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The Ohio State University, Ph.D., 1968
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GLACIOLOGICAL INVESTIGATIONS NEAR THE ICE SHEET
MARGIN, WILKES STATION, ANTARCTICA

DISSERTATION

Presented in Partial Fulfillment of the Requirements for
the Degree Doctor of Philosophy in the Graduate
School of The Ohio State University

By
Caspar Cronk, A.B.

* * * * * *

The Ohio State University
1968

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ACKNOWLEDGMENTS

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Studies in Geophysics. Professors Colin B. B. Bull and Howard J. Pincus

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## CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACKNOWLEDGMENTS</td>
<td>11</td>
</tr>
<tr>
<td>VITA</td>
<td>iv</td>
</tr>
<tr>
<td>TABLES</td>
<td>ix</td>
</tr>
<tr>
<td>FIGURES</td>
<td>x</td>
</tr>
<tr>
<td>INTRODUCTION</td>
<td>1</td>
</tr>
<tr>
<td>Definition of Certain Terms</td>
<td>5</td>
</tr>
<tr>
<td>Glaciological Principles</td>
<td>9</td>
</tr>
<tr>
<td>Location</td>
<td>17</td>
</tr>
<tr>
<td>Station History and Observations</td>
<td>23</td>
</tr>
<tr>
<td>Meteorological measurements</td>
<td>24</td>
</tr>
<tr>
<td>Accumulation and Ablation measurements</td>
<td>24</td>
</tr>
<tr>
<td>Pit studies</td>
<td>29</td>
</tr>
<tr>
<td>Ice temperatures and coring</td>
<td>29</td>
</tr>
<tr>
<td>Movement and strain measurements</td>
<td>30</td>
</tr>
<tr>
<td>Gravity measurements</td>
<td>31</td>
</tr>
<tr>
<td>Other studies</td>
<td>31</td>
</tr>
<tr>
<td>Australian studies</td>
<td>31</td>
</tr>
<tr>
<td>DETERMINATION OF THE MASS BALANCE</td>
<td>33</td>
</tr>
<tr>
<td>Determination of Specific Balance Values</td>
<td>35</td>
</tr>
<tr>
<td>Balance year</td>
<td>35</td>
</tr>
<tr>
<td>Reduction of stake measurements</td>
<td>36</td>
</tr>
<tr>
<td>Establishing the Equilibrium-line Position</td>
<td>46</td>
</tr>
<tr>
<td>MASS BALANCE IN RELATION TO MOVEMENT</td>
<td>51</td>
</tr>
<tr>
<td>The Mass Balance Near Stake 402</td>
<td>51</td>
</tr>
<tr>
<td>Strain Calculation and Vertical Velocity</td>
<td>51</td>
</tr>
<tr>
<td>Survey errors</td>
<td>58</td>
</tr>
<tr>
<td>Criticism of the calculation method, various assumptions</td>
<td>60</td>
</tr>
<tr>
<td>Velocity-versus-Depth Profile</td>
<td>63</td>
</tr>
<tr>
<td>Calculation of the velocity profile</td>
<td>69</td>
</tr>
<tr>
<td>Volume of Ice Passing Stake 402 Annually</td>
<td>76</td>
</tr>
<tr>
<td>Equivalent accumulation area</td>
<td>78</td>
</tr>
<tr>
<td>Calculation at S-1</td>
<td>83</td>
</tr>
<tr>
<td>Other Temperature Calculations</td>
<td>89</td>
</tr>
</tbody>
</table>
CONTENTS (Contd.)

SIGNIFICANCE OF TEMPERATURE AND VELOCITY CALCULATIONS ................................................................. 93

Basal Sliding ................................................................................................................................................. 93
Erosion related to bottom sliding .................................................................................................................. 95
Thule-type Marginal Moraine Formation ..................................................................................................... 97
The shear hypothesis ...................................................................................................................................... 97
The Weertman mechanism of basal freezing ............................................................................................... 101
Debris raised by compressive flow--Wilkes Station conditions ................................................................ 112
The Effect of Debris on the Flow Law--Surfaces of Apparent Shearing ...................................................... 125
Basal Conditions Inland from the Moraine at Thule, Greenland ................................................................. 130
Suggested Sequence for Moraine Formation ............................................................................................... 131
Age of the Windmill Island Ice-free Area and Comparison with Other Areas ................................................ 136

ANALYSIS OF STAKE AND RELATED MEASUREMENTS .............................................................................. 138

General Meteorological Conditions ................................................................................................................ 138
Seasonal Changes in Surface Conditions in the Wilkes Area ......................................................................... 143
Processes Contributing to Accumulation and Ablation ................................................................................ 146
Precipitation .................................................................................................................................................. 146
Blowing snow--Wind transport and deposition ............................................................................................ 147
Comparison of coastal winds at Wilkes with winds at S-2 ......................................................................... 151
Wind controlled accumulation and deflation near Wilkes Station ............................................................... 153
Condensation and evaporation ....................................................................................................................... 156
Blowing snow in relation to evaporation and condensation ......................................................................... 161
Superimposed ice ......................................................................................................................................... 162
Melt and run-off ............................................................................................................................................ 163
Relative contribution of the various processes to the specific balance--preparation of table ..................... 164

DISCUSSION OF THE BALANCE PROFILE ................................................................................................. 169

Variation in Specific Balance Along the Profile .............................................................................................. 169
Importance of Albedo Differences ................................................................................................................ 174
Deflation-versus-Melt and Evaporation ........................................................................................................ 175
# CONTENTS (Contd.)

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>The Balance Profile above the Marginal Moraine—Conclusions</td>
<td>176</td>
</tr>
<tr>
<td>Mass Balance in the Islands and on the Ramp</td>
<td>177</td>
</tr>
<tr>
<td>Resumé of Maudheim Results for Comparison</td>
<td>180</td>
</tr>
<tr>
<td><strong>CONSIDERATION OF TEMPERATURE MEASUREMENTS</strong></td>
<td>182</td>
</tr>
<tr>
<td>Possible Short Term Temperature Variation</td>
<td>182</td>
</tr>
<tr>
<td>Evidence from pits and cores</td>
<td>182</td>
</tr>
<tr>
<td>Temperature variation as shown by direct measurements</td>
<td>185</td>
</tr>
<tr>
<td>Evidence from 10-meter Ice Temperature—versus-Altitude Curve</td>
<td>186</td>
</tr>
<tr>
<td>Non-linearity of the 10-meter ice temperature—versus-altitude curve</td>
<td>193</td>
</tr>
<tr>
<td>Equilibrium Line Location from 10-meter Temperatures</td>
<td>197</td>
</tr>
<tr>
<td>Wilkes Station Temperature in Relation to the Temperature-versus-Altitude Curve</td>
<td>198</td>
</tr>
<tr>
<td><strong>CONCLUSIONS</strong></td>
<td>201</td>
</tr>
<tr>
<td>Ice Drainage in the Wilkes Area</td>
<td>201</td>
</tr>
<tr>
<td>Relationship of Exposed Rock Areas to the Extent of Ice in Surrounding Areas</td>
<td>208</td>
</tr>
<tr>
<td><strong>REFERENCES</strong></td>
<td>212</td>
</tr>
</tbody>
</table>
TABLES

Table | Page
--- | ---
1. Dates of Stake Emplacement and Abandonment | 27
2. Temperature, $\dot{\epsilon}$, and $\tau$, as a Function of Depth at Stake 402 | 72
3. Velocity as a Function of Depth and the Difference from the Surface Velocity as a Function of Depth at Stake 402 | 75
4. Balance Values at Stakes Near the Ice Sheet Margin in Water Equivalent | 79
5. Cumulative Balance Along a Strip 1 cm Wide Inland from Stake 402 | 82
6. Temperature, $\tau$, $\dot{\epsilon}$, and Difference from Surface Velocity, as a Function of Depth at S-1 | 87
7. Basal Temperatures Along the S-2 Trail, Calculated Assuming 3 Different Linear Temperature Gradients | 91
8. Velocity as a Function of Depth at Stake 402, and Calculated Velocity and Difference from Surface Velocity as a Function of Depth at the Model Equilibrium Line | 115
9. Values of $x$ and $t$ as a function of Depth for Three Initial Values of $a$ | 120
10. Estimates of the Relative Importance of the Processes of Accumulation and Ablation at Six Locations from the Ice Sheet Margin to S-2 | 165
# FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.</td>
<td>Air View of the Northern Portion of the Windmill Islands</td>
<td>3</td>
</tr>
<tr>
<td>2.</td>
<td>Air View Showing Wilkes Station and Exposed Areas of Clarke Island</td>
<td>4</td>
</tr>
<tr>
<td>3.</td>
<td>Diagram Illustrating the Relationship Shear Stress, ( \gamma = \bar{\rho}gh \sin \alpha )</td>
<td>11</td>
</tr>
<tr>
<td>4.</td>
<td>Relationship between Strain Rate, ( \dot{\varepsilon} ), and Shear Stress ( \gamma )</td>
<td>11</td>
</tr>
<tr>
<td>5.</td>
<td>Diagrams Illustrating Conditions for Extending Flow</td>
<td>16</td>
</tr>
<tr>
<td>6.</td>
<td>Wilkes Station Location Map</td>
<td>18</td>
</tr>
<tr>
<td>7.</td>
<td>Windmill Islands Map</td>
<td>19</td>
</tr>
<tr>
<td>8.</td>
<td>Topographic Map of Domal Area Inland from the Windmill Islands</td>
<td>20</td>
</tr>
<tr>
<td>9.</td>
<td>Topographic Profile Along the S-2 Trail from Grinell Nunatak to S-2</td>
<td>22</td>
</tr>
<tr>
<td>10.</td>
<td>Monthly Changes in Surface Level (Snow Accumulation as Measured) at Fl-2</td>
<td>37</td>
</tr>
<tr>
<td>11.</td>
<td>Three-stake Average of the Monthly Changes in Surface Level at Stakes 415, Fl-2, and 416</td>
<td>39</td>
</tr>
<tr>
<td>12.</td>
<td>Averages of Monthly Surface Changes for Six Groups of Stakes Representative of Different Portions of the Accumulation Profile from Stake 402 to BF-2</td>
<td>40</td>
</tr>
<tr>
<td>13.</td>
<td>Comparative Changes in Snow Surface Level during 1958 and 1959 for the Groups of Stakes Considered in Fig. 11</td>
<td>43</td>
</tr>
<tr>
<td>14.</td>
<td>Profile of 3-Stake Averages of Snow Accumulation over Coastal Portions of the S-2 Trail</td>
<td>44</td>
</tr>
<tr>
<td>Figure</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>--------</td>
<td>-----------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>15.</td>
<td>Profile of the Balance in Water Equivalent at Individual Stakes Over the Coastal Portion of the S-2 Trail</td>
<td>47</td>
</tr>
<tr>
<td>16.</td>
<td>Profile of 3-Stake Weighted Averages of Balance in Water Equivalent Over the Coastal Portion of the S-2 Trail</td>
<td>47</td>
</tr>
<tr>
<td>17.</td>
<td>Profile Showing Balance in Water Equivalent Over a 2-Year Period at Individual Stakes and for 3-Stake Weighted Averages</td>
<td>48</td>
</tr>
<tr>
<td>18.</td>
<td>Profile of Balance in Water Equivalent from Low 1958 to Low 1959 at 5-Mile Flags Along the S-2 Trail from Mile 10 to Mile 45</td>
<td>48</td>
</tr>
<tr>
<td>19.</td>
<td>Views of the Area Near Stake 402</td>
<td>52</td>
</tr>
<tr>
<td>20.</td>
<td>Diagram Showing Relative Positions and Velocities of Movement Stakes in the Vicinity of Stake 402 and the Moraine</td>
<td>53</td>
</tr>
<tr>
<td>21.</td>
<td>Sketch Showing the Relationship between Area Strain and Vertical Velocity</td>
<td>56</td>
</tr>
<tr>
<td>22.</td>
<td>Cross Section Showing the Probable Ice Thickness Near the Edge of the Ice Sheet</td>
<td>56</td>
</tr>
<tr>
<td>23.</td>
<td>Diagram Showing the Orientation of the Stress Axes and the Velocity Components</td>
<td>65</td>
</tr>
<tr>
<td>24.</td>
<td>Graphs of Shear Stress, $\tau$, and Strain Rate, $\dot{\varepsilon}$, as a Function of Depth at Stake 402</td>
<td>72</td>
</tr>
<tr>
<td>25.</td>
<td>Graph Showing the Horizontal Velocity as a Function of Depth at Stake 402</td>
<td>74</td>
</tr>
<tr>
<td>26.</td>
<td>Graph Showing the Difference between the Surface Velocity and the Velocity at Depth $y$ Parallel to the Surface as a Function of Depth at Stake 402</td>
<td>75</td>
</tr>
<tr>
<td>27.</td>
<td>Sketch Illustrating the Calculation of the Ice Volume Passing a Cross Section at Stake 402</td>
<td>77</td>
</tr>
</tbody>
</table>
FIGURES (Contd.)

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>28.</td>
<td>Sketches Illustrating the Calculation of the Volume of the Positive Balance</td>
<td>81</td>
</tr>
<tr>
<td></td>
<td>Along a Line Inland from Stake 402</td>
<td></td>
</tr>
<tr>
<td>29.</td>
<td>Graph of the Calculated Temperatures as a Function of Depth at S-1</td>
<td>86</td>
</tr>
<tr>
<td>30.</td>
<td>Graphs of Shear Stress, τ, and Strain Rate, ε, as a Function of Depth at</td>
<td>87</td>
</tr>
<tr>
<td></td>
<td>S-1</td>
<td></td>
</tr>
<tr>
<td>31.</td>
<td>Graph Showing the Difference between the Surface Velocity and the Velocity</td>
<td>88</td>
</tr>
<tr>
<td></td>
<td>at Depth y Parallel to the Surface, as a Function of Depth at S-1</td>
<td></td>
</tr>
<tr>
<td>32.</td>
<td>Sketches Showing Possible Velocity Distributions Across a Shear or Debris</td>
<td>100</td>
</tr>
<tr>
<td></td>
<td>Surface</td>
<td></td>
</tr>
<tr>
<td>33.</td>
<td>Sequence of Events After the Initiation of Several Closely Spaced Shear</td>
<td>102</td>
</tr>
<tr>
<td></td>
<td>Surfaces</td>
<td></td>
</tr>
<tr>
<td>34.</td>
<td>Idealized Distribution of Temperature within an Ice Sheet</td>
<td>105</td>
</tr>
<tr>
<td>35.</td>
<td>Diagram Showing the Relationship of the Position of the 0°C Isotherm to the</td>
<td>105</td>
</tr>
<tr>
<td></td>
<td>Edge of the Ice Sheet and the Resultant Changes in Thermal Conditions at the</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Base of an Ice Sheet</td>
<td></td>
</tr>
<tr>
<td>36.</td>
<td>Diagrams Showing the Incorporation of Debris into the Ice Sheet under</td>
<td>108</td>
</tr>
<tr>
<td></td>
<td>Non-steady State Conditions</td>
<td></td>
</tr>
<tr>
<td>37.</td>
<td>Sketch Showing a Possible Mechanism for the Inclusion of Disseminated Debris</td>
<td>111</td>
</tr>
<tr>
<td></td>
<td>Layers in Glaciers</td>
<td></td>
</tr>
<tr>
<td>38.</td>
<td>Diagram Showing the Orientation of the Axes and Velocity Components in the</td>
<td>115</td>
</tr>
<tr>
<td></td>
<td>Model</td>
<td></td>
</tr>
<tr>
<td>39.</td>
<td>Diagram Showing the Trajectory of Debris Included within the Ice</td>
<td>119</td>
</tr>
</tbody>
</table>
FIGURES (Contd.)

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>40.</td>
<td>Sketch Illustrating the Differences in Deformation between Clean Ice, Slightly Dirty Ice, and Debris Laden Ice</td>
<td>127</td>
</tr>
<tr>
<td>41.</td>
<td>Photo Showing the Curved Lineations on the Ramp</td>
<td>135</td>
</tr>
<tr>
<td>42.</td>
<td>Sketches Illustrating a Possible Mechanism for the Formation of Ramp Lineations</td>
<td>135</td>
</tr>
<tr>
<td>43.</td>
<td>Drift Densities Observed at Wilkes Station at 3 Levels Plotted as a Function of Reciprocal Wind Velocity, $V_z^{-1}$</td>
<td>150</td>
</tr>
<tr>
<td>44.</td>
<td>A Comparison of Winds at S-2 with Winds at Wilkes Station for the Month of November, 1957</td>
<td>152</td>
</tr>
<tr>
<td>45.</td>
<td>Profile Showing the Variable Depth of New Snow after a Single Storm as Measured by Soundings on November 19, 1958</td>
<td>157</td>
</tr>
<tr>
<td>46.</td>
<td>Profile Showing the Relative Contribution of Winter and Summer Accumulation to the Specific Balance</td>
<td>171</td>
</tr>
<tr>
<td>47.</td>
<td>Photos Showing Large Drifts in the Lee of Rock Prominances</td>
<td>178</td>
</tr>
<tr>
<td>48.</td>
<td>Stratigraphy of the Pit and Core at BF-2</td>
<td>183</td>
</tr>
<tr>
<td>49.</td>
<td>12-Month Running Average of Wilkes Station Temperatures from 1957 to 1965</td>
<td>187</td>
</tr>
<tr>
<td>50.</td>
<td>Graph Showing Differences between the 11-Meter and 16-Meter Temperatures as Measured at S-1 from January 1957 to January 1959</td>
<td>190</td>
</tr>
<tr>
<td>51.</td>
<td>Graph Showing 10-Meter Ice Temperatures along the S-2 Trail as a Function of Elevation</td>
<td>192</td>
</tr>
<tr>
<td>52.</td>
<td>Ice Thicknesses and Bedrock Elevations along a Traverse 300 Miles to the South from S-2</td>
<td>202</td>
</tr>
<tr>
<td>Figure</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>--------</td>
<td>-----------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>53.</td>
<td>Photo of Bedrock High Located in Front of a Moraine Arc to the North of Clarke Island</td>
<td>204</td>
</tr>
<tr>
<td>54.</td>
<td>The Effect of Sea Level Changes on the Position of the Antarctic Ice Margin</td>
<td>209</td>
</tr>
<tr>
<td>55.</td>
<td>Generalized Curve Showing Sea Level Changes from 10,000 Years BP to the Present</td>
<td>209</td>
</tr>
</tbody>
</table>

Map 1. Windmill Islands, Northern Area (from Hollin et al., 1961) in back pocket
INTRODUCTION

As part of the International Geophysical Year glaciological program at Wilkes Station, Antarctica (Lat. 66° 15.4' S, Long. 110° 31.5' E), a study was made of the mass balance of a small marginal portion of the continental ice sheet which covers eastern Antarctica. Around most of the Antarctic continent the edge of the continental ice sheet extends into the sea and the loss of ice, which has flowed from the central parts of the ice sheet, is primarily by calving directly into the sea. Where ice shelves have formed, in areas sheltered from the erosional effects of the sea, bottom melting may be important. Loss due to melting at the upper surface of the ice sheet is very slight at most localities.

Wilkes Station is located at one of the unusual areas where the ice sheet terminates on land. With no ice shelf present, the open sea provides a warming influence, and the occurrence of exposed rock causes additional warming. Here the major loss of ice is by melting at the glacier's upper surface during the summer.

A prominent feature near Wilkes Station is the ice-cored moraine system, similar to that found near Thule, Greenland. The moraines are composed of an ice core with a
thin cover of debris, and often form prominent ridges above the adjacent ice surface. The moraines are located some one to two kilometers inland from the ice terminus, at the top of an ice ramp, and at an elevation about 100 m above that of the ice edge. Individual moraines are often arcuate, but in general they are parallel to the ice edge (see Figs. 1 and 2, also Map 1).

The present dissertation is concerned with the determination of the mass balance of the peripheral portion of the ice sheet near Wilkes Station. A knowledge of the mass balance of the ice sheet in an area such as this is essential before the significance of ice-free areas and Thule-type marginal moraines can be understood, and before the conditions responsible for their development can be determined.

The determination of the mass balance involves the problems of determining the variation of accumulation and ablation with elevation, and their variation with distance from the coast. The detailed weekly stake measurements of accumulation and ablation yield an opportunity to estimate the relative importance of the various processes of accumulation and ablation, and to judge their variation with increasing distance from the coast.

The determination of the equilibrium-line position in the study area is complicated by the presence of a large area of superimposed ice. The location of the equilibrium line, a short distance inland from the moraine, is also
Fig. 1. -- Air view of the northern portion of the Windmill Island, showing Clarke Island, portions of Bailey Island, and the ramp and moraine at the margin of the ice sheet.
Fig. 2.—Air view showing Wilkes Station and exposed areas of Clarke Island (U.S. Navy photo).
considered in relation to the variation with altitude of the 10-m deep ice temperatures. The ablation area includes the moraine and the adjacent region upglacier from it.

In the ramp area, between the moraine and the ice margin, where an area of ablation might be expected, extreme variability of accumulation and ablation make it impossible to measure the net loss. Consequently, to determine the balance of the area in question it is necessary to calculate the vertical velocity and to compare it to the specific balance. Alternately the quantity of ice transported through a cross section at the equilibrium line can be calculated and compared to the accumulation upglacier from that cross section. For these calculations a knowledge of the variation of velocity with depth is required, and also the variation of temperature with depth. With a knowledge or estimate of the distribution of velocity and temperature, it is then possible to infer the local conditions at the base of the ice sheet. The problem of the formation of Thule-type marginal moraines can then be examined and their significance considered.

Definition of certain terms

The term accumulation refers to all processes by which material is added to the glacier surface. Ablation refers to all the processes by which material is lost from the glacier surface. Accumulation consists of precipitation,
including snow and any rain that freezes on contact with the glacier, additional snow brought in by the wind from adjacent areas, and hoarfrost, the product of sublimation onto the surface. The processes of ablation are melting, with the removal of the melt by run-off and evaporation, erosion and removal of material by the wind or deflation, and loss of material by sublimation off the surface. Material can also be melted or refrozen at the basal ice surface, but as yet this cannot be measured directly. Where a glacier terminates in water, the major loss may be by calving, and where ice shelves form, bottom melting is important. The difference between accumulation and ablation is termed the balance and may be either positive or negative. Specific balance (accumulation or ablation) refers to a measurement at a particular point on the glacier, while the term total balance refers to the balance over the total area of the glacier being considered. Where the meaning is made clear by the context, the terms specific or total may be omitted. It is clear that the measured accumulation or ablation, as normally determined at a stake, is more properly called the specific balance at the stake.¹

An ice sheet or glacier can be divided into areas which are defined by the ratio of the snow and ice which accumulates in one year, to that which is lost. That part of

¹For more complete discussion see Meier, 1962. The term balance here is equivalent to Meier's term budget.
a glacier with a positive balance, where total yearly accumulation is greater than total loss, is called the accumulation area. The area with a negative balance is called the ablation area. At the boundary between the areas of accumulation and ablation is a line where the annual accumulation is exactly equal to the annual ablation. On glaciers where all the accumulation is in the form of snow, this line is easily identified at the end of the melt season as the lower limit of the annual snow cover on the glacier. Traditionally this line is known as the firn line, firn being partly metamorphosed snow which is a year or more old. However, under certain circumstances the accumulation adjacent to the so-called "firn line" may be in the form of ice (superimposed ice) from the melting and refreezing of winter snow. Since the boundary between the accumulation and ablation areas is then within a region with ice at the surface, it is preferable to call this line the equilibrium line.

The determination of the balance at Wilkes Station is complicated by such an area of superimposed ice. Near the equilibrium line part or all of the winter's accumulation is melted. The water percolates down until it reaches the underlying ice, and since the ice temperature is below 0°C, the meltwater refreezes. In this way the entire thickness of winter snow may be converted to ice, and the resulting accumulation is superimposed ice rather than firn. The lowermost part of the accumulation area is then composed of
superimposed ice, and the uppermost part of the ablation area is also superimposed ice. The boundary between the areas of annual positive and negative balance cannot therefore be determined by inspection as is readily done where all accumulation is as snow, and accordingly the precise location of the equilibrium line can only be determined by direct measurements, usually on stakes.

Benson has subdivided the accumulation area into three facies (Benson, 1962, pp. viii-ix, pp. 24-25) based on changes in the physical properties of the upper firm layers. These changes are caused by differences in the intensity of melting, which varies primarily with air temperature and therefore with altitude. Above the firm line (equilibrium line) is the soaked facies where the previous winter's accumulation becomes wet throughout, the whole layer being raised to 0°C. Higher and further inland is the percolation facies which is characterized by ice glands and lenses, the result of the refreezing of melt water percolating in channels. In this area only the temperature of the percolation channels is raised to 0°C, while the mean temperature of the upper layers remains below this. The highest division is the dry snow facies where melting is negligible. In the area considered in this dissertation, the dry snow facies is not found, but a superimposed ice facies is present in the vicinity of the equilibrium line. The latter facies is not included in Benson's classification
which is based mainly on conditions in northwestern Greenland.

Glaciological principles

One of the chief aims of glaciological study is to determine the state of the mass balance, and to understand better the relationship between changes in the mass balance and changes in the climate. When the mass balance of a glacier is in equilibrium the total accumulation over the whole glacier is exactly equal to the total ablation. The excess of accumulation over ablation in the accumulation area, or total positive balance, is equal to the excess of ablation over accumulation in the ablation area, or total negative balance, and the volume of material transported from the accumulation area to the ablation area must be equal to each of these quantities. Stated more generally, in a glacier in equilibrium, the volume of material which flows through a given cross section in a year is equal to the total of accumulation less ablation in the area upglacier from that cross section. This volume is also equal to the total of ablation less accumulation in the area downglacier from that cross section. Most glaciers are very nearly in equilibrium, changes in their terminus position or thickness being due generally to a rather small, but sometimes systematically repeated, imbalance.
At each point on a glacier the ice thickness is determined by the flow law of ice, the relationship between shear stress and strain rate, and by the slope of the upper and lower surfaces of the glacier. It should be pointed out that the stress which is important in the determination of glacier movement is not hydrostatic stress, but is the shear stress parallel to the glacier bed. This is caused by the downslope component of the weight of the ice column at the point considered. For the simplified case with upper and lower surfaces of the glacier parallel to each other and with a rather small slope angle, if \( h \) is chosen perpendicular to the surface as is customary, the shear stress along the bed is given by \( \rho gh \sin \alpha \) (see Fig. 3).

If ice were an ideal plastic substance, the thickness of a glacier would be uniquely determined for a given slope. But since this is not the case, a range of thicknesses is possible. The reason for this becomes apparent from a comparison of the curves of shear stress-versus-strain rate for an ideal plastic and for a quasi-plastic substance such as ice (see Fig. 4). (The behavior described in glaciological literature as quasi-plastic is often referred to as quasi-viscous or viscoplastic.) The curve for a viscous liquid is also included for comparison. Note that for a plastic substance, all differential shearing would be located at the very base of the ice column where, for a certain thickness and slope, the shear stress just equals the yield stress.
\( \bar{\rho} \) - average density
\( g \) - gravitational acceleration
\( h \) - ice thickness (measured perpendicular to the surface)
\( \alpha \) - surface slope angle

Figure 3. Diagram illustrating the relationship shear stress, \( \tau = \bar{\rho}gh \sin \alpha \).

Fig. 4. Relationship between strain rate, \( \dot{\varepsilon} \), and shear stress, \( \tau \), for ice, A, a liquid of constant viscosity, B, and an ideal plastic, C (after Nye, 1951, p. 255).
Similarly in a quasi-plastic substance the shearing occurs primarily in the basal ice.

In a purely viscous material (a Newtonian liquid) a linear relationship exists between shear stress and strain rate. In this case no particular value for ice thickness is favored since the response of the strain rate to a change in stress, produced by a change in thickness (or slope), is the same at any thickness.

In a substance with an ideal plastic behavior, the flow law determines a maximum thickness for a slab of the material resting on a given slope. For thicknesses such that the shear stress is less than the yield stress, the strain rate remains zero so no deformation occurs. When the yield stress is reached, by an increase in $h$ or $\alpha$, the strain rate immediately becomes infinite; the material can support no stress in excess of the yield stress so, for a given slope, a unique thickness is maintained.

The response of a quasi-plastic substance is more like that of the plastic than that of the viscous liquid. At low stresses the strain rate is very low and a change in the shear stress causes only a very minor change in the strain rate. At high stresses a small change in shear stress produces a large change in the strain rate. A flow law of this kind favors a glacier thickness which will produce a shear stress falling along the knee of the strain rate curve as shown in Fig. 4. (For model calculation purposes, a
quasi-plastic flow law may be approximated by that of an ideal plastic, with proper choice of the yield stress as shown.) Should the ice thickness increase, the strain rate soon becomes very large. Even with a considerable increase in the amount of material added to the upper surface as accumulation, the increase in the strain rate, and therefore in the glacier's velocity, makes possible the removal of this larger volume of material with only a slight increase in glacier thickness. On the other hand, if the supply to the surface is decreased, a small decrease in thickness, and therefore in shear stress, causes a considerable drop in the strain rate. The smaller quantity of material is removed at a much slower rate so that only a little thinning of the glacier occurs. That thickness, which, for a given slope, produces a shear stress value that falls on the knee of the flow law curve, may be termed the equilibrium thickness.

On a real glacier the conditions change from point to point; accumulation and ablation tend to change the thickness and the bed slope may change as well. These changes produce different types of flow patterns within the glacier, depending on whether the glacier is tending to thicken or thin in comparison to the equilibrium thickness. The adjustments in thickness are made by changes in the vertical component of the velocity vector.

Three possible kinds of flow can be deduced from a consideration of the deformation of a non-compressible ideal
plastic and the results apply with only slight modification
to the deformation of ice (Nye, 1951, p. 554; Nye, 1952,
p. 87). The simplest kind of flow, which would occur in a
block of ice on a smooth slope with no accumulation or
ablation, is termed plug flow. In plug flow the velocity is
the same at all points and the movement of all particles is
parallel to the glacier bed so no vertical motion can occur
relative to the bed, and the thickness remains constant.

In extending flow the velocity increases from point
to point in a downglacier direction; in effect the glacier
is being stretched. Extending flow occurs on a glacier
under two sets of conditions, where the balance is positive
and in areas where the bed is convex.

An area where the bed is convex may be thought of as
a region of transition from an area with a given bed slope
to an area with a steeper slope. Over this region the
equilibrium thickness will decrease. For a constant dis-
charge rate (volume of material passing each cross section
in unit time is a constant) the velocity must increase where
the thickness decreases. In Fig. 5 volume A and volume B are
equal showing a constant discharge, but for \( h_2 \) less than
\( h_1 \), \( v_{e1} \) must be greater than \( v_{e1} \). Note that the flow
lines converge.

The addition of accumulation over a part of a glacier
might be expected to increase its thickness. However, in a
region of constant slope, a uniform equilibrium thickness is
found as a result of the flow law. With a constant thickness the added material produces an increase in the discharge rate in the downglacier direction. For a constant thickness the velocity must increase in the downglacier direction. In Fig. 5 the areas C and D, bounded by solid lines, are equal showing a constant discharge rate if there is no accumulation. With the addition of a uniform layer of accumulation the discharge rate will increase in the downglacier direction. The area of C, representing the discharge through section 1, must increase by a volume equal to c, while the area of D, representing the discharge at section 2, must increase by a volume equal to c + d.

In compressive flow the velocity decreases from point to point in a downglacier direction and accordingly the flow lines diverge. Compressive flow also occurs under two sets of conditions, where the balance is negative or where the bed is convex. Arguments similar to those used for extending flow can be used to show the necessity of compressive flow under the two sets of conditions.

It can now be seen that the maximum discharge rate is at the equilibrium line, separating the areas of accumulation and ablation. Only through this cross section must pass the whole positive balance of the accumulation area. This follows from the rule stated previously, that the flow per year through any cross section is equal to the total positive balance above that cross section, and also from
Fig. 5.—Diagrams illustrating conditions for extending flow.
the relationship of extending and compressive flow to accumulation and ablation.

The glaciological analysis which is presented in this dissertation is based on the preceding principles and models. The various components of the accumulation and ablation have been evaluated from stake and pit measurements and a velocity-versus-depth profile has been calculated from surface velocity data for a point near the equilibrium line. The temperature distribution must also be considered in the calculation of the velocity profile since the variation of velocity with stress depends upon the temperature. The calculations are based on the measurements and theories due to Glen and Nye.

The relative importance of the various elements of the mass balance is assessed for a portion of the ice sheet adjacent to an area of exposed rock, and reasons are suggested for the existence of such rock areas. Also discussed are possible mechanisms of origin of certain associated features, such as the marginal moraines and the dirt bands in the ramp area.

Location

Wilkes Station is located on the Budd Coast on the east side of Vincennes Bay at 66° 15.4' South, and 110° 31.5' East (see Figs. 6 and 7). The east shore of the bay lies in a north-south direction and along the middle part of
Fig. 6.—Wilkes Station location map (from Hollin et al., 1961, p. 3).
Fig. 7.—Windmill Islands Map (from Hollin et al., 1961, p. 33).
Fig. 8.—Topographic map of domal area inland from the Windmill Islands (after Morgan, 1966, p. 81).
this shore is an area of rock islands and peninsulas, called the Windmill Islands, which extend from beneath the edge of the ice sheet. Wilkes Station is on the northernmost of these peninsulas. North and south of the islands the ice sheet extends to the sea, but inland from the island area it terminates in an ice ramp. From the inner side of the exposed rock areas, at an elevation of about 40 m, the ramp rises in a distance of 1 1/2 km to a height of about 145 m where there is a system of ice-cored moraines, parallel to the edge of the ice sheet. The moraine system extends in a north-south direction for a distance of about 17 km, with only occasional breaks, and in many places forms prominent ridges. In contrast to most of the Antarctic, a true ablation area is present which includes a small area above the moraine and part of the ramp.

In cross section the coastal portion of the ice sheet has an approximately parabolic profile with only slight irregularities (see cross section, Fig. 9). The parabolic profile is modified at the ice margin in the island area by the ablation, and by the accumulation of drift snow on the ramp, largely due to the presence of the moraine. North and south of the Windmill Islands the ice sheet ends in a cliff varying from 5 to 20 m in height.
Fig. 9.—Topographic profile along the S-2 trail from Grinnell Nunatak to S-2 (after Hollin et al., 1961, p. 66).
Station history and observations

Wilkes Station was established in March, 1957, as part of the United States International Geophysical Year program and was operated as a United States base until February, 1959. Since then it has been a joint Australian and United States base under Australian command. During the first two years three glaciologists were on the base staff, including the writer in 1958. Other programs at the base included aurora observations, seismic and geomagnetic studies, ionosphere soundings, cosmic ray studies, and meteorology. The present dissertation presents an analysis of some of the glaciological observations, primarily made in the period from February 1958 to February 1959, though some of the measurements were initiated in 1957. Most of the data used has been published by the Ohio State University Research Foundation (Cameron, Løken, and Molholm, 1959; and Hollin, Cronk, and Robertson, 1961) and a more complete description of some of the measuring techniques is given in these publications. Where other data have been used the source is acknowledged.

In addition to the main base, a small subsidiary station, S-2, located 80 km inland, has been occupied intermittently since April, 1957 and various weather observations have been made at S-1, located 3 km inland from the moraine (see Figs. 6 and 7).
Meteorological measurements. At Wilkes Station a meteorological staff made regular three-hourly observations according to the U.S. Weather Bureau procedures. The measurements included temperature, pressure, humidity, windspeed and direction, cloud cover, precipitation, and visibility conditions. Upper air radio-balloon soundings were made twice a day when the weather permitted.

At S-2 regular meteorological observations were made seven times a day during much of 1957. The observations included temperature, wind speed and direction, relative pressure changes, cloud cover and visibility conditions. During 1958 observations were made only during occasional periods when the station was occupied for other work.

At S-1, which was visited only once a week, air temperature was measured by a thermograph, and for a short period in January 1959, wind speed was measured with a totalizing anemometer.

Daily observations of temperature, wind speed and direction, cloud cover, visibility, and precipitation were made during a traverse in September, October, and November, 1958. The route was South and then West from S-2 (see Fig. 6). The data collected on this traverse have not yet been published.

Accumulation and ablation measurements. Stake measurements to record accumulation and ablation in the area were initiated in 1957, several groups of stakes being
emplaced. However, some of the stakes put out in 1957 had to be replaced when they melted loose in the warm summer of 1957-1958. A few of the original stakes were broken. With the exception of some unsplit bamboo poles, used for trail markers or for strain measurements, the stakes placed in 1957 were 1 1/2 inch-diameter half-round dowels. All the 1957 stakes were implanted in the ice to a depth of 1/2 to 3/4 m with the aid of a SIPRE 3 inch corer. Only the bamboo stakes remained firm in the 1957-1958 summer and measurements on these give the only continuous record from 1957. The replacements in 1958 were split bamboo stakes emplaced to a depth of 1 m with a 1 inch auger. It was hoped that the smaller cross section of the bamboos would absorb less radiation so that the stakes would melt out less readily. This proved only partly successful, but by February 24 all stakes were solidly frozen in.

At sites where extreme melting was expected in the next summer, an additional stake was emplaced, drilled in to a depth of 2 m, and subsequently at these sites both stakes were measured. Although the 1958-1959 summer produced comparatively slight melting, and all stakes remained firm, it is felt that the deeply planted stakes would have remained sound even with severe melting. The split bamboos did to a large extent solve the melting problem, but they were subject to bending which must be allowed for when measuring accumulation or movement. When bending was severe the angle
was noted in the measurements. The accompanying table and maps give the location of the stakes and the dates during which they were used. If no termination date is given for a stake, it was still in use when the Australians took over in 1959 (see Table 1, and Map I, back pocket).

The measurements were made with a meter stick, measuring the distance between the snow surface and a reference mark on the stake. The measurements were recorded to the nearest half centimeter. Sometimes wind eddying around the stake caused local erosion or drifting so that a small hollow or drift formed around or behind the stake. The average surface level in the immediate area was then estimated and used for the measurements. The measurements are believed to be accurate within one centimeter. When the surface consisted of soft snow the thickness of unconsolidated material above a hard snow or ice surface was also recorded, since the soft snow usually blew away.

The technique used may be criticized because it does not allow for compaction, that is a lowering of the level of the original surface relative to the stake. However, where a stake is emplaced in ice and frozen solidly in there can be no compaction in the ice. On the other hand, the objection is valid for measurements in firm, although near Wilkes Station the error is probably slight. Very little compaction occurs in the top 1 to 1 1/2 m, judging from the densities measured at these depths in pits in 1957 and 1958.
TABLE 1.—Dates of stake emplacement and abandonment.  
Unless noted, stakes were still in use as of January, 1959.

<table>
<thead>
<tr>
<th>Stake</th>
<th>Date emplaced, etc.</th>
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<td>12 May, 1958</td>
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<td>402</td>
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<td>410</td>
<td>(S-1) 10 Feb. 1968</td>
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<td>17 Feb. 1958</td>
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<td>424</td>
<td>17 Feb, 1958 28 July, 1958, stake added</td>
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<td>429</td>
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<tr>
<td>430</td>
<td>12 Nov, 1957, last measured 21 April, 1958</td>
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<tr>
<td>BF-1</td>
<td>mid 1957, 1st measured 27 Jan, 1958</td>
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<tr>
<td>BF-2</td>
<td>28 July, 1958, stake added</td>
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<tr>
<td>F1-2</td>
<td>8 Feb, 1957, used with interruptions through 1958</td>
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<tr>
<td>S-2</td>
<td>trail stakes (5 mile intervals) 18 Oct, 1957 except for Flag 8, emplaced 7 Aug, 1958</td>
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<td></td>
<td>Moraine movement stakes, emplaced July, 1958</td>
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<td></td>
<td>A, superimposed ice stake near 402, 15 Dec, 1958</td>
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<td></td>
<td>B,</td>
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<td>M3, movement stake, ablation measured after 15 Dec, 1958</td>
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The pit measurements show no regular increase in snow density in the surface 2 m as would be expected if compaction had occurred. Accumulation has been in the form of wind-packed snow which may have an initial density greater than that caused by compaction at such shallow depths.

The biggest problem is in determining the proper density value for the conversion of measured accumulation to water equivalent. The conversion is most difficult in localities where accumulation is as superimposed ice with a cover of snow. (As long as the whole accumulation is in the form of ice, the measurement is unambiguous.) If part of the accumulation is snow, the level of the ice surface and the density of the snow must be determined. The problem of establishing the correct density to use in the computation will be discussed with the analysis of the stake data (see p. 45).

In addition to the stake measurements, the condition of the surface was usually recorded; whether it was ice, ice obviously derived from snow, snow which was icy (the crystals could easily be broken apart), crusted snow, wind-packed snow, wind-blown snow, or fresh snow. The wind-blown snow was loose, like fresh snow, but was in the form of small drifts. An estimate was also made of the percentages of the area near the stake covered by ice, old snow, and fresh snow, and any unusual features were noted.
Pit studies. The accumulation was studied in snow pits both in 1957 and in 1958. A 33 meter pit was dug in 1957 at S-2 and a complete analysis of accumulation at this location has been made (Cameron, 1963). Many other pits varying from 1/2 to 3 1/2 m deep were dug in the firn of the accumulation areas and in the ice of the superimposed ice and ablation areas. The data are presented in the reports by Cameron et al., 1959, and Hollin et al., 1961. The present dissertation is not primarily concerned with the analysis of the pit data so the measurements will not be described in detail. The accumulation values as determined in these pits are generally in agreement with the values as determined from the stakes.

Ice temperatures and coring. A series of ice temperature readings was made at S-1 where a set of thermohms was installed in 1957 to measure ice temperatures at depths between 1/2 and 16 m. A Leeds and Northrop Wheatstone bridge was used in the measurements and accuracy should be better than 0.1°C. In addition to the readings at S-1, temperature measurements were made in several 10- or 11-m holes drilled at locations along the trail between Grinnell Glacier on the coast near Wilkes Station and S-2. Several 10-m temperatures were also measured during the 1958 traverse. The cores from these holes were used for density measurements and to study the stratigraphy. Some sections
were made for the examination of crystal size and C-axis orientation with a universal stage.

**Movement and strain measurements.** Stakes were set out to measure movement or strain in three different areas. Absolute movement of the Vanderford Glacier was recorded from Haupt Nunatuk over a 2-year period, the stake positions being surveyed every 3 to 4 months (for details see Cameron, 1963, pp. 121-135).

A strain network (relative movement) was laid out in 1957 at S-2 but was only measured once a year. The present dissertation is not concerned with either of these areas in detail.

A third set of movement stakes was located near the ice margin, with stakes both above and below the moraine at the top of the ice ramp (see Map 1). It was thought that morainal material might have been brought to the surface along shear planes within the ice so the strain measurement location was chosen to include an area just inland from the moraine, exhibiting some possible shear planes which might still be active. Using a Wild T-2 theodolite the stake positions were triangulated from two stations located on the moraine, and from the same stations resection angles were measured to several permanent markers on Clarke and Bailey Islands. From the latter measurements the absolute movement of the moraine stations could be calculated. Relative vertical motion was measured with the same stakes but these
measurements were made by leveling with a transit. The stakes were split bamboos emplaced 2 m into the ice. The stakes in the strain network were first measured in late June and early July of 1958 and the measurements were repeated in December 1958. The relative vertical movement was measured at about the same dates.

**Gravity measurements.** While the resupply ship was at Wilkes in February, 1958, some gravity readings were made along the S-2 trail and from these a profile of the sub-ice rock surface was constructed (Hollin et al., 1961, pp. 51-66). This required a survey giving elevations of the stations. Within 13 km of the coast the positions were determined by a subtense bar and theodolite method. The traverse was continued along the S-2 trail using transits, and weasel odometers for distance measurement. Additional altitude checks were made using leapfrog altimetry methods.

**Other studies.** In addition to these glaciological observations, Hollin (unpublished) examined the glacial geology of the island area and Robertson (1960) studied the bedrock geology.

**Australian studies.** Since 1959, considerable glaciological work has been done in the Wilkes area by the Australians. Accumulation studies have been continued and the measurements have been extended 300 miles south of S-2. Seismic determinations of ice thickness have been made over the same traverse route. The problem of drift snow transport
has been carefully investigated and the temperature distribution within the top 30 m of the ice has been considered. Mass balance for the domal area inland from Wilkes has been evaluated and the whole area is probably very close to being in balance. This work has been reported by Budd (1966) for the period through 1961, and he includes references to other Australian work. Morgan (1966) has also studied the ice thickness using gravity methods.
DETERMINATION OF THE MASS BALANCE

The condition of the mass balance can be determined in several ways in accordance with the principles already set forth. Ideally the entire annual positive balance in the accumulation area and the entire annual negative balance in the ablation area should be measured and their totals compared. Simple as the idea is, the process of determining these values over a wide area is very difficult. Furthermore, at Wilkes it is impossible to determine exactly what part of the accumulation area is replenishing a given portion of the ablation area. That is not important in the study of a small glacier where balance measurements can be made over the entire surface of the glacier. However, where measurements can be made over only a small part of the glacier's surface, it is important that the material added to the glacier surface within the portion of the accumulation area studied, should eventually be lost within the portion of the ablation area studied. In other words, all flow lines which originate in the area where a positive balance is measured, should terminate in the area where a negative balance is measured. Unfortunately, in the Wilkes area, the configuration of the flow lines is not sufficiently well known, so sections of the accumulation and ablation areas cannot be properly related.
Another approach is to consider the quantity of material being transported through a given cross section at the equilibrium line and compare this quantity with either the accumulation upglacier from the cross section, or with the ablation downglacier from it. This method brings one back to the problem of determining the portion of the accumulation area which is furnishing material to the chosen cross section. In addition, to determine the flow through a cross section at the equilibrium line, one must know the change of velocity with depth, as well as the surface velocity at the position of the cross section. A method of calculating this relationship is given later (pp. 63-76).

Again with this approach, the mass balance can be precisely determined only if the configuration of the flow lines is completely known both at the cross section and extending from it to the accumulation or ablation area. However, if the flow past the equilibrium line can be calculated and if the value of accumulation is known at enough points in the accumulation area, the method can then give a means to estimate the portion of the accumulation area that is needed to supply material to a given cross section on the equilibrium line. If the accumulation area necessary to supply a given cross section at the equilibrium line extends far inland, then the mass balance in the adjacent ablation area will be due to conditions over a wide region. If the accumulation can be supplied by a very local area then the
balance in the ablation area reflects only local conditions.

A third approach may be used to ascertain the state of balance at a given location. If the vertical motion can be measured or calculated at a given point, where the specific annual balance is known, then these quantities can be compared. If the glacier is in equilibrium, a given amount of ablation should be balanced by an equal upward motion of the ice. If the vertical velocity calculated from the strain rate is to be compared with the specific balance, it must be assumed that the compressive flow is controlled by the ablation and not by irregularities in the bedrock surface. A calculation of this kind has been made for a point just above the moraine at stake 402.

**Determination of specific balance values**

Since any of the methods for determining the mass balance require a knowledge of specific balance values, the method of determining these quantities from the stake measurements will be described. An analysis of the accumulation and ablation patterns is presented in a later section (p. 169). However, since the concept of annual accumulation or ablation includes the idea of a balance year, the determination of this period of time will be discussed before the reduction of the stake data is described.

**Balance year.** Generally in the Antarctic there is little melting and the rate of accumulation is roughly constant, or at least almost always positive, and any 12 month
period can be used in determining annual accumulation. However, in the Wilkes Station area, where considerable melting occurs in the summer and where the duration of the melt season varies from year to year, the use of different twelve month intervals may give different values for the annual balance. The most consistent results are obtained by using the "natural balance year" though this may not be exactly a twelve month interval. The balance year is the interval of time between the lowest level of the snow surface of consecutive summers at a particular stake. At Wilkes the lowest surface at most stakes occurred in February or March but at some stakes as late as April, the date varying from stake to stake and from year to year. Use of such a balance year rather than an arbitrarily selected twelve month period is also recommended because this is the same interval that is interpreted from the snow stratigraphy studied in pit walls.

Reduction of the stake measurements. Using a representative reading for each month, the changes in surface level at each stake have been graphed for the period of records available. For an example the balance at Fl-2 is shown in Fig. 10. In choosing the representative monthly readings, those which show soft surface snow have been avoided. To reduce extreme local irregularities three stake running means have been calculated by the formula \( \frac{a + b + c}{3} \), where \( a \), \( b \), and \( c \) are readings at three neighboring stakes,
Fig. 10.—Monthly changes in surface level (snow accumulation as measured) at Fl-2. The dotted line indicates a period which may include a short interval when the stake melted loose. See Map 1 for stake location.
and the values graphed as shown in Fig. 11. No running mean value was calculated for the end pair of stakes. Means have also been calculated and plotted for larger groups of stakes, showing the average accumulation of snow during the year over larger segments of the measured area (see Fig. 12). The six divisions used are composed of the lower and upper profiles and four divisions along the S-2 trail between stake 402 and BF-2 (see Map 1).

The value of the measured annual accumulation or ablation at each stake has been determined in two ways from graphs, similar to Fig. 10, of accumulation-versus-time at individual stakes. For the 1958-1959 season the accumulation at each stake was evaluated for the period between the lowest summer reading in 1958 and the lowest summer reading in 1959, and also for the period from one February to the next. The first method gives the accumulation during the natural accumulation year. For the 1959-1960 season (Australian data) the accumulation year used was February to February since no data were available for March and April of 1960. These values of measured accumulation were plotted against position to show the pattern of snow accumulation on a profile along the lower part of the S-2 trail.

Accumulation values were determined in the same way for the three-stake means and the values plotted against position to show a smoothed pattern of snow accumulation over the same profile (see Fig. 14). The stake number associated
Fig. 11.—Three-stake average of the monthly changes in surface level at stakes 415, F1-2, and 416.
Fig. 12.—Averages of monthly surface changes for six groups of stakes representative of different portions of the accumulation profile from stake 402 to BF-2.
Fig. 12 (Contd.)
Fig. 12 (Contd.)
Fig. 13.—Comparative changes in snow surface level during 1958 and 1959 for the groups of stakes considered in Fig. 11.
Fig. 14.—Profile of 3-stake averages \( \frac{a + b + c}{3} \) of snow accumulation over coastal portions of the S-2 trail. The identification numbers refer to the center stake of each group.
with each mean value is that of the center stake of the group.

The annual accumulation of snow, determined for individual stakes has been converted to water equivalent (cm or g/cm²). From examination of various shallow pits in the coastal area, the average near-surface snow density has been estimated as 0.38 g/cm³ and the density of the bubbly superimposed ice has been estimated as 0.8 g/cm³. (The density of ice near the surface can vary from .65 to .85, g/cm³ (Budd, 1966, p. 75.) Because a portion of the accumulation is sometimes in the form of superimposed ice beneath the snow surface, soundings of the surface snow were used to determine the depth to the ice surface. Unfortunately these measurements were not made at all stakes, so where soundings were not made or where there is no record of an exposed ice surface, all accumulation is assumed to be snow.

Undoubtedly small errors exist in the conversion of snow accumulation to water equivalent, since only a limited number of density values are available and the density varies from place to place. However, the results using these values are consistent with the results from stakes farther inland, where the variations in density are not so marked.

The values of annual accumulation in g/cm² for individual stakes have been combined as running means by the formula \( \frac{a+2b+c}{4} \), and for the inland end stakes \( \frac{a+2b}{3} \) where \( b \) is the center stake of a group of three. At the coastal end,
stake 402 has been used as an individual value since it is
the only stake with marked ablation. Because changes from
snow to superimposed ice have been accounted for in the con-
version to water equivalent, the values of accumulation in
water equivalent indicate the balance at each stake much
more accurately than did the snow measurements. In calcu-
lating the running means the central stake of each group has
therefore been given a weighting higher than that of the
outer stakes rather than giving all stakes equal weighting
as was done for the snow accumulation means. The annual
accumulation rates in water equivalent, both for individual
stakes and for the running means, have been plotted against
positions in Figs. 15-18.

Establishing the equilibrium-line position

The profiles of annual balance plainly show that the
lower end of the profile near stake 402 is in the ablation
area while the upper end, above stake 414, is definitely in
the accumulation area. The center part of the profile is
complicated by the high accumulation at stake 411 and the
low accumulation at stake 413. These features are discussed
in a later section (p. 169). For the determination of the
equilibrium-line position the balance over these areas of
surplus and deficit has been averaged and the average value
of the balance is found to be positive. The 1958 profile
shows a marked positive balance over the whole area except
Fig. 15.--Profile of the balance in water equivalent at individual stakes over the coastal portion of the S-2 trail. The dotted line shows the balance at stake 404 assuming all accumulation is snow with no superimposed ice.

Fig. 16. Profile of 3-stake weighted averages \(\frac{a+2b+c}{4}\) of balance in water equivalent over the coastal portion of the S-2 trail. Averaged values for the upper and lower profiles are included. Stake 402 the only stake with marked ablation, is considered as an individual value.
Fig. 17.—Profile showing balance in water equivalent over a 2-year period at individual stakes and for 3-stake weighted averages. On the averaged profile stake 402 is considered as a single value.

Fig. 18.—Profile of balance in water equivalent from low of 1958 to low of 1959 at 5-mile flags along the S-2 trail from mile 10 to mile 45.
for stake 402. The 1959 profile generally shows negative balance below stake 410 though a few lower stakes show a slight positive balance.

The equilibrium line for 1958 was quite low, near stake 403 at about 177 m elevation, while the 1959 equilibrium line was near stake 409, at 256 m. The average of these gives an elevation of 216 m for the equilibrium line for the two year period, indicating a position near stake 406. The accumulation profile for the whole two-year period, however, shows the balance to be positive as low as stake 404. The only stake above this with negative balance is stake 413 in the area of low accumulation, already mentioned. Although the area between stakes 403 and 409 was one of negative balance in 1959 the ablation was so slight that it had little effect on the two-year total. The usual equilibrium line is probably in the vicinity of stakes 402 to 403 at 142 to 195 m elevation, but with only a very small positive balance for some distance above this. Budd (1966, pp. 78-79), using data from the period 1957-1962, points out that in the area of the coastal stake network the total change in the surface has been very slight over the period of measurement. The balance in individual years has fluctuated from positive to negative.

The evidence from the 10 m deep ice temperatures, which is discussed later (p. 197), is in general agreement
with the equilibrium-line position as determined by the stake measurements. The temperature data suggest a position of the equilibrium line somewhere between S-1, 262 meters, and the area of stake 402.
MASS BALANCE IN RELATION TO MOVEMENT

The mass balance near stake 402

The movement stake network (see Map I) near the ice margin coincided with the coastal end of the line of accumulation stakes (see Fig. 19). Stake 402, located slightly inland from the moraine and near the upper edge of the ablation area, was approximately in the center of a triangle formed by movement stakes M1, M4, and M5 (see Map I, Fig. 20). Therefore the mass balance at stake 402 can be determined by comparing the ablation at stake 402 with the vertical velocity at this point as calculated from the strain rate of the triangle formed by stakes M1, M4, and M5.

Strain calculation and vertical velocity

Using the coordinates of stakes M1, M4, and M5 (see Fig. 20) given in Hollin et al. (1961), p. 41), the area of triangle M1-M4-M5 was calculated for July 30–31 and for December 19, 1958. The difference between these areas divided by the initial area gives the area strain rate over the period considered, 133 days, and this can be converted to an annual strain rate. For this calculation a linear variation of velocity with distance along the sides of the triangle has been assumed. The area on December 19 was

51
Fig. 19.—Views of the area near stake 402, (a) in good weather, note icy surface, and moraine in background, (b) the same area with conditions of blowing snow.
Fig. 20.--Diagram showing relative positions and velocities of movement stakes in the vicinity of stake 402 and the moraine. LB, lower base, and UB, upper base, are located on portions of the moraine, while A is located at the debris outcrop on the ramp (see locations on Map 1).
smaller than on July 30-31, which indicates that the area is one of compressive flow. The results are as follows:

Area: July 30-31 14893.033 m²
Dec. 19 14888.094 "

Difference in area: ......................... 4.939 "

This gives for the annual strain rate:

$$\frac{4.939}{14893.033} \times \frac{365}{133} = 0.0091 \text{ yr}^{-1}$$

The errors in the survey are discussed on p. 58.

This calculation gives an area strain in the horizontal plane rather than the surface longitudinal strain of a two dimensional case, as is used in the calculations by Nye (1957, pp. 113-133) where surface strain perpendicular to the direction of flow is assumed to be zero. According to Nye's calculations, assuming two dimensional flow, and the upper surface parallel to the bed, the longitudinal strain rate is independent of depth. This is true not only when ice is considered as a plastic substance, but is also true when the calculations are made using the experimentally-determined flow law of ice (Nye, 1957, p. 120). The question of the dependence of the strain rate on depth is discussed on p. 60.

With the strain rate independent of depth and no bottom melting, the vertical velocity is the product of the strain rate and the ice thickness at that point. This can be seen by considering the vertical velocity due to
horizontal compression of a prism of incompressible material. Although ice is compressible, in the present case only plastic strains in a steady state situation are considered, so the assumption of no volume change is valid. If deformation is two dimensional the strain rate to be measured is the horizontal (or longitudinal) strain rate. If the deformation is three dimensional, the area strain at the surface should be measured (see Fig. 21). Simple calculations, assuming a change in only one or in both horizontal dimensions of the prism, yield the result:

\[ h_2 = h_1 \left[1 + (\text{strain rate}) (\Delta t)\right]^{-1} \]

where \( h_2 \) is the height measured an interval of time \( t \) after \( h_1 \) is measured.

When expanded and the higher order terms dropped, this gives:

\[ h_2 - h_1 = -h_1 [(\text{strain rate}) (\Delta t)]. \]

This is the change in the vertical dimension in unit time, or the vertical velocity.

The ice thickness at stake 402 is not known accurately, but gravity measurements at location C-1, only about 25 m away, indicate an ice thickness of 119 m (see Fig. 22). The thickness calculations are described in Hollin et al (1961,
Fig. 21.—Sketch showing the relationship between area strain and vertical velocity.

Fig. 22.—Cross section showing the probable ice thickness (determined from gravity measurements) near the edge of the ice sheet (after Hollin et al., 1961, p. 65).
A maximum error of 15 m or about 5% is given for B-3 located about 6 km inland from stake 402. A 5% error at stake 402 (or C-1) is about 6 m.

Using the calculated thickness of 119 m, the vertical velocity is:

\[
\text{Thickness at C-1 x strain rate} = \text{vertical velocity}
\]

\[
119 \text{ m} \times 0.0009 = 10.71 \text{ cm/yr}.
\]

The negative balance measured at stake 402 was 12.5 cm in 1958, and 10.5 cm from February 1959 to February 1960.

Though there appears to be a slight excess of ablation, the difference is within the possible error. This suggests that the glacier is very nearly in equilibrium.

A similar strain calculation has been made for the larger triangle ML-LB-UB (see Map I and Fig. 20). The calculated vertical velocity over this area is 6 cm/yr, somewhat less than that for the smaller triangle. The difference between 6 cm/yr for the larger and 10 cm/yr for

---

1Assuming a density of 2.65 g/cm\(^3\) for the bedrock, values of the Bouguer anomaly were determined at the stations in the ice free area. The Bouguer anomaly gradient in this area, 3 to 4 km wide, was extrapolated over a distance of about 1 km to the determine what the value of gravity would have been at stake 402 if all of the material between the surface and sea level were of density 2.65 g/cm\(^3\). The observed value of gravity is less than this extrapolated value and the difference is due to the replacement of rock of density 2.65 g/cm\(^3\) by ice of density 0.89 g/cm\(^3\). Using the formula for the attraction of an infinite slab, the ice thickness can be calculated, a deficiency of 1 mgal representing 13.6 m of ice.
the smaller triangle may be real, though it is within the possible errors as discussed in the next section. Assuming that the difference is real, two explanations are possible. First, the moraine area included within the larger triangle may be an area of greater negative balance, where the ice is becoming thin and beginning to stagnate, and the movement is insufficient to balance the ablation. This is unlikely as there is measurable motion well down the ramp and the moraine area does not appear to be stagnant.

The difference in calculated values of the vertical motion is probably related to less severe ablation in the moraine area, where surface debris protects the ice. In addition, drift snow accumulates in the lee of the ice-cored moraine ridges, forming major drift features which appear to be permanent. Compressive strain is greater between stake M1 and stakes M4 and M5 than it is between the latter stakes and UB and LB, which suggests that the greater negative balance is between stake M1 and stakes M4 and M5. Stake measurements show high negative balance in the same area (see Figs. 15-17).

Survey errors. From instrument stations on the moraine two sets of measurements were made to distant stations, including M1 and stations on the solid rock of the islands. Only a single set of measurements was made to the nearer movement stations, including M4 and M5. (One set is a pair of measurements, with the telescope direct and inverted.) With only these few measurements, a complete
statistical analysis has not been attempted. However, an assessment of the errors has been made. The average deviation from the mean values of the various measurements varied from 0.3 seconds to 6.9 seconds, and the average of all deviations is 4.4 seconds. The maximum probable errors have been estimated by calculating the position error caused at each point by an angular error of 7 seconds, the largest average deviation. The effect of these position errors on the strain rate is then considered. The position errors for each of the stakes M1, M4, and M5 are found to be 5 cm, 1 cm, and 1.5 cm, respectively. This gives an error in the M1-M4 strain of 12 cm, in the M1-M5 strain of 13 cm, and in the M4-M5 strain of 4 cm. For each of these strains the error is 0.0004 of the total distance. This proportional error should be the same for the strain calculated over a year and therefore can be used as a probable error in strain rate as well as in strain. The value ± 0.0004 has been used in the following calculations of the maximum probable error.

To determine the effect of the errors in the horizontal strains on the calculated value of the vertical strain, an equation has been used which relates the strains in three mutually perpendicular directions to the dilation, or volume strain. The dilation relationship states that the sum of strains in three mutually perpendicular directions is equal to the dilation: \( \varepsilon_1 - \varepsilon_2 - \varepsilon_3 = \Delta \), and this quantity is invariant for rotations of the coordinate system (Jaeger,
1956, p. 47). From this it follows that for an incompressible material (\( \Delta = 0 \)) the strain in a vertical direction is equal to the negative of the difference between any two horizontal strains measured perpendicular to each other.

Errors of 0.0004 in the horizontal strains (and strain rates) gives \( \pm 9.5 \) cm. as a maximum probable error in the vertical strain over a year, and therefore in the vertical velocity. Although this error is undesirably large, the conclusion that the marginal portion of the ice sheet in this region is in equilibrium is probably still valid.

Criticism of the calculation method, various assumptions. In calculating the vertical velocity, it has been assumed that the glacier bed is planar, thus ignoring the relationship between bottom topography and compressive or extending flow. Although the assumption is not completely valid, the bed relief in this area, as calculated from gravity measurements, is relatively slight so that the assumption of a flat bed does not seem to be unreasonable (see Fig. 22).

A more important criticism involves the question of whether the strain rate is in fact independent of depth, as calculated in Nye's theory. The calculations include the assumption that the surface slope and the bed slope are the same, which implies that the ice thickness is constant. Paterson (1962) has modified Nye's theory, allowing the bed slope to differ from the surface slope. Paterson's studies
show that the strain rate can vary with depth, and this has been verified by bore hole measurements on the Athabasca Glacier in British Columbia. However, on the Athabasca Glacier strain interpretations are complicated by the effect of the valley walls.

The situation at stake 402 is uncomplicated by valley walls and tributary glaciers, but the surface slope and bed slope undoubtedly differ somewhat from each other. Unfortunately, the value of the bed slope cannot be determined and the ice velocity at the bottom is unknown so there are insufficient data to apply Paterson's modified theory.

On the Athabasca Glacier the strain rate usually decreased at depth but in a few places the reverse was true. Where the surface strain rate was low, the change in strain rate with depth was small. Since the strain rate, recorded at stake 402 is small, it is unlikely that there is significant change in the strain rate with depth. If the strain rate decreases with depth, a given prism will undergo less horizontal compression and the value of the vertical velocity calculated for stake 402 will decrease slightly.

As mentioned previously, the vertical velocity at 402 was not measured directly, since the vertical movement survey was never tied to bedrock. However, the change in elevation difference between point M1 and point M4 was measured and is recorded in the Wilkes Station data (see Hollin et al., 1961, p. 42). The difference in elevation
between these points increased by 17 cm between July and December. It is unlikely that the lower point, M4, had a negative vertical velocity, so, assuming M4 has remained at a constant elevation, M1 appears to have a vertical motion of at least 17 cm. This value is high, in comparison to that calculated from the area strain rate but is within the possible error. It suggests that the strain rate does not decrease at depth, but could even increase at depth, or that the value of ice thickness calculated from the gravity measurements is too low. If either of these is true the vertical velocity calculated for 402 would be increased, and there may be more upward motion than is compensated for by ablation.

Another possibility is that part of the vertical motion is related to differential movement along discrete inclined shear surfaces. The formation of ice-cored moraines has often been attributed to movement of this kind (Goldthwait, 1951, Schytt, 1955, Bishop, 1957). In this case vertical movement in the vicinity of the shear, but only in this vicinity, would be equal to the product of the horizontal movement and the tangent of the dip angle of the shear plane. Note that in contrast, the vertical strain calculated in compressive flow is representative of the strain over the whole area of the measured compression. Also, if the movement is on a single shear plane, the horizontal strain should all take place across a single line, while in compressive
flow the strain should be evenly distributed over the area measured. Closely-spaced movement stakes might distinguish between the two types of movement, but from the present data the strain measured at the surface may be due to either type of movement.

The horizontal strain between M1 and M4 is 40 cm; the vertical strain between M1 and M4 and also between M1 and M5 is 17 cm as measured by leveling with a transit. If the motion were assumed to be on a single plane where inland ice is riding up over stagnant marginal ice, the plane of movement would make an angle with the horizontal of about 19° ($\tan^{-1} \frac{17}{50} = 19°$). The suspected shear planes which were observed showed dip angles of 25° to 30° (Hollin et al., 1961, p. 229). To investigate shear movement, trenches were dug through several assumed shear planes during 1958 but no signs of very active movement were observed. Pegs emplaced across a major dirt band in 1957, when remeasured in 1958, showed apparent changes in position of very small magnitude and of questionable significance. It seems likely then that most of the strain is due to compressive flow, with perhaps a minor component due to movement along discrete shear planes.

**Velocity-versus-depth profile**

In order to estimate the quantity of ice passing through the cross section at stake 402, the ice velocity must be calculated at various depths. The calculations are based on Nye's theory and follow a method developed by Nye.
in 1959 (p. 495). It should be pointed out that the model used for the calculations is not exactly the same as the physical situation at Wilkes. At stake 402, minor strain was measured in the direction perpendicular to the velocity vector, where in Nye's model, surface strain perpendicular to the velocity vector is zero. The strain rate used in the calculations is the measured area strain rate which is the same as the vertical strain rate but with the sign reversed. The calculation gives a velocity profile which would be found in a glacier showing only two-dimensional flow, but with the ice thickness and vertical strain rate the same as that found at stake 402. The volume of ice calculated to pass through the model cross section should be the same as that transported through the cross section at stake 402. The equations required for the calculation are as follows (Nye, 1957, p. 121, p. 129, 1959, p. 494).

\[
\begin{align*}
(1) \quad \bar{\rho} y \sin \alpha &= \frac{T}{\xi} \sqrt{\xi^2 - r^2} \\
(2) \quad u &= \pm rx - 2g \sin \alpha \int_0^y \frac{\dot{\varepsilon}}{T} \bar{\rho} y \, dy + u_o
\end{align*}
\]

As shown in Fig. 23, the \( x \) direction lies in the surface of the glacier and in the direction of flow, the \( y \) direction is perpendicular to the \( x \) direction and toward the glacier bed. The point \( x_o, y_o \) is an arbitrary point under consideration on the glacier surface.

\( \bar{\rho} = \) the average density of ice and firm above depth \( y \)
\( \alpha = \) surface slope of the glacier, \( 4^\circ \) at M1.
Fig. 23.—Diagram showing the orientation of the stress axes and the velocity components.
\( r = \text{longitudinal strain rate in yr}^{-1}, \text{measured at the surface but assumed uniform with depth. (The strain rate measured and used is an area strain which amounts to the assumption of two dimensional flow.)} \)

\( \tau = \text{effective shear stress (see note 1, p. 71 for definition)} \)

\( \dot{\varepsilon} = \text{effective strain rate} \)

\( u = \text{velocity at depth} \)

\( u_0 = \text{surface velocity} \)

The parameters \( \dot{\varepsilon} \) and \( \tau \) are connected by a flow law of the type \( \dot{\varepsilon} = f(\tau) \), determined experimentally by Glen in 1955. The flow law of ice is dependent on temperature and is of the form \( \dot{\varepsilon} = (\frac{\tau}{A})^n \) where \( A \) is the temperature dependent constant and \( n \) is 3.2 to 4.2 depending on the analysis of the data.

The value \( n = 3.2 \) was obtained using the minimum observed creep rates. The value \( n = 4.2 \) corresponds to the quasi-viscous or steady-state creep rate. This value was obtained by removing the portion of the creep rate due to transient creep as calculated using the assumption that ice obeys Andrade's law (Glen, 1955, pp. 529-530). By this method a steady state creep rate was calculated for the experimental runs at low stresses, where, even at the termination of the run, the creep had not reached a steady state condition. Glen suggests that the value \( n = 4.2 \) is the better choice to represent the flow law of glacier ice. Nye also uses the value \( n = 4.2 \) since it is in better agreement with
flow law determinations from bore hole and tunnel experiments (Nye, 1953; Nye, 1957, p. 129). Accordingly, the value 4.2 has been used in the present calculations.

In the determination of the flow law, Glen was using higher stresses than those considered in the present calculations. Bukovitch and Landauer (1960), made some measurements at low stresses and obtained a hyperbolic sine relation for the flow law, but preliminary strain measurements at low stresses (≈ 0.05 bars by Dooley, 1964) suggest that Glen's power law is also applicable at these low stresses. Analysis of the closure rates in various bore holes and tunnels is also in agreement with a power law (Nye, 1957, 1959, Mathews, 1959). Paterson (1962) gives a complete review of flow law investigations and concludes that a power law is most consistent with the data. The temperature constants determined by Glen (1955, p. 528) have been used for the present calculations.

Since the flow law depends on the temperature, the relation of temperature to depth must be known. At 402 a linear relationship has been assumed using a gradient of 1°C per 40 m. This value has been obtained by assuming a heat source located at the glacier bed equal to the average geothermal heat plus the heat of friction (Robin, 1955, pp. 524–526). Since a large portion of the differential motion in the ice is near the glacier bed, there is some justification for calculating the frictional heat as if it were all
at the glacier bed. In assuming a linear relationship of
temperature and depth, the effect of the mass transfer on the
temperature distribution is not considered. These effects
have been treated by Robin (1955) and by Jenssen and Radok
(1963). The effect of accumulation is to lower the tempera-
ture in the upper parts of the glacier below the value
expected for a linear relationship. This is because accumu-
lation causes a downward velocity in the glacier which carries
cold surface accumulation downward. In an ablation area the
effect is reversed (see Benfield, 1948-49). The effect of
lateral flow is similar to the effect of accumulation, since
relatively cold material from higher up the glacier is
brought downward. Under some circumstances the effect due to
lateral flow may be more pronounced than that due to accumu-
lation and temperatures at moderate depth may be lower than
the surface temperature. This situation is discussed briefly
on p. 104. Stake 402 is in the ablation area, but close
to the equilibrium line, so the effect of ablation would be
to raise the ice temperature very slightly, while the effect
of lateral motion should be to lower the ice temperatures.
Since these effects are of opposite sign, the deviations from
the linear relation will in part cancel each other. The
accuracy of the data does not justify the calculations which
would be necessary if these effects were to be treated fully.
The temperature measured by a thermohm at 11 m depth is -7.75°C, and, having assumed a linear gradient, the temperature at the glacier bed is -5.05°C. Above 11 m, annual temperature variations are significant.

**Calculation of the velocity profile.** The method for determining the velocity at various depths is one of successive approximations and the procedure is as follows (Nye 1959, p. 494). First choose a value of depth \( y \) and read the appropriate temperature. Then estimate a value for \( \dot{\varepsilon} \) which by the flow law relation determines \( \tau \), using the constant for the selected temperature. The value of \( \rho gy \sin \alpha \) can then be determined from equation (1) (p. 64) and thus a value of \( y \) determined. In general the value of \( y \) thus determined will be different from that originally selected. Therefore a better value for \( \dot{\varepsilon} \) can be estimated until the sequence is self-consistent. Five or six repetitions is usually sufficient. With \( \dot{\varepsilon} \) and \( \tau \) determined for a specific depth, these values can be used in equation (2) (p. 64) and the velocity calculated at the given depth.

The data for the flow law which is used in the following calculations is from the laboratory work by Glen (1955, p. 528). The flow law as given by Glen is in the form

\[
(3) \quad \dot{\varepsilon}_c = k \sigma^n
\]

where \( \dot{\varepsilon}_c \) is the compressional strain rate, \( \sigma \) is the compressive stress, \( n = 3.2 \) to \( 4.2 \) and the value of \( k \) varies
with the temperature. The law may also be given in the form

\[ (4) \quad \dot{\varepsilon}_c = \sigma^n \exp\left(\frac{-Q}{RT}\right) \quad \text{(Glen, 1955, p. 532)} \]

The constant \( k \) in (3) is expressed in (4) as \( \exp(-Q/RT) \) where \( B \) is a constant, \( Q \) is a heat of activation, \( R \) is the gas constant and \( T \) is the absolute temperature. Since

\[ k = B \exp(-Q/RT), \quad \log k = \log B - \frac{Q}{RT} \]

and a plot of \( \log k \) vs. \( 1/T \) should lie on a straight line. Glen determined the value of \( k \) at \(-0.02\), \(-1.5\), \(-6.7\), and \(-13^\circ C\). The values of \( \log k \) vs. \( 1/T \) for the three lower temperatures do lie on a straight line. The value of \( k \) for \(-0.02^\circ C\) is much higher, which Glen suggests is an indication of partial melting.

Steinemann (1958b, pp. 263-264) found that qualitatively, temperate ice (at \( 0^\circ C \)) behaves mechanically in the same manner as cold ice. The parameters describing the behavior of "dry" temperate ice can be extrapolated from experimental results on ice at sub-zero temperatures. Rheologically, "wet" temperate ice should be considered as an aggregate of a ductile phase and a highly viscous phase. The mechanical properties predicted for "dry" temperate ice will be increasingly modified as the proportion of liquid present increases. The present calculations are not affected by this problem, since the deforming ice is always at sub-zero temperatures.

The values of \( k \) used in the present calculations have been determined from Glen's values of \( k \), using a plot of \( \log k \) versus \( 1/T \).
Since Glen's results are given in terms of compressive stress and strain rate, the values must be converted to effective or octahedral shear stress and effective strain rate, since these are the parameters applicable to glacier strain and used in Nye's formulas. The following conversions are used (see Nye, 1953, p. 486).

\[ \gamma_{\text{oct.}} = \frac{1}{\sqrt{3}} \sigma, \quad \dot{\varepsilon}_{\text{oct.}} = \frac{\sqrt{3}}{2} \dot{\varepsilon}_{\text{comp.}} \]

and therefore

\[ c = k (\sqrt{3})^{4.2} \cdot \frac{\sqrt{3}}{2} \]

where \( c \) is the temperature dependent constant in Nye's statement of the flow law, analogous to Glen's \( k \). The flow law used is then of the form \( \dot{\varepsilon} = c \gamma^n \)

Computation forms were made up to facilitate the successive approximations for \( \dot{\varepsilon} \) and \( \gamma \). The values determined for \( \dot{\varepsilon} \) and \( \gamma \) are tabulated below and shown graphically in Fig. 24. With these values equation (2) can be solved for \( u \) at the given depths. The term \( r_x \) (p.64) is zero because the position of \( M_1 \) is taken as the origin of the

\[ u = \pm r_x - 2g \sin \alpha \int_0^y \frac{\dot{\varepsilon}}{\gamma} \rho_y \, dy + u_o \]

---

Octahedral shear stress is the shear stress in a plane which intersects the principle stress axes at unity, i.e., the plane of a face of an octahedron having each corner on a principle axis. The effective strain rate is the strain per unit time in a direction normal to the octahedral plane.
TABLE 2.—Temperature, $\dot{\epsilon}$, and $\gamma$, as a function of depth at stake 402.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Temp (°C)</th>
<th>$\dot{\epsilon}$ (yrs⁻¹)</th>
<th>$\gamma$ (bars)</th>
</tr>
</thead>
<tbody>
<tr>
<td>31</td>
<td>-7.25</td>
<td>0.00104</td>
<td>0.382</td>
</tr>
<tr>
<td>53</td>
<td>-6.7</td>
<td>0.0015</td>
<td>0.401</td>
</tr>
<tr>
<td>81</td>
<td>-6.05</td>
<td>0.004</td>
<td>0.512</td>
</tr>
<tr>
<td>101</td>
<td>-5.5</td>
<td>0.012</td>
<td>0.621</td>
</tr>
<tr>
<td>121</td>
<td>-5.00</td>
<td>0.028</td>
<td>0.742</td>
</tr>
</tbody>
</table>

Fig. 24.—Graphs of shear stress, $\gamma$, and strain rate, $\dot{\epsilon}$, as a function of depth at stake 402.
coordinate system. The integral has been evaluated graphically by plotting the values of \( \frac{\dot{e}}{T} \rho y \) against depth and then determining the area under the curve. For the higher values of \( y \), the integral has been evaluated graphically from a logarithmic plot, since the relationship is almost logarithmic. The velocity was then obtained by multiplying the integral for the appropriate depth by \(-2g \sin \alpha\) and subtracting this product from the value of the surface velocity.

The surface velocity at stake 402 was not measured so it has been estimated in the following manner. The velocities at stake M4 and M5 have been combined to give an average value for the velocity along the lower side of the strain triangle. This value was then averaged with the velocity at stake M1, which yields a value of 136.5 cm/yr for the surface velocity at stake 402.

The value of the velocities at given depths, and the velocity differences between the surface and the given depth, \(2g \sin \alpha \int_{0}^{y} \frac{\dot{e}}{T} \rho y \, dy\) are tabulated below and presented graphically in Figs. 25-26. The value of 121 m has been used arbitrarily in the calculations instead of the 119 m depth determined for C-1 by gravity measurements. (A 20-m interval was used for the deeper calculation; 11 m is the upper end of the extrapolated temperature profile.) The velocity difference between the surface and 119 m as read from the graph is 109 cm/yr, indicating a basal ice
Fig. 25. Graph showing the horizontal velocity as a function of depth at stake 402.
TABLE 3.—Velocity as a function of depth and the difference from the surface velocity as a function of depth at stake 402.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Velocity (cm/yr)</th>
<th>Difference from surface velocity (cm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface</td>
<td>136.5</td>
<td>0</td>
</tr>
<tr>
<td>31</td>
<td>134.8</td>
<td>1.70</td>
</tr>
<tr>
<td>53</td>
<td>130.9</td>
<td>5.56</td>
</tr>
<tr>
<td>81</td>
<td>117.7</td>
<td>18.77</td>
</tr>
<tr>
<td>101</td>
<td>88.8</td>
<td>47.71</td>
</tr>
<tr>
<td>121</td>
<td>12.5</td>
<td>124.02</td>
</tr>
</tbody>
</table>

Fig. 26.—Graph showing the difference between the surface velocity and the velocity at depth y parallel to the surface as a function of depth at stake 402. For the described method of calculation, a power-law relationship is indicated by the graph.
movement of 27.5 cm/yr. At a depth 122.5 m the basal velocity would be zero. Note that it has already been suggested, p. 62, that the ice thickness calculated from the gravity data may be too low. This small value for velocity $u$ at the base of the glacier suggests that if basal sliding does occur it can be of only minor importance.

The only well-considered mechanism for basal sliding, that of Weertman (1957, 1964), requires basal ice at the pressure melting point, since part of the sliding is by pressure melting at the upglacier side of small obstacles. (Weertman's model of basal sliding is more fully described later, pp. 93-95.) At stake 402 the basal temperature is calculated to be $-5^\circ$C, so no basal melt can occur. Elsewhere, no basal sliding has been observed where tunnels have reached the bed of glaciers with basal temperatures below the pressure melting point (Goldthwait, 1960, Holdsworth, personal communication).

**Volume of ice passing stake 402 annually**

In order to determine the volume of material passing through a given cross section of unit width in a year, the area marked "A" in Fig. 27, between the velocity curve and the initial reference line must be determined. Since the difference between the surface velocity and the velocity at depth, when plotted with respect to depth, approximates an exponential function, the equation of this curve can easily be ascertained from the logarithmic plot. To determine the
Fig. 27.—Sketch illustrating the calculation of the ice volume passing a cross section at stake 402.
area under this curve, marked "B" in Fig. 27, the equation need only be integrated between the limits 0 and 119 m. This value can then be subtracted from the volume which would pass the unit width cross section if the velocity were constant from top to bottom, area of "A" and "B." The equation of the velocity difference–versus–depth curve is:

\[
\text{vel. diff, cm/yr} = 0.338 \exp \left(4.77 \times 10^{-4}\right) \text{(depth in cm.)}
\]

The calculations are as follows.

\[
\int_{0}^{119} 0.338 \exp \left[(4.77 \times 10^{-4})y\right] \, dy = 236607 \text{ cm}^3
\]

136.5 \times 11900 = 1624350 cm\(^3\), for vel. constant with depth

1624350 cm\(^3\) - 236607 cm\(^3\) = 1387743 cm\(^3\)

The volume passing a unit (1cm) cross section in a year is 1.3877 m\(^3\) of ice. Using an ice density of 0.89 g/cm\(^3\), the same as that used for the gravity determination of ice depth, this is equivalent to 1.235091 m\(^3\) of water.

**Equivalent accumulation area.** The value for the quantity of water equivalent passing through a cross section one cm wide at stake 402 can be compared to the value of accumulation along a strip one cm wide extending inland from stake 402. Such a comparison shows whether a large or only a small portion of the accumulation area is required to maintain the flow through the cross section considered. The value of accumulation along such a one cm wide strip has been calculated in the following manner. From Table 4, which
<table>
<thead>
<tr>
<th>Smoothing Method</th>
<th>1958 Low</th>
<th>1959 Low</th>
<th>1958 Low</th>
<th>1959 Low</th>
</tr>
</thead>
<tbody>
<tr>
<td>Individual Stakes</td>
<td>8.0</td>
<td>7.6</td>
<td>7.6</td>
<td>7.5</td>
</tr>
<tr>
<td>Individual Stakes, Jan. 1958</td>
<td>7.5</td>
<td>7.1</td>
<td>7.0</td>
<td>6.9</td>
</tr>
<tr>
<td>Individual Stakes, Jan. 1959</td>
<td>7.0</td>
<td>6.7</td>
<td>6.6</td>
<td>6.5</td>
</tr>
<tr>
<td>Individual Stakes, Feb. 1958</td>
<td>6.5</td>
<td>6.2</td>
<td>6.1</td>
<td>5.9</td>
</tr>
<tr>
<td>Individual Stakes, Feb. 1959</td>
<td>6.0</td>
<td>5.7</td>
<td>5.6</td>
<td>5.5</td>
</tr>
<tr>
<td>Individual Stakes, March 1958</td>
<td>5.5</td>
<td>5.2</td>
<td>5.1</td>
<td>5.0</td>
</tr>
<tr>
<td>Individual Stakes, March 1959</td>
<td>5.0</td>
<td>4.7</td>
<td>4.6</td>
<td>4.5</td>
</tr>
<tr>
<td>Individual Stakes, April 1958</td>
<td>4.5</td>
<td>4.2</td>
<td>4.1</td>
<td>4.0</td>
</tr>
<tr>
<td>Individual Stakes, April 1959</td>
<td>4.0</td>
<td>3.7</td>
<td>3.6</td>
<td>3.5</td>
</tr>
<tr>
<td>Individual Stakes, May 1958</td>
<td>3.5</td>
<td>3.2</td>
<td>3.1</td>
<td>3.0</td>
</tr>
<tr>
<td>Individual Stakes, May 1959</td>
<td>3.0</td>
<td>2.7</td>
<td>2.6</td>
<td>2.5</td>
</tr>
<tr>
<td>Individual Stakes, June 1958</td>
<td>2.5</td>
<td>2.2</td>
<td>2.1</td>
<td>2.0</td>
</tr>
<tr>
<td>Individual Stakes, June 1959</td>
<td>2.0</td>
<td>1.7</td>
<td>1.6</td>
<td>1.5</td>
</tr>
<tr>
<td>Individual Stakes, July 1958</td>
<td>1.5</td>
<td>1.2</td>
<td>1.1</td>
<td>1.0</td>
</tr>
<tr>
<td>Individual Stakes, July 1959</td>
<td>1.0</td>
<td>0.7</td>
<td>0.6</td>
<td>0.5</td>
</tr>
<tr>
<td>Individual Stakes, August 1958</td>
<td>0.5</td>
<td>0.2</td>
<td>0.1</td>
<td>0.0</td>
</tr>
<tr>
<td>Individual Stakes, August 1959</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Note: The table values are smoothed balance values at stakes near the 1% sheet margin.
gives accumulation values in g/cm², the average annual value of accumulation over the two-year period has been chosen and the differences in accumulation between adjacent stakes have been tabulated. Stake positions have been projected onto a line through stake 402, and approximately perpendicular to the moraine system. The distances between these projected stake positions have been measured on the map of the northern island area, Map 1. Referring to Fig. 28a, it is apparent that the volume of the positive balance between two stakes is numerically equal to the sum of the rectangular area and the triangular area. The distances between projected stake positions, the specific balance at each stake, and the volume of net annual positive or negative balance between stake positions, together with a running total of the volume of the positive balance above the equilibrium line is given in Table 5.

The total volume of material in water equivalent passing a one cm wide cross section at the equilibrium line has been determined by adding the volume of water ablated between stake 402 and the equilibrium line to the annual volume of material passing stake 402. By successively summing the volumes of accumulation from stake position to stake position, one can determine how long a strip inland from the equilibrium line is needed to furnish a volume of material equal to that passing the equilibrium line. In
Fig. 28.--Sketches illustrating the calculation of the volume of the positive balance along a line inland from stake 402, (a) illustrates the situation over an interval of positive balance, (b) illustrates the approach used over intervals where the balance changes from positive to negative.
TABLE 5.—Cumulative balance along a strip 1 cm wide inland from stake 402.

<table>
<thead>
<tr>
<th>Stake</th>
<th>Distance between stakes</th>
<th>Annual balance at stake (from 2 yr. average)</th>
<th>Balance difference between stakes</th>
<th>Cumulative balance inland from stake 402</th>
</tr>
</thead>
<tbody>
<tr>
<td>402</td>
<td>380</td>
<td>-11.4</td>
<td>11.0</td>
<td>-0.22452</td>
</tr>
<tr>
<td>403</td>
<td>16</td>
<td>-0.4</td>
<td>0.4</td>
<td>-0.00032</td>
</tr>
<tr>
<td></td>
<td>equilibrium line 284</td>
<td>0</td>
<td>7.2</td>
<td>0</td>
</tr>
<tr>
<td>404</td>
<td>325</td>
<td>7.2</td>
<td>3.7</td>
<td>0.10224</td>
</tr>
<tr>
<td>405</td>
<td>250</td>
<td>3.5</td>
<td>0.5</td>
<td>0.276115</td>
</tr>
<tr>
<td>406</td>
<td>190</td>
<td>3.0</td>
<td>2.7</td>
<td>0.413615</td>
</tr>
<tr>
<td>407</td>
<td>300</td>
<td>5.7</td>
<td>1.0</td>
<td>0.496265</td>
</tr>
<tr>
<td>408</td>
<td>320</td>
<td>4.7</td>
<td>3.7</td>
<td>0.652265</td>
</tr>
<tr>
<td>409</td>
<td>340</td>
<td>8.4</td>
<td>2.0</td>
<td>0.861865</td>
</tr>
<tr>
<td>410</td>
<td>500</td>
<td>6.4</td>
<td>13.7</td>
<td>1.113465</td>
</tr>
<tr>
<td>411</td>
<td>480</td>
<td>20.1</td>
<td>14.5</td>
<td>1.775965</td>
</tr>
<tr>
<td>412</td>
<td>430</td>
<td>5.6</td>
<td>6.6</td>
<td>2.392765</td>
</tr>
<tr>
<td>413</td>
<td>430</td>
<td>-1.0</td>
<td>3.5</td>
<td>2.441715</td>
</tr>
<tr>
<td>414</td>
<td>570</td>
<td>2.5</td>
<td>1.4</td>
<td>2.473940</td>
</tr>
<tr>
<td>415</td>
<td>310</td>
<td>3.9</td>
<td>4.9</td>
<td>2.656340</td>
</tr>
<tr>
<td>Fl-2</td>
<td>370</td>
<td>8.8</td>
<td>1.2</td>
<td>2.853190</td>
</tr>
<tr>
<td>416</td>
<td>250</td>
<td>7.6</td>
<td>2.6</td>
<td>3.156590</td>
</tr>
<tr>
<td>BF-1</td>
<td>300</td>
<td>5.0</td>
<td>1.9</td>
<td>3.314090</td>
</tr>
<tr>
<td>417</td>
<td>470</td>
<td>6.9</td>
<td>1.8</td>
<td>3.492590</td>
</tr>
<tr>
<td>BF-2</td>
<td>470</td>
<td>5.1</td>
<td>3.774590</td>
<td></td>
</tr>
</tbody>
</table>
these calculations the equilibrium line is taken to be between stakes 403 and 404.

The value of material passing through a one cm wide cross section at the equilibrium line as calculated by the above method is 1.459 m³/yr. It can be seen from Table 5 that the distance between the equilibrium line and a point between the projected positions of stakes 410 and 411, about 2 1/2 km, is sufficient to supply the volume passing the equilibrium line. Where plug flow is assumed, velocity constant with depth, the required distance is from the equilibrium line to a point just short of stake 411. This is a surprisingly short distance to furnish all the ice that passes the equilibrium line. It is obvious that most of the drainage is to one side or the other of the island area, and that only a tiny portion of the total accumulation is discharged through the moraine and island area. The main drainage of this area is through the Vanderford Glacier to the south, and also through the Peterson Glacier, see Figs. 6 and 7 and Map 1. The relation of these observations to the formation of the moraine will be discussed later (p. 201).

Calculations at S-1

A second set of temperature-versus-depth profiles and velocity profiles were calculated at S-1, about 2 km inland from the equilibrium line. The calculations were made to compare conditions at S-1 with those at stake 402. Because
no surface velocity measurements were made at S-1, the longitudinal strain rate had to be calculated from the accumulation, assuming that there is a downward velocity at the surface exactly equal to the accumulation. Therefore, only the difference between the surface velocity and the deep velocity can be determined. Note that no conclusions about the glacier's equilibrium can be drawn, since equilibrium has been assumed to calculate the strain rate. The temperature distribution has been calculated by Robin's method (Robin, 1955) since S-1 is in the accumulation area and there is nothing to offset the so-called Robin effect. The temperatures are probably even lower than calculated since no allowance is made for colder material being introduced by lateral flow. This effect is probably slight because the velocity is very low.

The equation developed by Robin (1955, p. 525) is as follows. Temperature difference between height $h$ and the ice surface is given by:

$$\theta_H - \theta_h = (\frac{d\theta}{dh})_{\text{bottom}} \sqrt{\frac{2hk}{A}} \left[ \text{erf} \left( \frac{A}{\sqrt{2hk}} . h \right) \right]_h^H.$$

$H =$ thickness of the ice sheet. (Total ice thickness is 203 m but a thickness of 192 m is used for calculating the temperature profile since significant annual temperature variations related to surface temperature changes extend to a depth of 11 m.)

$h =$ height above the base of the ice.
\( \theta = \text{temperature} \) (\( \theta_h = \text{temperature at h above base} \))

\( \dot{A} = \text{the accumulation rate in cm/yr} \)

\( k = \text{thermal diffusivity of ice} \) (\( 1.18 \times 10^{-2} \text{cm}^2/\text{sec} \))

Error function \( \text{erf} \ x = \int_{0}^{x} e^{-y^2} \text{dy} \)

\( \frac{d\theta}{dh} \) at the bottom is calculated to be \( 1°C/44m \), assuming only geothermal heat flow. The worldwide average heat flow has been used since no measurements have been made in Antarctica.

This formula was developed for the center of an ice cap where there is no lateral flow in any direction and no large contribution from frictional heat. It can be used as an approximation at S-1 because the velocity there is low.

The temperature profile as calculated by this method is given in Fig. 29. Since the annual accumulation at S-1 is only 12.3 g/cm\(^2\) and the depth 203 meters, the temperature curve departs only very slightly from a linear slope of \( 1°C/44 \text{m} \). The 11 m temperature of -7.6°C was determined from thermohm measurements continued over a two year period (Hollin et al., 1961, p. 201). The calculated basal temperature at S-1, -3.25°C, is still well below the melting point.

Using these temperatures, \( \gamma \), \( \dot{e} \) and the velocity differences from the surface can be calculated in the same manner as at stake 402 and are given in Table 6. In Fig. 30 \( \gamma \) and \( \dot{e} \) are plotted as a function of depth and the velocity differences are plotted in Fig. 31. At S-1 the velocity difference is only 117.2 cm/yr at a depth of 203 m,
Fig. 29.---Graph of the calculated temperatures as a function of depth at S-1.
TABLE 6.—Temperature, $\gamma$, $\dot{\varepsilon}$, and difference from surface velocity, as function of depth at S-1.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Temp ($^\circ$C)</th>
<th>$\gamma$ (bars)</th>
<th>$\dot{\varepsilon}$ (yrs$^{-1}$)</th>
<th>Difference from Surface Veloc. (cm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11</td>
<td>-7.6</td>
<td>(&quot;surface temp&quot;)</td>
<td>0.3338</td>
<td>0.00072</td>
</tr>
<tr>
<td>63</td>
<td>-6.43</td>
<td></td>
<td>0.3692</td>
<td>0.00139</td>
</tr>
<tr>
<td>113</td>
<td>-5.3</td>
<td></td>
<td>0.4799</td>
<td>0.0055</td>
</tr>
<tr>
<td>163</td>
<td>-4.16</td>
<td></td>
<td>0.5663</td>
<td>0.0127</td>
</tr>
<tr>
<td>193</td>
<td>-3.48</td>
<td></td>
<td>0.5949</td>
<td>0.0165</td>
</tr>
</tbody>
</table>

Fig. 30.—Graphs of shear stress, $\gamma$, and strain rate, $\dot{\varepsilon}$, as a function of depth at S-1.
Fig. 31.—Graph showing the difference between the surface velocity and the velocity at depth $y$, parallel to the surface, as a function of depth at S-1.
while at stake 402 the velocity difference was 109 cm/yr at 119 m and it would have been 124 cm/yr at 121 m. The less pronounced decrease of velocity with depth at S-1 is related to the difference in slope which is only 1° 54' at S-1 compared with 4° at stake 402. The values of the velocity difference imply that the surface motion at S-1 is at least a minimum of 117 cm/yr, but since the basal ice is well below the melting point there is probably only slight basal slip, if any, and the surface velocity is likely to be little more than 117 cm/yr. The equation of the curve for the velocity difference is:

\[
\text{vel. diff (cm/yr)} = 0.4294 \exp (2.763 \times 10^{-4}) (\text{depth in cm}).
\]

This can be compared with that for stake 402 on p. 78.

Considering the increase in ice thickness between stake 402 and S-1, it is surprising that the differences between surface and basal velocities at the two locations are so similar. This implies that the surface velocities are similar.

**Other temperature calculations**

In addition to the calculations at stake 402 and S-1, some estimates of bottom temperatures have been made at various distances inland. These estimates were made assuming a linear relation of temperature with depth, so that the Robin effect has not been included. At S-1 this effect is
very slight, but it becomes more important as accumulation and ice thickness increase. Robin (1955, p. 527) calculated the effect of various accumulation rates on the temperature profile at the center of an ice sheet, where no addition of heat is produced by ice movement. The two models shown by Robin are for ice thicknesses of 400 m and 3000 m, and for accumulation rates varying from 0 to 512 cm/yr. For accumulation of 8 to 16 g/cm², similar to that near Wilkes, the departure from a linear temperature-versus-depth relationship is noticeable even with an ice sheet thickness of 400 m, and it is very pronounced for an ice sheet thickness of 3000 m. The differences in temperature between the surface and basal ice which are calculated using a linear temperature relationship are maximum values of the temperature difference required to remove a given quantity of heat from the base of a glacier. The Robin effect reduces the temperature difference which is required to remove a given quantity of heat. A gradient of 1°/44 m is required to remove the average geothermal heat. A basal ice movement of 5 m/yr increases the necessary gradient to 1°/35.6 m, assuming a yield stress of .75 bars, and that all the frictional heat is liberated at the glacier base.

The estimates of the required differences between surface and basal ice temperatures, calculated using only the linear temperature relationship, are given in Table 7. The 11-m ice temperatures at the same locations are also
TABLE 7.—Basal temperatures along the S-2 trail, calculated assuming 3 different linear temperature gradients. Table shows temperature differences and gradients required to remove heat from base of ice sheet.

<table>
<thead>
<tr>
<th>Location (miles from coast)</th>
<th>Ice Thickness (in m)</th>
<th>Temperature at 11m depth (°C)</th>
<th>1°C/44m (geothermal heat only)</th>
<th>1°C/35.56m (geothermal + frictional heat of 5m/yr)</th>
<th>1°C/29.85m (geothermal + frictional heat of 10m/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S-2 (Mile 50)</td>
<td>1037</td>
<td>-19.3</td>
<td>23.32</td>
<td></td>
<td></td>
</tr>
<tr>
<td>FL-2 (Mile 40)</td>
<td>1134</td>
<td>-17.75</td>
<td>25.52</td>
<td></td>
<td></td>
</tr>
<tr>
<td>FL-4 (Mile 30)</td>
<td>723</td>
<td>-16.0</td>
<td>16.18</td>
<td>20.02</td>
<td></td>
</tr>
<tr>
<td>FL-6 (Mile 20)</td>
<td>571</td>
<td>-14.4</td>
<td>12.73</td>
<td>15.75</td>
<td>18.76</td>
</tr>
<tr>
<td>FL-8 (Mile 10.6)</td>
<td>369</td>
<td>-11.86</td>
<td>8.14</td>
<td>10.07</td>
<td>11.99</td>
</tr>
</tbody>
</table>
tabulated. If the temperature differences, calculated to be necessary to remove the heat produced at the glacier base, are greater than the difference between 0°C and the 11 m temperature, then the basal ice temperature will be raised to 0°C. Even using the approximation of a linear temperature relationship, and allowing for frictional heat equivalent to 5 m/yr of basal ice movement, the sub-zero basal ice temperatures extend as far inland as Flag 8, more than 10 km from the equilibrium line. Calculations by Budd (1963, pp. 107, 141) indicate probable surface velocities between S-2 and the coast of 3.5 to 4.0 m/yr. If only geothermal heat is considered, sub-zero basal ice temperatures extend about 50 km inland. Taking into account Robin's effect, the required difference between surface and basal ice temperatures is considerably reduced, and the basal ice temperature should remain below 0°C for a greater distance inland than indicated by the present calculations.

The importance of the basal ice temperatures in relation to whether or not basal sliding will occur is discussed more fully in the next section.
SIGNIFICANCE OF TEMPERATURE AND VELOCITY CALCULATIONS

Basal sliding

In Weertman's model of basal sliding in a glacier (1957, 1964), the movement of the ice over obstacles on the bed takes place through two processes. The first process is one of stress concentration on the upglacier side of obstacles, leading to greater strain rates and therefore accelerated plastic flow. The magnitude of stress concentration increases with increasing obstacle size. The larger the obstacle size, therefore, the greater will be the sliding velocity due to this mechanism.

In the second process, ice melts on the high pressure side of an obstacle and refreezes on the low pressure side. Latent heat energy is transported with the melt water from the upglacier to the downglacier side of an obstacle. For the continued operation of the process, the latent heat freed in the freezing process must be returned to the upglacier side of the obstacle by conduction. For the pressure melting mechanism, the smaller the obstacle size, the larger will be the sliding velocity.

Basal sliding requires the operation of both mechanisms. At small obstacles the stress concentrations are small and the sliding velocity due to the stress concentration
mechanism is negligible. The heat conduction through large obstacles is very small, so these limit the sliding velocity due to the pressure melting mechanism. According to Weertman's first model (1957), the observed rate of basal sliding is controlled by a critical obstacle size for which the sliding rates due to the two mechanisms are equal.

Kamb and LaChapelle (1964) have observed sliding motion at the base of the Blue Glacier. The basal ice was at the pressure melting point. At the base of the glacier they found a layer of comparatively clear bubble-free ice (0 to 3 cm thick) which could be readily distinguished from the overlying ice by texture and structure. They have called this layer the regelation layer. From careful observations on this layer, and from some simple experiments, Kamb and LaChapelle conclude that plastic flow due to stress concentrations is of relatively slight importance for sliding past obstacles of the size which controls the thickness of the regelation layer. Weertman (1964) has also recalculated the critical obstacle size and sliding velocity using an improved model with more reasonable assumptions. These results are in better agreement with the Blue Glacier observations.

Barnes and Tabor (1966) have investigated the hardness of ice, using a spherical indenter, over a temperature range from about -10°C to 0°C. Below -1.2°C their results are comparable in form to those of Glen, though the
stresses applied were 50 to 100 times greater than those used by Glen. At -1.2°C and above, pressure melting was observed, with a marked decrease in hardness. Furthermore, when tests were made on specially prepared bubbly ice, the ice in the zone effected by pressure melting became transparent in comparison to the surrounding bubbly ice. This supports the contention of Kamb and LaChapelle that the clear basal layer of ice is a regelation layer.

The regelation mechanism is essential then for the occurrence of basal sliding, and the basal ice must be at the pressure melting point. When the ice temperature is below the pressure melting point, there is no mechanism which allows the basal ice to slide over small obstacles. Goldthwait (1960, p. 56) observed no basal sliding in the sub-zero ice of the Red Rock ice cliff, Nunatarssuaq, northwest Greenland. Even moss and lichen on the rock under the ice were undisturbed.

Near Wilkes Station, the area in which the basal ice temperature remains below 0°C extends inland from the ice margin for a considerable distance, on the order of 20 to 50 km. It is inferred therefore, that the ice sheet is frozen to its bed in the coastal area near Wilkes, and that no sliding can occur.

Erosion related to bottom sliding. Erosion by glacial action is of two main types. The glacier bed can be abraded by rock tools held in the basal ice, or
alternately, large blocks of rocks are loosened and removed from the underlying surface by the ice. The latter process is called plucking. Plucking is particularly favored at locations where the rock is already fractured by faulting or by closely spaced joints.

It appears that the basal ice must be at the pressure melting point for either process to operate effectively. Since abrasion is a scraping process it is almost certainly related to bottom sliding, which requires ice at the pressure melting point. Wherever plucking is observed, scouring is invariably observed in adjacent areas, again indicating basal temperature at the pressure melting point. Locally, plucking may be the dominant form of erosion, particularly in the lee of an obstacle where, with a decrease in pressure, the ice probably freezes to the rock, though in adjacent areas the ice is probably melting.

On the other hand, striations have been observed on the bed of a glacier with sub-zero basal temperatures (Holdsworth, personal communication). Such striations are probably formed by large blocks of rock carried in the ice slightly above the bed, but with the bottom edges of the blocks occasionally touching the bed. Under such circumstances, however, erosion must be extremely slow.

Consideration of erosion associated with former continental ice sheets leads to one of two conclusions. Either the erosion took place at a very considerable
distance from the ice margin, the marginal ice being frozen to the bed, or at the margins of the ice sheets the basal ice was at the pressure-melting point. Where former ice sheets terminated on land in temperate regions (such as Ohio or northern Germany) abundant evidence of melt suggests that the upper surface was at 0°C. The ice therefore was probably at 0°C throughout, and erosion could occur right up to the ice margin.

**Thule type marginal moraine formation**

**The shear hypothesis.** Two main hypotheses have been advanced for the formation of moraine systems such as those near the edge of the ice sheet at Wilkes Station and at Thule, northwest Greenland. The first hypothesis is that of shearing (Goldthwait, 1951; Bishop, 1957), the moraines being referred to as shear moraines. According to this hypothesis material is brought to the surface along discrete failure surfaces. The surfaces are presumed to extend from the base of the ice sheet where they are tangential to the bed, to the upper surface of the glacier, which they intersect at an angle. If the upper and lower surfaces of the glacier are parallel and failure occurs by shear according to the Von Mise criterion, this angle should be 45°, provided the angle of internal friction is zero (Nye, 1951, p. 265).
The moraine material is assumed to be scraped into the shear surfaces at the bottom of the glacier and carried along the shears to the surface by differential ice movement. As the ice melts, the debris is freed, collects on the ice surface, and soon becomes thick enough to form a protective cover. The areas with the protective cover of debris then melt less rapidly than neighboring debris free areas and are left as ice-cored ridges forming topographically high features. Ostrem (1959, p. 229), working in northern Sweden, has found that ablation is inhibited by a debris cover only 1/2 cm thick. 1.8 to 2.6 m of debris reduces the wastage to an unmeasurably small rate.

Several objections have been raised to this hypothesis as more has been learned about the distribution of debris in the ice. Observations in tunnels in the Thule area of Greenland show that the mode of occurrence of debris varies widely. It may form solid layers of stone and sand up to 0.5 m thick, or may form layers 1 to 2 m thick of slightly dirty ice containing finely disseminated sand or dirt particles. The layers of slightly dirty ice are rather common (Weertman 1961, p. 966). Even if debris can be scraped into shear surfaces, the formation of either 0.5 m of almost solid rock and sand, or the formation of a meter or more of slightly dirty ice seems unlikely.

Goldthwait (1951, p. 569) has described shear planes in the Barnes Ice Cap which swell around boulders and till
clots, forming "augen" structures. Similar features have been observed in the Wilkes area (Cameron et al., 1959, pp. 12, 14). The geometry of such structures is more suggestive of plastic deformation than of failure by fracture. Should a shear fracture be formed with such an "augen" geometry, it is difficult to imagine motion along such a fracture, transporting debris toward the surface. At the surface, the "overthrusts" sometimes observed, where clean ice appears to be over-riding dirty ice, are in many cases due to differential ablation (McCall, 1960, p. 49).

Investigations of movement along debris surfaces exploded on the sides of tunnels have produced no evidence for the presence of discrete shear planes. Butkovitch and Landauer, and Hilty (Weertman, 1961, p. 968) observed no discrete shearing motion (a velocity discontinuity) across debris layers, though there was differential flow in the ice (see Fig. 32). Abel (Swinzow, 1962, p. 223), however, found that bands of included debris formed zones of increased differential flow. This is possibly due to a difference between the flow law of clean ice and the flow law of ice with included debris, dependent on the quantity of debris in the ice. Abel's observation does not show a velocity discontinuity along a discrete shear plane.

In the shear hypothesis, each debris horizon is believed to have been scraped into the ice along a shear surface. The formation of closely-spaced debris bands would
Fig. 32.—Sketches showing possible velocity distributions across a shear or debris surface, (a) a velocity discontinuity across a shear surface; (b) the debris surface does not effect the velocity profile; (c) a debris zone produces an increase in the strain rate, but does not produce a discontinuity in the velocity profile.
require that closely-spaced shear planes can exist, and yet there is no apparent reason why yield should cease along one plane of weakness, and a new plane of weakness form only a fraction of a centimeter away. If the closely-spaced shears are active simultaneously, then, because velocity decreases toward the glacier edge, the inner shears would catch up to and overrun those nearer the edge (see Fig. 33). In summary, the shear hypothesis does not seem adequate to explain the observed phenomena.

The Weertman mechanism of basal freezing. Weertman (1961, pp. 270-272) has given a critique of the shear hypothesis and has proposed an alternative, involving a freezing mechanism for the incorporation of debris bands into the ice. The temperature of the upper surface (10 m depth) of an ice sheet is controlled by the mean air temperature, as long as the air temperature remains for most of the year below 0°C. (The thermal conductivity of ice, \( K = 5.3 \times 10^{-3} \text{ cal/cm/sec/deg C} \), determines the rate at which heat energy can be transferred through the ice for any particular value of the thermal gradient.)

The temperature at the base of the ice is determined by a balance between the sources of heat at the glacier bed (geothermal heat, which averages about 38 cal/cm²/yr over the Earth's surface, and frictional heat due to movement, 18 cal/cm²/yr for 10 m/yr at .75 bars) and the rate at which heat can be conducted away by the thermal gradient in
Fig. 33.—Sequence of events after the initiation of several closely spaced shear surfaces; (a) shear surfaces just beginning to raise debris from the bottom; (b) the shear surfaces have caught up with each other by shearing over the bed; (c) a late stage where the shear surfaces have joined each other and only a single shear surface can exist at the bed (from Weertman, 1961, p. 969).
the ice above the bed. The maximum temperature of the basal ice is the pressure melting point, approximately 0°C. If the thermal gradient above the basal ice is insufficient to remove the heat energy at the same rate as it is produced, then melting will occur. Thermal energy exceeding the quantity that can be removed by the thermal gradient supplies the heat needed to melt the ice.

Lliboutry (1966) has shown that differential movement in the ice near the bed can raise the temperature to the melting point in a zone tens of meters thick. In this case no heat can be transferred through the isothermal layer, so all of the geothermal heat, plus frictional heat from bottom sliding is used to melt basal ice. Within the isothermal layer, all frictional heat produced by differential flow is used for melting, so this zone is composed of "temperate" or "wet" ice. Only at the upper surface of the isothermal layer can the heat produced by shearing be conducted away.

When considering the vertical variation of temperature within an ice sheet, the simplest approach is to assume a linear temperature variation (thermal gradient constant). This assumption is a simplification which implies negligible flow and negligible accumulation. The effect of accumulation is to carry cold surface snow downward, thus reducing the ice temperature in the upper portion of the ice sheet below that which would be found if the temperature variation with depth were linear. The
effect of ablation is the opposite to that of accumulation. Lateral flow introduces cold material from higher regions and therefore reduces the temperature at a given depth below that obtained assuming a linear temperature variation. Away from the central portion of an ice sheet, negative temperature gradients are often found near the surface (see Fig. 34). Scott Kane (personal communication) measured a positive temperature gradient of +2.01°C/100 m at the Pole of Inaccessibility near the center and highest part of the Antarctic ice sheet. At the South Pole there is already a negative gradient of -0.94°C/100 m.

A negative gradient may form where cold ice is transported outward, lowering the temperature at a point within the glacier below the mean annual air temperature of the surface above that point. This in effect creates a heat sink within the glacier and heat energy will be conducted toward the low temperature area from both above and below. Near-surface temperature gradients in such an area are of limited use in inferring temperature conditions at the ice base.

A polar type glacier can be divided into several regions, depending on whether the basal ice temperature is at 0°C with basal melting, at 0°C with basal freezing, or below 0°C. A thermal gradient of about 1°/40 m is required at the glacier base to remove only the geothermal heat. Where the ice is relatively thin, near the edge of the ice
Fig. 34.—Idealized distribution of temperature within an ice sheet (after Robin, 1955, p. 531).

Fig. 35.—Diagram showing the relationship of the position of the 0°C isotherm to the edge of the ice sheet and the resultant changes in thermal conditions at the base of an ice sheet (from Weertman, 1961, p. 971).
sheet, a difference between the surface and bottom temperatures of only a few degrees is sufficient to produce the necessary gradient. Accordingly the basal ice is frozen to the bed.

In the thicker central portion of an ice sheet the gradient cannot be maintained, even though the surface temperature is lower than that near the ice sheet margin (according to the lapse rate of about $1^\circ/100$ m) and the basal temperature is at the pressure melting point. Heat which cannot be removed by the thermal gradient above the bed is used in melting the basal ice. The melt water produced at the glacier base is forced outward by the pressure gradient and eventually it reaches the marginal regions of the ice sheet, where the temperature of the basal ice should be below the pressure melting point. However, the water produced by melting further inland then freezes to the basal ice, releasing additional heat energy and maintaining the temperature at the pressure melting point.

The rate of freezing is controlled by the rate at which the latent heat being released can be removed by conduction. The width of the zone where basal freezing occurs is such that water can be frozen to the ice at the same rate at which it is transported across the inner margin of the zone. The transfer of water from the inner regions of an ice sheet to the margins serves to remove from the interior some of the excess heat (in the form of
latent heat), and bring it to a region where it can be removed from the glacier by conduction. Where there is no more water to freeze to the base of the ice and liberate latent heat, the ice sheet becomes frozen to its bed and the melting point isotherm descends below the ice-rock interface (see Figs. 35 and 36).

The effect of accumulation or lateral flow is to increase the thermal gradient above the bed, thereby reducing the difference in temperature between the top and bottom surfaces of a glacier which is required for the removal of a given quantity of heat energy from the glacier base. This will tend to increase the size of the region where the ice is frozen to the bed at the expense of both of the other zones.

According to Weertman, the incorporation of dirt into the bottom of a glacier occurs under special conditions of freezing at the glacier bed. If a glacier is resting on material that is loose or can be wedged loose by freezing, the material can be frozen into the bottom of the glacier. This occurs whenever the melting point isotherm shifts downward from a position coincident with the lower ice surface, to a position in the material below. Such a vertical change in the position of the melting point isotherm also implies a lateral change in the position of the point where the isotherm intersects the base of the ice. Under steady-state
Fig. 36.—Diagrams showing the incorporation of debris into the ice sheet under non-steady state conditions (from Weertman, 1961, p. 972).

(a) To the left of point $x_2$, the $0^\circ C$ isotherm descends into the debris which makes up the bed. To the right of point $x_2$ the bottom is at the melting point and water is being frozen to the bottom surface. Water flowing from the right permits the bottom to be kept at $0^\circ C$ but the water supply is exhausted at point $x_2$.

(b) Less water is flowing from the right and the point of descent of the $0^\circ C$ isotherm has shifted to the right.

(c) A greater supply of water now flows from the right. Water now freezes to the $0^\circ C$ isotherm position given above, and pushes up into the ice the debris already frozen onto the bottom as already shown. Point $x_2$, marking the limit of the flow of water, has moved to left.

(d) A number of debris layers shown fixed within the ice after several repetitions of the illustrated cycle.
conditions this position would not change, and debris could not be included in the basal ice.

In a glacier or an ice sheet, however, it appears that such steady-state conditions occur rarely, if ever. Changes in thickness and surface slope affect the shear stress at the base and it in turn affects the frictional heat produced at the bed. Such changes cause a shift in the position of the boundary between the regions where the ice is frozen to the base and where refreezing is taking place. A shift inland of this boundary causes the melting point isotherm to move down into the debris beneath the ice and the debris is then frozen to the bottom of the ice sheet. A shift of the boundary toward the edge of the ice sheet extends the area in which water is freezing onto the ice bottom. If debris has already been frozen to the bottom, new ice will form beneath the frozen-on debris, because this is where the ice is now at the melting point. The debris is then incorporated into the basal ice with more ice below it (see Fig. 36). Repetition of this cycle, Weertman contends, causes the closely spaced dirt bands observed in the Thule ramp tunnel.

However, Weertman's proposal of frequent shifts in the position of the melting point isotherm, seems unnecessary to account for the observed phenomena. Such rapid changes in the thermal regime of the basal ice are difficult to explain. Sustained changes in mean annual air temperature could alter
both the boundary conditions of the ice sheet and the position of the intersection of the melting point isotherm with the bed, but for short-term surface variations the thermal inertia of the ice precludes frequent shifts of the isotherm position. Major changes in snow accumulation may alter the ice velocity and thus the frictional heat generation but it is unlikely that such major changes would be rapid enough to produce the closely spaced dirt bands that are observed.

It seems possible to explain the observed phenomena more simply. If, within the zone where water is freezing onto the basal ice, the glacier bed is composed of alternating areas of cleaned hard rock and loose debris, then one might expect the formation of bands of dirt and clear ice as in Fig. 37. It seems reasonable to assume that if water is being frozen onto the base of a glacier in an area where debris is available, some of the loose material that is at the freezing interface will be incorporated into the basal ice. Near the base the vertical component of velocity is controlled primarily by the rate at which the new ice forms at the glacier base. (The component of vertical velocity due to compression is very small near the base, approaching zero.) The dirty bands come from areas where the ice flows over available debris and the clear ice forms where the ice flows over cleaned resistant rock. Similarly, a wide area of loose material may give rise to a relatively thick band of slightly dirty ice. It is likely that the Weertman
Fig. 37.—Sketch showing a possible mechanism for the inclusion of disseminated debris layers in glaciers. Layers of slightly dirty ice are formed where bottom freezing occurs over basal areas with morainal debris. Bands of clear ice form where bottom freezing occurs over basal areas which are free of debris.
process, with a shift in position of the melting point isotherm, is responsible for the larger masses of fairly concentrated debris that are observed (Weertman, 1964, p. 968).

If the margin of the ice sheet is in the ablation area, compressive flow causes the flow lines to rise from the ice base to the upper surface. Debris frozen into the base of the glacier is brought to the surface and may eventually form ice-cored moraines. It should be pointed out that water melted in the inland areas may sometimes escape beneath major outlet glaciers where rapid motion produces sufficient heat to extend the region of bottom melting to the glacier margin. This has been observed in northeast Greenland even in February with temperatures at -37°C (Hamilton, 1958, p. 172). In adjacent regions, although the basal ice at the margin of the ice sheet is frozen to its bed, no zone of bottom freezing will occur inland from the margin if all this bottom melted water from further inland has escaped beneath the outlet glacier.

Debris raised by compressive flow—Wilkes Station conditions. The conditions near Wilkes Station appear to be the same as those postulated by Weertman for the formation of marginal moraines. The base of the ice sheet is frozen in the marginal area and almost certainly is melting further inland. It is probable that the melting-point isotherm does not rise to coincide with the ice base for at least 15 to
20 km inland from the ice margin, so that any debris present near the margin must have been incorporated into the ice further inland than that. However, material incorporated into the bottom of the ice will not rise appreciably relative to the bottom unless it is in an area of compressive flow, where the flow lines rise. Except in small areas where the bed is concave, compressive flow occurs only in the zone of ablation. Accordingly, appreciable upward transport of debris occurs only downglacier from the equilibrium line.

To determine whether this mechanism for bringing debris to the surface is reasonable and to determine how great a lateral distance is required, calculations of flow paths have been made, based on Nye's equations for the velocity distribution (Nye, 1957a). In these calculations a simplified model has been used, based on the conditions at stake 402 where the depth is 120 m, and the calculated vertical strain rate is .0009/yr. In the model, deformation is restricted to two dimensions, so the longitudinal strain rate is equal to the vertical strain rate. The slab of ice, long enough so that the solution is not influenced by edge effects, is assumed to be under uniform longitudinal compression, independent of depth. A constant thickness of 120 m is maintained by ablation which is assumed to be exactly equal to the upward velocity. The uniform strain rate throughout the slab produces a uniform vertical velocity at
all points on the upper surface and therefore the ablation rate is the same at all points.

At the upglacier end of the slab, the velocity conditions are made equivalent to those at the probable average equilibrium-line position for the years 1957 to 1959 in the Wilkes area, which by calculation is 396 meters upglacier from stake 402. The velocity-versus-depth profile, as calculated for stake 402, has been adjusted to what it would be at the equilibrium line. Since the difference in velocity between stake 402 and the equilibrium line (in the model) is due only to the longitudinal strain rate, the velocity-versus-depth profile at the equilibrium line is assumed to be the same as that at stake 402 with a constant factor added to it. This constant is equal to the product of the strain rate and the distance between stake 402 and the equilibrium line.

\[ 396 \text{ m} \times 0.0009 \text{ yr}^{-1} = 0.3564 \text{ m/yr} \]

Table 8 shows the velocity at various depths at stake 402, the corresponding velocities at the equilibrium line in the model, and the difference between surface velocity and velocity at depth \( y \) at the model equilibrium line. Given this data, the problem is to determine \( x \) in terms of \( y \) along the path of a particle moving according to Nye's velocity equations (see Fig. 38). The model equilibrium line is taken as the origin of the \( x, y \) system. The necessary
TABLE 8.—Velocity as a function of depth at stake 402, and calculated velocity and difference from surface velocity as a function of depth at the model equilibrium line.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Velocity at 402 (.396 x .0009)</th>
<th>Velocity at model equilibrium line</th>
<th>Diff. from Surface Velocity (cm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surf.</td>
<td>1.365 + .356 = 1.721</td>
<td>0</td>
<td></td>
</tr>
<tr>
<td>31</td>
<td>1.348 + .356 = 1.704</td>
<td>.017</td>
<td></td>
</tr>
<tr>
<td>53</td>
<td>1.309 + .356 = 1.665</td>
<td>.056</td>
<td></td>
</tr>
<tr>
<td>81</td>
<td>1.177 + .356 = 1.533</td>
<td>.188</td>
<td></td>
</tr>
<tr>
<td>101</td>
<td>.888 + .356 = 1.244</td>
<td>.477</td>
<td></td>
</tr>
<tr>
<td>119</td>
<td>.275 + .356 = .631</td>
<td>1.090</td>
<td></td>
</tr>
<tr>
<td>121</td>
<td>.125 + .356 = .481</td>
<td>1.240</td>
<td></td>
</tr>
</tbody>
</table>

Fig. 38. Diagram showing the orientation of the axes and velocity components in the model.
equations are those for horizontal and vertical velocity.

\[ v = \pm r (y - h) = \pm r (h - y) = \frac{dy}{dt} \quad (5) \]

\[ u = u_o - 2g_x \int_0^y \rho \frac{\dot{y}}{r} \, dy - rx = \frac{dx}{dt} \quad (6) \]

\[ v = \text{velocity perpendicular to the surface} \]
\[ u = \text{velocity parallel to the surface} \]
\[ h = \text{thickness of glacier} \]
\[ y = \text{depth to point of calculation} \]
\[ g_x = g \sin \theta, \text{the component of } g \text{ acting parallel to the bed.} \ (\text{For other symbols, see p. 66}) \]

If the longitudinal strain rate and thickness are assumed constant, the vertical velocity \( v \) at a given depth is a constant, independent of \( x \). Therefore the following relation between time and depth can be written, which will remain true throughout the area.

\[ y = h - a(1 + r)t \quad \text{or} \quad t = \frac{\log \frac{h - y}{a}}{\log (1 + r)} \quad (7) \]

where \( a \) is the initial distance from the bottom at \( t = 0 \) \((a = [y - h] \text{ at } t = 0)\), \( t \) is the time in years, \( h \) = glacier thickness in meters, \( r \) = longitudinal strain rate in years\(^{-1}\), and \( y \) = depth to the particle in question after \( t \) years of movement from the initial position, a distance \( a \) above the bottom.
Several approaches can be used to solve the problem. The most direct is to find, from Nye's expressions,

\[ \frac{dx}{dy} = \frac{dx}{dt} \cdot \frac{dt}{dy} \]

and then to integrate, so that

\[ x = \int_0^y (\frac{dx}{dt} \cdot \frac{dt}{dy}) \, dy \]

Another approach is to use Nye's equation for shear strain (Nye, 1959)

\[ \frac{d^2x}{dt \, dy} = \dot{\gamma} = -2 \sqrt{\frac{\dot{\varepsilon}_x^2 - r^2}{\gamma}} \]

and integrate twice. In both of these cases the integration becomes rather involved. The method used is less direct but leads to a simple solution.

We let \( u_1 \) be the velocity in the \( x \) direction at a given depth \( y \) on the line \( x = 0 \), or, to express the same relation in terms of time, at \( t = 0 \). Nye's expression for velocity at depth \( y \) at \( x = 0 \) (see p. 64 and p. 71) is

\[ u_1 = u_0 - 2g_x \int_0^y \frac{\dot{\varepsilon}_x}{\gamma} \rho y \, dy \]

Let \( u_2 \) be the velocity in the \( x \) direction at a point located a given distance \( x \) down the glacier.

\[ u_2 = u_1 - rx \]

The sign is minus since the area being considered is one of compressive flow. The annual decrease in the velocity, \( u \) of a particle moving down the glacier is equal to the
compressive longitudinal strain rate multiplied by the
distance which the particle would have moved in a year at the
initial velocity. Therefore, a given number of years after
the velocity at depth \( y \) was \( u_1 \),

\[ u_2 = u_1 (1 - r)^t. \]

Equating these two expressions for \( u_2 \):

\[ u_2 = u_1 (1 - r)^t = (u_1 - rx) \]

\[ x = \frac{u_1 [1 - (1 - r)^t]}{r} \]

Substituting the value of \( t \) in terms of \( y \) from (8) gives

\[ x = \frac{u_1 [1 - (1 - r)^{\frac{\log (h - y)}{a \log (h - y)}}]}{r} \]

From this equation values of \( x \) in terms of \( y \) along
a flow path can be calculated directly, for a particle start-
ing at a given distance above the bottom. The results for
three different values of \( a \) are given in Fig. 39 and
Table 9. Note that the initial distance of a particle above
the glacier bottom has comparatively little effect on the
point at which the material reaches the glacier surface, but
it does affect considerably the time required for a particle
to reach the surface. The path of a particle intersects the
surface of the model at approximately 45°, which is the
angle formed between the surface and Nye's slip-line direction.
The slip lines would form in the direction of maximum
Fig. 39.—Diagram showing the trajectory of debris included within the ice. The positions and times in transit have been shown for three different assumptions of initial distance above the glacier bed. The lower diagram shows the trajectory without vertical exaggeration. The shape of a slip line in an ideal plastic is included for comparison.
TABLE 9.--Values of $x$ and $t$ as a function of depth for three initial values of $a$.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>$x$ (m)</th>
<th>$t$ (yrs)</th>
<th>$x$ (m)</th>
<th>$t$ (yrs)</th>
<th>$x$ (m)</th>
<th>$t$ (yrs)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface</td>
<td>1904.29</td>
<td>6092.33</td>
<td>1909.05</td>
<td>7110.88</td>
<td>1910.64</td>
<td>7881.39</td>
</tr>
<tr>
<td>20</td>
<td>1887.23</td>
<td>5889.65</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>50</td>
<td>1851.28</td>
<td>5501.92</td>
<td>1859.14</td>
<td>6511.73</td>
<td></td>
<td></td>
</tr>
<tr>
<td>70</td>
<td>1771.07</td>
<td>5119.15</td>
<td></td>
<td></td>
<td>1785.34</td>
<td>6908.21</td>
</tr>
<tr>
<td>90</td>
<td>1564.69</td>
<td>4551.31</td>
<td>1580.57</td>
<td>5569.87</td>
<td></td>
<td></td>
</tr>
<tr>
<td>110</td>
<td>1025.07</td>
<td>3330.08</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>115</td>
<td>771.17</td>
<td>2559.57</td>
<td>822.49</td>
<td>3578.13</td>
<td>839.59</td>
<td>4348.64</td>
</tr>
</tbody>
</table>
differential shearing (Jaeger, 1956, p. 144), the direction showing maximum movement between adjacent planes (the direction of the maximum shear stress). The slip-line directions also show the orientation of potential fractures by shear (Nye, 1952a, p. 88), if it is assumed that the internal friction is negligible.

In Fig. 39 the upper portion of the curve showing the path of a particle, as calculated above, is very similar to the equivalent portion of the curve showing slip-line direction. Differences in the lower part of the model are due to the fact that the slip-line directions of Nye and Jaeger have been calculated on the assumptions of an ideal plastic using the "von Mise" yield criterion (Jaeger, 1956, p. 93). They assume that at stresses below the yield stress no deformation occurs. In the present analysis, the flow paths have been calculated using values of $u_1$ which are based on the experimentally determined flow law of ice ($\dot{\varepsilon} = k\sigma^n$, see p. 69).

In an ideal plastic, all deformation within the ice block is due to a longitudinal compression (or tension), $\sigma_x$ (Nye, 1952) exactly equal to the yield strength. All movement related to lateral motion is limited to the basal interface, the only place where the shear stress, $\tau_{xy}$, reaches the yield strength (Nye, 1952, p. 87; 1957, pp. 118-119).

When the experimental flow law is used, the values of shear stress, $\tau_{xy}$, are such that appreciable differential flow occurs throughout the lower portion of the ice mass.
The greatest rate of change in longitudinal velocity, versus depth, $du/dy$, (Nye's $\gamma$, 1957) is in the bottommost layers. Near the upper surface $du/dy$ becomes very small as the value of $\gamma_{xy}$ decreases. In this near-surface region the conditions are similar to those in the ideal plastic model and the slip-line direction is not distorted by major shearing within the ice as it is in the basal portion of the ice.

When the vertical component of the movement is considered as being due to compressive flow throughout the ice mass, rather than to movement along discrete shear or fracture planes, the direction of slip between individual grains or along basal plates is still controlled by the direction of maximum shear stress. The deformation then takes place along an infinite number of slip planes spaced infinitely close together. These slip planes all agree with the true direction of maximum shear stress, but only in the upper portion of the ice does the true shear direction approximate the slip line direction as calculated for an ideal plastic material.

In the calculations, either for uniform compressive flow or for discrete movement along shear planes, the material is assumed to be isotropic and to have negligible internal friction. The first assumption is certainly an idealization, as the laminar structures often found in glaciers may be expected to alter the direction of yielding, either by fracture or flow, if they are approximately but not
precisely parallel to the direction of maximum shear stress. Such laminar structures in a glacier are due to changes in crystal size, changes in bubble content and orientation, changes in dirt content, or change in crystal orientation relative to that of neighboring areas. It is assumed that the smaller the angle between a surface of weakness and the slip-line direction, the more likely is deformation to occur on that surface, either by fracture or flow. The effect of internal friction is that the angles between the fractures and the maximum compressive stress direction are less than the 45° angle calculated for the maximum shear stress, the extent of the change being controlled by the hydrostatic pressure as well as by the internal friction. The slip-line direction for yielding in plastic flow is unaffected by internal friction (Jaeger, 1956, p. 91).

The dirt bands observed in the Wilkes Station area are at an angle of 25° or 30° to the horizontal (30° or 35° to the surface). This fairly low angle may be due in part to the effect of internal friction, implying some failure by fracture, or to the presence of anisotropic structures, or it may be related to the form of the ice cross section near the ice margin. The glacier has a roughly parabolic profile in the marginal area rather than being of the uniform thickness assumed in the models.

According to the above calculations for the flow path of a particle, the horizontal distance required to bring
debris from the base to the top of the model glacier is about 2 km. The effect of strain in the third dimension, as is observed at Wilkes, is to decrease this distance. The distance between the 1958 and 1959 equilibrium line and the major moraine is slightly less than one km but of the same order of magnitude as that predicted by the model.

In a more realistic model $h$ and $r$ should be allowed to vary with $x$. However, it would not be correct to take $h$ and $r$ from measurements on an ice sheet with a marginal moraine already present, since the presence of the moraine alters the value of $h$ and probably greatly affects the value of $r$. A better approach is to estimate a reasonable variation of $h$ with $x$ and of $r$ with $x$, based on the variation of ablation rates before a moraine has been formed. In the formulae calculated above, the thickness and strain rate should be average values of $h$ and $r$ over the distance between the equilibrium line and the point $x$ in question. Then, given $h = f(x)$ and $r = g(x)$, to determine $x$ for a given $y$, a system of successive approximations is required. First, a value of $x$ is chosen, thus determining $h$ and $r$. These values of $h$ and $r$ are then used in equation (9) and this gives a value of $x$ which is probably different from that originally chosen. The initial choice of $x$ is corrected and the process repeated until the initial and final values of $x$ are the same.
This more realistic approach did not seem worth attempting in the present example, because only limited data are available about the sub-ice topography and the variation of ablation with position. In the absence of moraine material, r would almost certainly increase downglacier from the equilibrium line, since ablation is greater at lower elevation and the increasing slope promotes run-off. Such an increase in r is not noted at present, because the moraine material protects the ice from ablation and forms an obstacle in the lee of which snow drifts form. If r formerly did increase down glacier, the flow lines would have risen more rapidly than in the model calculation and the distance between the model equilibrium line and the marginal moraine would be decreased. Judging then from the calculations made, it is expected that the equilibrium line when the marginal moraine was first being formed was at least as far inland as it is now, but probably not much further.

The effect of debris on the flow law--
surfaces of apparent shearing

Once debris has been incorporated into the ice, probably by some form of the Weertman freezing-on mechanism, it appears that this debris can be brought to the surface by compressive flow. It has been shown that even with a rather small longitudinal strain rate, compressive flow can accomplish this within reasonable intervals of both distance and time. This process should produce debris surfaces,
perhaps intermittent, within the ice, and these surfaces should be practically coincident with the flow-line direction. The possible influence of this material on subsequent deformation must be considered.

Field evidence suggests that ice which includes a small percentage of debris will deform more readily than clean ice. Abel, as already noted (p. 99), found increased differential movement across certain bands of included material. Swinzow (1962, p. 226) describes the deformation of debris bands with varying debris content in comparison with the deformation of the surrounding ice (see Fig. 40). The bands with a small percentage of fine debris act as zones of weakness, while bands with a high content of coarse material deform less easily than the surrounding ice.

Swinzow (1962, p. 225) suggests a change in the flow law dependent on the percentage of debris included in the ice. For a given applied stress, the strain rate for clear ice should be multiplied by a factor "i" which increases as the proportion of ice decreases, provided the included particles do not come into contact with each other.

\[ \dot{\varepsilon}_{\text{silty ice}} = \dot{\varepsilon}_{\text{ice}} \cdot "i" \] (for a specified stress)

When the included particles come into contact with each other, they will act as reinforcing, provided they deform less readily than the ice, and the strain rate for a given stress will be less than the strain rate for pure ice. A
Fig. 40.—Sketch illustrating the differences in deformation between clean ice, slightly dirty ice, and debris laden ice. Dotted lines represent the tunnel shape before deformation, the solid line, after deformation. In (a) the band of silty ice forms a zone of weakness, and the band protrudes into the tunnel. In (b) the included rocks and pebbles are in contact and the clean ice deforms more readily than the band of coarse debris.
variation in the flow law along laminar debris bands should produce anisotropic conditions. The strain rate measured across a block of ice containing a debris band will vary depending on whether the measurement is parallel to or perpendicular to the debris band.

The mechanism responsible for the observed change in the flow law has not been investigated, but some possibilities are suggested. The presence of inclusions, harder than the ice, may lead to stress concentrations and therefore accelerated plastic flow. If the ice temperature is as high as -1°C, stress concentrations may cause pressure melting and so produce regelation flow. In hardness tests, Barnes and Tabor (1966) produced pressure melting and recrystallization at temperatures down to -1.2°C. The applied stress was on the order of 100 bars, higher than to be expected in nature, though Robin (Barnes and Robin, 1966) has suggested that stress concentrations on the order of 50 bars may be possible at the base of a glacier in the vicinity of irregularities.

The increased deformation due to debris inclusions may also be related to changes in the thickness of the liquid layers at grain boundaries. Nakaya and Matsumoto (1953) demonstrated the presence of a liquid-like layer on ice. More recent work by Hosler, Jensen and Goldshlak (1957, see also Jellinek, 1964) confirmed this. In a water-vapor saturated atmosphere, evidence of a liquid film
was found as low as -25°C. Ice prepared from a 0.1% NaCl solution gave evidence of a thicker liquid layer (Nakaya and Matsumoto, 1953, p. 5). Apparently the soluble NaCl is concentrated at grain boundaries, thus effecting the liquid layer.

It is suggested that the presence of morainal debris may also affect the thickness of the liquid layer. Though much of the debris is probably of low solubility, the more soluble components may be concentrated at the grain boundaries. The expected effect would be less than that found for NaCl. If the change in deformation properties is related to a change in the liquid layer, an increase in the importance of intergranular movements is expected.

At the base of the Meserve Glacier, Antarctica, Holdworth (personal communication) has observed dirty ice which shows greater deformation than the overlying clear ice. In this case the debris-carrying ice also shows a concentration of soluble salts.

Although the presence of zones or surfaces of weakness seems likely, due to the inclusion of moraine material, the total strain observed should not be affected, provided that the glacier has established an equilibrium profile. The strain rate actually measured is an average over the distance of the measurements. Even with a change in the stress-strain rate relationship, caused by the presence of debris, the total movement should not be affected, as it is
ultimately controlled by the negative balance in the area. However, a lower stress would be sufficient to maintain the movement. Accordingly a decrease in thickness and in slope might be expected in those portions of an ice sheet where the stress-strain rate relationship has been modified by the presence of included debris.

**Basal conditions inland from the Moraine at Thule, Greenland**

In 1966 a deep drill hole at Camp Century, roughly 125 miles inland from the moraine at Thule, penetrated the base of the ice at a depth of about 1425 meters (Lyle Hansen, personal communication). The basal ice was found to be at a temperature of -13°C. This implies that there can be no basal sliding, and that no morainal material is being incorporated into the ice base at this location. However, morainal debris is already present in the basal ice, apparently having been frozen-in, somewhere further inland. The ice within 15 m of the bed appears slightly dirty, containing suspended silty material. Beneath what is described as the "ice bed-material interface" the material is mostly sandy, but includes a wedge of coarse pebbles in a silty matrix, and beneath the wedge is an ice lens almost 10 cm thick. Till-like material is found beneath the ice lens.

Velocities at and near the bed have not been determined. However, movement is almost certainly present
throughout the lower portion of the ice which is only slightly dirty. If the 10 cm thick ice "lens" is in fact a continuous layer of wide extent, some movement may occur within it. In that case the true glacier base would be at the contact between the ice lens and the till material and the overlying sand and gravel should then be considered as englacial material. It appears likely that relatively thin sandy layers might occur above thicker ice lenses, in which case movement would also certainly occur in the ice lenses.

The evidence from the Camp Century drilling is consistent with the mechanism of moraine formation discussed here. Although the basal temperature at Camp Century is well below the pressure melting point, moraine material has already been incorporated into the ice at some point further inland. This material will be raised towards the surface as soon as the ice enters a region of compressive flow.

**Suggested sequence for moraine formation**

The following sequence of events is suggested to explain the formation of a marginal moraine. Before the appearance of a moraine, the upper surface of the marginal portion of the ice sheet has a cross section approximating one limb of a parabola and is clear of debris. The basal ice at the edge of the ice sheet is frozen to the bed, but in the central part of the ice sheet basal melting almost certainly occurs. The steepness of the ice front depends
partly on the amount of forward motion of the upper layers relative to the lower ones. If a rather steep ice cliff should form (see Swinzow, 1962), except under special conditions of velocity distribution and differential ablation, it would probably soon be converted to a gradually sloping tongue, due to drift accumulation in the lee of the steep front.

Before marginal moraine formation, the subglacial temperature conditions of the ice sheet are such that either no debris is incorporated into the ice, or debris in the ice is not raised to the upper surface. This implies that the base of the glacier is all at the melting point, or is completely frozen to the bed in the region supplying the future moraine area, or there is no material available on the glacier bed, or that no region of compressive flow exists to raise the debris within the ice.

In the Wilkes Area, the most reasonable supposition is that before the time of moraine formation, the ice sheet was of greater extent, probably terminating in a calving ice cliff. Under such circumstances the zone of basal melting might have extended to the sea, and in any event, there would have been no ablation area where compressive flow could raise debris to the surface. It is likely that moraine-forming conditions began to develop as soon as the ice sheet margin had retreated to, and begun to stabilize, near its present position.
With the change in conditions at ice base, material is incorporated into the basal ice and carried along the flow lines to the surface. The concentration of included morainal material may vary from finely disseminated silt to masses of sand and gravel as much as 1/2 m thick. The uneven distribution of included material may lead to large variations, both in the rate of moraine formation and in the distribution of material at the surface. Initially the presence of debris at the surface promotes increased melting, but soon a protective layer of debris is formed, ablation is reduced, and the debris-covered area becomes a ridge.

Once a marginal moraine is formed, the conditions at the ice margin are altered. It is possible that later, dirt-laden "shear zones" will appear downglacier from the initial moraine, due to the decrease in r across the moraine area, caused by the reduced ablation, and by an increase in r downglacier from the moraine. At Wilkes, limited outcrops of relatively unweathered moraine material occur in the ramp area, downglacier from the major moraine ridge. On the other hand retreat of the ice margin and equilibrium line may cause new "shears" to form upglacier from the original moraine.

Once formed, marginal moraines tend to perpetuate themselves and to maintain the position of the outer margin of the ice by encouraging the accumulation of drifts down-
glacier from the moraine. A pronounced increase in ablation in the moraine area and further inland seems necessary to initiate a retreat.

In the Wilkes Station area snow drift accumulation behind the newly formed moraine ridge and subsequent differential ablation has probably produced the closed or partially closed, oblong dirt bands which are visible on the ramp downglacier from the moraine ridge (Fig. 41). Dirt, blown from the newly exposed moraine and from local rock outcrops, formed layers on the surface of a large snow drift, accumulating in the lee of the growing debris ridge. Several dirty surfaces formed as the moraine ridge increased in size. The increase in ice thickness due to snow drift accumulation produced increased movement in the middle of the ramp. Where the slope has bulged outward, its surface no longer has the proper shape to remain in the lee of the moraine. Ablation of the former drift snow in these uplifted areas has cut through the upraised portions of the dirt bands giving the lineations now observed (see Fig. 42).

Although marginal moraines can form only at the ablating edge of a glacier, it is likely that their formation may actually cause some thickening and possibly some slight advance of the marginal ice. At any rate, for the formation and preservation of such a moraine, the position of the ice margin must be rather stable. For the formation of a moraine such as that at Wilkes Station, the conditions
Fig. 41.--Photo showing the curved lineations on the ramp (U.S. Navy photo).

Fig. 42.--Sketches illustrating a possible mechanism for the formation of ramp lineations. (a) Layers of drift snow, with debris on the summer melt surface, have formed on the ramp in the lee of the moraine. (b) The ramp surface is bulged outward by renewed movement. (c) Subsequent ablation on the ramp has cut through the upraised portion of the old drift snow surfaces.
cannot change very much for a period probably on the order of at least six thousand years (see Table 9, p. 120).

**Age of the Windmill Island ice-free area and comparison with other areas**

The inferred stability of the position of the ice margin during moraine formation implies that the rock areas of the Windmill Islands have been exposed for at least a similar period of time. The date of the ice retreat from the Windmill Islands has not been investigated directly, but at Bunger Hills, a similar area about 300 miles to the west, the Russians have studied this problem (Rozycki, 1960). From the morphology of raised beaches, they have deduced that the ice retreated from the Bunger Hills area several thousand years ago at the beginning of the post-glacial climatic optimum. This would be about 7,500 years ago according to Flint (1957, pp. 377-380). Also in the ice-free areas near McMurdo Sound, Pévé, from a variety of evidence, has concluded that fresh-looking moraines, some of them ice-cored, within a mile of the present ice front are at least 6,000 years old (Pévé, 1962, p. 100). The ice has retreated only a rather small distance during that time.

In the Windmill Islands raised beaches, apparently similar to those found at Bunger Hills, are found up to 30 m above present sea level, indicating that the area was ice-free during a period when either the sea level was...
higher or the land was still depressed. A coraline algae from a 23 m beach has been dated at 6,040 \pm 250 yrs (Cameron, 1963, p. 7) which is in good agreement with the moraine evidence for the age of the area. In the southern part of the area, cavernous weathering of igneous rocks also suggests a considerable period of ice-free conditions.
ANALYSIS OF STAKE AND RELATED MEASUREMENTS

General meteorological conditions

Before presenting a detailed analysis of the stake measurements and estimates of the relative importance of the various processes which contribute to accumulation or ablation, a general review of the Antarctic meteorological situation is presented.

The general circulation pattern of the Antarctic results from the thermal interaction between the cold surface of the Antarctic continent, the comparatively warm ocean waters which surround it, and the atmosphere (Gousev, 1960, p. 233). The circulation shows a stratification of two layers. In the lower thin layer cold air slides off the continental slopes toward the sea, while in the upper layer warm maritime air moves inland. As a result of this circulation an inversion of as much as 25°C is often noted in the lower layers of the atmosphere above the continent. The inversion extends to an altitude of 200 to 300 m over the continent, and to an altitude of 1000 to 2000 m in the coastal districts.

Over east Antarctica there is a semipermanent high-pressure area, due to the presence of the cold continental air mass. This Antarctic high is surrounded by a
semipermanent low pressure zone, which is approximately at the Antarctic front, the line where the boundary surface between the cold continental air mass and the surrounding and overlying warmer air mass intersects the ocean surface. Pressure troughs and ridges along the low-pressure zone appear to originate as wave phenomena along this surface (Gousev, 1960, p. 235). Additional cyclonic centers originate in lower latitudes over the southern oceans and then migrate south along several southeastward-trending tracks, finally entering the zone of low pressure surrounding the Antarctic (Astapenko, 1960, p. 249, Fig. 4). Thereafter they may continue eastward in this zone or major trough until they lose their identity, become quasi-stationary or move slowly westward, or they may move inland over the continent (Gibbs, 1960, p. 49). At several positions along the trough, cyclones may tend to become stationary, and it is primarily from these positions that depressions move southwestward or southward over the continent. One of these positions is located on the coast of Wilkes Land, to the west of Wilkes Station, between Wilkes and Bunner's Oasis, or Bunner Hills (Astapenko, 1961, p. 250).

The strong katabatic, or gravity induced, winds, a characteristic of many Antarctic coastal stations, are related to the general circulation though their direction is not in general parallel to the isobars, as is approximately the case for normal cyclonic circulation. (Conse-
quently, surface wind observations are often of little use for synoptic analysis.) The katabatic winds are caused by radiation cooling of air on the slopes of the ice-sheet, and the movement of this cold air under the influence of gravity.

The average rate of radiation cooling per unit area in the Antarctic is equivalent to the typical rate of cooling on a clear night in middle latitudes. Under these conditions in middle latitudes, gradient winds, related to cyclonic circulation, of up to 15 knots have been suppressed and replaced by local katabatic winds (Ball, 1960, p. 10). Gradient winds of this velocity or more are undoubtedly suppressed on occasion in the Antarctic in the area of the sloping continental margins.

The direction of katabatic winds is generally in the direction of maximum slope, though some leftward deflection may occur due to the coriolis acceleration. Local irregu-
larities in the slope may also considerably affect the wind speed and direction. Where a topographic trough exists, a strong funnel effect with unusually high winds may occur (Taubert, 1960, p. 53; Ball, 1960, pp. 14-16). Where katabatic winds are particularly favored, the average wind speed may be 2 1/2 times that of areas not influenced by katabatic winds (Taubert, 1960, p. 54).

Formerly, strong katabatic winds were believed to be characteristic of the entire east Antarctic coastline. However, observations during the International Geophysical
Year and later show that the high-speed seaward winds are concentrated in certain areas. It is not by chance, though, that Antarctic stations have been placed in locations with strong katabatic winds, for it is the strong seaward winds that break up the ice near the coast and allow ships to approach most easily in these locations (Dzerdzeevskii, 1960, p. 50). In other areas coastal winds may have a generally landward direction (Morita and Murakoshi, 1960, p. 30).

Winds with a seaward direction may be produced by two different processes (Dzerdzeevskii, 1960, p. 49). Purely katabatic winds originate as described above, generally under conditions of weak barometric pressure gradients. In general, the wind direction is not along the isobars and may often be perpendicular to them. The thickness of air flow during purely katabatic winds is not great, and the extent of penetration over the sea, beyond the effect of the continental slope, is very limited. A diurnal variation may be noted, particularly in the warm season, when upslope thermal currents may develop during the day at the continental margin.

Seaward winds may also be related to cyclonic activity over the sea. Since circulation about a low pressure center is clockwise in the southern hemisphere, a strong seaward flow of air occurs along the western edge of a cyclonic center. As a center passes a coastal station in
an eastward direction, the station first records winds from the north or northwest, or at least a reduction of katabatic winds. If cyclonic circulation is strong enough, considerable warm maritime air may be carried inland, and if sufficient moisture is present, precipitation may result, as the maritime air is forced upward by the continent. As the low pressure system passes, a calm, or easterly winds are observed, or if the system is not very strong, katabatic winds may return or increase. After the center of the low pressure system has passed, a strong south wind is observed, because at this time the direction of cyclonic circulation is in the same direction as the continental slope. Under these conditions the depth of continental air moving coastward is considerable, and the winds may extend a considerable distance out to sea. In areas where topographic features favor strong katabatic winds, extremely high winds may be expected when a cyclonic center lies just to the east.

Wilkes Station is located in an area where katabatic winds occur frequently, though they are not so severe there as at Cape Denison or Port Martin. Gravity winds seem to be more pronounced at the Vanderford Glacier where a topographic trough exists. During the 1958 spring traverse, persistent winds were encountered in the region south of the Vanderford where the Vanderford trough extends inland. Unusually high surface snow densities (0.4 to 0.6 g/cm³) and some bare ice areas were recorded in this area, as well as
considerable sastrugi development. In the area immediately adjacent to Wilkes Station the katabatic winds are from the east or southeast, because the coastline here, along the east side of Vincennes Bay, runs north-south.

A consideration of the varying wind conditions at S-2 and at the Base Station (see pp. 151-153, and Fig. 44) indicates that coastal conditions are related primarily to cyclonic circulation, while at S-2 the winds are normally katabatic. Only rarely does the influence of cyclonic circulation extend as far inland as S-2.

Seasonal changes in surface conditions in the Wilkes area

Because the 1958 measurements have been most used in the detailed analysis of accumulation and ablation, the surface conditions throughout the year will be described briefly.

When the new party arrived on January 25, 1958, rapid melting was in progress on the ramp, and in the area just inland of the marginal moraine. On the ramp, numerous open stream channels were filled with rapidly flowing water and near the moraine (elev. 142 m.) the surface was one of coarse granular crystals. Stake measurements show this area to be in the ablation zone, and since the surface generally remains as ice even in winter, the loosened crystals probably were formed from solid ice by radiation melting along crystal boundaries. Further inland, particularly at
the south end of the two cross profiles (see Map 1), large areas of smooth ice were covered by a thin sheet of flowing melt water (elev. about 375 m).

By late February melting had almost ceased and only on the warmest days was any melt water produced. The lower area of the ice sheet, near stake 402 (alt. about 160 m.), seemed much icier than it had been at the height of the season. Apparently the crystals that had been loosened by radiation melting had refrozen to form a solid ice surface. The most icy surface conditions were recorded in March when all of the partially melted portions of the surface had refrozen, but new accumulation was not yet present. Inland from F1-2 (elev. 342 m.) and along the upper profile, some stakes never showed ice at the surface, even in summer, but only very icy snow.

Melt water channels were observed along the edge of the ice sheet, extending inland to the area between stakes 403 and 404 (elev. 170-175 m) and sheet flow of water was observed in higher areas. Some shallow parallel depressions which crossed the upper profile nearly at right angles may have been related to drainage, but alternatively may have been due to wind effects. Bare ice areas also occurred to the east of the upper profile but no ablation measurements were made there.

In general, significant accumulation did not occur until May, and most of the accumulation was in July of
later. The new snow tended to blow off the bare ice areas, and consequently the more icy areas stayed free of snow longer than other areas.

Often precipitation occurred during a calm or with little wind as the eastern edge of a cyclone passed Wilkes. With the return of the usual wind as the low pressure center moved westward, much of the new fallen snow was removed leaving only remnants of the new snow in drifts standing up to a foot high. A snowfall remained on the surface only if a protective crust could develop before the reoccurrence of high winds. In the spring such a crust developed due to melting at the surface during the day and subsequent refreezing at night. Occasionally an accumulation of wind-packed snow was compact enough to withstand the wind without a further protective crust.

The melt of the 1958-59 summer was much less than that of the year before. Even in December 1958 and January 1959, bare ice was exposed only at the edge of the ice sheet in the area of stake 402. Only part of the winter snow in the superimposed ice zone was changed to ice, so that snow remained on the surface over most of the area that had been exposed ice in the 1957-58 summer. The conditions in 1959 were probably intermediate between those of 1957 and 1958.
Processes contributing to accumulation and ablation

Many processes are involved in the deposition and removal of snow, some of which are very important while the effect of others is almost negligible. The relative importance of the various processes may also vary from place to place. The accumulation processes in the Wilkes Station area are precipitation, blowing snow, and condensation, while the ablation processes are melt and run-off, erosion by wind or deflation, and evaporation. The formation of superimposed ice is a secondary accumulation process, as it involves only a change in the form of the material already supplied to the surface, usually with only minor movement of the material. The importance of the various processes will be discussed separately.

Precipitation. Precipitation, normally as snow, is the initial source of most of the accumulation in the Antarctic, though the snow may be shifted from place to place several times, or may be melted and refrozen before it is permanently incorporated into the ice sheet. In 1958-59, in the area of Wilkes Station, significant accumulation was almost always due to a definite snow fall.

1The term sublimation should properly be used for the changes between solid and vapor in either direction. The terms "evaporation" and "condensation" are used here for these changes because it is felt that these terms indicate less ambiguously whether it is the process of adding material, or of removing material that is being referred to.
usually under calm conditions, though much of the loose snow was later removed by high winds.

An inspection of surface weather maps shows that practically all the precipitation recorded at Wilkes Station or at S-2 was associated with the passage of low pressure systems along the circum-Antarctic low pressure trough. This is shown particularly clearly at S-2 where precipitation invariably occurred only with winds from the northeast to northwest, instead of the usual east and southeast, and with low wind speeds. In other words, precipitation occurred when a cyclonic center was slightly to the west of S-2 and maritime air was brought inland along the leading edge of the cyclonic system. Periods with precipitation at S-2 were less numerous than on the coast and occurred only when the center of a low pressure system passed south of the edge of the continent or along it. Accumulation recorded when there was no evidence of cyclonic influence was always associated with high winds and blowing snow, and probably was not due to precipitation.

**Blowing snow**—wind transport and deposition. The quantity of snow transported past any given location depends on the wind velocity and on the availability of new snow. Because wind velocity varies with height above the ground, the variation of drift snow density with height for given wind conditions must be considered. Investigations made by the Australians at Wilkes Station in 1959 (Dingle
and Radok, 1960; Budd, 1966) showed a logarithmic vertical wind profile according to the equation

\[ V_z = \left( \frac{u_*}{k} \right) \log_e \left( \frac{z}{z_0} \right) \]

where \( V_z \) is the wind velocity at height \( z \), \( u_* \) is the wind velocity gradient, \( k \) is a Prandl constant, and \( z_0 \) is the roughness parameter, or the height of the top of the surface layer of zero velocity. These results agree with those of other workers (Bagnold, 1936; Lilquist, 1956; Loewe, 1956; Lister and Taylor, 1961). For a logarithmic wind profile the calculated equation for variation of drift density with height is

\[ n_z = n_h \left( \frac{z}{h} \right)^{-\frac{w}{ku_*}} \]

where \( n_z \) and \( n_h \) are the drift densities at height \( z \) and \( h \) respectively and \( w \) is the fall velocity (Dingle and Radok, 1960, p. 77, for further background see Loewe, 1956, pp. 125-132). From this a relationship of drift density to wind velocity is derived,

\[ \log n_z = \log n_h - \left( \frac{w}{k^2} \right) \left[ \log_e \left( \frac{z}{z_0} \right) \log(z/n) \right] (V_z^{-1}) \]

which can be stated, after simplifying, as

\[ n_z = n(\infty) \left( 10^{V_z} \right)^{-1} \]

The meaning of \( n(\infty) \) will be explained below.
Previous investigators have presented empirical power-law relations which approximate the given exponential law over the limited velocity range investigated in the particular study. However, an important difference in the two kinds of laws may be noted. With a power law, as \( V \) becomes very large the drift density also increases indefinitely, but with an exponential law, as \( V \) becomes very large the drift density approaches a finite value. This value is the \( n(\infty) \) given above. Furthermore, data from Wilkes Station (see Fig. 43) suggests that for very high wind velocities the drift density at all levels, at least up through 4 m, approaches the same limiting value (about 350 g/m\(^3\)). The results of the experimental work at Wilkes by Dingle and Radok are in agreement with their theoretical calculations, briefly summarized above.

Provided therefore that sufficient snow is available, either as precipitation or already on the ground, the quantity of drift snow transported should agree with the theoretical relationships. If the wind velocity (at some given reference level) decreases in the direction of the wind path there will be deposition, and if the wind velocity increases, there will be deflation. In an area where the lateral wind velocity gradient is zero, the amount of snow that is deposited from the blowing snow layer will be equal to the quantity which is added as precipitation.
Fig. 43. Drift densities observed at Wilkes Station at 3 levels plotted as a function of reciprocal wind velocity, $V_z^{-1}$. Note that at all 3 levels the drift density approaches a maximum value of about 350 gm/m$^3$ with increasing wind velocity (from Dingle and Radok, 1960, p. 83).
In the discussion above, a sufficient supply of snow has been assumed. Frequently the controlling factor for the occurrence of blowing snow is the availability of loose snow. An icy surface or a well-developed crust may prevent erosion and even rather high winds may transport very little snow.

Comparison of coastal winds at Wilkes with winds at S-2. A comparison of the winds at Wilkes Base Station with the winds at S-2 has been made for a three-month period when the S-2 records are complete (see Fig. 44 as an example). The winds at the coastal station shift frequently in direction and speed, with winds frequently from the north and west as well as from the east and south. At the same time, the wind at S-2 remains predominantly from the southeast, varying to east and south, and only rarely from the north or west. The wind velocity at S-2 is almost always greater than at the base.

It appears that the winds along the coast are related mostly to cyclonic circulation or local thermal effects, while the winds at S-2 are generally katabatic. The influence of only a few of the cyclonic centers extends as far inland as S-2. Westerly winds are unusual at both locations, since they can occur only when a cyclonic center passes to the south of the observing station.

Measurements of wind velocity were also made at S-1 during a short period in January of 1959 (see Hollin et al.,
Fig. 44.—A comparison of winds at S-2 with winds at Wilkes Station for the month of November, 1957. Periods with precipitation are denoted by an "s" beside the mark showing wind direction.
1961, p. 68). The velocity at S-1 is generally slightly greater than at the base.

In the above comparison, a marked decrease in wind velocity is observed between S-2 and Wilkes Station. The wind velocity at S-1 is only slightly higher than at the base. From the available data, however, it is impossible to determine whether the decrease is gradual or whether it takes place over a rather short distance. As the slope increases between S-2 and the coast, the wind velocity may increase slightly, and then decrease sharply near the coast.

The decrease in wind velocity should cause some deposition between S-2 and the coast during times of blowing snow. However, since it is not known whether the decrease is gradual or localized, the deposition might be either widespread or occur in a limited area.

Wind controlled accumulation and deflation near Wilkes Station. The Wilkes Station records show many times when unconsolidated snow from a previous storm was removed during periods of high wind. Less frequently the accumulation recorded on the ice sheet was more than the precipitation measured at the base, and one case of major accumulation was recorded on the ice sheet when no precipitation was recorded at base. In all of the latter cases, blowing snow was recorded for at least part of the period in question. Two examples are given below.
During the period from August 11 to 19 only 2.5 cm of accumulation were recorded at base and 7 observations of a trace of precipitation. Twelve observations of blowing snow with winds up to 56 knots were also recorded. (These observations are from the 3-hourly Weather Bureau observations and the method of observation is not stated.) During the same period 9 representative stakes on the ice cap showed an average of 11 cm of accumulation each, and all stakes showed at least 5 cm gain in snow.

From November 17 to 24 a trace of precipitation was recorded at base on 5 occasions and blowing snow with winds up to 36 knots on 6 observations. At the same time 8 out of 9 stakes on the inland ice showed gains of from 1 to 6 cm with an average over the 9 stakes of more than 2 cm of snow. The snow must have been deposited in both cases during the periods of blowing snow.

The more usual effect of high winds was very uneven deflation and deposition, deflation usually predominating. The erosion was uneven, controlled by local irregularities in the hardness of the surface. Sculptured drift forms, up to 30 cm high, were sometimes observed when a partially consolidated snow fall was partly removed. Local accumulation during periods of high wind was often caused by the presence of a shifting drift at a particular stake location. This kind of temporary accumulation is shown by the weekly surface changes at individual stakes; a week with an
inconsistently high accumulation at a given stake was almost always followed by a comparable loss during the next week.

The small areal extent of these irregularities in short-term accumulation is shown by the data from locations where two stakes were measured. On a pair of stakes a meter or so apart, the magnitude of the change in the difference between readings in successive weeks was sometimes as much as 5 cm of snow, but usually much less.

The larger scale irregularities in accumulation were probably caused by local wind effects and preferred areas of deposition in individual storms, possibly related to a sharp decrease in wind velocity over a short distance. An effective sudden decrease in wind velocity at a given level can occur if the wind stream suddenly rises away from the surface, as suggested by Ball (1960, p. 70). Given a thin sheet of air flowing rapidly down a slope, if the velocity exceeds a certain critical value, depending on the slope and the intensity of the inversion, the sheet of moving air will lift away from the surface. Upslope circulation of thermal currents may also be responsible for lifting a katabatic wind. A theoretical model of the behavior of katabatic winds in the region of sharp deceleration near the coast has been worked out by Ball (1956).

On several occasions at Wilkes a phenomena was observed which may have been caused as the moving air layer
lifted away from the surface. A thick cloud of blowing snow was observed over the ice sheet but it terminated sharply with a roll-like appearance, either near the marginal moraine or somewhat further inland, in the vicinity of BF-2. Considerable deposition might be associated with such an abrupt termination of the blowing snow area. In a few cases, soundings of new snow depths after storms showed local areas of high accumulation which might be related to this phenomena (see Fig. 45). More localized unevenness in the depth of new snow was due to old hard drifts which had been hidden beneath an even surface of new snow.

Condensation and evaporation. Condensation or evaporation at an ice (or water) surface occurs whenever a vapor-pressure gradient exists above the surface, the water vapor moving from the region of high vapor pressure to that of lower vapor pressure. The rate of vapor transfer depends among other things on the steepness of the vapor-pressure gradient. The vapor pressure over an ice surface depends on the ice temperature, while that of the air mass depends on the specific humidity of the air as well as on the temperature. Wind is often important in causing a high rate of evaporation or condensation, because it continually changes the air next to the surface, thus maintaining a steep vapor-pressure gradient. Winds also shift air masses from one region to another and may bring in air with either relatively high or low vapor content.
Fig. 45.--Profile showing the variable depth of new snow after a single storm as measured by soundings on November 19, 1958.
When considering evaporation or condensation, the nature of the ice sheet surface must be considered, whether it is new snow, old snow, or ice, since both the physical and thermal properties of these materials vary considerably. Because snow of low density is quite permeable, a vapor-pressure gradient, dependent on temperature, can exist within the snow and, if the permeability extends to the surface, can be removed to the outside air. Also the effect of wind may penetrate beneath the snow surface.

Because evaporation or condensation either uses or liberates large amounts of heat energy, the rate at which this energy can be supplied to or removed from the surface is very important. As the thermal conductivity of ice is about 10 times that of average snow, the heat available for evaporation (or removed during condensation) by conduction through ice is about 10 times that available by conduction through snow for equivalent thermal gradients (Williams, 1963, p. 35). This effect is partly offset by increased transport of heat by air moving within the snow.

Short-wave length radiation also contributes more energy to an ice surface than to a snow surface, because the albedo of an ice surface is much less than that of a snow surface. The quantity of turbulent heat transfer depends primarily on wind conditions and thermal gradient over the surface, and should be roughly equivalent over ice and snow.
If the vapor pressure gradient produces an evaporation rate which requires more heat energy than can be supplied to the surface, the temperature of the surface is lowered, thus reducing the vapor pressure gradient.

The experiments of Fitzgerald (1886, p. 610) show that under the same conditions, evaporation from a snow surface is about half of that from an ice surface. The differences in conductivity and in the albedo of the surface are responsible for the difference.

During 1958 no equipment was available at Wilkes for the measurement of the vapor-pressure gradient above the ice-sheet surface, and therefore no calculations of the vapor transfer can be made. However, from stake measurements and observation of the surface, the importance of evaporation or condensation can be estimated.

Condensation seemed to be quite rare and probably produced very little accumulation. During 1958 the only observation of hoar crystals forming on the ice and snow surface was on August 4. The air temperature at the time was much higher than on the previous few days, so it is likely that the ice-surface temperature was lower than the air temperature. On three occasions rime was noticed on the guy wires of the S-1 shelter and on the stakes, but was not noted on the ice surface. Temperature data suggest that the ice surface was warmer than the air, so probably
had a higher vapor pressure, with consequent loss by evaporation.

In most areas of the Antarctic, evaporation is normally considered to be of minor importance, but it can be significant in local situations, particularly in bare-ice areas. Mellor reports for Mac-Robertson Land an average daily evaporation rate in winter of 0.5 mm, and a summer evaporation rate of at least 2 mm per day (Mellor, 1959, pp. 528-529). An estimate of the evaporation in the Wilkes area can be obtained from areas with ice at the surface. Any lowering of an ice surface, as shown by stake measurements, during periods when no melting occurs must be due to evaporation from the ice surface.

To analyze the stake figures for evaporation, all the readings when ice was recorded at a stake have been plotted. Where a drop in the ice surface was recorded during a period when the S-l temperatures were well below 0°C and no evidence of melt was observed, it has been assumed that the decrease was due to evaporation. These measurements indicate evaporation values of up to 1 cm of water equivalent during certain weeks. This approaches a maximum figure rather than an average, since during some periods ice was recorded for consecutive weekly readings but the surface showed no change, or rarely, even a net gain. On the other hand, even during weeks which showed a
loss, undoubtedly periods occurred when favorable evaporation conditions did not exist.

**Blowing snow in relation to evaporation and condensation.** When considering evaporation and condensation, the occurrence of blowing snow is important because it alters the conditions of the vapor-pressure gradient. MacDowell (1960, pp. 156-157) has pointed out the importance of evaporation of blowing snow. At the surface of small snow grains a relatively high vapor pressure exists due to the short radius of curvature. Evaporation occurs provided the vapor pressure at the surface of snow grains is higher than that of the surrounding air mass. This evaporation produces a layer of air which may even be supersaturated relative to a plane surface of ice. Such a layer with high specific humidity will greatly reduce evaporation from the glacier surface, and may even cause condensation onto the surface.

One observation suggests that this process may be related to the formation of wind crusts. On June 24 a wind crust was observed on a snow surface. Close observation revealed that the crust continued across an adjacent bare ice surface as a thin layer of milky ice and the continuity of the layer suggests that the whole layer may have been added by condensation. During the previous week long periods of blowing snow had been recorded.
In addition, as a consequence of the heat required for the high evaporation in a blowing snow mass, a heat sink is created in the blowing-snow layer. Therefore, during times of blowing snow, the extrapolation of temperature and vapor-pressure gradients from readings at two levels is questionable.

Superimposed ice. Under a combination of special conditions superimposed ice may be an important form of accumulation in the limited area between the zone of snow accumulation and the peripheral zone of ablation. The ice temperature immediately below the surface must be well below 0°C so that melt water can be refrozen, and the winter snow fall must be small enough so that the heat energy available in the summer months can convert all the snow to slush. Otherwise the accumulation will be in the form of thick lenses of ice inter-layered with soaked firn. The formation of superimposed ice is a gradual process continuing throughout the summer season. As the snow cover melts, the water trickles down, raising the temperature of channels in the lower layers of snow to 0°C. When the melt water reaches the impermeable ice surface it refreezes, forming superimposed ice. Given the proper temperature and snow conditions, the area must also be poorly drained so that the melt water does not flow away. Accordingly superimposed ice is generally not found on the relatively steep slopes of mountain glaciers.
In the Wilkes area additions to the ice sheet in the form of superimposed ice were not apparent from the surface observations until the end of the season. In January 1958 the surface material was granular and by late summer slush covered much of the lower ice sheet and ramp. When all the water in the surface layers froze, the surface took on an icy appearance and slight additional melt on warm days later smoothed much of the roughness. Locally, surface melt, even slightly inland of BF-2, occurred as late as the end of March and early April, as evidenced by the formation of smooth sheets of frozen melt water.

Melt and run-off. The occurrence of run-off requires two conditions, first the necessary melting and second, sufficient slope to allow the melt water to flow away. Where the temperature of the ice immediately below the surface is lower than 0°C, the surface slope determines whether the melt water forms superimposed ice or is removed as run-off. The snow cover is generally converted to a slushy condition before appreciable drainage occurs.

In the Antarctic, the proper conditions for run-off occurs only rarely, but when they do this is usually the major cause of ablation. At Wilkes Station, ablation is important only where run-off is important and the amount of run-off varies greatly from year to year. In the moraine area, the same area where the slope steepens, the presence
of dirt greatly increases the amount of melting. In the ramp area the importance of the run-off is shown by the well-developed streams which flow in the summer.

Relative contribution of various processes to the specific balance --preparation of table

From an analysis of all the weekly surface changes, an estimate has been made of the contribution of each process to the total of accumulation and ablation recorded at selected stakes. The figures given are not very accurate as the compilation involves much estimation and some assumptions, but it is believed that the resulting table does point out the importance which blowing snow may have at some places, and the local differences in accumulation. The estimate has been made for only a limited number of locations along the trail to S-2, the locations having been picked for their spacing along the trail with regard to different types of accumulation facies. Stakes which show very unusual behavior have not been used.

Stake 402 is in the ablation area, S-1 is in the superimposed ice area, and stake 416 is at the inner edge of the superimposed ice area where the final accumulation is sometimes as ice lenses. For these three stakes weekly measurements are available so the true total of material both added and removed from the surface, or total mass exchange, is more nearly approximated than at stakes with
TABLE 10.—Estimates of the relative importance of the processes of accumulation and ablation at six locations from the ice sheet margin to S-2.

<table>
<thead>
<tr>
<th></th>
<th>402 (mile 15)</th>
<th>S-1 (mile 45)</th>
<th>416 (mile 15)</th>
<th>Fl-6 (mile 45)</th>
<th>Fl-1 (mile 45)</th>
<th>S-2 (mile 45)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Precipitation</td>
<td>12.6</td>
<td>15.5</td>
<td>15.5</td>
<td>15.5</td>
<td>15.5</td>
<td>15.5</td>
</tr>
<tr>
<td>Blowing Snow (in)</td>
<td>4.0</td>
<td>19.3</td>
<td>21.9</td>
<td>4.3</td>
<td>4.0</td>
<td>6.8</td>
</tr>
<tr>
<td>Sublimation and evaporation (in)</td>
<td>1.2</td>
<td>0.5</td>
<td>0.5</td>
<td>0.2</td>
<td>0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>Runoff (in)</td>
<td>1.2</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Total in</td>
<td>19.0</td>
<td>35.3</td>
<td>37.9</td>
<td>19.8</td>
<td>19.6</td>
<td>22.4</td>
</tr>
<tr>
<td>Blowing snow (out)</td>
<td>14.6</td>
<td>18.7</td>
<td>14.8</td>
<td>6.2</td>
<td>4.8</td>
<td>2.0</td>
</tr>
<tr>
<td>Sublimation and evaporation (out)</td>
<td>6.0</td>
<td>4.0</td>
<td>3.0</td>
<td>2.0</td>
<td>2.0</td>
<td>2.0</td>
</tr>
<tr>
<td>Runoff (out)</td>
<td>9.2</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>Total out</td>
<td>29.8</td>
<td>22.7</td>
<td>17.8</td>
<td>8.2</td>
<td>6.8</td>
<td>4.0</td>
</tr>
<tr>
<td>Specific balance</td>
<td>-10.8</td>
<td>12.6</td>
<td>20.1</td>
<td>11.6</td>
<td>12.8</td>
<td>18.4</td>
</tr>
<tr>
<td>Sum (in and out)</td>
<td>48.8</td>
<td>58.0</td>
<td>55.7</td>
<td>28.0</td>
<td>26.4</td>
<td>26.4</td>
</tr>
</tbody>
</table>
less frequent measurements. Mile 20 and mile 45 are typical of the accumulation area along the trail east of mile 10. Here stake measurements are only available about once a month so the totals of accumulation and ablation are not so well represented. However, the accumulation is more regular than near the coast so the difference from the true total mass exchange is probably not great. S-2 is included because it is the location farthest inland where stake data are available and good pit observations are also available. The accumulation figure used at S-2 is the average of the 11 stakes in the deformation network. The total mass exchange is estimated from the measurements on the stakes along the S-2 trail.

In the conversion of the measurements of snow accumulation to water equivalent, the choice of a proper density value is a problem. In the inland areas an average density has been taken from the pit data. This is an oversimplification but is valid for the portion of accumulation that is permanent, the specific positive balance. It leads to an overemphasis of the comparatively low-density new snowfalls which are later removed, and thus to a slight overestimate of the total mass exchange.

When converting the measurements of the three coastal stakes to water equivalent, the superimposed ice accumulation has been allowed for as follows. The thickness of cover over the superimposed ice was taken from soundings to the
hard surface, and from the pits nearest the stake localities. Since the ice is newly formed ice, a rather low density value of 0.7 g/cm\(^3\) has been used except at stake 402 where the ice is less porous. From the available measurements, the density of the snow above the ice has been estimated to be 0.38 g/cm\(^3\). Superimposed ice is not considered as one of the accumulation mechanisms in the table since it is a secondary conversion of material which has already been supplied to the surface.

In compiling the table the precipitation has been assumed constant over the whole area. The value used is intermediate between the accumulation recorded along the S-2 trail, and that recorded at S-2. At stake 402 a slightly lower value for precipitation is used because at this stake part of the precipitation was probably never deposited at the surface. Note that the stake observations cannot distinguish between precipitation and blowing snow. The accumulation at S-2 is suspected to be abnormally high since erosional features were more unusual there than at other localities.

A value for the variation in precipitation with distance from the coast might be arrived at as follows. The total moisture in a given layer of air above the surface could be calculated from the radio-sonde data. If the surface rises, the air layer will be forced upward and will be cooled adiabatically. The amount of condensation that
this cooling would cause could then be calculated. However, in the present case, too many variables are unknown to make such a calculation worth while. The proper thickness of the air layer to be considered is not known, the thickness of the katabatic wind is not known, and the amount of lateral loss cannot be determined. Since a valid value of the variation of precipitation cannot be determined, a constant value has been used.

This selected value for precipitation gives a slight loss due to blowing snow at mile 20 and mile 45, suggesting a slightly increasing wind speed between S-2 and the coast. The other 5-mile flags show a small positive balance while S-2 shows an anomalous gain from blowing snow. The coastal stakes, except for stake 402, show a gain from blowing snow, as expected if the wind speed decreases near the coast. At stake 402 little snow, either from precipitation or blowing snow, is deposited and what is deposited is easily removed from the smooth ice surface.

Evaporation and run-off have been estimated by considering the losses from ice surfaces in conjunction with the concurrent S-1 temperatures.
DISCUSSION OF THE BALANCE PROFILE

Variation in specific balance along the profile

Profiles showing the variation in specific balance across the margin of the ice sheet have been prepared as described on pp. 36-46. Individual stake values and running-mean values of both snow accumulation and water equivalent have been plotted as a function of distance from the edge of the ice sheet. As expected, an area of low accumulation is found at the edge of the ice sheet where the ablation is high, but in addition an area of little gain or even loss occurs a short distance inland from S-1. At stake 413 (see Figs. 15-17) the annual balance is very low or sometimes negative, while nearby at stake 411 the positive balance is unusually large. This irregularity is apparent both in the profiles based on running means and in the profiles of individual stake values.

Before examining the possible causes of this irregular pattern of the specific balance, it is necessary to consider possible sources of error. The most likely source of error is that due to incorrect assumptions of the density of the positive-balance layer, and the neglect of changes in density during the year. Where the positive-balance layer
for the year is snow and the density increases so much that
the average is greater than the value used (0.38 g/cm$^3$) to
calculate the water equivalent, then the true positive
balance is greater than that calculated. At some of the
locations where measurements indicate a loss of material
from the surface during the year, the surface recorded at
the beginning of the year had been ice. No increase in
density could have occurred at these locations, so the loss
must have been genuine. It is concluded that at least the
major irregularities in the profile of the specific balance
are real.

In an attempt to find the cause of this pattern of
specific balance, each year's records have been divided
into two periods, in order to examine the differences in the
patterns of winter and summer accumulation (see Fig. 46).
As expected, accumulation was found to be more in the fall
and winter while ablation was generally greater than
accumulation during the spring and summer. At the edge of
the ice sheet, in the area of annual loss, moderate
accumulation occurs in winter, while in spring and summer
there is a marked excess of ablation. This area of negative
balance, near stakes 402 and 403, is hence due perhaps more
to intense ablation than to lack of accumulation. The high
summer ablation in this area is related to the steeper
surface slope, because the melt water runs off instead of
forming superimposed ice. The somewhat low accumulation is
Fig. 46.—Profile showing the relative contribution of winter and summer accumulation to the specific balance.
related to the fact that snow is easily blown from the generally icy surface in the area.

The pattern of rather large positive balance near S-1 and an area of low or negative balance slightly inland is apparent in both halves of the year, which suggests that this pattern is caused as much by uneven distribution of accumulation as by uneven ablation. In some situations an area of high accumulation may be caused by the orographic effect favoring precipitation at a certain height (Benson, 1962, pp. 35-38; Swithinbank, 1959, p. 135), but the anomalies under consideration here are too local to have been caused in this way. They are perhaps related to slight local changes in wind velocity caused by minor changes in the surface slope of the ice, or to a local decrease in wind velocity at the surface where the katabatic wind layer has lifted away from the surface (Ball, 1960, p. 70).

After several storms which deposited a layer of uncompacted snow rather than the more usual wind-packed snow, soundings were taken to determine the depth of new snow on the hard ice surface. The loose snow was deposited with rather low winds and it was expected that much of the snow would later be removed by higher winds. These measurements do show certain areas of higher-than-average accumulation. If the precipitation is constant over the area or varying only slowly with altitude, the areas of high and low accumulation must be due to wind effects
influencing deposition, or to transport of material after it has first reached the ground. Some of the areas of high accumulation after a given storm coincided with the areas of large annual positive balance, while none of the areas of high accumulation for a given storm fell within an area of low or negative balance. However, the locations of high accumulation vary considerably from storm to storm.

Examination of the accumulation profile (Figs. 14-17), either for snow or for water equivalent, shows that the excess of positive balance in the S-1 area over the normal balance along the trail is approximately equivalent to the deficit in the balance relative to the normal value a short distance inland from S-1. This suggests that there may have been transport of material from the area of low or negative balance to that of abnormally high positive balance, either as a result of wind transport, or the movement of meltwater down the slope a short distance before it refroze. Local topographic irregularities should be expected in either case, causing eddies and variation in wind velocity, or increased slope for water movement.

Furthermore, the irregularities in the bed, determined from the gravity survey, should be reflected by slight surface irregularities (Nye, 1959, pp. 503-505; also Robin, 1967). No such irregularities were detected in the survey along the trail, but there was a subjective impression that the trail
seemed to flatten above S-1 and then to rise more steeply between Fl-2B and BF-2.

Furrow-like depressions tranding down slope were recorded in April of 1958 near the upper profile (north and south of BF-2). They may have been incipient melt-water channels, although no flow was ever observed in them, or they may have been formed by wind action. No such depressions were noticed in the area around S-1, which seemed to be more level. At the same time bare ice areas were observed inland from the upper cross-profile and at the end of the summer there was considerable sheet-flow of water on the south ends of both the upper and lower profiles. Late in the season ice formed from melt water was recorded in the same areas.

**Importance of albedo differences**

Any changes in the nature of the ice sheet surface will generally be accompanied by a change in the albedo of the surface. New snow may reflect up to 95% of incoming short-wave length radiation (Liliquist, 1956, pp. 91-92) while an old ice surface with some dirt present, usually reflects less than 40% and sometimes as little as 20% (Mellor, 1964, p. 93). In addition, the portion of energy that is absorbed is transmitted further in ice than it is in snow. On the other hand both snow and ice act almost as ideal black body radiators with respect to long-wave length
energy. When energy is lost from the surface as long-wave length radiation it is radiated only from the surface few centimeters.

If a local area in the marginal region of an ice sheet has relatively low snow accumulation or has considerable superimposed ice accumulation, the layer of snow over the ice is thinner than usual. Even if the ice is not exposed, the portion of the absorbed energy which reaches the ice is then transmitted to a greater depth than usual. Energy which is transmitted beyond the depth from which long-wave length energy is radiated from the surface, cannot be lost directly by radiation and is available to warm the upper layers of the ice sheet. Where ice is actually exposed at the surface, there is a sharp decrease in albedo (compared with snow), more energy is absorbed and is therefore available for heating, evaporation, and melting. Conditions with ice near, or at the surface are most likely where accumulation is low, and accordingly the areas of low or negative balance may be accentuated due to the greater amount of energy absorbed in such areas.

**Deflation versus melt and evaporation**

In the 1958-1960 period, wind probably was not solely responsible for the anomaly above S-1, in that removal of material at stake 413 has continued beyond the hard layers of the 1958-59 summer and also below the ice layer of the
1957-58 summer. It seems unlikely that mechanical wind erosion alone could do this. Melt or evaporation must be responsible for the loss of what was formerly icy snow or ice. At stake 413 the reduction of the surface down to the level of April 1958 may have been primarily by deflation, but the removal of material below this level must have been by melt or evaporation. However, though the material was probably not removed by direct mechanical action of the wind, the presence of a high wind, without blowing snow, would be favorable to either melt or evaporation since it can maintain steep temperature and vapor-pressure gradients.

The balance profile above the marginal moraine--conclusions

In the ablation area at the margin of the ice sheet, ablation in the form of melting is the most important factor causing the negative balance. High winds also reduce accumulation because snow is not easily retained on the smooth icy surface. The steeper slope is important since without it much of the melt water would be converted to superimposed ice instead of being removed. Irregularities in the values of the specific balance further inland are probably caused by irregularities in the wind velocity which produce local areas of increased erosion or accumulation. Local variations in the wind velocity may well be related to irregularities in the surface slope. Variations in surface slope may also cause migration of material in
the form of melt water. Once an area of low or negative balance has been established, albedo differences may help to maintain the low accumulation. Melt or evaporation is probably important in the removal of firmly consolidated snow or ice.

Mass balance in the islands and on the ramp

The island area itself consists of about 75 km$^2$ of exposed bedrock scattered over a total area of 450 km$^2$ (see Fig. 7, also Map 1). The topography of the rock exposures in the island area is quite rugged on a small scale, with a relief of up to 125 m. Permanent snow drifts, in fact formed largely of ice, are found on the seaward side of many of the rock prominances, the leeward side with respect to katabatic winds. The largest of these long lee drifts rise gradually to a height of 20 m or more (see Fig. 47). Many of the rock exposures are kept snow-free year round by the high winds and by relatively high ablation rates when even a small portion of rock is exposed to solar radiation.

The island area is bounded on the east by the ice sheet which terminates here in an ice ramp. From the eastern-most rock exposures, at an elevation of about 40 m, the ramp rises in about 1 1/2 km to a height of 140 m where the marginal moraine is found. On the coastal side of the more prominent portions of the moraine ridge, large snow
Fig. 47.—Photos showing large drifts in the lee of rock prominances, (a) in the southern portion of the Windmill Islands, and (b) in the lee of an ice-cored moraine ridge near Clarke Island.
drifts have formed. The ramp surface has a generally smooth profile, sometimes slightly convex where large drifts have formed, sometimes slightly concave where melting has been more prevalent. In summer and fall the surface of the ramp is crossed by many stream channels up to 2 m in depth, the development of which varies from year to year depending on the intensity of melting.

Stakes 00 to 06 and stake 430 were located in the island area and on the ramp. All of them were implanted in areas of snow or ice, between rock exposures or on the ramp (see Map 1). Additional measurements of the winter's accumulation were made in November 1958 by sounding the snow depth to the old firn or ice surface (see Hollin et al., 1961, pp. 105-106).

The processes of accumulation and ablation in the island area and on the ramp are the same as on the ice sheet but the variations both locally and annually are much greater. The area is one of heavy melting and great run-off, but also of high accumulation as the drifts behind prominences are replenished by winter accumulation and the summer stream channels are filled in. In the 1958-59 season many of the ramp island stakes showed a net gain but the next year showed a net loss. In general these stakes showed much greater changes than did the ice sheet stakes, and the balance at individual stakes varied greatly, depending on their location.
The large areas of bare rock are indirectly responsible for increased ablation, inasmuch as these areas absorb large amounts of short-wave energy which is reradiated as heat energy. Melt-water pools and even small lakes frequently form in the rock areas and considerable evaporation occurs locally from these water surfaces. Heat from the rock areas is carried by advection to the ramp and marginal snow. The warm air forms thermal winds, particularly in summer, which rise along the ramp and cause increased melting on the ramp and probably even inland from the moraine.

At present the snow and ice areas in the islands seem to be approximately in equilibrium, summer losses being replaced by snow drift formation in the winter. A slight lichen trim line can be seen in some rock areas (Cameron, 1963, pp. 149-150; Cameron et al., 1959, p. 22) but it indicates only a small change in the surface level of the snow drifts. The trim line does not indicate any significant change in the ice sheet thickness.

**Resumé of Maudheim results for comparison**

Maudheim, the base for the Norwegian-British-Swedish Expedition (1950-1952), was located on an ice shelf on the coast of Queen Maud Land approximately at Latitude 71° South, Longitude 11° West. Inland from the ice shelf, the continental ice surface rises over a series of rounded steps.
Such a location is more typical of much of the Antarctic margin than is the Wilkes area. The rate of accumulation varied slightly according to the season (Swithinbank, 1957, p. 64), but with practically no summer ablation, the annual accumulation could be measured over almost any twelve-month period. Only enough melting took place to form some soaked firn and some ice crusts.

The accumulation measurements may be divided into two parts. Stake measurements were made frequently throughout the year on the generally level ice shelf near Maudheim. Also, at four times during a sixteen-month period, stakes were measured along a trail going 300 km inland and ascending to an elevation of 1460 m (Swithinbank, 1959, p. 125). On the ice shelf, wind usually had little effect on the accumulation, and on the average, equal amounts of snow were deposited and removed by the wind. However, even the slight topographic irregularities present caused some differences in the accumulation (Swithinbank, 1959, pp. 54-55). The measurements along the trail inland show the effect of the katabatic winds in relation to the topography. On the slopes accumulation was lower than normal but the snow not deposited, or removed from there, appeared to have been dumped at the base of the slope (Swithinbank, 1959, p. 135). In addition a precipitation shadow effect was recognized on the inland side of some of the hills.
CONSIDERATION OF TEMPERATURE MEASUREMENTS

Possible short term temperature variation

Evidence from pits and cores. The consideration of the formation of the marginal moraine indicates that the margin of the ice sheet has remained in approximately its present location for a long time (p. 136). Accordingly no major change in equilibrium-line position is likely to have occurred. However, temperature and accumulation may vary enough to show short-term changes in the surface position of the boundary between different accumulation facies. The equilibrium line at Wilkes does not shift readily because its position is controlled largely by the sharp increase in ablation at the edge of the glacier, where the slope becomes steep enough to carry away melt water.

To investigate the lateral shifting of a facies boundary, one can investigate the vertical record in the vicinity of the facies change. The stratigraphy from the walls of a few pits and from cores in the vicinity of BF-2, near the 1957-58 summer boundary between superimposed ice and firn, suggests a recent shift in the position of the facies boundary. The measurements were made in July and August of 1958, when the surface was new snow (see Fig. 48, 182
**Fig. 48.--Stratigraphy of the pit and core at BF-2**

(from Hollin et al., 1961, p. 213.)
also see Hollin et al., 1958, pp. 213-214). Below the 1957-58 summer surface (actually April 1958) the pits show alternating layers of soaked firn or depth hoar crystals, and ice. Not counting the new surface snow, this banded zone is roughly 50 cm in thickness. Below this level almost continuous ice is recorded with only occasional bands of firn down to the total depth of the core at 10 m. The top of the continuous ice is probably 2 yrs old or more.

The few bands of firn that are preserved represent years of unusually low temperature or high accumulation. The fact that occasional firn layers are preserved shows that the intervening ice is superimposed ice and not ice formed by compaction of firn.

The much greater proportion of firn layers in the uppermost meter suggests that there has been a shift in the facies boundary and at the present time (1960) only part of the snow cover in the BF-2 area is converted to superimposed ice, whereas formerly all accumulation in this area was changed to superimposed ice. It could be argued that melt in the next season might saturate lower firn layers to form superimposed ice, but this is unlikely since the lower firn layers are separated by apparently continuous ice layers several centimeters thick. The same sort of firn and ice layering in the uppermost meter was observed in February 1958, particularly in the area of the upper profile, when one-inch holes were drilled to place new accumulation stakes.
It should be remembered that firn remained at the surface in the upper profile and BF-2 region throughout the 1957-58 summer, a summer which was rather warm with pronounced melt, compared to the next year. From the pit profiles, however, it appears that formerly the summers were probably even warmer.

The change in facies at BF-2 implies a change either in temperature, accumulation, or slope, or a combination of these. The possibility of a change in slope can be eliminated at once, since forward movement is so slow here, probably only a few meters per year, that the position of this element of ice could have changed little in relation to the surface profile of the ice sheet margin. An increase in accumulation could accomplish the change, but if the interpretation of pits farther inland is correct, particularly the carefully-studied pit at S-2 (see Cameron et al., 1959, p. 48), no noticeable change has occurred in the annual accumulation during the last few years. (For the past 20 years the 5-year means of accumulation have been 13 ± 0.5 cm of water.) The most probable reason for the change is, therefore, a recent lowering of the temperature.

Temperature variation as shown by direct measurement. To show more realistically the variation in air temperature, 12-month running means have been calculated from the monthly air temperature averages published by the U.S. Weather Bureau. In effect these are the successive annual averages,
using all possible 12-month intervals. Over the period studied here, 1957-59, there is a decrease in the annual temperature average of 3°C from the warmest to the coldest 12-month period, followed by a partial warming up in 1959 (see Fig. 49). Fluctuations of the same magnitude continue through 1965 with maximum and minimum 12-month averages of -7.7°C and -11.3°C.

**Evidence from the 10-meter ice temperature curve**

In a location where the air temperature does not greatly exceed 0°C at any time, and where surface melt does not transfer heat energy to lower layers by percolation, the temperature at a given depth should be a function only of time, the air temperature, and the thermal diffusivity of the material. The amplitude of the annual temperature range decreases exponentially with depth, and at a depth of 10 m in firn or ice does not normally exceed 1 or 2°C, the mean temperature being the same as the mean annual air temperature. The data from Byrd and Pole stations show that in a firn region agreement is fairly close between 10 m and deeper ice temperatures and annual air temperatures. Exact agreement with the average air temperature of a particular year cannot be expected since the 16 m temperature is an average over several years, while the average air temperature in successive 12-month intervals can vary appreciably.
Fig. 49.—Twelve-month running average of Wilkes Station temperatures from 1957 to 1965.
Fig. 49. (Contd.)
During the 1957-58 period a series of holes was drilled in conjunction with pit studies along the S-2 trail, to study the snow stratigraphy and to obtain ice temperatures at a depth of 10 m or more. The holes were spaced at 10-mile intervals between mile 10 and S-2 and were more closely spaced nearer the coast.

In addition, a more detailed set of observations was made at S-1. In 1957 thermohms were placed at the surface and at depths of 1/2, 2, 4, 7, and 16 m and readings were taken almost every week thereafter. The 11-m temperature was not constant, showing fluctuations of close to a degree, but the 16-m temperature showed very little change over a 2-yr period, and the apparent fluctuations are believed to be primarily due to instrumental errors. The 10- and 11-m temperatures at locations other than S-1 have been corrected to the equivalent 16-m temperature using the 11- and 16-m data from S-1. To obtain the correction a graph was made, using the data over the 2-yr period, 1957-59, showing the difference between the 11-m temperature and the 16-m temperature, with the 16-m temperature assumed constant (see Fig. 50). Corrections for temperatures at other locations were determined by selecting on the graph the difference between the 11- and 16-m temperatures for the date when the temperature was taken. Small errors undoubtedly occur since the S-1 area is ice, while the inland sites are in firn and the difference in thermal properties may slightly
Fig. 50.—Graph showing differences between the 11-meter and 16-meter temperatures as measured at S-1 from January 1957 to January 1959.
alter the form of the temperature curve. However, since the curve is nearly linear in the 11- to 16-m region, such errors should be small. The difference between an 11-m and a 10-m correction is also believed to be too small to make an interpolation necessary.

Cameron and Bull (1962) have used a somewhat more sophisticated method to correct the S-1 temperatures. Having applied a correction determined from the apparent temperature obtained with a calibration resistance, the 16-m temperatures were then fitted to a sinusoidal curve. The amplitude of the temperature fluctuations about the mean value (assumed to be constant in the present work) is less than the probable error of 0.2°C given by Cameron and Bull (1962, p. 179, also Cameron, 1963, p. 97). Accordingly the error due to the assumption of a constant 16-m temperature should be minor.

The 10- or 11-m ice temperature for each of the locations between S-2 and the coast, corrected assuming a constant temperature at 16 m, has been plotted against the altitude (see Fig. 51). This curve gives a straight line relationship for the inland locations indicating a lapse rate of slightly more than 1°C per 100 m (dry air lapse rate is 0.977°C per 100 m, Humphreys, 1940, p. 30). Nearer the coast, a pronounced warming starts at about mile 10. This departure from a linear relationship is most pronounced at S-1 (about 2.5°C difference) and then decreases at M1 near
Fig. 51.—Graph showing 10-meter ice temperatures along the S-2 trail as a function of elevation.
the moraine, and on the Grinell Glacier near the shore.

Non linearity of the 10-meter ice temperature-versus-altitude curve. The higher 10-m ice temperature at S-1 shows at once that a relatively large quantity of heat energy has been transmitted to the ice sheet in this area. Either the mean annual air temperature is considerably higher than the temperature predicted by the inland lapse rate, or there is an excess of heat supplied to the surface over what is lost.

The mean air temperature is almost certainly not so high as would be indicated by the 10-m temperature, but it may well be slightly higher than the temperature predicted by the lapse rate. On the coast the air is warmed by contact with the water and also by long-wave length radiation from the low albedo rock areas. The warm air rises along the ramp, particularly in summer and the effect of the warm air, judging by the degree of melting observed, may be felt as far inland as S-1 or even mile 10.

The high ice temperatures on the curve appear to be caused by an excess of heat supplied to the surface. The major processes of heat transfer at the ice surface are turbulent heat transfer, which may either add or remove heat depending on the temperature gradient between the air and the ice surface, short-wave length radiation and condensation, both of which supply heat to the surface, and
long-wave length radiation and evaporation, which remove heat from the surface.

The effect of turbulent heat transfer is to prevent the ice surface and air temperature from differing greatly, the amount of energy transferred depending on the wind speed and on the temperature gradient above the surface. Turbulent heat transfer is particularly important when the air temperature is above 0°C, because the ice surface cannot exceed 0°C and the result is a very steep thermal gradient above the surface. Evaporation and condensation remove or add heat only at the very surface, where their effects can be compensated for by an increase in the turbulent heat transfer. Heat loss due to long-wave length radiation is also only from the upper few centimeters.

The distribution of energy added by short-wave length radiation varies dependent on the material. In snow, particularly new snow, the energy is concentrated in the surface few centimeters (Gerdel et al., 1954, p. 4), while in ice it is distributed over some tens of centimeters. Furthermore, due to the albedo difference between ice and snow, several times as much energy is supplied to ice as is supplied to snow, given the same amount of radiation incident at the surface. Of the processes mentioned above, only short-wave length radiation on ice can supply energy at a deep enough level to effect the 10-m temperature.
In an area where appreciable melting occurs, the movement of melt water alters the heat distribution. Heat energy, transported in the form of latent heat, can be moved from the surface by the downward percolating melt water to a depth where it can effect the 10-m ice temperature. Cameron, in his discussion of the 10-m temperature-versus-altitude curve, attributes the variation from a linear relationship primarily to the movement of heat energy by melt water (Cameron, 1963, pp. 80-88). (In determining the temperature curve, Cameron has used the 10-m temperatures as measured, without correction, but derives the same lapse rate.)

Though it is important at S-1, the present author would put less emphasis on the role of melt water transport of heat. Ice is exposed at the surface part of the time, at other times the snow cover is relatively thin, and therefore melt water can transport heat only to the level of the ice surface. Under such conditions of exposed or only thinly-covered ice, energy from short-wave length radiation can be transmitted to depths beyond the effects of radiation cooling, and should be effective in raising the 10-m ice temperature. If ice is exposed only in the warm summer months and covered by snow during the winter, a relatively large amount of heat will be supplied to the low-albedo ice in summer, but an insulating snow cover in winter will reduce heat losses. The higher conductivity of
ice compared to snow favors downward heat transport rather than transport toward the surface.

In the lower altitude portion of the temperature-versus-altitude curve, stations M3 and Grinell Glacier, the 10-m ice temperature is lower than that at S-1. At these stations, located near the edge of the ice, the surface slope is comparatively steep and melt and run-off are important. Much of the energy supplied to the surface is used for melting snow and ice, and the melt water formed is removed as run-off, all the energy used for melting being removed with it.

Another factor, which may increase the loss of energy through run-off, is related to the difference between clean and dirty ice. (By dirty ice is meant ice with disseminated fine particles of morainal material, probably in the silt size, but without a covering of debris on the surface.) On the Penny Ice Cap, in Baffin Land, temperatures at 60 cm depth in clean ice were higher than those at the same depth in dirty ice (Ward, 1952, p. 12). At Wilkes it was noted that stakes implanted in dirty ice remained frozen firm, while stakes in a similar location but in clean ice melted loose. In clean ice melting takes place along crystal boundaries, loosening the crystals to a depth of 10 cm or more. With the low albedo of ice, considerable energy is absorbed and melt can occur down to the depth of penetration. This ice has an opaque appearance.
In dirty ice the radiation that penetrates the surface is absorbed by the dirt particles near the surface and is used for melting in the surface layers. Melting is controlled by the location of dirt particles rather than being along crystal boundaries, and the ice is comparatively transparent. In dirty ice, where all the absorbed radiation energy is concentrated in a narrow layer near the surface, considerable melting occurs, and because in the Wilkes Station area the dirty ice is found in the steeper marginal area, the melt is removed. With dirty ice a larger proportion of the absorbed radiation is removed in run-off than with clean ice, and where in addition there is a steep surface slope, the 10-m temperature should be lower than in an equivalent area with clean ice. This accounts for the lower 10-m temperatures in the area of stake 402 and on the ramp.

**Equilibrium line location from the 10-meter temperatures**

As has been shown above, the decrease in the 10-m temperature is caused by the removal in the run-off of a portion of the radiation energy that has been absorbed by the ice surface. On the temperature-versus-altitude curve, the point where the 10-m temperature begins to decrease, should mark the location where run-off becomes a significant item in the mass balance. The ablation zone along the margin of the ice sheet is limited to the area where run-off is
important, other forms of ablation being of only minor
importance, and therefore the location where the 10-m
temperatures begin to decrease should indicate the position
of the equilibrium line. The Wilkes temperature data
indicate an equilibrium-line position between S-l and M3
(M3 is near stake 402) which agrees with the equilibrium-
line position from stake data.

Temperatures from depths of 10-m or greater were not
taken at enough stations near the coast to determine whether
S-l is the location with the highest 10-m temperature, so
the exact position where temperatures begin to decrease
cannot be determined. It would be interesting to make a
series of more closely spaced 10-m temperature measurements
between BF-2 and M3 to determine more closely the shape of
the temperature-versus-altitude curve in this region.
Perhaps this is a feasible technique for determining the
equilibrium-line position at any time of year in a super-
imposed ice area, without the need for stake measurements
over a period of a year or more.

Wilkes Station temperature in relation
to the temperature-versus-altitude curve

The linear portion of the temperature curve (Fig. 51)
when extrapolated to 10-m elevation, that of Wilkes Station,
gives a temperature of about -7 1/2°C. In comparison, the
12-month running means of temperatures at Wilkes Station
range from -7.7°C to -11°C. The average temperature over the
whole period is \(-9.9^\circ C\), about \(2.5^\circ C\) colder than the temperature predicted by the lapse rate. This negative temperature difference is surprising in view of the proximity of warm ocean waters and large areas of low albedo rocks, both of which should have a warming influence on the air temperatures.

When the positions of the 12-month mean temperatures are plotted on the lapse-rate curve, the lowest temperature of \(-11^\circ C\) indicates an altitude of more than 350 m, while even the highest temperatures of \(-7.7^\circ C\) is not as high as the station temperature predicted by the lapse rate. Were the equilibrium-line position controlled only by temperature, and the temperature varied only according to the lapse rate, it is evident that the accumulation zone would extend to sea level. Again this demonstrates the importance of surface slope in controlling the position of the equilibrium line. Although the Wilkes annual temperature is generally lower than that predicted by the lapse rate, a relatively small number of summer days with temperatures above freezing, added to the local effects of radiation from rock area, appears to be sufficient to keep the area ice free.

The discovery that the average annual temperature at Wilkes Station is about \(2.5^\circ C\) below that predicted from the lapse rate, together with the pit data from the BF-2 area, is suggestive of a slight very recent cooling in the area,
though the evidence is certainly not conclusive. At any rate, it appears that the equilibrium-line position is rather insensitive to changes in annual temperature, though the nature of accumulation just above the equilibrium line may change between firn and superimposed ice.
CONCLUSIONS

Ice drainage in the Wilkes area

From the study of conditions in the area near Wilkes Station some general conclusions can be drawn about the conditions leading to the formation and maintenance of ice-free areas, and also about the conditions necessary for the formation of Thule-type ice-cored moraines. Of greatest importance is the topography of the sub-ice surface. For 35 miles or more inland from the Windmill Islands, the bedrock elevation, as determined by gravity measurements, is above sea-level, so the ice is comparatively thin and consequently ice movement is rather slow.

The major drainage of ice is either to the north or to the south of the area near Wilkes. The Vanderford Glacier to the south is the nearest and most important drainage channel in the general area. The Australian seismic measurements (see Fig. 52) show a sub-ice valley inland from S-2, probably an extension of the Vanderford system, and the 1958 traverse in the area inland from the Vanderford Glacier, crossed a surface depression with crevasses, indicating comparatively rapid movement. The segment of the ice sheet immediately inland from the Windmill Islands is relatively slow moving, particularly
Fig. 52.—Ice thicknesses and bedrock elevations along a traverse 300 miles to the south from S-2 (after Jewel, 1962, Plate 3).
when compared to the adjacent main stream of the Vanderford Glacier. On a smaller scale a similar situation exists on the inland side of each moraine arc (see Fig. 53). A bedrock high beneath a moraine arc causes a decrease in ice thickness which in turn leads to decreased ice movement, so ice movement in the central portion of a moraine arc is extremely slow.

From calculation of ice thickness along the S-2 trail, combined with surface slope measurements, the value of $\tau$, the shear stress at the bed, has been calculated ($\tau = \rho gh \sin \alpha$). Over the distance 30 miles inland from the moraine, the thickness remains well below the equilibrium thickness (see p. 13). $\tau$ varies between 0.5 and 0.7 bars, well below that portion of the stress-versus-strain rate curve (Figs. 4, also p. 12) where a slight increase in thickness due to an increase in accumulation will cause an extremely large increase in strain rate. In the marginal area with low basal shear stress, a slight thickening would cause only a small increase in the strain rate. Further inland where the shear stress is higher, drainage is to either side of the island area. Here an equivalent increase in accumulation would cause a comparatively large increase in strain rate. Accordingly, for an increase in accumulation, the marginal areas can thicken somewhat with only a small effect on the ice margin, while most of the increased accumulation is removed to either side.
Fig. 53.—Photo of bedrock high located in front of a moraine arc to the north of Clarke Island (U.S. Navy photo).
On the other hand, it is interesting to note that Nye's method of calculation (1959) gives reasonable results in the calculation of the velocity-versus-depth profile at stake 402 even though a low stress portion of the Glen flow law is used. The calculated difference between the surface velocity and that at the glacier bed is practically equal to the surface velocity as measured by triangulation to bed-rock stations on Bailey and Clarke Peninsulas. This indicates, as was expected, that little or no bottom sliding occurs at stake 402, where the basal ice temperature is less than 0°C.

From the velocity-versus-depth profile, the volume of material flowing past the equilibrium line can be calculated. The equivalent volume of accumulation can be furnished by a rather small surface area, again emphasizing that the ice drainage is primarily to one side or the other, rather than through the region near the marginal moraines.

In addition to the peculiar ice thickness conditions required for the existence of an exposed rock area, the formation of a marginal moraine requires special conditions of temperature at the base of the ice. Inland, the basal ice temperature is at the pressure melting point, and sub-glacial melting undoubtedly occurs, while at the ice margin the glacier is frozen to its bed with basal ice temperatures below 0°C. Just inland from the frozen margin,
material for the moraine is frozen into the basal ice along with basal melt water from further inland.

Another requirement for the formation of a marginal moraine is the maintenance of the position of the ice margin with very little change over a long period of time. The length of time to bring material from the glacier bed to the surface is on the order of 6000 years, and for this period the position of the ice margin should be almost static, in order to initiate a marginal moraine.

Once a rock area has been uncovered, and a marginal moraine has been formed, several factors operate which tend to keep the position of the ice margin static. Accumulation probably may vary considerably from year to year, since precipitation is related to the passage of low pressure systems, and is then subject to redistribution by wind, sometimes with considerable unevenness related to the topography. However, if an appreciable increase in accumulation were to occur, the direction of the major drainage would probably continue to be to the sides of the ice free area. Because the major drainage to the sides takes place in the accumulation area, the velocity must increase in the direction of the flow, which implies an increase in the basal shear stress as well. It is in the areas where the shear stress is already highest, that an increase in ice thickness, due to higher accumulation, will cause the greatest increase in movement. The major characteristics
of the present drainage pattern therefore probably would be little changed by an increase in accumulation.

The equilibrium-line position is controlled primarily by the increase in slope near the ice margin, and in addition sufficient heat to cause appreciable ablation is available only near the ice margin. Reradiation from exposed rock areas warms the air, producing winds which ascend the ramp, but this influence is probably felt at most only a few miles inland of the moraine. Extreme melt and run-off is also favored by the presence of disseminated dirt in the ice, and this condition exists only near the moraine.

On the other hand, once a moraine has developed to the extent where it has topographic form, snow carried by the wind accumulates downglacier from it, in the lee with respect to katabatic winds. In spite of severe melting on the ramp in any particular melt season, during the following accumulation season drifts build up again to the height of the moraine, thus maintaining the ramp. The height of the moraine itself varies little.

Since it seems that similar conditions must exist for the formation of any marginal moraine similar to that at Wilkes, it is probably safe to assume that the area inland from such a moraine is relatively inactive and should be free of crevasses. Accordingly, moraines of this type
indicate locations which should make safe access routes to the inland ice.

The Wilkes area is interesting also because of the importance of superimposed ice accumulation. Sufficient heat is available to melt a part or all of the annual snow cover for a moderate distance inland from the moraine, but with cold ice below, the melt water is refrozen with only slight lateral displacement.

Though sufficient heat is available for melting, the ablation area does not extend inland since the melt water is not removed. Because the equilibrium line is in the superimposed ice zone, its position is hard to determine, but an estimate of its position can be made by an examination of the curve of the 10-m ice temperature-versus-altitude curve. The equilibrium line is located approximately where the 10-m ice temperature values begin to decrease with decreasing elevation of the drill sites. The reason for this is that ablation is primarily due to run-off, and heat energy absorbed by the surface is then removed in the melt water.

**Relationship of exposed rock areas to the extent of ice in surrounding areas**

In general, the position of the margin of the Antarctic ice sheet is controlled by sea level (see Fig. 54, also Hollin, 1962). Ordinarily the ice sheet terminates at
Fig. 54.—The effect of sea level changes on the position of the Antarctic ice margin (from Hollin, 1962, p. 187).

Fig. 55.—Generalized curve showing sea level changes from 10,000 years BP to the present (after Fairbridge, 1961, p. 99).
the place where the water is deep enough to float the ice. Shelves exist only where there are islands or shoals which keep the floating ice from breaking up and drifting away. With the marginal position thus determined around most of the Antarctic continent, the surface shape, and therefore thickness of ice, is then controlled primarily by the flow law. The general slope of the surface is altered locally by major drainage channels, such as the Vanderford Glacier, which are related to irregularities in the sub-ice topography.

The surface configuration in the area of the Windmill Islands, controlled largely by the position of the ice margin in surrounding areas, is such that the ice, lying on the relatively high sub-ice topography, is comparatively thin, and so is relatively inactive. Drainage to the sides is so effective that the ice thickness inland from the Windmill Islands never has the opportunity to increase appreciably. The equilibrium extent of the ice sheet, controlled by conditions in surrounding areas, has left the rock area of the Windmill Islands exposed, and the existence of a land terminus, linked with the proper temperature conditions inland from the area, has been responsible for the marginal moraine formation. Inasmuch as the extent of the ice sheet in the surrounding areas is determined by sea level, it is likely that the present conditions in the vicinity of Wilkes Station were
initiated by the rise of sea level at the close of the Pleistocene.

Dating from moraines in the ice free areas around McMurdo Sound (Pévé, 1962) and from a raised beach in the Windmill Islands (Cameron et al., 1959) gives a minimum age of 6000 yrs ago for the date of ice retreat (see p. 136). Russians (Rozychi, 1960) suggests that Bunger Hills has been ice free for "several thousand years" and that ice probably vacated the area at the beginning of the post-glacial climatic optimum. Flint (1957, p. 377) dates the warm hypsithermal time as between 7,500 and 4000 years ago. A date between 8000 and 10,000 years ago for the major rise in sea level (see Fig. 55) is in fair agreement with a period on the order of 6000 years for moraine initiation, with a few thousand more years allowed for further moraine development.
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MAP 1.

WILKES STATION
GLACIOLOGICAL DATA 1958-1959
WINDMILL ISLANDS NORTHERN AREA

KEY
--- OUTLINE UNCERTAIN

LAKE

Δ POSITION FIXED BY TRIANGULATION
Θ POSITION FIXED BY SUBTENSE SURVEY
* POSITION FIXED BY OTHER MEANS
R ROCK
A MORAIN SECTION

SCALE

YARDS 400 0 1 2 3 4 MILE

PROJECTION: MERCATOR
COMPILED BY: C. CRONK AND J. HOLLIN 1960
DRAFTED BY: G. VAN NIEL - INSTITUTE OF POLAR STUDIES
THE OHIO STATE UNIVERSITY

INSET: TRAIL GAPS AREA
MAP INCLUDES MOVEMENT SURVEY POINTS
(SEE MOVEMENT SURVEY F)

SCALE

0 100 200 300 400 YARDS

LIGHT BLUE ICE (SEE TEXT)
INSET: TRAIL GAPS AREA
MAP INCLUDES MOVEMENT SURVEY POINTS
(SEE MOVEMENT SURVEY F)
SCALE
0  100  200  300  400 YARDS

SCALE
0  100  200  300  400 YARDS

ISOLATED MORAINES

3 SHEAR PLANES

GREY ICE

DARK BLUE ICE

LIGHT BLUE ICE (SEE TEXT)

B13 . 403

H) LOWER BASE MORaine

C) LOWER BASE

D) LOWER BASE

C) LOWER BASE

NORTH GAP

SOUTH GAP

TRENCH

UPPER BASE

GREY ICE

DARK BLUE ICE

LIGHT BLUE ICE (SEE TEXT)

LINEATIONS

C) GREY

DARK BLUE ICE

LIGHT BLUE ICE

SCALE
0  100  200  300  400 YARDS

SCALE
0  100  200  300  400 YARDS