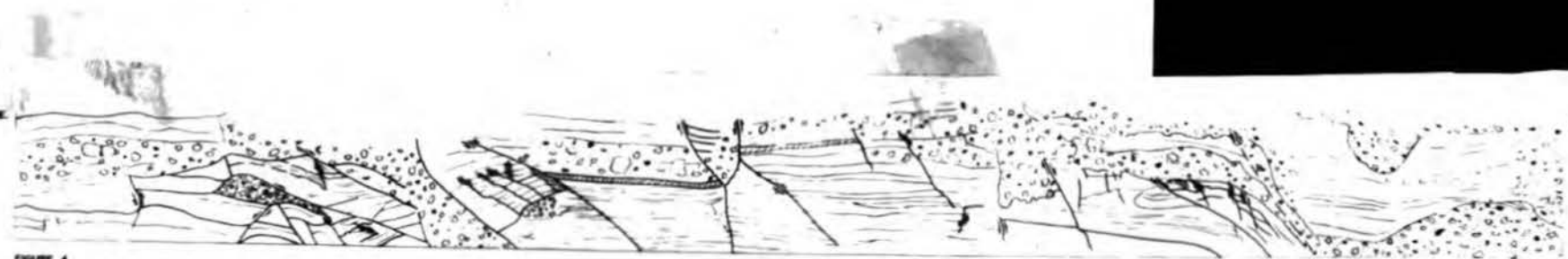
Diagrammatic Section of Cutting 'R' Malanna

SCALE

Near the eastern end of Cutting Q is a normal, westerly dipping fault striking 325°. In the body of the cutting the beds are horizontal. They are strongly jointed in places the main faults trending north and showing horizontal slickensides indicating west side north movement. At the western end is a normal fault downthrowing to the east.

In Cutting P is a normal fault near the western end dipping 190° at 74°. In Cutting O, the beds are almost horizontal although in places they dip west and form a small monocline. They are disturbed by a small thrust dipping west and by joints mainly

trending 340° but with some at 10°. The dip is west in Cutting N and there are minor faults. Joints are common striking 120° and there are some parallel to the cutting which show horizontal slickensides. The Permian sandstone near the end of Cutting M dip 230° at 50° and this steep dip probably indicates a fault down-throwing south-west in the vicinity. Tertiary conglomerates in the next cutting west (L) appear to be horizontal. This may show that the fault is pre-conglomerate but the cuttings are quite a few yards apart so that the drag dip might die out east of Cut L.

The Permian rocks then show numerous normal faults mainly trending north-westerly and forming small horsts and graben. With these are associated minor west dipping thrust faults. The normal faults trend 330°, 340°, 325°, 10° and 280°. Joints trending 300°, 325°, 340°, 0°, 10°, 10°, and north-easterly occur. On the 325° set and the 0° set there is evidence of dextral movement and there is horizontal movement also on the north-easterly set. The faults are consistent with tension from SW - NE and some of the joints fit this pattern also. However, some joints do not seem to be related to this tension and a more detailed analysis is required. One case of bedding plane slip with movement down to the west or north-west was noted.

The major structures are quite inconsistent with faulting produced by an ice sheet moving from east to west. What thrusts are present dip west. The only evidence for ice thrust is the bedding plane slip but this could also occur with normal faulting

of dipping beds. Thus structures in the Permian rocks at Malanna used by David to support his hypothesis of glaciation in the area are inconsistent with this but consistent with normal faulting.

The Tertiary beds also show variations in the dip and some faulting. (see figure 5b) In the northern cuttings they are horizontal or nearly so. In Cutting E they dip $230°$ at $26°$ and in cut D they dip $85°$ at $28°$ so that a syncline may be inferred trending roughly $340°$ between these cuttings. In Cutting A the beds at the north end dip $50°$ at $40°$ and those in the south end dip $240°$ at $6°$. An anticline trending about $325°$ would appear to be present. In Cutting F there is a fault with small displacement and in Cutting A two normal faults forming a graben. The northern one dips $260°$ and $45°$ and the southern one $50°$ at $50°$. This graben would have a trend of about $335°$ and this is close to the main fault direction in the Permian rocks. Approximate co-incidence of the fold axes with the fault directions suggests a genetic connection.

SOME ASPECTS OF THE GEOLOGICAL HISTORY

The Palaeozoic history of the Malanna area has been dealt with elsewhere (Gill and Banks, 1950, and Banks and Ahmad, in press). It is relevant to the present discussion only that the Lower and Middle Palaeozoic sediments were folded into plunging folds and then overlain unconformably by more or less flat-lying Permian sediments. Dolerite intrusions occurred in the Jurassic. At some

later time the Firewood Siding and Eden Faults developed. The
Henty Surface cuts across these faults and thus is later. This
surface cuts across the Lower and Middle Palaeozoic rocks despite
their differential resistance to erosion. It is, therefore, an
erosional surface. Thin layers of gravel occur on it in places
but there is no evidence that it is a stripped surface. It is
certainly not a stripped pre-Permian surface as the Henty Surface
extends inland to the foot of the West Coast Range where it has a
height of 1100 feet approximately. This range is probably a monad-
nock range as postulated by Bradley, (1954, p. 195). The top of
this range is probably part of the stipped pre-Permian surface
(Bradley, 1954, p. 195) as shown by the Permian at Mount Sedgwick
(Edwards, 1941) Mount Read and Mount Dundas, where the base of the
Permian is at over 3000 feet. The Henty Surface is now dissected
by stream valleys up to 700 feet deep but some of the original
surface remains so that it is not likely to be as old as Early
Tertiary. The surface was deeply dissected by rivers prior to the
formation downstream from the Henty River road bridge of the
terminal moraines. These moraines are dissected

little except for the gorge cut through them by the Henty River and are thus probably not very old. Both the glacier responsible for these moraines and the glacier which occupied the Linda Valley are distributaries of the minor ice cap which occupied the West Coast Range between Mount Sedgwick and Mount Tyndall. The advance of the glacier occupying the Linda Valley has been dated as about 26,000 years, i.e. about equivalent to the beginning of the Wisconsin Glaciation, and the moraines below the Henty River road bridge might well be almost contemporaneous. They could probably safely be considered as Upper Pleistocene. Thus the Henty Surface was formed well before the Upper Pleistocene.

The relationship of the sediments in the railway cuttings to the development of the Henty Surface cannot be established on evidence so far available in this area. They may have been deposited in lowlands in a pre-Henty surface of considerable relief. They may have been formed in a graben in this landscape delineated by the Firewood Siding and Eden Faults. The fault postulated between Cuttings L and M. probably preceded the deposition of the gravels in Cutting L because of their horizontal disposition. At a later stage the Henty Surface would have been eroded in the Lower and Middle Palaeozoic, Permian and Cainozoic sediments. On the other hand the Cainozoic sediments may have been deposited as the Henty Surface was developing, streams stripping material off the higher country and depositing in the valleys until a surface, partly erosional and partly depositional, was formed. Yet again .

the sediments may have resulted from uplift of the Henty surface
along a fault line between Cuttings L and M with formation of
alluvial fans and fluviatile plains against the steep fault scarp.
The sediments may have gradually filled up the fault lowland until
they reached a profile of equilibrium related to the Firewood
Siding Surface. No evidence is yet available to choose between
these alternatives.

The surface from which the Cainozoic sediments were derived
must have had a considerable slope, at least in places, to account
for the large particle size in the gravels. Boulders up to five
feet in diameter in bedded deposits with a sandy matrix imply a
considerable velocity and volume of water and thus a steep gradient.
Variations in the competence of the depositional currents are
shown by the occurrence of interbedded gravels, sands and clays.
There is evidence of a systematic variation in current competence
in the succession - gravel, sand, clay (with or without lignite) -
which is developed completely or incompletely eight times in the
sediments of Cutting E and three times each in Cutting D and
Cutting A. The recurrent increase in competence represented by
gravels, in many places associated with local disconformities, may
be due to recurrent increases in rainfall following periods of
lower rainfall, or to recurrent uplift due to faulting or lowering
of base level. On several occasions peaty swamps were present.
The cross-bedding in the sands suggests two sources, one to the
north or north-east of Cutting L and the other somewhere between
Cuttings E and F. Thus cross-bedding dipping in a generally

southerly or south-westerly direction occurs in most cuttings but cross-bedding dipping north or north-west occurs in Cuttings F, I and K. The presence of dolerite boulders in all cuttings suggests a northerly derivation, the dolerite mass near Firewood Siding probably being the source. The Owen Conglomerate boulders in the older deposits and superficial gravels in and south of Cutting E suggest a partial derivation of sediments in these cuttings from the east, the nearest Owen Conglomerate being about a mile and three quarters south-east of Firewood Siding. Some of these sediments are older than 32,000 years and some of them are older than the Firewood Siding Surface. In all cases it is probable that they are older than the last phase of the Pleistocene glaciation. Subsequent to deposition of the sediments in Cuttings A to E folding and faulting occurred. The trends of the folds and faults are almost parallel and this suggests genetic connection. They may well be due to slumping on a clay bed down a slope trending about 340° and dipping south-west. This roughly parallels many of the faults in the Permian rocks, and the slope may have been a fault scarp.

At some time after the development of the Henty Surface it was uplifted. Partial erosion of this surface produced the Firewood Siding Surface, cut in Lower Palaeozoic, Permian and Cainozoic sediments (at least those in Cutting L). The Henty River appears to have been a meander, ~~later entrenched,~~ after in this surface and ~~thus~~ uplift, was entrenched in it ~~was flowing during development of the surface.~~ At the time of development of the surface the Henty River locally had a height of

about 400 feet relative to present sea-level and must have been higher further upstream. As the terminal moraines on the Henty River have a height of only about 250 feet above sea-level at their downstream termination and there is no evidence of displacement or tilting of the Henty Surface between Malanna and the Henty River road bridge the moraines must be later than the development of the Firewood Siding Surface. Thus this surface is ~~probably~~ pre-Upper Pleistocene. The surface falls from about 400 feet at the Eden Fault to 350 feet at Cutting L, a distance of about two miles so that there is a fall of about 25 feet per mile. The base level controlling this surface is not known. "Beach terraces" at a level of about 400 feet above present sea-level are recorded by Twidale (1957, p. 12) from just north of the Pieman River, about sixteen miles north of the Malanna area, where they are incised into a higher plateau. No evidence is quoted for a marine origin for these "beach terraces", nor near Malanna are any marine deposits known at this level. It is not, of course, clear that the "beach terraces" and the Firewood Siding Surface are in any way related but there is some parallel in the relationship of the two features to a higher surface and sea-level. The Firewood Siding Surface postdates both the sediments in the railway cuttings and the Henty Surface. Since development of the Firewood Siding Surface there has been rejuvenation and the knick-point on the Badger River has moved upstream about three quarters of a mile.

For fuller and more accurate reconstruction of the Cainozoic history and palaeogeographies, detailed sedimentation studies will

have to be made and some method of correlation evolved which is
applicable to the terrestrial sediments. This may well involve
comparison of quantitative pollen analyses of the lignites. In
addition more information on rock distribution and detailed study
of contour maps of this and neighbouring areas will be needed.

SUMMARY AND CONCLUSIONS

After development of the Firewood Siding and Eden Faults in
a terrain of Palaeozoic sediments and dolerite, erosion produced the
Henty Surface. Before, during, or after formation of this surface
a thickness of at least a few hundred feet of terrestrial gravels,
sands, silts and lignites was deposited. The grainsize of some of
the gravels indicates considerable gradient for the transporting
streams. The succession, gravels, sand, silt or clay, is repeated
completely or incompletely at least eight times.

Uplift of the Henty Surface was followed by erosion which
finally produced the Firewood Siding surface. This latter surface
developed after deposition of some of the sediments. Further
uplift resulted in erosion of the Firewood Siding surface. After
this uplift glaciation affected the West Coast Range and the Henty
River valley as far downstream as the road bridge about eight miles
east of the Malanna area at a height of about 250 feet above sea-
level. This glaciation was probably equivalent to the Wisconsin
in the Northern Hemisphere. The uplifted Henty and Firewood
Siding Surfaces are locally being dissected by streams which, at
least in the Lower and Middle Palaeozoic rocks, are strongly

structurally controlled.

In the Malanna area there is no evidence of Pleistocene glaciation, whether by ice-sheet or otherwise, and the physiographic sedimentational and structural evidence advanced by David (1926) in support of glaciation in this area were misinterpreted. More detailed studies indicate that all the facts known about the area suggest a Cainozoic history involving faulting, deposition of terrestrial sediments by streams, or in lakes and swamps and erosion by streams. It is concluded therefore that there is no longer any justification for retaining the term "Malanna Phase" for a Pleistocene glacial phase based on this area. Use of the term should be discontinued. Should future work indicate an early phase of Pleistocene glaciation in Tasmania a new geographic term should be used.

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LOCALITY INDEX.

	Latitude	Longitude
Badger River	41°59'	145°14'
Firewood Siding	41°59'	145°16'
Geologists' Creek	42°00'	145°16'
The Gooseneck	41°52'	145°35'
Henty River	42°02'	145°18'
Henty Siding	42°02'	145°15'
Malanna	42°01'	145°13'
Mount Professor	41°59'	145°22'
Mount Road	41°50'	145°30'
Mount Sedgwick	42°00'	145°35'
Pieman River	41°50'	144°55'
West Coast Range	(41°44'	(145°33'
	42°18'	145°38'