THE SHERMAN GLACIER ROCK AVALANCHE OF 1964: ITS EMLACEMENT
AND SUBSEQUENT EFFECTS ON THE GLACIER BENEATH IT

DISSERTATION

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The Sherman Glacier rock avalanche in the Chugach Mountains of south-central Alaska fell during the great Alaska earthquake of 1964. The $10.1 \times 10^6 \text{ m}^3$ of jointed and fractured rock and about $2 \times 10^6 \text{ m}^3$ of snow and ice may have been partially fluidized in the shaking that first dislodged snow and a small cirque glacier before a flood of rock rubble slid down all sides of a peak now called Shattered Peak.

The avalanche lost half of its available energy in the first 30 sec of the 3.5 min avalanche, largely through sliding with high friction ($\mu = 0.60$) down the surface of rupture. By this time, it was mechanically fluidized by the high kinetic energy of its clasts, and flowed in supercritical laminar flow as a complex Bingham plastic with a yield stress of 0.02 bar and with such high viscosities ($1 \times 10^6 \text{ poise for } 0.02 < \tau < 0.1 \text{ bar and } 4 \times 10^7 \text{ poise for } \tau \geq 0.1 \text{ bar}$) that it behaved largely as a thin flexible sheet, with deformation principally confined to shear at the base. It thus, essentially slid across a snow-covered, gently sloping surface with low friction ($\mu = 0.11$) to cover 8.25 km$^2$ of the ablation area of Sherman Glacier with an insulating blanket of...
rock debris that averages 1.65 m in thickness.

Melting of ice in the area now covered by debris has been reduced to about 20 per cent of its pre-avalanche value, that ranged between -5 Mg m$^{-2}$a$^{-1}$ at about 420 m elevation, to about -8 Mg m$^{-2}$a$^{-1}$ at about 200 m. This has increased the net mass balance of the glacier from very negative to very slightly positive, largely as a result of two years of heavy winter snowfall (1964-65 and 1969-70) and of the mass of the avalanche with the 1963-64 winter snow preserved beneath it.

Measured stratigraphic net balance at a point at about 430 m elevation on the glacier correlates almost perfectly with a function of selected climatological parameters (winter snowfall and temperature, and summer temperature) measured at nearby Cordova F.A.A. for the period from 1964 to 1971; this enables a 30-year record of net mass balance to be calculated. Correlation of a smoothed climatic record from Cordova F.A.A. with a smoothed record from Sitka Magnetic (at Sitka Alaska, 430 km southeast of Sherman Glacier) provides a 100-year record of net balance for glaciers in this region of coastal Alaska that is in excellent agreement with the known history of behavior of Sherman Glacier over the last three-quarters of a century.

A study of flow of the glacier from 1964 to 1971 suggests that over this period, the glacier has responded only to changes in climate during the previous twenty years, and apparently has not yet responded to any effect of the earthquake, or of the rock avalanche. The debris apparently has slowed the rate of thinning of the debris-covered portion of the glacier, so that the effect of the avalanche to 1971 may have been only to slow or prevent changes that otherwise would have
occurred.

A review of the effect of debris on other glaciers shows that the glacier eventually will advance as a result of the avalanche. The predicted advance of about 630 m may not have begun until 1972 and should continue for about 50 years at a rate that will not exceed 20 m a⁻¹.
Nothing can be more certain than the fact, so well stated by Charpentier in his 10th section, that the glacier does not owe its increase to the snow of avalanches, nor indeed to any snow which falls on the greater part of its surface.

J.D. Forbes

4th letter on glaciers
Edinburgh New Philosophical Journal
January 1843
PREFACE

After the great Alaska earthquake of 27 March 1964, President Johnson wrote: "It is important we learn as many lessons as possible from the disastrous Alaskan earthquake." More than 80 major rock avalanches fell during the earthquake, and most fell on glaciers, presenting and excellent opportunity to learn of rock avalanches, their mode of transport, and their effects on glaciers.

In the years after the earthquake, many studies of them have been made, and many lessons learned. My studies at Sherman Glacier, in the Chugach Mountains of south-central Alaska, were not initially directed towards determining effects of the earthquake, although the continued monitoring of the glacier's response was a part of the work that I undertook. In 1969, I initiated a program of surface strain measurement to determine the origin of a train of wave ogives at the base of a major icefall on Sherman Glacier. The program was ended abruptly in 1970 by a winter of record snowfall that left the ogive system still beneath more than 10 m of snow in late summer. My major efforts in that summer were redirected towards methods of evaluating mass balance of a glacier in which accumulation exceeded a measurable depth for available tools, and ablation in some areas was hidden from view beneath debris, and in others exceeded the limits of measurement (-11 m).
This redirection of effort has led me to examine most closely the studies made by others of effects of debris covers on glaciers, and in turn, to examine the mechanism of emplacement of the debris cover on Sherman Glacier. In many areas of these examinations, I have found deficiencies in the earlier studies. Some of them I have overcome with my own observations, but many became apparent only after my field study had ended, and I have attempted to overcome them by reinterpreting the data of others, or by developing methods of estimation that make use of the limited available data.

Beyond the many lessons learned of rock avalanches and their effects on glaciers, are the more important lessons taught by the failures of others: that each problem should be approached with an open and enquiring mind that seeks, not to justify an existing model, but to improve the model, because each new event has new evidence to offer that leads to a better explanation of the processes involved; that few events are unique so that their temporal behavior can not be learned from the behavior of similar earlier events; and I have learned not to expect processes to perform miracles, nor to invoke processes that involve consequences inconsistent with the product.

In all my work I have attempted to develop a product that I believe will stand as close an examination as I have given to the work of others, but, as always, my facts are no better than Forbes' facts: they are consistent with my experience, and correct to the best of my knowledge.
ACKNOWLEDGMENTS

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I am especially indebted to Dr. Čedomir Marangunić for making available his field data for the first four years of this study, and to Austin S. Post, U.S. Geological Survey, who provided an extensive collection of aerial photographs of Sherman Glacier.
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Studies in Glacial Geology. Professors Richard P. Goldthwait and Valter Schytt

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Studies in Computer applications to Geology. Professor Charles E. Corbató and Dr. Peter J. Morgan

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xxiii
INTRODUCTION

The Problem

A major earthquake in Alaska in 1964 triggered innumerable landslides. Many fell on glaciers, and several, of the rock-avalanche type, covered such large areas that they might have a significant effect on the regimes of the glaciers beneath them. One such landslide was the Sherman Glacier rock avalanche (Fig. 1) that covered 8.25 km² of the ablation zone of Sherman Glacier, south-central Alaska (Fig. 2).

Effects of this rock avalanche on mass balance and flow of the glacier were determined during field seasons from 1965 to 1971 by studies of accumulation, ablation, and surface motion.

Some basic questions that arose at the time of the earthquake were:
1) what is the mechanism of transport of large rock avalanches;
2) what are the immediate effects of rock avalanches on the ice beneath them; 3) what is the long-term response of a glacier to a thick debris cover; 4) does this type of event have significance in the geologic record; and 5) does this type of event pose a significant environmental hazard?

Answers to these questions have been sought by a number of people working on several rock avalanche deposits and glaciers since 1964. Work by R. L. Shreve (1966, 1968a, 1968b), and by Ć. Marangunić (1968, 1972) have provided several answers to the question of avalanche trans-
Fig. 1. Sherman Glacier: before and after the 1964 earthquake. (1963 and 1964 photography by Austin S. Post)
Fig. 2. Map of Sherman Glacier showing approximate positions of movement markers and survey stations.
port mechanism. Investigations by Č. Marangunić (1968, 1972), and C. Bull (1969) (see also Bull and Marangunić, 1967 and 1968, and Marangunić and Bull, 1966 and 1968) on Sherman Glacier, and by J. Reid (1969) on Slide Glacier, have provided some answers to other questions, but some were left unanswered in these studies, and many of their conclusions are no longer consistent with available data.

With additional information gathered since the work of Marangunic, I have been able to substantially revise the interpretation of the mode of emplacement of the rock avalanche and of the subsequent effects of the debris cover upon Sherman Glacier. My studies have largely been confined to determining the long-term responses of a glacier to emplacement of a thick debris cover upon its surface. I have made a detailed analysis of factors influencing ablation of ice beneath debris covers to determine the effect of the avalanche cover on mass balance of Sherman Glacier. I have also developed a long-term record of mass balance for the glacier from some direct measurements, from a 30-year record of weather at nearby Cordova Airport, and from a 100-year record of weather at Sitka, Alaska, some 430 km away. These studies reveal the contribution of the rock avalanche to the varying mass balance of Sherman Glacier and permit a more precise analysis of the response of the glacier to a rock avalanche upon its ablation zone.

The subjects dealt with in this work fall into four categories: 1) emplacement of the avalanche; 2) the effect of this on the mass balance of the glacier beneath it; 3) the mass balance of the glacier; and 4) its response to these changes in balance. These subjects are
treated in the four chapters:

Chapter I  The Sherman Glacier rock avalanche
Chapter II Effects of a debris cover upon ablation at Sherman Glacier
Chapter III Mass balance at Sherman Glacier
Chapter IV Response of Sherman Glacier to change in mass balance.
CHAPTER I
THE SHERMAN GLACIER ROCK AVALANCHE

This chapter constitutes the text of a manuscript submitted as Chapter 14: The mechanics of rock avalanches: as determined from the Sherman Glacier rock avalanche of 1964, and other earthquake-induced rock avalanches on Alaskan glaciers: in Voight, B., ed, Geology and mechanics of rockslides and avalanches: to be published as Memior 143 of the Geological Society of America. For this reason, its format differs slightly from subsequent chapters, in that its internal cross-referencing is by page number, and not by section number, as in later chapters.
Introduction

South central Alaska lies in one of the world's more active seismic regions, with a long history of earthquakes of large magnitude. Since 1898, 24 major earthquakes of magnitude 7.0 or greater on the Richter scale have occurred (Table 1). Tremendous tectonic distortion, uplift, and faulting have left the rock of the Chugach and Kenai Mountains almost as a Chinese block puzzle. As is common in youthful mountainous terrain exposed to intense glacial erosion, oversteepened mountainslopes abound. It is therefore not surprising that, in this region, many large landslides have been triggered by earthquakes. In just 5000 km$^2$ of land area around the epicenter of the Great Alaska earthquake of 1964, Hackman (1965, 1968, p. 44-45) identified 2,036 slides that included 78 involving rock debris. Landslides and avalanches were, however, not merely localized around the epicenter; uncounted thousands occurred over a very broad area of about 130,000 km$^2$.

More than 80 major rock avalanches fell during the 1964 earthquake, presenting an excellent opportunity to learn of rock avalanches and their mode of transport. Many studies of these avalanches (Field, 1968; Reid, 1969; Ragle and others, 1965; and Bull and Marangunić, 1968) have dealt primarily with effects on glaciers beneath the avalanche deposits, but this is of little concern here. Several studies of morphology and mode of emplacement of some of the larger rock avalanches were made (Shreve, 1966 and 1968a; Marangunić and
Table 1. Major Alaskan earthquakes within 160 km of glacierized areas

<table>
<thead>
<tr>
<th>Date</th>
<th>Richter Magnitude</th>
<th>Lat. (°N)</th>
<th>Long. (°W)</th>
<th>General Location</th>
<th>Nearest glacierized mountains</th>
</tr>
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<td>8.2-8.3</td>
<td>60°</td>
<td>142°</td>
<td>Icy Bay</td>
<td>Chugach, St. Elias</td>
</tr>
<tr>
<td>10 September 1899</td>
<td>7.8</td>
<td>60°</td>
<td>140°</td>
<td>Yakutat Bay</td>
<td>Chugach, St. Elias</td>
</tr>
<tr>
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<td>8.5-8.6</td>
<td>60°</td>
<td>140°</td>
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</tr>
<tr>
<td>27 August 1904</td>
<td>7.8-8.3</td>
<td>64°</td>
<td>131°</td>
<td>Mt. McKinley</td>
<td>Alaska Range</td>
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<tr>
<td>15 May 1908</td>
<td>7.0</td>
<td>59°</td>
<td>141°</td>
<td>Yakutat Bay</td>
<td>Chugach, St. Elias</td>
</tr>
<tr>
<td>19 September 1909</td>
<td>7.4</td>
<td>Seward area</td>
<td></td>
<td>Kenai Peninsula</td>
<td>Kenai</td>
</tr>
<tr>
<td>31 January 1912</td>
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<td>61°</td>
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<tr>
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<td>Kenai, Aleutian Range</td>
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<tr>
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<td>7.1</td>
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<td>137°</td>
<td>Chichagof Island</td>
<td>St. Elias</td>
</tr>
<tr>
<td>24 October 1927</td>
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<td>139°</td>
<td>Dry Bay</td>
<td>St. Elias</td>
</tr>
<tr>
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<td>146° 30'</td>
<td>Hinchinbrook Island</td>
<td>Chugach</td>
</tr>
<tr>
<td>27 April 1933</td>
<td>7.0</td>
<td>61° 12'</td>
<td>150° 42'</td>
<td>Cook Inlet</td>
<td>Chugach, Kenai, Alaska Range</td>
</tr>
<tr>
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<td>Chugach</td>
</tr>
<tr>
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<tr>
<td>22 July 1937</td>
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<td>64° 42'</td>
<td>146° 42'</td>
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<td>Alaska Range</td>
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<tr>
<td>3 November 1943</td>
<td>7.3</td>
<td>61° 48'</td>
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<tr>
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<td>146° 06'</td>
<td>Mt. Deborah</td>
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<td>Blying Sound</td>
<td>Kenai</td>
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<tr>
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<td>151°</td>
<td>Kenai Peninsula</td>
<td>Kenai</td>
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<td>Lituya Bay</td>
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</tr>
<tr>
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<td>61° 06'</td>
<td>147° 42'</td>
<td>Prince William Sound</td>
<td>Chugach, Kenai</td>
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</table>

From Field (1968)
This work reviews these studies, attempts to resolve their differences, and presents new data that enables a new mechanism of emplacement of rock avalanches to be put forward.

Two of the more significant difficulties encountered in the study of large catastrophic landslides have been the questions of the mode and rate of dissipation of the immense amount of energy that is released in the collapse of whole mountainsides. The Sherman Glacier rock avalanche of 1964 was a particularly valuable event from the point of view of seeking answers to these questions. Many features, relating to the deformational history of this avalanche, were preserved with exceptional clarity in the deposit that it left atop the broad flat surface of Sherman Glacier. This work discusses these features, and how they relate to the mode of emplacement of the avalanche.

Most of the large rock avalanches that fell on glaciers in Alaska during the Alaskan earthquake (Table 2) left deposits that are apparently similar to that of the Sherman Glacier rock avalanche, and they were probably emplaced by the same mechanism. Several of them, however, left deposits that have a very different morphology. An alternate mechanism of emplacement that may have been utilized in these avalanches is also discussed.

This research was supported by National Science Foundation Grants GP-4396, GA-409, GA-983, and GA-11752, by the Graduate School, and by the Research Foundation, both of The Ohio State University.
Austin S. Post, U.S. Geological Survey, supplied many of the aerial photographs utilized in this analysis.

**Comment on terminology**

Although thousands of landslides occurred during the 1964 earthquake, few attracted much attention. One group that aroused considerable interest was that of the large landslides whose deposits appeared exceptional in being spread over an area very large compared with the area of the source. This apparently efficient dispersal is the outstanding characteristic of this group, and the term rock avalanche is used here as a convenient name for this class of landslide. No material, or mechanism of transport is expressed or implied in the use of the term avalanche. Mudge's (1965, p. 1034) classification of rockfall- and rockslide-avalanches is rejected as unnecessary because most very large landslides involve entire peaks or mountain slopes so that sliding is always dominant.

**Rock avalanches triggered during the Great Alaska Earthquake**

The Great Alaska earthquake, centered at 61.04 ± 0.05° N., 147.73 ± 0.07° W., at a depth of about 33 km (Krouskopf, 1968, p. 9), began at 03 hr 36 min 14.0 sec, 28 March 1964 GMT (5:36 pm, 27 March, local time), reached a magnitude estimated to have been between 8.4 and 8.6 on the Richter scale, and lasted perhaps 3 to 4 min. Because of the many major aftershocks, ten of magnitude greater than 6.0 occurred within 24 hr of the initial shock and approximately 12,000
of magnitude greater than 3.5 occurred within 69 days (Press and
Jackson, 1965), few of the multitude of reported landslides can be
positively attributed to the initial shock of the earthquake; how-
ever, this is not surprising in such a sparsely populated region.

Seventy-nine major rock avalanches that produced 51 deposits
of area greater than 0.5 km$^2$ fell on glaciers between August 1963
and August 1964 (Table 2). At least several others covered areas of
0.5 km$^2$ during this interval, but did not fall on glaciers. Many
other avalanches and landslides of smaller size occurred; although
very numerous, little is known of these, aside from Hackman's (1965)
study of the abundance of slides, of all kinds, generated by the
earthquake within a small area around the epicenter; 95 percent of
these were snow avalanches, without perceptible debris. Of the
larger rock avalanches, the Puget Peak avalanche was observed and
filmed in motion (Hoyer, 1971), and the Sherman Glacier rock avalanche
was heard during the earthquake, but the results of the more remote
events were not observed until days, and even months after the earth-
quake. Nevertheless, it is most probable that it was the initial
major shaking that triggered most of the observed landslides and
avalanches, because the numbers of new landslide deposits observed
in August 1964 greatly exceeded those observed in previous and sub-
sequent years of observation (Tables 3 and 4).

There is good evidence that at least one of the largest land-
slides (The Sherman Glacier rock avalanche) did not fall at the onset
of the initial shaking, but fell moments later, after massive snow
<table>
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<th>Identify-</th>
<th>Lat.</th>
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<th>No. of</th>
<th>Area</th>
<th>Length</th>
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<td></td>
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</tr>
<tr>
<td></td>
<td></td>
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<td>31°</td>
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<td>3</td>
<td>N</td>
</tr>
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<td></td>
<td></td>
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<td>1,370</td>
<td>1</td>
<td>61°13'</td>
<td>147°14'</td>
<td>1</td>
<td>1</td>
<td>1.5</td>
<td>SW</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2</td>
<td>13°</td>
<td>147°16'</td>
<td>2</td>
<td>1</td>
<td>1.5</td>
<td>ESE</td>
</tr>
<tr>
<td>Ramney</td>
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<td>1</td>
<td>61°11'</td>
<td>147°34'</td>
<td>1</td>
<td>1</td>
<td>2</td>
<td>SE</td>
</tr>
<tr>
<td>Serpentine</td>
<td>26</td>
<td>1</td>
<td>61°09'</td>
<td>146°16'</td>
<td>1</td>
<td>0.5</td>
<td>2.5</td>
<td>S</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2</td>
<td>60°61'</td>
<td>146°19'</td>
<td>1</td>
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<td>2</td>
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</tr>
<tr>
<td>Surprise</td>
<td>70</td>
<td>1</td>
<td>61°02'</td>
<td>146°31'</td>
<td>2</td>
<td>3</td>
<td>3</td>
<td>ESE</td>
</tr>
<tr>
<td>Harrisman</td>
<td>49</td>
<td>1</td>
<td>60°56'</td>
<td>146°28'</td>
<td>2</td>
<td>1</td>
<td>2.5</td>
<td>N</td>
</tr>
<tr>
<td>Pigot</td>
<td>16</td>
<td>1</td>
<td>60°54'</td>
<td>143°30'</td>
<td>1</td>
<td>0.5</td>
<td>4</td>
<td>E</td>
</tr>
<tr>
<td>Twentymile</td>
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<td>1</td>
<td>60°57'</td>
<td>146°38'</td>
<td>1</td>
<td>2</td>
<td>2.5</td>
<td>W</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2</td>
<td>56°</td>
<td>143°30'</td>
<td>3</td>
<td>1.5</td>
<td>1.5</td>
<td>1 W, 2 NW</td>
</tr>
<tr>
<td>Contact</td>
<td>10</td>
<td>1</td>
<td>60°28'</td>
<td>148°28'</td>
<td>4</td>
<td>3</td>
<td>1.5</td>
<td>NE</td>
</tr>
<tr>
<td>Unnamed</td>
<td>11</td>
<td>1</td>
<td>59°48'</td>
<td>149°57'</td>
<td>3</td>
<td>1.5</td>
<td>1.5</td>
<td>W</td>
</tr>
<tr>
<td>Unnamed</td>
<td>4</td>
<td>1</td>
<td>59°42'</td>
<td>145°03'</td>
<td>2</td>
<td>2.5</td>
<td>1.5</td>
<td>W, SW</td>
</tr>
<tr>
<td>Unnamed</td>
<td>4</td>
<td>1</td>
<td>59°44'</td>
<td>150°13'</td>
<td>1</td>
<td>0.5</td>
<td>1.5</td>
<td>E</td>
</tr>
</tbody>
</table>

* Area covered by dust
+ Area from 1:10,000 map by H. Brecher (unpublished)

From Post (1968)
avalanching had begun (Marangunic and Bull, 1968, p. 389). It was thus dislodged by the continued shaking, as were the larger snow avalanches reported by LaChapelle (1968).

Studies of most of the rock avalanches have been confined to occurrence, location, and area. Profiles of four of the avalanche paths were prepared by Post (1967, 1968) where adequate maps were available; Reid (1969) studied Slide Glacier, but his work was primarily concerned with the glacier itself; Hoyer (1971) undertook a detailed study of the Puget Peak avalanche, one of those that did not fall on a glacier; Marangunic (1968, 1972), Marangunic and Bull (1968), Plafker (1968), and Shreve (1966 and 1968a) made detailed studies of the largest avalanche on Sherman Glacier; Post (1967, 1968) made photographic interpretation of general characteristics of the larger avalanches; and Johnson and Ragie (1968) made an aerial photographic analysis of the Allen II rock avalanche.

Only two studies of source areas were undertaken; Plafker (1968) made a brief ground and an extensive photogrammetric study of source areas of two of the larger rock avalanches at Sherman Glacier; Marangunic and Bull (1968) also studied the same two source areas.

Rock avalanches not triggered by the earthquake of 1964

Avalanches before August 1963

Eleven rock avalanches of area greater than 0.5 km² fell on glaciers between 1945 and August 1963 (Table 3); none of these approaches the magnitude of the largest avalanches of 1964.
Table 3. Rock avalanche deposits on glaciers since 1945 and prior to 1964 Earthquake

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Area (km²)</th>
<th>Year</th>
<th>Lat. (°N)</th>
<th>Long. (°W)</th>
<th>Area (km²)</th>
<th>Length (km)</th>
<th>Direction traveled</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cascade</td>
<td>181</td>
<td>1945?</td>
<td>59° 07'</td>
<td>135° 47'</td>
<td>3</td>
<td>2.5</td>
<td>SE</td>
</tr>
<tr>
<td>John Hopkins</td>
<td>310</td>
<td>1961?</td>
<td>58° 47'</td>
<td>137° 10'</td>
<td>2</td>
<td>2</td>
<td>NE</td>
</tr>
<tr>
<td>Margerie</td>
<td>130</td>
<td>1961</td>
<td>58° 56'</td>
<td>137° 13'</td>
<td>2.5</td>
<td>3</td>
<td>SE</td>
</tr>
<tr>
<td>Netland</td>
<td>39</td>
<td>152?</td>
<td>59° 26'</td>
<td>137° 34'</td>
<td>2.5</td>
<td>2.5</td>
<td>NW</td>
</tr>
<tr>
<td>Smith</td>
<td>18</td>
<td>1955?</td>
<td>61° 16'</td>
<td>147° 48'</td>
<td>0.5</td>
<td>1.5</td>
<td>E</td>
</tr>
<tr>
<td>Bryn Mawr</td>
<td>23</td>
<td>1960?</td>
<td>61° 15'</td>
<td>147° 52'</td>
<td>1</td>
<td>3</td>
<td>ESE</td>
</tr>
<tr>
<td>Vassar</td>
<td>13</td>
<td>1958?</td>
<td>61° 13'</td>
<td>147° 53'</td>
<td>1</td>
<td>1.5</td>
<td>ESE</td>
</tr>
<tr>
<td>Barry</td>
<td>78</td>
<td>1960</td>
<td>61° 11'</td>
<td>148° 07'</td>
<td>4</td>
<td>3.5</td>
<td>SE</td>
</tr>
<tr>
<td>Serpentine</td>
<td>16</td>
<td>1963?</td>
<td>61° 07'</td>
<td>148° 16'</td>
<td>0.5</td>
<td>2.5</td>
<td>W</td>
</tr>
<tr>
<td>Surprise</td>
<td>57</td>
<td>1963</td>
<td>61° 02'</td>
<td>148° 31'</td>
<td>0.5</td>
<td>1.5</td>
<td>SE</td>
</tr>
<tr>
<td>Pigot</td>
<td>21</td>
<td>1945?</td>
<td>60° 54'</td>
<td>148° 29'</td>
<td>1</td>
<td>3</td>
<td>E</td>
</tr>
</tbody>
</table>

From Post (1968)
Post (1967) also notes older, undated deposits on Chistochina Glacier in the Alaska Range, on Casement Glacier in Glacier Bay, and several on Margerie Glacier. Field (1964, p. 16) reports another probable rock avalanche on Ferris Glacier that was first seen in 1912, but does not report its area. Rock and soil avalanches, rock slides, and earth flows occurred over a large area during displacement along the Fairweather fault in 1958 (Tocher, 1960), but only two were large - a smaller one on La Perouse Glacier and a larger one, now known as the Lituya Bay Avalanche of 1958. The latter is noted for the giant wave that it induced in Lituya Bay and not for having removed a small section of the terminus of Lituya Glacier. Tarr (1909, p. 83) noted a "great area of moraine" on Hayden Glacier, a feeder of Malaspina Glacier, "with a form somewhat like a delta, apparently representing an enormous avalanche, possible precipitated during the 1899 earthquake".

By all reports, avalanching was far more important during the 1899 earthquake than during the 1964 earthquake, yet few large rock avalanche deposits were noted after 1899 whereas more than 80 were reported after 1964. The advent of aerial reconnaissance is a probable cause of this anomaly. Publications relating to the 1899 earthquake make little distinction between rock avalanching and snow avalanching in the addition of mass to glaciers. Tarr and Martin (1912, p. 48) note that "Downfall of rock and snow during the first half of September 1899 was enormous in amount and spread over a wide area". The mountain face on the east side of Yakutat Bay was
### Table 4. Rock avalanches more recent than 1964 Earthquake

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Area (km²)</th>
<th>Year</th>
<th>Lat. (°N)</th>
<th>Long. (°W)</th>
<th>Area (km²)</th>
<th>Length (km)</th>
<th>Direction traveled</th>
</tr>
</thead>
<tbody>
<tr>
<td>Allen</td>
<td>230</td>
<td>1965?</td>
<td>60° 47'</td>
<td>144° 56'</td>
<td>7.5</td>
<td>7.5</td>
<td>NNE</td>
</tr>
<tr>
<td>Fairweather</td>
<td>260</td>
<td>1965?</td>
<td>58° 53'</td>
<td>137° 40'</td>
<td>8.5</td>
<td>10.5</td>
<td>WSW</td>
</tr>
<tr>
<td>Blossom</td>
<td>8</td>
<td>1965</td>
<td>60° 03'</td>
<td>140° 05'</td>
<td>1.5</td>
<td>1.5</td>
<td>E</td>
</tr>
<tr>
<td>Marvin</td>
<td>310</td>
<td>1965</td>
<td>60° 06'</td>
<td>140° 07'</td>
<td>1</td>
<td>3</td>
<td>W</td>
</tr>
</tbody>
</table>

From Post (1968)
"entirely changed" in 1899 (op. cit., p. 49). From a ship anchored off the coast the clouds of "dust and snow" could be seen rising from the St. Elias Range for 180 km (Capt. Durkee, reported in Tarr and Martin, 1912, p. 50). A Dr. Cox at Yakutak Bay in 1899 was kept uneasy by, among other things, the "roar of great landslides down the sides of the mountains every little while" (op. cit., p.17).

Avalanches after August 1964

Post (1968, p. 296) reported that four large rock avalanches fell in the Chugach Mountains between August 1964 and August 1965 (Table 4). Two of these were as large as the largest precipitated by the 1964 earthquake, and they left apparently similar deposits. No data have been published subsequently. The pre-1964 earthquake frequency of about one rock avalanche of large size every two years would suggest that perhaps another five have fallen to date, but Post (personal communication) has seen none. They are by no means a rare event in Alaska's mountain ranges, although the earthquake may have precipitated a 160-year supply.

THE SHERMAN GLACIER ROCK AVALANCHE

Several rock falls and rock avalanches of various sizes fell in the vicinity of Sherman Glacier during the 1964 earthquake (Fig. 3). Only the largest avalanche has been studied in detail. Accounts of many of the characteristics and possible mechanisms of this major rock avalanche are given in varying detail by a number of workers
Fig. 3. Mosaic of the Sherman Glacier rock avalanche and other adjacent landslides induced by the 27 March 1964 great Alaska earthquake. Width of view is about 8 km (from 1967 aerial photography).

**Geologic setting**

The geology of the Sherman Glacier valley has been described by Plafker (1968) and by Marangunić and Bull (1968). Bedrock exposed around Sherman Glacier consists of a highly deformed slightly metamorphosed sedimentary sequence, probably belonging to the Orca Group (lower Tertiary), and granitic igneous rocks that locally intrude this sequence. The sedimentary rocks vary from thin to very thick beds of well indurated, gray, poorly sorted, fine- to medium-grained sandstone interbedded with thin, finely laminated beds of dark gray siltstone. All have been locally metamorphosed by small intrusions of light-gray biotite hornblende granite.

Closely spaced joints, faults, and tight folds are ubiquitous through the bedded rocks. Regional strike roughly parallels the trend of Sherman Glacier, with moderate to high dips to the north, and local overturning. Steeply dipping faults commonly strike roughly parallel to the bedding and exert a strong control on regional and local topography.

Marangunić and Bull (1968) suggest that the geological structure of the ridge and mountain top from which the avalanche fell is a broad asymmetrical, gently westerly plunging syncline that parallels
the valley of Sherman Glacier (Fig. 4). This structure contrasts with that shown by Plafker (1968, Plate 1) (Fig. 7) which seems more in accord with structure observable on aerial photographs. The actual structure, however, is sufficiently complex that both interpretations are acceptable simplifications from the differing perspectives.

Shreve (1966, p. 1639-1640) has described the avalanche source as a steeply dipping, pervasively jointed, tabular block of hard, unbedded, and coarse-grained graywacke, and hard black non-fissile argillite, but the block also included a small hanging glacier, snow, till and talus. This block, as much as 120 m thick, 550 m wide, and 750 m long (modified from Plafker 1968, Plate 2), formed the crest and western upper flank of a former mountain summit, the remnants of which are now called Shattered Peak (Figs. 1 and 3).

The avalanche scar appears to follow a preexisting set of hematite-coated, intersecting, curved fault surfaces that closely parallel the steeply dipping bedding, (Plafker 1968, Fig. 3). The generally highly-polished avalanche scar dips steeply (at 45° - 75°) to the west, toward Andres Glacier (Fig. 3), tributary to Sherman Glacier.

Avalanche paths

Collapse of Shattered Peak gave rise to avalanches down all three sides of the horn, onto Andres, Eliza and Saddlebag Glaciers (Fig. 3 and 6). The center of mass of the debris lay to the west of the present summit and most debris slid in an initial WNW
Fig. 4. Geological sketch of the source of the Sherman Glacier rock avalanche. A smaller avalanche also fell from Pyramid Peak (from Marangunić and Bull, 1968, Fig. 3). Bedding and faulting are greatly simplified.
Fig. 5. Shattered Peak: a view of the avalanche scar from Andres Glacier six years after the earthquake. The smooth fault surfaces that form the upper boundaries of the scar are visible above the apron of avalanche snow that now obscures the original "break in slope" of the avalanche path.
Fig. 6. Map of the Sherman Glacier rock avalanche (from an unpublished 1:10,000 line map of Sherman Glacier compiled by H. Brecher, The Ohio State University, from 1967 aerial photography; flow features from a 1:10,000 orthophotograph by H. Brecher and from aerial photographs).
direction down the surface of rupture. This led the disintegrating rock mass to hit an-up-to-150-m-high spur on the western side at Andres Glacier 1 km west of the avalanche origin (Fig. 6). About a quarter of the mass swept over the spur to reach Sherman Glacier, while the remainder was deflected to the east, northeast, and north (Fig. 2). The greater portion of debris, about half, then flowed directly down the local slope on Andres Glacier following closely the base of the spur, before spreading across and down Sherman Glacier (Fig. 6 and 7).

At the foot of the up-to-150-m-high spur, the avalanche struck a 40-m-high rock cliff, but most of the remainder of the terrain must have been covered by snow that early in Spring. On stoss slopes of the spur much of the mantle of snow was stripped, along with vegetation and slabs of root-bonded humus, but on lee slopes, plant communities were left largely undisturbed, save by a scattering of boulders. Not all soil, however, was stripped from the stoss slopes; where some plants have since pushed through the debris in many places.

Marangunic and Bull (1968) observed the glacier surface beneath the avalanche debris at a few hand-dug holes, along edges of the avalanche, and at margins of newly opened crevasses. They found about 2 m of winter snow overlain by debris in most places. The snow rested on a clearly identifiable glacier ice surface. Locally snow was absent or only a few centimeters thick. This snow blanket-ed and bridged numerous crevasses and smoothed many minor
1 Farthest point of travel
2 Northernmost point of travel
3 Maximum height climbed
4 Original slab
5 Break in slope

Fig. 7. Profiles of the avalanche path (from a 1:10,000 map by H. Brecher). The dip of the bedding of Shattered Peak is from Plafker, 1968, Plate 1.
topographic irregularities of centimeter scale, but features with a relief of a meter or so, such as medial moraines, were still present in the avalanche path and are still visible in the surface relief of the avalanche deposit (Fig. 1).

Locally the avalanche descended vertical faces up to 100 m high in the lee of the spur, but for the most part it travelled over slopes of between 2 and 6°. Avalanche flow generally conformed to regional slope within broad limits (Fig. 6); only in a few places did it flow directly across slope. Significant upslope flow (relative to present topography) occurred in overtopping the spur and over about 600 m of glacier in the lee of the spur, and minor upslope flow occurred at the north wall of the valley and in the southwestern corner of the avalanche.

Regional topography beneath the avalanche apparently exerted strong control over the avalanche path, but it was not the only factor controlling the direction of motion: lateral friction, momentum, and regional topography of the moving upper surface of the avalanche were other controls.

Disintegration of the initial slab

Very little is known of the disintegration of the initial slab: Shreve (1966, p. 1642) believes that fragmentation of the block had occurred before the debris crossed a major break in slope at the foot of the scar (Fig. 7); Marangunić and Bull (1968, p. 388), however, suggest that fragmentation occurred at this break, when
flexure of the slab occurred. Shreve, Plafker, and Marangunic' and Bull, believed that the entire block sheared off as a unit along the present scar surface, and then disintegrated during the fall. However, debris poured off Shattered Peak in all possible directions, suggesting that the peak could have fallen apart in situ. Perhaps the pervasively fractured pile of rubble that was Shattered Peak behaved thixotropically and was liquified by the violent ground motion during the earthquake and flowed rather than slid. One peak in the vicinity had its summit completely shattered in the shaking but very little of it fell, and the shattered pile of debris still remained more or less intact months after the earthquake (Fig. 8).

The impact on Andres Glacier also caused extensive fracturing of the avalanche mass, and of the eastern face of the overtopped spur: one portion of the spur was brecciated (Marangunic' and Bull, 1968, p. 388); the glacier, however, could not be observed to determine what effect the impact had had on it.

The mass certainly must have continued to disintegrate in transport. However, the reported decrease in grain size with distance from the source (Marangunic' and Bull, 1968, p. 388) has not been demonstrated to be a function of distance of transport, and indeed was not substantiated (see Marangunic' and Bull, 1968, p. 388). Debris that swept over the spur contains none of the enormous boulders that characterise most of the debris that did not (Fig. 3), and thus may be more fragmented. The former travelled further, but "distance travelled" may not have contributed as much
Fig. 8. A shattered pile of rubble that was not precipitated as a major rock avalanche during the 1964 earthquake. This unnamed peak lies in the Chugach Mountains of south-central Alaska (1965 photography by Austin Post).
to fragmentation as did the irregular topography along the path of flow.

**Thickness and Volume**

Estimates of average thickness of rock-avalanche debris on Sherman Glacier vary. Plafker (1968, p. 380) estimated, through photogrammetrical studies of the peak (from photographs taken before and after the earthquake: 1950, and 1964) that $25.6 \times 10^6 \text{ m}^3$ of rock, snow, and ice was lost from Shattered Peak. He estimated that this would spread to a mean thickness of about 3 m over the area of the debris without compensation for the change in density. He used a density of 2.69 Mg m$^{-3}$ for the rock in his calculation of mass, and this agrees well with Marangunic's (1972, p. 27) measurement of 2.7 Mg m$^{-3}$ for the graywacke. Marangunic (1972, p. 79) also measured a bulk density of the debris of 2.0 Mg m$^{-3}$ two years after the avalanche. Thus, for the measured density change of 25 percent, the photogrammetric estimate should be corrected to 4.0 m average thickness of debris. Shreve (1966, p. 1639) estimated thickness to be 3 to 6 m, but did not state how this estimate was made. Marangunic and Bull (1968, p. 390) made a number of direct measurements and gave a mean thickness of 1.3 m, but pointed out that many individual blocks exceed this. This wide range in estimates of a mean thickness is perplexing and warrants some discussion, especially as Marangunic (1968, p. 92) comments on the nearly uniform thickness of 1.0 to 1.2 m over large areas of the debris.
Plafker (1968, p. 378) suggests that ±7 per cent would be a conservative estimate of precision of his photogrammetry. It is a reasonable estimate of precision for a multiplex plotter, that was used, but is an optimistic assessment of the quality of the 1950 stereophotographs of the peak now known as Shattered Peak. It also does not take into account the probability of systematic errors; that is it is not an assessment of accuracy. Accuracy can only be assessed by comparing Plafker's maps with other independently derived maps: field-checking is impossible in this case.

The "4000-ft" contour along the west face of the peak now known as Shattered Peak on the Cordova C-3 Quadrangle (U.S. Geol. Surv. 1:63,360 series, Alaska, 1953) is about 430 ft long. On Plafker's Plate 2, the same contour is 730 ft long, a +70 percent difference. In the east-west direction the difference is about -12 percent. In addition, the gradient of the west face is nearly 5° greater on his plate than on the C-3 Quadrangle map. These differences could be accounted for if the stereoscopic model of the 1950 photography was tilted when Plafker's maps were made, because it is unlikely that such a significant tilt is present in a large area of the C-3 map. Correction for the tilt about a north-south axis would elongate the triangular contours around the former peak in Plafker's Plate 2 to conform with those of the Cordova C-3 sheet. If the same tilt was also present in his post-earthquake map, Plafker's volume estimate would still be correct,
but this does not appear to be the case. A 1:10000 unpublished
map, compiled from 1967 1:20000 aerial photographs, by H. Brecher
(unpublished) also includes the west face of Shattered Peak, but
only at a 25-m contour interval. Comparison of the two maps
suggests that both probably are accurate in scale and orientation
over that face.

Correction of Plafker's profiles across the scar for this
apparent tilt would remove much of the former puzzle of missing
mass where no rock had fallen, and would greatly reduce the volume
estimate for the avalanche. I have made an approximate correction
that suggests that Plafker's estimate is about 80 percent too large
and should be reduced by at least $11.4 \times 10^6 \text{ m}^3$. Plafker (1968,
p. 378) also estimated that not more than $1.5 \times 10^6 \text{ m}^3$ of snow and
ice was included in the avalanche. This was based on the thickness
of the remaining ice veneer along the southern margin of the scar
5 months after the avalanche, but the mass that fell also included
an entire cirque glacier. This glacier probably contained more than
$2 \times 10^6 \text{ m}^3$ of ice and snow. Thus, the photogrammetric estimate of
original volume of rock should be reduced to $12 \times 10^6 \text{ m}^3$, although
Plafker's photogrammetry requires revision to obtain a precise
value. This corrected volume would spread to an average thickness
of 1.96 m for the 25 percent decrease in bulk density had it all
flowed in the one avalanche, but debris fell in three directions.

The reliability of Shreve's estimate of 3 to 6 m cannot
readily be assessed; he gives neither the number, nor location of
his observations, nor what was observed. His estimate greatly exceeds most of Marangunic's and my own observations at exposures in crevasses and pits. Marangunic measured 15 thicknesses of debris, widely distributed over the avalanche surface, largely from the sites of his fabric studies (Marangunic and Bull, 1968, Fig. 9). Thickness ranges from zero to 5 m and is most variable on Andres Glacier where these extremes were observed. If the extreme values are excluded from an assessment of mean thickness, the debris has a remarkable uniform thickness ranging between 1 and 2 m over most of the glacier, and averages 1.65 m. This average suggests an original rock volume of $10.1 \times 10^6$ m$^3$ which is in remarkable agreement with Field's estimate of $10 \times 10^6$ m$^3$ (Ragle and others, 1965, p. 40).

Relief on the debris surface locally exceeds this mean thickness. At one boulder, the debris is at least 16 m thick, but thickness at isolated boulders should not be utilized in assessing an average debris thickness. This one boulder has a volume of about 2000 m$^3$ and perhaps a small part of the $1.9 \times 10^6$ m$^3$ missing between the corrected photogrammetric estimate and my estimate could be due to the volume of isolated boulders, but the apparently missing volume is almost exactly the volume of the other avalanches from the same peak. The areas of these avalanches, however, are poorly known, and their thickness unknown. Grooves in the avalanche surface in some places exceeded 2 m depth in the summer of 1965, but differential
melting of ice beneath debris of different thickness has caused them to rapidly deepen with time, and their initial depths are unknown. In other localities, the avalanche flowed in preexisting channels on the glacier surface and the relief on the debris surface partially reflects this topography. On 1965 aerial photographs, one debris ridge cut by a crevasse contains visible glacier ice that stands higher than the debris surface in an adjacent groove.

Composition of Avalanche Debris

About 85 per cent of the avalanche debris is graywacke, about 10 per cent is argillite, and the remaining 5 per cent is composed of soil, glacier-ice fragments, snow, and small amounts of vegetation (Marangunić and Bull, 1968, p. 388). Some vegetation (grasses, mosses, lichens and ferns) still grows on boulder sized fragments of the former surface of the mountain.

There is a significant amount of the original weathered mountain surface that is retained as the surface of the avalanche deposit; much of it is retained on large boulders that are big enough to protude through the rubble. However, a small portion of original surface is retained on much smaller boulders on the avalanche surface, indicating that much of the original block surface was retained as an outer skin of the moving avalanche, and that little mixing or block rotation occurred.

Fragments of the weathered mountain surface are rare in the lee of the overtopped spur, whereas debris such as vegetation
scraped from the spur is abundant across this debris surface; this observation contrasts with Shreve's (1966, p.1641) statement that such debris is present only at the distal edge. More mixing occurred along this path.

The general lack of mixing during transport is also shown by textural and lithological zonation in the debris (Fig. 3). Shreve (1966, p.1640) described "a three dimensional jig-saw puzzle effect ..., in which individual blocks are shattered yet undisaggregated". These shattered blocks are but one end of a series that grades to subparallel, roughly transverse textural and compositional bands of disaggregated material with gradational boundaries (Fig. 3 and 6). Darker bands were formed from argillaceous beds, while bands of large boulders originated from disintegration of massive graywacke horizons. Some bands of fine texture were formed by disaggregation of till or talus while others are friable argillite. Lithological and textural bands in the debris in the lee of the overtopped spur parallel the path of movement of the avalanche (Fig. 3); in marked contrast to the transverse banding elsewhere. These textural and compositional bands are a product of lithology and its distribution within the parent block, and the mechanical significance of the contrast depends critically upon this unknown distribution. Strata whose strike in the parent block was parallel to the direction of motion would now be dispersed as longitudinal bands in the avalanche, whereas strata whose strike was transverse to motion would have dispersed as transverse bands.
In their study of internal composition of the avalanche, Marangunic and Bull (1968, p. 392) also made measurements of orientation of the major axis of many boulders at 16 sites in the debris. These data are presented as plotted points in their fig. 9, but were later reduced to more useful contoured fabric diagrams (ten are presented in Fig. 9). All fabrics are strong and statistically significant. Several are exceptionally strong (for example see Fig. 9g). In general, long axes parallel flow direction, but transverse fabric modes are also present.

In several fabrics, the transverse mode produced by rolling clasts, is dominant (fig. 9d, and f) while the longitudinal mode, that probably is due to the orientation of clasts by the longitudinal stretching of the debris, dominates others (fig. 9a,b,c,h, and i). The longitudinal-stretching mode appears in fabrics obtained from near the surface of the debris, while the transverse, rolling mode appears in basal-debris fabrics as a response of the debris to a gradient in basal shearing stress, which is the cause of the rolling motion. These observations confirm Heim's conjecture that the basal-friction-induced rolling of debris at depth, decreases towards the top of "Trummerstroms", where flow dominates (Heim, 1882, p. 83).

Texture of the deposit

Clast size on the avalanche deposit ranges widely (Fig. 10): boulders to 20 m in diameter occur, and rock flour is present.
Fig. 9. Contoured fabric diagrams for clasts at selected sites in the Sherman Glacier rock avalanche (lower hemisphere projections of long axes; contour interval is one standard deviation of a uniform fabric of the same number of clasts; contouring by a computer program by C.E. Corbató following the method of Kamb (1959); data provided by Ć. Marangunić). Also shown are the arbitrarily selected flow lines used to evaluate flow of the debris and the location of the datum profile (see Internal Strain).
Fig. 10. Range of clast sizes in the Sherman Glacier rock avalanche. The boulder is 17 m high: the largest single fragment in the avalanche. The avalanche source, Shattered Peak, forms the background.
Most of the finer size grades that may have been on the surface have long been removed by rain and melting snow, and they are now concentrated at depth in the deposit. Marangunić (1968, p. 91) found the modal size of debris to be about 5 to 10 cm, but there is some question as to how this value was obtained (see Marangunić and Bull, 1968, p. 388) and even more question as to how an accurate value could be obtained.

There are regional and local variations in apparent modal size in the debris that are probably due to parent lithology. Marangunić (1968, p. 91) also suggested that modal particle size decreases distally, a result of disintegration of clasts in transit; but an adequate assessment of this question has not been made. Because the entire mass of the avalanche probably was in simultaneous motion (p. 71), it is improbable that a distal decrease in particle size occurred, and this would be consistent with Shreve's (1968b, p. 28) more detailed study of particle size, on the Blackhawk Slide, in which he found no change.

Porosity and permeability of the material are high but decrease with depth in the debris and change with time. Standing bodies of water were present on the debris surface in 1969, 1970, and 1971, but only where there was limited drainage through underlying ice and it probably is the low permeability of the underlying ice that inhibits drainage (c.f. the Madison Canyon Slide that dammed a lake; Hadley, 1964). Drainage of basal layers of the debris is often poor, and moisture content of the near-basal debris is near
saturation but it is not known if this is only because of the underlying melting ice, or of an inherent low permeability of the debris. No attempt has been made to measure porosity or permeability of the debris, even though permeability is a central factor in most suggested flow mechanisms.

In addition to the compositional and textural bands previously mentioned, the surface of the avalanche shows a number of other features that might be significant in analysis of its mode of emplacement. These include grooves, fissures, cones, lateral ridges, distal rims, and folds (Fig. 6).

Grooves

A most striking feature of the Sherman Glacier rock avalanche is the pattern of subparallel, shallow, V-shaped longitudinal grooves that cover the surface from the point of impact of the avalanche with Andres Glacier to the furthest points of travel (Fig. 1 and 3). Grooves are wider and deeper near the distal margin, where some are 10 to 15 m wide and up to 2.5 m deep. These tend to separate major lobes near the debris margin, but smaller grooves are abundant within lobes and even within major grooves.

Both Shreve (1966, p. 1641) and Marangunić and Bull (1968, p. 390) note that major grooves separate trains of different rock type while the latter also note that minor grooves slightly offset transverse compositional bands. Aerial photographs show that major grooves also offset these bands; larger grooves have much larger offsets (Fig. 3).
Grooves (and equally the ridges between them) are most useful features in the analysis of emplacement of the avalanche for they clearly show direction of flow and, by their cross-cutting relationships, sequences of flows within the debris.

Shreve (1966, p. 1641) proposed that larger grooves were formed by shear between debris trains travelling at different speeds, whereas smaller grooves are splits caused by divergent flow. Marangunic' and Bull (1968, p. 391), on the other hand, believe that all grooves represent shear. The grooves delineate some form of flow domain, and the offsetting of bands at grooves indicates that shear was present. However, many grooves are crossed by transverse fissures and folds in the debris with no offset, and no apparent change in grooving either. Folding and fissuring, in general, appear to be later features in the evolution of the surface in these places, although locally, fissures are apparently distorted by shear at some grooves.

Although grooves are prominent surface features of the debris, it is not known that all were originally at the surface. Some grooves and ridges may have formed at the base during the avalanche only to rise to the surface by collapse of debris when the avalanche stopped. These grooves and ridges would probably have a more subdued form than primary surface features. Such distinctly different forms have not been identified and may not be present.
Transverse fissures

Fissures transverse to flow direction (Fig. 3 and 6) occur as swarms of parallel, irregularly sinuous, deep V-shaped or flat-bottomed trenches 3 to 15 m wide, 1 to 5 m deep, and 160 m long (the full range is about 30 to 200 m in length) that are separated from each other by distances that are two to three times their width (Shreve, 1966, p. 1642). Most are concave towards the distal edge and terminate at grooves bounding major debris streams within the avalanche, with ends oriented about 45° to direction of flow. Shreve considers that their shape indicates that basal drag was negligible compared with lateral drag which "supports a conclusion that the debris slid rather than flowed". Basal drag probably is not a factor in their formation and the fissures support no such conclusion, although the conclusion is about 99 per cent true (p. 88).

Marangunic and Bull (1968, p. 391) distinguish two types of fissures: those concave toward the distal edge; and those en echelon ac a 45° angle to the direction of flow. They attribute both types of fissures to tension within the flowing debris, interpreting en echelon fissures to be formed in zones of shear between major flows, and concave fissures to be formed by shear within flows.

I have only been able to identify two clearly en echelon fissures, and it may be that Marangunic and Bull have mis-identified deformed "concave fissures". These concave fissures are one of several keys to the mechanics of the avalanche; they indicate (a) the state of
stress within the debris avalanching, and (b) the nature of the flow. The fissures are analogous to transverse crevassing in glaciers and, as was demonstrated by Nye (1952), curvature of fissures results from a longitudinal tensile stress modified by lateral shear.

The occurrence of transverse fissuring instead of continuous simple stretching indicates that debris near the upper surface of the avalanche was capable of brittle failure while in motion. This is consistent with flow of debris as a Bingham plastic (p. 75) and far from indicating that drag from "beneath was negligible compared with that from the sides" (Shreve, 1966, p. 1642), transverse fissures may result from a major subsurface drag of a thin rigid "crust" lying on an extending "fluid" substrate of more rapidly deforming debris (p. 92).

Lengths of transverse fissures provide a measure of the widths of debris through which a nonhydrostatic stress was transmitted during the avalanches. The smallest units were streams about 30 m wide and the longest streams under longitudinal tension were about 200 m wide; but these were not the widest units (see below).

En echelon fissures and transverse folds

Flow with longitudinal compression and lateral shear can produce longitudinal crevassing in glaciers; the crevasses make angles of 45° to flow at the margin of the glacier but curve to be more nearly parallel to flow towards the center. A similar state of stress could produce the en echelon fissures discussed by
Marangunic' and Bull. In only one locality, the plumose easternmost segment of the avalanche debris where it debouches from Andres Glacier, has a suitably oriented set of fissures been observed by me on aerial photographs (2 fissures). Longitudinal compression would thus appear to be extremely uncommon within the flow streams, although it must have been significant during the impact with the spur opposite Shattered Peak.

Although materials under compression can fail in tension when laterally unconfined, as fissures imply, they more frequently fail by buckling, or by shearing when there is lateral constraint. Toward the northwestern distal margin of the avalanche is a region of transverse, equally spaced undulations in the debris (Fig. 3, 6 and 11) previously unreported because it closely resembles topography produced by collapse of debris into subjacent glacier crevasses (Fig. 3). Orientation of these ridges is, however, an impossible one for glacier crevasses in this area because aerial photographs taken in 1963 show that crevasses in this region had a quite different orientation. The ridges are folds in the overlying debris, a result of buckling of the debris sheet in compression. The wavelength of folding is about 15 m and the amplitude is unknown but was probably only about 1 m (since 1964, different rates of melting of ice beneath the crests and troughs of the folds have caused the amplitude to grow with time). Folds extend over a length of about 400 m and a width of 250 m, making this region the largest area of avalanche that can
Fig. 11. Transverse folds in the debris in the northeastern corner of the avalanche deposit. Fold amplitude accentuated by nine years of differential melting of ice beneath wave crests and troughs, and by oblique lighting. (1973 aerial photography by Austin S. Post)
be shown to have acted as a more-or-less rigid unit during emplacement, and the only area that can be clearly shown to have behaved as a thin flexible sheet, in that it transmitted a non-hydrostatic stress. This region of folding occurs immediately below a region of locally steeper gradient which suggests (a) that the compression was a response to change in slope and (b) that the debris did not travel far after compression.

Failure by buckling indicates that, at this location, debris was in longitudinal compression and stressed above its yield stress.

Cones

Shreve (1966, p. 1642) noted a "most puzzling" occurrence of cones of tightly packed finer-grained debris that are piled on top of some boulders (Fig. 12) and which "must form by a ... mechanism as yet unknown" (Shreve, 1966; 1968b, p. 39). Marangunic' and Bull (1968, p. 391) suggested that some of these cones are glacial deposits or talus, often with stratification and soil development, originally present on the mountainside. The abundant fragments of former mountain surface now at the surface of the debris are not merely represented by weathered lichen-covered rock, but also by fragments of the surficial deposits of talus, etc., preserved in transport, generally atop of large boulders. It is, however, significant that no large amount of glacier ice from the small cirque glacier was ever noted on the debris surface, which supports Marangunic's (1972, p. 84) conclusion that the glacier slid off the mountain top before the rock avalanche.
Fig. 12. Dirt cone on the surface of the avalanche deposit on Sherman Glacier. The debris in this cone may represent a sample of the interior of the avalanche, trapped atop a boulder during flow and thinning of the avalanche.

(Photograph by C. Marangunic)
Distal rims and lateral ridges

Shreve (1966, p. 1641) suggested that the edges of the Sherman Glacier rock avalanche are "bounded by lateral ridges and distal rims typical of the Blackhawk type" (of landslide). They are described as being 3 to 15 m high and 15 to 150 m wide, chaotically hummocky in topography and, in places, imbricate in structure. Each of the multiple minor lobes of the western margin is noted to have low lateral ridges and a prominent distal rim.

Marangunic and Bull (1968, p. 393) noted a narrow rim in debris at the distal edge, but they also noted rimless distal edges. They refer to a "bulldozing" action, of debris lying over dirty avalanche snow with boulders, at the rimmed edges. In a discussion of mechanisms of emplacement they state "An early settling along the edges would probably have caused pressure ridges to be produced by the material pushing behind, and no such phenomenon has been observed, except perhaps where the distal edge reached the northern hillside ... and ... at the eastern lateral edge" (and also on Andres Glacier). Shreve's "imbricate structures" (Shreve, 1968b, p. 40) are the "pressure ridges" of Marangunic and Bull (1965, p. 393).

Marangunic and Bull (1968) do not explicitly state that the distal rims are raised, although this may have been implied by their their use of rim; however Shreve's (1966) Fig. 4 that might be expected to show a very prominent raised distal rim because it is emphasized in the caption, shows nothing that is raised, except
the entire avalanche that now sits on a pedestal of glacier ice (ice beneath the debris is insulated from the atmospheric heat source and melts more slowly than adjacent unprotected ice). Aerial photographs of the avalanche debris taken on 30 May, 1964 before summer melting began show a prominent raised distal rim only along the eastern edge. This edge also has been called a lateral ridge (Marangunić and Bull, 1968, p. 393) but it is more closely akin to a distal rim although it was utilized as a lateral boundary by later flow units in the avalanche (see also p. 91). The May 1964 photographs show that the margins initially were not chaotically hummocky to any great extent, and were not raised, except in isolated hummocks on the ends of some ridges (between grooves). An apparent raising is present at one locality along the northern margin where a debris train thins behind the margin. Hummocks became more prominent and chaotic with time as ice within and beneath the debris has melted. On 30 May, 1964 there was no raised rim in the region of Shreve's (1966) fig. 4.

Marangunić and Bull, and Shreve, did not visit Sherman Glacier until the summer of 1965, after melting had considerably altered the margins of the avalanche, and washing of the surface by rain had subdued the sharpness of the original topography so that their observations are not necessarily of features that were present at the end of March 1964 and which were recorded on aerial photographs in May of that year.

A sharply upcurved distal rim of a completely different form and origin to that referred to above has developed with time at the
western edge; this feature occurs in the exaggerated vertical relief provided by 1969 stereopairs of aerial photographs and had become plainly visible from the ground by 1971. I have found this feature to be an edge effect of the heat-balance regime at the debris-ice interface, and it is not discernible on aerial photographs taken at the end of summer of 1965.

In the May 1964 photographs, many of the multiple minor lobes along all margins terminate without ridges or hummocks. Others terminate at hummocks that are often smoothly continuous with ridges behind them, and present no chaotic appearance. A low lateral ridge is present only on one side of one minor lobe, and this occurs where the lobe followed along a trough in the glacier surface. Ridges that bound some flow units within the debris mass near the northern distal margin are probably the equivalent of lateral ridges. Lateral ridges up to 30 m high, such as have been described for the Blackhawk slide in California (Shreve, 1966b, p. 27) are absent, but the Sherman Glacier rock avalanche is nearly an order of magnitude thinner than that avalanche, and perhaps would have ridges only about 3 m high as Shreve (1966, p.1641) estimates. This however is inconsistent with his estimate of the mean thickness of the avalanche (see p.29).

Avalanche snow

In the same May 1964 stereoscopic photographs, snow and ice can be seen in the sides of some debris piles along the eastern and
northern margins of the avalanche, but this was extensively melted and covered by slumping debris by the end of the first summer, when Sherman Glacier was again photographed from the air. In a sequence of six sets of aerial photographs of the avalanche, a marked change in surface topography can be traced from year to year. Some of this change is due to differences in melting of ice beneath debris of different thickness, some is due to opening of glacier crevasses, some is due to rain wash, some to movement of ice beneath, but most of the change in the first five years was from compaction and melting of entrained and buried snow and ice.

The internal composition of the avalanche debris was examined by Marangunic and Bull (1968, p. 389). In pits on Andres Glacier mixed snow and debris was found under a layer containing ice fragments. They interpret this stratigraphy to suggest that a sequence of small avalanches consisting mostly of snow and ice preceded the major one, whereas the major avalanche was believed to have consisted almost exclusively of rock, without appreciable snow: "If snow had been included ... it would certainly have been preserved ...". At this time, they believed that melting beneath the debris was only 50 kg m$^{-2}$ a$^{-1}$ (Bull and Marangunic, 1968, p. 312), whereas I have subsequently measured close to 500 kg m$^{2}$ a$^{-1}$ of melting beneath the debris and have estimated that over broad areas it may rise as high as nearly 2000 kg m$^{-2}$ a$^{-1}$, so that the lack of snow 2 years after the avalanche is not conclusive evidence that it was never present.
However, Marangunic and Bull (1968, p. 392) later refer to a "bulldozed edge" and boulders in the snow at the distal edge where "a mixture of dirty avalanching snow, clean snow and boulders is invariably found". Although they observed (p. 393) "no mixing of snow and avalanche material in the supposedly bulldozed eastern lateral edge", Marangunic (1972, p. 56) subsequently identified such mixing in this region.

Close examination of the sequence of aerial photographs shows that snow apparently was extensively involved in the avalanche. Relative relief between ridges and grooves decreased rapidly during the first year although this trend quickly reversed in later years, as glacier ice has melted more quickly beneath the thinner debris below the grooves. All margins contained mixed snow and debris, with snow visible in early 1964, and collapse features, reflecting the melting of snow, visible in later photographs. May 1964 photographs show that the snow surface on Sherman Glacier to the east of the debris was detectably higher than the adjacent debris surface immediately west of the eastern avalanche margin. This height difference was absent at the end of the summer of 1964. The evidence suggests that snow was, to some extent stripped and molded by the passage of the avalanche, and not left completely undisturbed as others (Marangunic and Bull, 1968; and Shreve, 1966) have inferred.

Velocity and friction in the avalanche

Several attempts have been made to estimate the speed of the Sherman Glacier rock avalanche. Following Heim (1932, pp. 114-119),
Shreve (1966, Table 1) estimated a minimum speed of 52 m sec\(^{-1}\) for the Sherman Glacier rock avalanche from the minimum kinetic energy needed for the avalanche to overtop the spur beside Andres Glacier. Marangunić (1972, p. 89) suggested that rock avalanches might behave like dust-snow avalanches (as was earlier noted by Heim, 1882, p. 82), and thus a minimum sliding velocity could be calculated through a relationship suggested by Voellmy (1955). Voellmy suggested that high speed dust-snow avalanches are initiated when fine powder snow avalanches become airborne by overriding the air being pushed in front of them. When the transfer of kinetic energy from the avalanche to the air compresses the air to a pressure equal to or greater than the normal stress at the base of the avalanche, the overriding avalanche becomes fully supported by the compressed air. Thus Voellmy suggested that

\[
\frac{1}{2} \rho_{air} v^2 \geq \rho_{aval} g h
\]

where \(\rho_{air}\) and \(\rho_{aval}\) are densities of air and avalanche respectively, \(g\) is the acceleration of gravity, \(h\) is the thickness of the avalanche and \(V\) is the velocity of the air, and also approximately that of the avalanche. Marangunić calculated that the minimum sliding velocity at which the Sherman Glacier rock avalanche could have become airborne, if it had only been overriding air, was 226 m sec\(^{-1}\) or nearly twice the maximum speed attainable through a frictionless fall from Shattered Peak to Andres Glacier.
Following Kent's (1966) suggestion that rock avalanches may be gas fluidized by air passing up through them, Marangunić (1972, p. 89) used the fluidization theory of Leva (1959, p. 64) to obtain a minimum fluidization-by-air-velocity of 300 m sec\(^{-1}\) by using Voellmy's suggested mechanism (above) for supplying the air.

Through friction, the avalanche could not have reached the maximum freefall velocity of 120 m sec in the 760 m fall; for this reason, Marangunić suggested that the fluidizing medium overridden by the avalanche could not have been air alone. Because of the presence of snow on the glacier surface and snow-and-ice avalanche deposits beneath the rock debris, Marangunić suggested a more dense and viscous fluidizing medium of snow and air that could have accomplished fluidization with a fall velocity of only 95 m sec\(^{-1}\). This velocity is dependent on estimates of thickness, density, and particle size of the debris within the moving avalanche, and on the density and viscosity of the mixture of snow and air that was being overridden. Only the first two parameters have been estimated with any precision, and Marangunić also notes that Leva's work, developed empirically for industrial materials, may be inapplicable to the very coarse, heterogeneous debris.

Marangunić (1972, p. 93) suggested that other big rock avalanches were gas fluidized by air, alone, and thus should have had velocities that were about three times his estimate for the Sherman avalanche (95 m sec\(^{-1}\)) and they should be an order of magnitude thicker. This latter appears to be strikingly the case (see for example Shreve,
1968b, p. 27, 33 and 36). Marangunić suggested that particle size in the other air-fluidized avalanches was smaller, in order to account for the fact that velocity estimates for these generally ranged only around 50 m sec$^{-1}$. He noted that a choice of a modal particle diameter of 2.5 cm (such as that estimated by Shreve for the Blackhawk Slide in California, the only other avalanche for which data was available), instead of 7.5 cm as used for the Sherman Glacier rock avalanche, would give almost exact agreement between velocities calculated through his theory and the two estimates of velocities of rock avalanches from the reports of eye witnesses (45 m sec$^{-1}$ for the Frank Slide, McConnell and Brock, 1904, p. 8; and 50 m sec$^{-1}$ for the Elm Slide, Heim, 1932, p. 93). It should be noted, however, that Marangunić had no reliable estimate of modal particle size for debris at Sherman Glacier, and Shreve's more detailed estimate for the Blackhawk Slide "discriminates against the smaller clasts" (Shreve, 1968b, p. 28). It is perhaps even more significant that the basis for the estimates from the reports of eyewitnesses can be questioned.

Heim's (1932, p. 93) estimate for the Elm Slide stands in marked conflict with eyewitness reports that he had earlier presented (Heim, 1882, p. 87, and p. 95). One of Heim's probably more reliable witnesses to the Elm Slide, the local Sheriff, who watched the avalanche from start to finish from a position of safety, reported that it lasted several minutes, yet at 50 m sec$^{-1}$ it should have been over in less than a minute. Another witness, Kasper Zentner, a
cobbler who avoided the landslide only by fast footwork, turned and ran after he had watched the avalanche turn at the end of its free flight through the air, and begin to flow down the valley toward him. By running as fast as he could he was able to avoid the debris by about a meter. He made repeated measurements, as best he could over the debris, of the distance he had run and estimated that he had gone 290 to 300 paces in about 40 sec. These two reports are consistent with a mean velocity for the Elm Slide of only about 20 m sec\(^{-1}\). This value agrees well with the estimate of velocity of the avalanche (30 m sec\(^{-1}\), Heim, 1932, p. 114-119) at the end of its free flight through the air when it should have attained its maximum velocity before surging up the opposite slope.

Thus we can see the difficulty of trying to match a theoretical estimate to one that is supposedly based on observation. A better approach is to determine the velocity of the avalanche directly from the deposits that it left. Consideration of the forces acting on a particle in motion suggests a method that uses the geometry of particle paths. Forces acting on a particle in motion across a smooth slope are friction and gravity. Friction acts only in a direction opposed to motion and cannot produce a curved path. Gravity has a component in the downslope direction and any particle in motion across a slope is subject to a gravity force in the downslope direction and will consequently describe a curved path on the slope until the motion is parallel to the slope direction. Whether or not this is also the behavior of an aggregate of particles will not be considered here.
Simple mechanics requires that a particle of mass $m$ and velocity $v$ move in a circular path (which is a reasonable approximation to the curved path if we consider only a very small length of path) of radius $r$ when subject to a radial force $F_r$ where

$$F_r = \frac{mv^2}{r}$$

The force producing curvature is a gravity force $F_g$ acting downslope (slope $\alpha$)

$$F_g = mg \sin \alpha$$

and has a component in the radial direction that is equal to $F_r$. Thus

$$F_r = \frac{mv^2}{r} = mg \sin \alpha \cos \theta$$

where $\theta$ is the angle between downslope direction and a normal to the curved path. Hence

$$v^2 = rg \sin \alpha \cos \theta$$

For very small lengths of path, friction plays no role in the model, because the curvature is a result of forces acting perpendicular to the velocity vector whereas the force of friction is parallel to this vector.

In the southwestern sector of the avalanche debris on Sherman Glacier, about 1.5 km from the distal margin, flow lines expressed
as grooves, are aligned directly downslope. Within 500 m downflow, local slope makes an angle of about 25° to flow direction, and the grooves begin to describe an arc with a radius of 1100 m (Fig. 3 and 6) until, about 700 m further, flow is again aligned down slope. Local slope in the region is about 1 in 30. The speed of the avalanche at beginning of curvature was therefore about 12 m sec⁻¹. This value is an estimate for a particular point in time and space and not an average for an entire avalanche.

Implicit in applying a single particle analysis to an aggregate of particles, such as avalanche debris, is the assumption that particle-particle interactions are statistically uniform on either side of each particle. In the general case, this assumption would seem to be justified.

Following Heim (1932), Shreve (1966, p. 1640) estimated a "coefficient of friction" for the Sherman Glacier rock avalanche of 0.22, being the tangent of the mean slope from source to distal margin (this friction model has been treated somewhat more generally by Scheidegger, 1973, p. 232) who mistakenly cites Shreve’s Sherman value as 0.19 in his Table 1 and Fig. 2, however see below). A better estimate of a mean coefficient of friction (0.19) can be made from the tangent of the mean slope of the locus of displacement of the center of mass of the avalanche from Shattered Peak to Sherman Glacier, but this estimate is a mean value for the entire avalanche which initially moved over rock before reaching the glacier to move over snow and ice.
In addition, friction estimates based on this rigid-block model for avalanches combine both internal and external friction.

From the velocity recorded by the debris margin \( (V_1 = 0 \text{ m sec}^{-1}) \), and the velocity calculated above \( (V_2 = 12 \text{ m sec}^{-1}) \), a coefficient of friction \( (\mu) \) for the avalanche moving over snow and ice can be calculated using these velocities, the distance travelled \( (s = 1 \text{ km}) \), and the loss in height \( (100 \text{ m}) \), using the relationships:

\[
2as = V_1^2 - V_2^2
\]

\[
a = g \sin \alpha - \mu g \cos \alpha
\]

where \( g \) is the acceleration due to gravity and \( \alpha \) is the surface slope. The debris apparently was decelerated by friction at \(-1.1 \text{ m sec}^{-2}\), which gives an effective coefficient of basal kinetic friction of 0.11.

Kinetic friction also can be estimated for several other localities at the western margin of the avalanche where narrow streams of debris were just able to move beyond the general limits of the remainder of the avalanche debris. In the region of these localities, the slope of the surface was close to the limiting slope-angle where friction is almost equalled by the component of gravity down the slope. The tangent of the limiting angle, at which debris just comes to rest, is the coefficient of kinetic friction.

One stream of debris against the southern edge of Sherman Glacier, came to rest after it had travelled 450 m while dropping 50 m
which gives a coefficient of kinetic friction \( \mu \) of 0.111, while about 500 m to the north, another stream dropped 30 m in 275 m (\( \mu = 0.109 \)). Thus the coefficient of kinetic friction of the Sherman Glacier rock avalanche at the end of its run over ice and snow was 0.11. It is probable (see p. 83) that, at this stage of the avalanche, internal deformation, and hence loss of energy through internal friction, was negligible, and that these friction values are of simple basal kinetic friction.

Robin and Barnes (1969, p. 972) found coefficients of sliding friction for a steel ball ploughing across a smooth ice surface that ranged from 0.095 to 0.15 depending on temperature, with a range of from 0.10 to 0.13 near the melting point. Weast (1969, p. F17) lists the coefficient of friction of waxed hickory skis sliding slowly on wet snow as 0.14. Whether or not either of these situations is a good analogy to rock debris moving over ice and snow will not be considered.

Because the avalanche path was largely snow over ice at a temperature near 0°C (initially perhaps -3°C but warmed to the melting point by frictional heating, see p. 98), the coefficient of kinetic friction could hardly have been higher than the value calculated above over most of the path. If it was constant over the entire path, the velocity of the avalanche at any point along the path may be calculated through the simple equations of motion, viz.
\[ v_1^2 - v_2^2 = 2as \]

\[ v_2 = v_1 + at \]

\[ s = v_1^2 + \frac{1}{2}at^2 \]

The distance from where the avalanche struck the spur by Andres Glacier to the westernmost distal lobe is about 3.5 km and the average slope is 4.5°. If the debris had a frictional acceleration of -1.07 m sec\(^{-2}\) and a gravitational acceleration downslope of 0.80 m sec\(^{-2}\) (down a slope of 4.7° for the path of the center of mass of the thinning debris which was about 30 m thick by the spur, Fig. 14a), it traversed the 3.5 km in 163 sec and had an initial velocity of about 44 m sec\(^{-1}\). From conservation of energy, it can be calculated the avalanche needed a minimum velocity of 37 m sec\(^{-1}\) to overtop the spur (the spur is 100 m high at this point but the debris was at least 30 m thick), so it apparently had some 7 m sec\(^{-1}\) to spare.

At the foot of Shattered Peak, Andres Glacier is 1.3 km wide and surface slope across the glacier is close to horizontal. If the debris thinned by about 20 m in the crossing, it probably crossed this glacier in about 23 sec with an initial velocity of about 67 m sec\(^{-1}\). A frictionless fall from Shattered Peak would have given the debris a velocity of 96.5 m sec\(^{-1}\) at the base of the peak for the 475 m drop in the center of mass (\(S = \frac{1}{2}gt^2\), and \(v = gt\)).
From the relationship between kinetic energy and velocity \( (E = \frac{1}{2} mv^2) \), about 52 per cent of available energy apparently was lost in the initial fall to Andres Glacier. In the 475 m drop, the debris slid about 780 m down a 40° slope. If the energy loss was due to friction at the base, then the apparent coefficient of kinetic friction was 0.41, and the debris accelerated down the slope at 3.22 m sec\(^{-2}\) to reach Andres Glacier in 30 sec after sliding was initiated \( (S = \frac{1}{2} a t^2 \text{ where } a = g \sin 40° - \mu g \cos 40°) \). The expected coefficient for rock sliding on rock is in the range of about 0.45 to 0.60, which suggests that this early phase of the avalanche involved a gradual transition from a simple sliding block to a mass moving by a combination of sliding and rolling (p. 35).

Thus, the Sherman Glacier rock avalanche travelled a little more than 5.7 km in 216 sec at an average speed of 26 m sec\(^{-1}\), and its average speed was very similar to my estimate (20 m sec\(^{-1}\)) from reports of witnesses to the Elm Slide, and to Müller's (1964, p. 198) estimates (25-30 m sec\(^{-1}\)) for the Vaiont slide in Italy.

In overtopping the spur, the debris in the Sherman Glacier rock avalanche shattered a part of a 40 m high face and stripped vegetation that lay beneath a snow cover, so that 0.11 is probably an inappropriate coefficient of basal friction for that 400 m of path. Only about a quarter of the mass went over the spur. If this was the topmost quarter of the block on Shattered Peak, it had already fallen 675 m to Andres Glacier to reach a maximum speed of 83.5 m sec\(^{-1}\), and had a speed of 66.6 m sec\(^{-1}\) at the foot of the spur.
If it barely overtopped the spur, the average coefficient of friction on the stoss side would have been 0.31, which may be an appropriate mean coefficient of friction for rock sliding on, and stripping, a snow-covered hillside.

The center of mass of the avalanche debris now lies about 2.5 km from the base of Shattered Peak so that the mean coefficient of friction for 780 m at 0.41 and 2500 m at 0.11 is 0.18 which is close to the expected average value (0.19, p. 57) for the observed present distribution of debris.

In the preceding paragraphs, coefficients of kinetic friction beneath the Sherman Glacier rock avalanche were estimated to be 0.11, and 0.41 which seem reasonable values for a substance sliding over wet snow and rock, respectively. If, just as a reasonable estimate, the debris had lost about half of its initial energy in falling from Shattered Peak it would have hit Andres Glacier at 68 m sec⁻¹. If the debris front had then slid in contact with snow it could have stopped moving when it reached the present position of the avalanche margin, only if the effective mean coefficient of friction had been about 0.11.

Shreve (1966, p. 1642) considered the possibility that dry snow or a slurry of snow and air could have been a basal lubricant and must have been involved (Shreve, 1968c, p.1654 used a basal layer of snow to provide a gradient in permeability that he found necessary for an air layer to persist) but instead he proposed that the Sherman
Glacier rock avalanche was air-layer lubricated in the manner that he had postulated for the Blackhawk slide of California; that is, it fell upon and trapped a cushion of compressed air upon which it traversed gentle slopes with little friction. He did not, however, suggest a value for the friction and made no attempt to estimate it. The coefficient of kinetic friction of greased surfaces is about 0.05, which suggests that the coefficient for a surface lubricated by an almost frictionless layer of air should be considerably less.

An estimate of the coefficient of kinetic friction of an air layer can be made if one considers a very simplified model of an avalanche moving as a flat plate over a thin fluid layer of air above a flat substrate. In the simplest case of Newtonian flow, the effective coefficient of kinetic friction must be variable because the force \( F \) required to keep two surfaces of area \( A \) moving with relative velocity \( V \) when they are separated a distance \( d \) by a fluid of viscosity \( \eta \) is

\[
F = \frac{AV}{d\eta} = ma
\]

This formula is also a reasonable approximation in turbulent flow if Newtonian viscosity is replaced by kinematic eddy viscosity. The effective acceleration (a) of the Sherman avalanche (mass m) due to friction on an air layer might thus have been about \(-1.65 \times 10^{-5} \text{ m sec}^{-2}\) for extreme values of parameters \((m = 1 \times 10^{10} \text{ kg}, V = 100 \text{ m sec}^{-1}, A = 8.25 \text{ km}^2, d = 0.1 \text{ m})\) for an order of magnitude estimate,
\(\eta_{\text{air}} = 2 \times 10^{-5} \text{ newton sec m}^{-2}\) if deformation did not become turbulent. This is equivalent of a coefficient of kinetic friction of about \(1.7 \times 10^{-6}\). At velocities of the order of \(100 \text{ m sec}^{-1}\), however, deformation of the air layer must have been turbulent (Reynolds number of \(5 \times 10^5\)) and an empirical kinematic eddy viscosity should be used. Even if the effective viscosity rises two to four orders of magnitude, which would be a reasonable estimate for the change, the effective coefficient of friction only varies by an equivalent amount and is still very low, and for most purposes, effectively zero.

If the avalanche had been lubricated by a layer of air at any time it would have accelerated over the slopes of the glacier surface and would have been unable to stop within the known confines of the avalanche, because, when it stopped being airborne, it could only have slid on snow (at \(\mu = 0.11\) it would have to start sliding at the foot of the peak for the velocity estimated above.*). Thus, the Sherman Glacier rock avalanche could not have been lubricated by a layer of air as envisioned by Shreve: it must have slid on a snow-covered surface along its entire path, as did the Puget Peak avalanche (Hoyer, 1971, p. 1282). Shreve (1966, p. 1641) noted that a basal layer of scraped-up snow, mud, or alluvium was present in all five of the "landslides of the Blackhawk type" that he had studied; so it is apparent that they all slid on their substrate.

* Note that \(\mu = 0.11\) was the value for the last 275 m of travel.
Internal strain and internal friction during avalanching

At 5:36 pm 27 March 1964 a block of rock on Shattered Peak was about 400 m long, 250 m wide and 100 m thick (Marangunić, 1972, p. 79); within the next 3 to 4 minutes of continued shaking the block fell off the mountain, and, 3.5 minutes later, lay spread over an area of 8.25 km$^2$ to a depth of about 1.65 m. Thus, strain rate within the debris was about 0.04 sec$^{-1}$ (this is studied in more detail below) and the debris apparently responded to the stress of avalanching as a viscous "fluid". Because the peak had previously survived intact for thousands of years, it is clear that internal friction was greatly reduced below that of the original block. The presence of transverse fissures and folds demonstrating brittle and flexible behavior shows that this "fluid" followed no simple viscous flow law.

Deformation of the debris during the Sherman Glacier rock avalanche can be studied by means of a simplified plastic model of avalanches. In this model, the deforming block fell to the base of Shattered Peak on Andres Glacier, then spread laterally under its own weight. Because of the steep northerly slope of Andres Glacier, southward flow was blocked by the glacier so that lateral spreading of the debris forced most of the avalanche to flow down the glacier to spread over the broad surface of Sherman Glacier.

If we consider an avalanche model in which volume remains constant, that is, in which dilation to some optimum porosity has already occurred, then average thickness is a simple function of the area covered by the avalanche. We have already determined an
effective coefficient of friction for the avalanche and can therefore compute an effective stopping force over any part of the glacier surface from surface slope and the friction coefficient. Because we know the final velocity of the avalanche margin (0.0 m sec\(^{-1}\)), the final mean thickness of the avalanche (1.65 m), and the total area covered (8.25 km\(^2\)), we can back calculate the velocity of the margin, the mean thickness, and area covered by the avalanche for any chosen position of the avalanche margin prior to its coming to rest. Not only can this be done for the entire avalanche, but it can also be done for individual debris trains as delineated by surface grooves and ridges which are apparently a measure of flow direction, and hence define flow lines.

Because debris trains are too numerous to be studied individually, I have arbitrarily divided the Sherman Glacier rock avalanche into eight groups of debris trains (Fig. 9) and traced the deformational history of the seven of these trains that flowed down Andres Glacier to Sherman Glacier. These seven sets of debris trains were divided along clearly visible and reproducible flow lines on a 1:10,000 orthophotograph. The orthophotograph includes only a small portion of Andres Glacier, and flow lines are only poorly traceable there anyway; for these reasons the estimation of widths of flow units was started at an arbitrarily selected datum near the 450 m contour on Andres Glacier (Fig. 9) and flow lines were assumed parallel in the 900 m length of rock avalanche south of this datum.
Widths and elevation-drop of the sets of debris trains were then measured at 200 m intervals along the flow lines from this datum to the margins of the deposit. From these data, mean thicknesses of the trains were calculated for successively 200 m shorter lengths of the trains until the datum thickness was reached for each set of trains (Fig. 13). No attempt was made to compensate for a systematic variation in thickness that must have existed along the flow lines during deformation of the "plastic" debris. Only one compensation was made in the calculation for the difference in motion between that of a rigid block and that of a plastic block: this was, that after the initial calculations of thickness at various avalanche margin positions, the surface slope of the glacier was modified to take into account the drop in height of the center of mass of a moving column of debris as the column thinned, for the purposes of calculating the component of gravitational acceleration.

The crosscutting relationship of grooves and ridges near the datum (Fig. 1) suggests that the easternmost debris trains had stopped flowing before the western debris trains ceased to move at the datum section, so that it is uncertain if the thicknesses in Fig. 14a represent a simultaneous profile. However, the calculated distribution of thicknesses across the datum line, at the time that the avalanche crossed this line, becomes a remarkably uniform level surface (Fig. 14b) when plotted as elevation of the avalanche surface by addition of the transverse profile of Andres Glacier at this section. This level surface suggests that the debris thickness,
Fig. 13. Variation in thickness with time at the datum profile in the Sherman Glacier rock avalanche, for selected sets of debris trains (see Fig. 9 for locations of the numbered trains).
Fig. 14. a) Calculated thickness of debris across a datum profile near the 450 m contour on Andres Glacier (Fig. 7).
b) Elevation of the surface of the front of the Sherman Glacier rock avalanche as it passed the datum line.
The glacier surface elevation is from a 1:10,000 map by H. Brecher.
at that point in time, was controlled by fluid properties of the flowing loose debris. It also suggests that the profile is one of simultaneous thickness of incompressible debris, because it is unlikely that the timing and compressibility varied in such a way as to produce this remarkable result.

Because of the lack of lateral gradient on the surface of the avalanche, it is probable that the flow units maintained constant widths with time in this region. Thus, the use of the still present grooves and ridges to define flow lines early in the avalanche is probably valid even though the features themselves may not have been present at that time.

Two of the units (5 and 6, Fig. 9) end beyond the limits of contouring on the orthophotograph so that their velocities can not be calculated with any certainty (particularly as they stopped with an apparently positive net acceleration over the last 200 m of travel; this is because they rode up the north wall of the valley with a higher coefficient of basal friction). The calculated velocities, and hence estimated times, for the other units probably contain slight systematic errors because of a longitudinal gradient in thickness along flows that would have influenced the debris velocity. The error due to the gradient in thickness of the debris is most significant for the early stages of the avalanche when the gradient in thickness was greatest. In later stages the error should be insignificant. The assumption of incompressibility introduces
no significant systematic error even if porosity grow continuously throughout the avalanche (p. 76), because the debris density-thickness product is used in calculation, and the total change in porosity was only about 25 per cent. However, if porosity had increased continuously, the transverse slope of the debris surface at the datum section would not have been horizontal because compression changes thickness in direct proportion to the thickness.

The calculated velocity profile across Andrea Glacier at the datum line (Fig. 15) is consistent with the calculated thicknesses and suggests also that the data are for a near-simultaneous profile. In addition, the duration of flow of the units (the time of ending of change in thickness in (Fig. 13), suggests that simultaneous flow through the datum section is also consistent with the observed crosscutting relationships of units 6 and 7 (Fig. 9).

The change in thickness at the datum section shows an expected rapid decrease in the rate of thinning of the debris with time (Fig. 13). When this thinning is expressed as vertical strain rate (Fig. 16) significant discontinuities are apparent in the curves.

The cause of these discontinuities appears to be a change in the rheology of the flowing debris. If the vertical strain rate (which is the effective strain rate if there is no lateral strain and the debris is incompressible) at the datum profile is plotted against basal shear stress at this profile (Fig. 17) it is apparent that the debris responded differently at high shear stress, than at low shear stress.
Fig. 15. Velocity profile of the Sherman Glacier rock avalanche when it crossed the 450 m contour line on Andres Glacier.
Fig. 16. Variation in vertical strain rate with time for four debris trains in the Sherman Glacier rock avalanche. (units as numbered in Fig. 9)
Fig. 17. Variation in effective strain rate with change in effective stress within the Sherman Glacier rock avalanche (for location of numbered flow units, see Fig. 1).
At low stress, below 0.1 bar, the debris behaves as a simple Bingham plastic and obeys a visco-plastic flow law with a viscosity of between $0.4 \times 10^6$ and $1.6 \times 10^6$ poise and a yield stress of 0.02 bar (see also p. 85). The probability that debris-trains vary slightly in thickness from each other and also along their lengths means that the calculated basal shear stresses and strain rates may have small systematic errors and the actual values of viscosity and yield stress are uncertain, but probably of the correct order of magnitude. The existence of a yield strength for the debris requires that the eventual thickness of the debris at any point over the avalanche deposit, be a function of the surface slope (p. 85). This systematic variation in thickness has not been investigated, but the calculated yield strength of $1.6 \times 10^3$ Newton m$^{-2}$ (p. 86) requires that the debris thinned to about 0.83 m at the datum section rather than to the mean of 1.65 m. The effect of this variation is to introduce systematic errors into the calculated thickness, strain rates and shear stresses, and hence into the viscosities and yield stress, but again the estimated results are probably of the correct order of magnitude.

At higher stresses, above 0.1 bar, one debris train exhibits an apparently sudden increase in viscosity from about $1 \times 10^6$ poise to about $40 \times 10^6$ poise, while another behaves as a dilatent fluid and has a smoothly increasing viscosity with increasing stress. It appears that the apparent viscosity of the avalanche debris
becomes dependent on the strain-rate itself at shear high strain-rates, although this increase in apparent viscosity might also be a function of increasing normal stress due to thicker debris, or if the debris is a truly thixotropic material, it may represent a decrease in the apparent viscosity with time as the debris structure progressively breaks down.

The high shear stresses were experienced early in the avalanche when gradients in thickness change in time and space were at their greatest values. Thus a small part of the apparent change in rheology may be due to the systematic errors in estimation of strain rate and stress but it is not entirely due to these errors. Dilation of the debris during the avalanche and the growth of porosity and permeability may also have caused some of the change. This dilation is a necessary consequence of strain of a granular substance and some degree of permeability is necessary to permit an inflow of air to fill the growing void space.

**Dilation during avalanching**

Average density of rock in the debris is 2.69 m$^3$ m$^{-3}$ (Plafker, 1968, p. 378) while the bulk density is only 2.0 Mg m$^{-3}$ (Marangunic' 1972, p. 79). Thus, the avalanche brought about a 25 per cent dilation of the rock volume, which apparently inhaled a net $2.5 \times 10^6$ m$^3$ of air. The dilation probably accounts for the rarity of bruising of boulder edges during transport that was noted by Marangunic' and Bull (1968, p. 383), but it may also account for much more.
If the avalanche was unable to trap and compress a layer of air beneath it, all of the air that was drawn in would have to be inhaled from above, where pressure is atmospheric. Thus, air pressure in the avalanche was below atmospheric pressure during most of the time of travel (Fig. 18). Because some settling of debris occurred as the avalanche came to rest, when the packing of the debris collapsed from being more-or-less cubic to permit internal deformation, to being more closely tetragonal, somewhat more than \(2.5 \times 10^6\) m\(^3\) of air was inhaled during flow, but the additional air was expelled as the avalanche stopped and settled (Fig. 18). The observation of expulsion of air at the end of an avalanche is thus no indication of high internal or basal air pressure during the avalanche (cf. Kent, 1966, p. 82).

The rate of inhalation of air is controlled by three factors that may also play a limiting role in the strain-rate and spreading of avalanches: thickness, porosity and permeability. The rate of growth of porosity through strain in fragmentation is limited by the rate at which air can fill void spaces, and this is limited through Darcy's law by permeability of the debris which increases with time in the avalanche and by debris thickness that decreases with time (Fig. 13). By limiting the rate of growth of void space, the partial vacuum between fragmenting clasts may have maintained the high viscosity of the debris that decreased with time at a rate determined by the growth of permeability in the avalanching mass.
Fig. 18. Possible variation in basal pressure within the Sherman Glacier rock avalanche: (a) in the dual-viscosity Bingham plastic model proposed here; (b) in the air-layer lubrication model previously proposed. The time scale in the air-layer model has been adjusted to that of the Bingham plastic sliding and rolling on snow and ice.
Darcy's law

\[
q = \frac{k A \Delta P}{\eta L}
\]

relates the rate of flow of a fluid \((q \text{ cm}^3 \text{ sec}^{-1})\) to the permeability \((k \text{ darcy's})\), the cross-sectional area of the permeable medium \((A \text{ cm}^2)\), the pressure drop along the path of flow \((\Delta P \text{ bar})\), the viscosity of the fluid \((\eta \text{ poise})\), and the length of path of flow \((L \text{ cm})\).

The Sherman Glacier rock avalanche inhaled about \(2.6 \times 10^6 \text{ m}^3\) of air in about 200 sec over a time averaged area of \(2.75 \times 10^6 \text{ m}^2\). Thus, air flowed into the avalanche at an average speed of \(4.6 \times 10^{-3} \text{ m sec}^{-1}\) (at this rate the Newtonian viscosity of air is valid, but early in the avalanche, the rate may have been very much higher, and the flow turbulent, and an eddy viscosity should be used). Shreve (1968c, p. 654) suggested that 1 darcy is a reasonable estimate of permeability of poorly sorted rock avalanche debris. The viscosity of air is about \(1.7 \times 10^{-4} \text{ poise}\). The average pressure gradient \((\Delta P/l)\) in the Sherman Glacier rock avalanche was thus about \(7.7 \times 10^{-5} \text{ bar m}^{-1}\). The yield stress of the debris is 0.02 bar (p. 75 and p. 86) so that an internal air pressure difference of 0.02 bar between the surface and the basal debris might inhibit flow. This pressure difference would occur at a depth of 2.6 m in the debris at the estimated permeability of 1 darcy and the mean inflow of air.
A discontinuity in vertical strain-rate (Fig. 16) occurs at about 15 sec for flow unit 1, 45 seconds for 2, and 30 sec for 6. These times correspond to estimated debris thicknesses of 12 m, 3 m, and 2 m for the respective flow units (Fig. 13).

Variation in permeability or air flow may account for the differences between these estimates of the limiting thickness for the flow transition and the predicted average for the whole avalanche (2.6 m). Air flow of $5.6 \times 10^{-3}$ m sec$^{-1}$ or permeability of 0.8 darcy would account for the 3 m thickness, and air flow of $3.7 \times 10^{-3}$ m sec$^{-1}$ or permeability of 1.2 darcy would account for the 2 m thickness. The 12 m thickness would require a much smaller air flow ($0.9 \times 10^{-3}$ m sec$^{-1}$) or a larger permeability (5 darcy). Textural variations visible over the debris suggest that minor permeability variations are probable but no measurements have been made. It is possible that a much higher permeability of debris against the spur could be a result of the tortuous path of some of this debris in sliding up the side of the spur and down again before flowing down Andres Glacier. This might have caused greater fragmentation and hence possibly greater porosity, which is the most significant factor in determining permeability. It is also possible that this backflow of debris may have continued to feed debris into the westernmost debris stream during the avalanche, so that the estimated thicknesses for that stream are over-estimated, in which case the horizontal debris-surface profile (Fig.14b) is incorrect.
In addition to the 25 per cent dilation in volume of the debris indicated by the change in bulk density, there is also a dilation indicated by transverse fissures (p. 41). The rigid brittle behavior of the debris in fissuring may have occurred at a late stage in the avalanche, when the shear stress within it had dropped below the yield stress of 0.02 bar.

The relative clast motion in the debris that is evident from the obvious strain and dilation would, by itself, have reduced the internal friction within the avalanche because the coefficient of sliding friction can be as much as 30 per cent less that that of static friction. In addition, Hsü (1975, p. 136) has suggested, actually following Trawinski (1953), that fine debris between larger blocks may contribute to the loss of internal cohesion in flowing rock debris. Botterill and Vander Kolk (1971, p. 70) however, note that although the addition of fines reduces the observed viscosity of gas fluidized debris, the action is probably not through fines acting as "ball bearings" as Trawinski suggested, but rather is through the decrease in permeability as fines enter and clog pore space.

Heim (1882, p. 82 and 83) noted that the essential difference between a simple blockfall and a debris flow was that, whereas an isolated block travelled in a zig-zag bounding path through elastic impacts with its surroundings, an aggregate of blocks, in which individual blocks are small compared to the mass of the whole,
behaved quite differently, because the individual stones are confined to bouncing back and forth between their neighbors and only the outer stones may fly out. In this way, kinetic energy is exchanged between particles by elastic collisions, and the aggregate behaves as a liquid, and flows in continuous curves rather than following a zig-zag course.

I suggest that the greater part of the loss of internal cohesion in flowing rock debris is through the statistical separation of the individual clasts through elastic rebounds from these constant collision with neighbors. This separation would greatly reduce internal friction, and probably even maintain at least the minimum dilation to cubic-closest-packing that is necessary for strain to occur in particulate media. The role of fines in such an "elastic-energy dilated" medium is uncertain because the fines themselves must always be present as a result of attrition of clasts in such a high energy environment, (probably the highest energy environment of all natural processes of sediment transport) and a rock avalanche of uniform sorting can never be observed. By filling pore space between larger clasts, fines serve to decrease permeability and hence decrease the rate of flow of whatever gas is present in the void space. In addition, or perhaps alternatively, by decreasing the porosity of the debris, fines decrease the mean-free-path of clasts and thus increase the frequency of elastic collisions. In this model of loss of internal cohesion and reduction of internal friction, the
elasticity of the clasts and hence of the parent rock must be the controlling factor in the reduction of internal friction, and the apparent viscosity of the flowing debris should be a function of the elasticity of the rock, and the amount of energy stored as elastic energy at any one time in the avalanche.

The internal friction is the factor controlling the apparent viscosity of the debris at low shear stress (but above 0.02 bar), and it is possible that a transition from kinetic to static friction is the cause of the 0.02 bar shear strength.

From the relationship between force, viscosity, depth, and velocity (p. 63), and the calculated shear strength, the top of a debris column 1.65 m thick should move $3.3 \times 10^{-3}$ m sec$^{-1}$ faster than the base. This vertical gradient in horizontal shear strain-rate of 0.02 sec$^{-1}$ means that clasts 0.05 m in diameter, slide past each other at only 0.001 m sec$^{-1}$. This speed may be too small to keep the clasts moving relative to each other so that deformation stops. This process or perhaps just the loss of a net storage of elastic energy may account for the sudden stop observed in some rock avalanches (Buss and Heim, 1881, pp. 142-147). This "sudden" stop, however, that observers have remarked upon, may be accounted for in other ways, not related to the mechanism of avalanches. That is, a body experiencing uniform deceleration simultaneously experiences successively greater relative reductions in velocity so that to the observer it appears to slow at an increasing rate.
Because

\[ V_2 = V_1 + a t \]

if follows that

\[ \frac{V_2 - V_1}{V_1} = \frac{a t}{V_1} \]

and therefore, for equal increments of time, the relative rate of decrease in velocity increases as the velocity decreases.

Botterill and Van der Kolk (1971) note a succession of changes in behavior of a gas-fluidized solid as the rate of flow of gas increases. At low flow of gas, the solid becomes a Bingham plastic that dilates as gas flow increases, with consequent decrease in apparent viscosity. Above an optimum gas flow, the fluidity of the plastic is such that it appears to boil as bubbles of gas rise to the surface. At even higher rates of gas flow, the entire mass is lifted off from its bed on a layer of gas whose pressure is able to support the column.

The Sherman Glacier rock avalanche apparently behaved as a simple Bingham plastic and thus perhaps was gas fluidized. Marangunic (1972, p. 89) suggested that the Sherman Glacier rock avalanche was fluidized by a pressurized air-snow slurry rising through the debris from a layer of trapped air and snow that the avalanche had been able to override because of its high velocity. He found that an avalanche velocity of 95 m sec\(^{-1}\) was needed to initiate this
process and it may not be a possible mechanism at the velocities calculated above. He also reported no disseminated snow and ice through the debris that could represent a deposit from this slurry; on the contrary he identified sharply defined contacts between stratified avalanche snow and rock avalanche debris on Andres Glacier, where little melting could have taken place prior to his observation. In addition, the presence of a layer of trapped air implies that the fluidization of the debris was already beyond the stage at which bubbles of air rise violently to the surface, yet there is no indication in the deposit of the fluidity that this implies. Instead, it appears to have slid as a thin flexible sheet with very high viscosity, with internal shear inclined at about 45° to its base (that is, flattening under its own weight), with deformation principally confined to sliding at the debris-snow interface, and with little turbulence.

Yield strength of the avalanching sheet

If an avalanche comes to rest when the shear stress at the base if reduced to the shear strength of the moving debris, then its yield strength $(k)$ can be simply calculated from debris thickness $(h)$, density $(\rho)$, and surface slope $(\alpha)$ through the formula

$$k = \rho g h \sin \alpha$$

where $g$ is the acceleration due to gravity. For an average segment
of debris on Sherman Glacier, \( \sin \phi \) is 0.05, and the product of thickness and density is 3.3 Mg m\(^{-2}\). Thus, the yield strength of the Sherman Glacier rock avalanche was about \( 1.6 \times 10^3 \) Newtons m\(^{-2}\) (about 0.02 bar).

Yield strength can also be determined through the sizes of boulders transported by the avalanche (Johnson, 1970, p. 486). One boulder carried by the Sherman Glacier rock avalanche was about 18 m high and was buried about 3 m in a thick ridge of debris. The approximate formula for yield strength,

\[
k = c \left( \frac{\gamma_r - \gamma_d}{n} \right)
\]

(Johnson, 1970, equation 13.1.2), where \( c \) is the half height of the boulder, \( \gamma_r \) and \( \gamma_d \) are unit weights of the rock and debris and \( n \) is the reciprocal of the fraction buried, gives a yield strength of about \( 4.5 \times 10^3 \) Newtons m\(^{-2}\) which is in reasonable agreement with that from surface slope. Johnson (1970, p. 490) points out that yield strengths calculated from sizes of supported boulders are higher than that from surface slope because boulders are not simply supported by the yield strength of the debris. The hardening may also represent an increase in strength with increasing normal stress due to the weight of the boulder.

The depth of burial utilized above is an estimated one. No measurements of boulder sizes and depths of burial have been made on the Sherman Glacier rock avalanche. Because ice melts more rapidly beneath loose debris than beneath solid boulders, because of advection of heat by air flow, percolation of rain water,
and condensation of moisture in the permeable debris, depths of burial of boulders have decreased with time and should have been measured in the first year after the avalanche to obtain accurate estimates of the debris strength. In addition, yield strength could have been calculated from the sizes and shapes of grooves, ridges, and cones on the debris surface, but none were measured soon enough after the avalanche to be of use, and they change rapidly with time, through rainwash and melting of the snow that was involved.

For comparison with the yield strength of the Sherman Glacier rock avalanche, the strength of a rock avalanche of more "normal" thickness, the Blackhawk slide, whose thickness varies from 15 m to 30 m for variation in size of the surface slope of 0.063 to 0.025 (Shreve, 1968b, Plate 1), ranges from $14 \times 10^3$ to $18 \times 10^3$ Newtons m$^{-2}$ (about 0.2 bar) and Johnson’s (1970, p. 489) calculations of yield strengths of a water-saturated debris-flow range from $1.7 \times 10^3$ to $5 \times 10^3$ Newtons m$^{-2}$. The Sherman Glacier rock avalanche thus appears to have had a strength more typical of mud flows than of rock avalanches.

**Turbulence**

A notable characteristic of the Sherman Glacier rock avalanche is the apparent lack of mixing shown by transverse lithological and textural zonation, by the "jigsaw puzzle effect", and by debris cones and abundant pieces of old surface of the mountain still present on the surface of the avalanche. If turbulence was present in the avalanche, it was certainly minimal.
The velocity distribution with depth in the Sherman Glacier rock avalanche (Fig. 19) can be calculated from the shear stress, which is the cause of deformation, and the equation for deformation in a viscous substance (p. 63). Thus

\[ v = \rho g h d \sin \alpha / \eta \]

where the velocity \( v \) of the surface, relative to the material at depth \( h \), is given by the product of the depth, the density \( \rho \), the gravitational acceleration \( g \), the thickness \( d \) over which deformation takes place (\( d = h \) for Newtonian viscosity, but \( d < h \) for Bingham viscosity, because of the non-deforming layer at shear stress \( < 0.02 \) bar), the sine of the surface slope \( \alpha \), and the reciprocal of the viscosity \( \eta \).

For debris 24 m thick with surface slope of 6° at the datum profile on Andres Glacier, the expected velocity relative to the base is 0.25 m sec\(^{-1}\) at the top of the higher viscosity layer for an apparent viscosity of 4 x 10\(^7\) poise. The stress in the top 3 m is below the critical stress for the viscosity transition (Fig. 17) and the layer should deform with a viscosity of 1 x 10\(^6\) poise, so the expected increase in velocity over this layer is 0.13 m sec\(^{-1}\) because it is confined to the layer where the shear stress is > 0.02 bar. At shear stresses < 0.02 bar, no shear strain occurs. This provides for a velocity distribution as shown (Fig. 19). The velocity distribution in debris of other depths and surface slope is similar.
Fig. 19. Predicted velocity distribution with depth in a Bingham substance with dual viscosities and a yield stress of 0.02 bar. Viscosity is relative to the moving base. Absolute velocity of the base is about 54 m sec$^{-1}$. 
Reynold's numbers

\[ \text{Re} = \frac{\rho \cdot v \cdot d}{\eta} \]

for the various deforming layers in this velocity profile (Fig. 19) are 0.0 for the non-deforming surface layer, \(6 \times 10^{-3}\) for the 2.2 m thick slowly deforming middle layer, and \(3 \times 10^{-3}\) for the more slowly deforming lower layer, and this latter value decreases linearly as the layer thickness decreases. At the base of the debris where the absolute velocity is about 54 m sec\(^{-1}\), there is a very high shear-strain rate that probably takes place within a mean clast diameter, if many clasts are rolling, and may occur across a very narrow zone if the clasts have gathered a basal sole of ice. Here, the Reynolds number should not be greater than 0.08 if air from the snow provided sufficient airflow for the lower viscosity to be used. At this scale of layer thickness, however, the viscous model of deformation of the particulate medium is inappropriate, and the rolling clasts are also inconsistent with the model of viscous deformation.

At the initiation of motion, when density was 2.69 Mg m\(^{-3}\), thickness was 100 m and surface slope was perhaps 40°, the velocity due to viscous deformation could not have been greater than 43 m sec\(^{-1}\) if the viscosity was \(4 \times 10^7\) poise, and thus, the Reynolds number was 3.

At Reynolds' numbers below 10 to \(10^3\), flow is invariably laminar; thus flow within the Sherman Glacier rock avalanche must
have been laminar at all times. This was not true of the air flow around the moving debris, where Reynolds number ranged between $10^6$ and $10^8$ for most of the time of travel for the range of velocities and size of boulders rising above the surface of the avalanche.

The Froude number

$$ F = \frac{v}{(gh)^{1/2}} $$

for the Sherman Glacier rock avalanche was at a maximum ($\approx 4$) at the base of Shattered Peak when velocity and thickness were greatest and did not drop below 1 until a speed of about 4 m sec$^{-1}$ was reached, in the last 15 m of flow of the avalanche. Thus, flow of the debris apparently was supercritical almost throughout the avalanche and became subcritical only in the last 6 seconds of flow. The transition, however, occurred more or less smoothly and simultaneously over the entire avalanche, and no "hydraulic jump" occurred, except perhaps in one region where an early halt occurred. This is the eastern rim of the avalanche where the debris was halted by uphill flow and possibly by deep soft snow. An irregular hummocky surface developed on the debris along this rim and this may express the turbulence developed in a transition from supercritical to subcritical flow in this region.

Other factors preventing the development of turbulence except in localized areas may have been the very large moment of inertia
of the avalanche and the apparent "hardening" of the debris at high strain rates.

**Development of surface features**

The Bingham plastic behavior of debris during the Sherman Glacier rock avalanche gave it a crust whose properties assist in explaining the origin of the conspicuous surface features of the avalanche deposit: transverse fissures and folds, longitudinal grooves and ridges, and some debris cones.

At the datum profile, referred to above (Fig. 9), lateral strain apparently was zero throughout the avalanche, and hence, for incompressible debris, the longitudinal strain rate varied directly with vertical strain rate (Fig. 16), being equal in magnitude but opposite in sign (for conservation of volume). This longitudinal extending strain in the deforming debris put the surface rigid layer in longitudinal tension, and the loose debris, which has essentially no tensile strength, yielded by brittle failure in the form of transverse fissures up to 2 m deep. In other areas of the avalanche, there was significant extending lateral strain that also stressed the avalanche crust, which yielded to form longitudinal grooves up to 2 m deep with ridges between them. These grooves and ridges are absent from regions of zero or compressive lateral strain, but such regions were rare in the spreading debris.
In addition to the widespread regions of longitudinal and lateral tension, there was also a region of significant longitudinal compression above the yield strength of the debris (the region of folding marked as 1 in Fig. 6) where the surface profile of the glacier surface is concave (Fig. 20). As is characteristic of viscous multilayer media supporting rigid flexible crusts, the debris yielded in compression by buckling, with the surface rigid layer folding in a characteristic wavelength ($\approx 15$ m) determined by the thickness of the layers and their mechanical properties, with the viscous subsurface layer deforming to fill the crests beneath the buckled surface. The theory of this type of behavior has been analyzed by Biot (1960). The stress at the time of buckling can be calculated, without reference to this theory, from the density of the debris, and its deceleration over the path travelled (i.e. from the time rate of change of momentum). This calculated stress (Fig. 20) is independent of the debris thickness and gives perhaps the best estimate of the yield strength of the debris. At the moment of folding, the rigid upper layer of debris transmitted a nonhydrostatic compressive stress of 0.0169 bar for a coefficient of basal friction of 0.11 and a debris density of 2 Mg m$^{-3}$. The excellent agreement of this value of the yield strength of the debris with the estimates by other techniques (p. 75 and p. 85) tends to confirm that the debris had a yield strength in motion of about 0.02 bar, a coefficient of basal friction of about 0.11, and a final mean thickness of about 1.65 m (for a density of about 2.0 Mg m$^{-3}$).
Longitudinal compressive stress resulting from the time-rate-of-change of momentum of the avalanche traversing the surface profile in the region of folding. If the debris transmitted the nonhydrostatic stress before buckling, it had a yield strength in compression of 0.0169 bar for a coefficient of sliding friction of 0.11.
A further manifestation of Bingham flow is the effect of a yield strength on lateral shear in the flowing debris. Lateral shear strain is evident in the calculated velocity profile across the datum line on Andres Glacier (Fig. 15), in the displacement of lithological and textural zonation where these cross grooves, and in the deformation of transverse fissures.

In those regions where the lateral shear stress was less than 0.02 bar, no lateral strain occurred in the debris surface and the lateral shear stress was transmitted through the rigid crust. At intervals across the width of the avalanche (with spacing defined by the gradient in subcrustal lateral shear strain rate) the shear stress accumulated to the shear strength and lateral shear strain occurred in many discrete longitudinal zones of shear separated by undeformed zones in the debris. The zones of shear probably coincided with longitudinal tension grooves because the tensional separation of clasts facilitates the shear deformation and the local thinning also would concentrate the lateral shear stress at grooves.

Some ridges, as was mentioned earlier (p. 49), may be lateral ridges bounding flow units. These are analogous to mudflow levees and other ridges common along the sides of debris flows. Johnson (1965) suggested that these ridges develop as dead flow areas during Bingham flow within a channel and this mechanism is appropriate to the closing stages of the Sherman Glacier rock avalanche.
when isolated streams of debris were still in motion between static debris buttresses near the distal margins along the western and northwestern edge of the avalanche. Shreve's (1966b, p. 39) mechanism of formation of lateral ridges by escape of air is in keeping with development of high shear stress between lateral ridge and flowing avalanche but similar ridges will develop in other flow models, so that the presence of lateral ridges is not diagnostic of any particular mechanism of flow.

When clasts at the margins of flow fall off the sides of the debris sheet, they quickly lose energy to their surroundings and come to rest. Static friction must be overcome to move a particle at rest and momentum must be transferred from some moving clasts to move it. This rapid loss of energy at the edge soon produces a static lateral wall which can prevent further lateral spreading. In addition, this static lateral wall must dilate before internal strain can occur and this requires even more energy than that to initiate motion of a single particle.

The Sherman Glacier rock avalanche lacks prominent lateral ridges because its plumose shape has a lack of lateral margins, and internally, most of its units were in simultaneous flow during avalanching and failed to develop ridges between them. Good lateral ridges or levees developed along static debris buttresses along the northwestern margin, but these ridges were rapidly subdued during the summer of 1964.
The plumose shape is merely the consequence of the almost ubiquitous lateral extension at the advancing front of the avalanche.

Heim (1882, p. 101) found debris cones similar to those reported from Sherman Glacier on boulders in the Elm slide in Switzerland, and suggested that they were the remains of a surficial layer of fine slate, the rest of which had fallen between the larger blocks as the latter had spread apart. This is correct for many cones such as the xenolithologic cones described by Shreve (1966, p. 1642) that may form as flowing debris thins about large boulders. In the Bingham flow model (p. 75), a cone would form by thinning of the loose debris, when the rigid surface layer of the avalanche is lowered to intersect the top, or bearing surface, of an underlying boulder. This boulder may not always be exposed at the surface of the avalanche if it is smaller than the final local thickness of the surrounding deposit. Such a model for the development of cones would suggest that the maximum height of a cone is limited by the yield strength of the flowing debris, and that the debris in the cone is a sample of the surrounding interior of the avalanche. The presence of xenolithologic cones that are apparently of a different lithology to surrounding surface boulders thus suggests horizontal stratification within the avalanche so that debris below the surface may be of a different rock type than that observed to cover the surface. In this model of cone development
the observed finer nature of debris in cones relative to the
surrounding debris (Heim, 1882, p. 101; Shreve, 1966, p. 1642;
and Marangunic and Bull, 1968, p. 391) is consistent with the
observation that the interior of the Madison Canyon Slide contains
very much more finer debris than is visible on the outer surface
(Hadley, 1964, p. 119).

Energy dissipation at the sliding interface

The $2.67 \times 10^{14}$ Joules of energy available to the Sherman
Glacier rock avalanche, through loss of potential energy, is
sufficient to have heated the entire mass of avalanche by $12^\circ$C.
But the rock mass could not have been heated alone, the avalanche
was of such short duration, thermal conductivity of rock so low,
and most deformation took place at the base, so that only an
insignificant amount of heat could have been gained by the rock
debris. Instead, most of the energy was lost through friction of
sliding at the base and most of the energy became heat in a
narrow basal zone. Some energy was needed to break up the original
block, but, in fragmentation, most of the elastic energy reappears
as kinetic energy in the dispersing debris, and very little is
converted directly to heat.

If all of the energy became heat at the base, it could have
melted 70 kg m$^{-2}$ of snow (0.12 m at density 0.6 Mg m$^{-3}$). If heat
was generated at the interface too rapidly to be dissipated by
melting ice, some heat would have been used to generate steam,
but in the presence of ice near its melting point, it is most improbable that the heat was dissipated in any way other than in melting ice.

Earlier (p. 76 ) I suggested that the rate of inhalation of air can be a limiting factor in the spreading of rock avalanches. The behavior of snow in this high energy environment may help to explain why the Sherman Glacier avalanche is one order of magnitude thinner than other rock avalanches of similar volume (see Shreve, 1968b, p. 27-33).

The evolution of vapor from water heated in a partial vacuum by grinding of rock against rock could have facilitated the separation of clasts that is necessary for strain to occur, and this could allow the rock mass to spread over a greater area than a dry avalanche might have spread over. However, the maximum basal shear stress in the Sherman Glacier rock avalanche was about 0.65 bar. If water vapor was to contribute all of the gas needed to permit deformation at this stress, water would have to be heated to more than 73°C (saturated vapor pressure of water at 73°C is about 0.35 bar). For the process to have even been a dominant effect at the yield stress (0.02 bar), the water would have to be heated to more than 89°C (at 89°C the saturated vapor pressure of water could have contributed half of the necessary 0.98 bar). At 0°C the vapor pressure of water (0.006 bar) could only have contributed 0.6 per cent of the required gas pressure.
Because of the very high viscosity of the debris in motion, very little deformation occurred within the debris and most occurred at the base. Thus, very little energy could have been dissipated within the flowing avalanche, and very little heating should have occurred there, making the above mechanism a very improbable explanation for the observed difference in thickness. The concept is, however, applicable to the flow of hot, gas-evolving nuees ardentes. The presence of an internal mechanism to bypass the normal low permeability of such poorly sorted debris would give rise to a lower apparent viscosity at high strain-rates than would otherwise be the case, permitting such avalanches to spread more thinly, and more rapidly, with fluid flow more apparent. Even in nuees ardentes, it is expected that the yield strength of flowing tephra would be insufficient for the tephra sheets to maintain high pressure air layers beneath them.

Significant snow was not present in some of the other large rock avalanches such as the Elm or Frank slides, or so far as is known, in the Blackhawk or Silver Reef slides. The Elm and Frank slides had significant basal layers of mud (Shreve, 1968b, p. 37) and the Elm slide crossed small streams and a pond while the Frank slide crossed a river, and some steam could have been produced. Shreve (1968, p. 37) doubts that significant water was present at the base of the Blackhawk slide because it slid down a permeable alluvial fan. Rock avalanches from Little Tahoma Peak (Crandell
and Fahnstock, 1955, p. A6) did contain significant snow. Although they are only some 2 to 4 m thick at the end of their runs, their average thickness is probably some 14 m. These avalanches were, however, laterally confined in stream channels.

Snow had two other possible roles in the Sherman Glacier avalanche. The density of winter snow at Sherman Glacier is about 0.6 Mg m\(^{-3}\) when measured in late summer and is probably as low as 0.45 Mg m\(^{-3}\) in late March when the avalanche fell. Thus the substrate over which the avalanche rode may have been as much as 50 per cent air. Some of this air may have been released by compaction as the avalanche rode over it, and some was certainly released when the 70 kg m\(^{-2}\) of snow was melted. This air could contribute to the air flow that facilitated the strain within the debris, but it could not have provided a lubricating air layer in the sense of Shreve's model because the finite bearing strength of the snow must necessarily have prevented compression of the air to a pressure capable of lifting the debris off the snow substrate. In addition, rolling of clasts is to some extent dependent on the mechanical properties of the substrate through the torque applied to the clasts. The soft snow on the surface of Sherman Glacier may have inhibited rolling to some extent, because of its own low shear strength and its high compressibility, and because it may have been scraped up by the moving mass to clog the base of the avalanche.
Role of thickness in rock avalanches

Somehow or other, the Sherman Glacier rock avalanche was able to continue deforming until its thickness dropped below a critical value perhaps defined by the permeability and shear strength of the debris, (or by the normal stress, or by all three), and this reduced the apparent viscosity of the debris by an order of magnitude, allowing the debris to continue flowing to another critical thickness defined only by the shear strength. Perhaps, the compressibility and high air-content of snow was a sufficient factor in the early stages of the avalanche, when the debris exerted a maximum load on the thick snow cover in Andres Glacier, to have permitted the Sherman Glacier debris to cross some threshold value of thickness while it still had enough energy to continue its motion. This mechanism was unavailable to the other major avalanches discussed by Shreve.

On the other hand, the cause of the thickness difference may lie with the thicker avalanches. They may have travelled fast and far enough that they reached gentle slopes before having spread below the critical thickness, and thus were slowed to a halt before the viscosity transition occurred. The thicker avalanches are thus an order-of-magnitude thicker because their very thickness was the controlling factor in their rate of deformation.

Thickness \((h)\) controls the rate of inflow of air \((q = k \Delta p)\), the basal friction \((F_r = \mu \rho g h \cos \alpha)\), the basal shear stress \((\tau = \rho g h \sin \alpha)\), the basal normal stress \((p = \rho g h \cos \alpha)\), and
the basal rolling torque \( T = \frac{1}{2} (F + \tau) \), and these in turn control the duration of the avalanche and the rate of deformation within it (i.e. thicker avalanches are thicker because they are thicker).

**Mode of emplacement of other Alaskan rock avalanches**

Although no other rock avalanches resulting from the 1964 earthquake has been studied in detail, the processes shown to be active during the Sherman Glacier rock avalanche appear equally applicable to other avalanches, including such different forms as the very lobate Waxell Ridge avalanche (Fig. 21a) that apparently behaved as one large spreading lobe, and the highly digitate Allen II avalanche (Fig. 21b) that separated into a number of streams. The same variety of form was duplicated in later avalanches in the lobate Allen IV avalanche and the digitate Fairweather Glacier avalanche seen in 1965 (Fig. 21 and 22 of Post, 1967, and 1968, and also Johnson and Ragle 1968, Fig. 3).

The different forms might indicate whether debris fell simultaneously, or in a number of discrete falls over a prolonged period; in which case, the Waxell Ridge, Allen IV, and Sherman Glacier rock avalanches fell as single massive units, while the Allen II and Fairweather Glacier avalanches each occurred as a number of smaller separate avalanches, separated by significant, though not necessarily very long, intervals of time. This is illustrated by the crossing debris trains in the Allen II avalanche (center of Fig. 21b), by the irregular path of the earlier of these two trains (the one with dexteral curvature) where its sinisteral curvature was deflected off
Figure 21: Earthquake-induced rock avalanches on Alaskan Glaciers; a study in contrasts:

(a) The lobate Waxell Ridge avalanche in the Badgley Icefield. Note the powerful control of slope of the substrate on direction of flow of the avalanche in the middleground of the photograph.

(b) The digitate Allen II avalanche on Allen Glacier. Note the large boulder sitting atop the lower end of the sharply hooked debris stream. This stream narrows towards its distal end, is cut by a later stream, and came to rest with the boulder travelling upslope. Note also the earlier deflection of a sharply sinistral curvature of this stream by earlier flows to its left. The final sharply dextral curve in its path is apparently a result of disintegration of the rolling boulder which changed the amount and direction of taper of the effectively conical boulder from being small and to its left to being large and to its right.

(Photographs by Arctic Institute of North America)
an earlier debris train, and by other cross-cutting relationships that are more evident on vertical aerial photographs (Johnson and Ragle, 1968, Fig. 2).

Johnson and Ragle (1968, p. 372) noted a number of features characteristic of the Allen II type of avalanche with numerous divergent debris trains: (1) the debris streams have no lateral constraints such as sidewalls; (2) the trajectories of some flow streams are smoothly and uniformly curved; (3) some flow streams do not expand laterally along their length, indeed, two narrow toward their distal ends; (4) some flows have giant blocks perched on their extreme distal ends; (5) the most regular flows have no distal rims, transverse zones or other irregularities; and (6) the length-to-width ratio of many flows is very high (>10:1). They suggested that these features are the product of an equilibrium process where a balance between gravity forces and basal friction is established in which material feeds into the heads of debris streams at a constant rate from an amorphous pile of debris at the foot of the ridge from which the rock fell. In essence, this latter suggestion is similar to my own, that a part of the driving force of motion was the loss of potential energy of the debris while it thinned during longitudinal and lateral extension in flow but, in my mechanism, the pile of debris itself flattens as it slides on its base, rather than merely feeding debris into streams from a constant location at a constant rate. If the viscosities estimated for the Sherman Glacier avalanche are
typical for such motion, internal flow was the less significant part of the total motion so that Johnson and Ragle's "amorphous pile" should also have slid bodily down the slope had it fallen in one massive rockfall. In addition, their flow model does not account for the absence of lateral extension of the advancing debris front.

I suggest an alternative mechanism for rock avalanches of the Allen II type that may better account for their observed characteristics. This mechanism requires the presence of very large boulders rolling at the leading edge of the advancing avalanche. If these boulders are prisms, they will roll in straight paths, but if they are pyramids (or truncated pyramids), they will roll in curved paths with a radius of curvature related to their taper (Fig. 22).

The maximum normal force \( F_n \) at the surface of a rolling boulder (density \( \rho_r \), radius \( r \)) is at the top (Fig. 23) where

\[
F_n = \rho_r \omega^2 r
= 4 \rho_r v^2 / r
\]

where \( \omega \) is the rotational velocity and \( v \) is the forward velocity. When this normal force exceeds the tensile strength of the rock, the surface of the boulder will disintegrate and it will leave a trail of debris in its wake equal in width to the length of the axis of rotation. The debris would probably be laid as a thin carpet, perhaps thinner than the critical thickness for internal flow (i.e. basal shear stress \(< 0.02 \text{ bar}\)) so that no lateral spreading should
Fig. 22. The path followed by a rolling clast whose shape changes from a cylinder to a truncated cone by differential disintegration along its length.
Fig. 23. Forces acting on a cylinder rolling down an inclined plane. The basal friction \( F_f \) and the component of gravity down the slope \( F_g \) provide the torque that produced rolling. The angular rotation due to rolling causes a tensional normal force \( F_n \) and reaches a maximum value \( F'_n \) at the surface of the cylinder where the absolute velocity is highest. A rolling boulder may fly apart if this normal force exceeds the tensional strength of the rock. Crushing at the base of the rolling boulder may also be involved in the disintegration.
occur. This carpet of debris "unrolled" from the rolling boulder would still have forward momentum and should slide, but soon would be brought to rest by basal friction and by friction against static lateral walls that would develop almost immediately after passage of the rolling boulder.

As it disintegrates, the boulder may fragment in such a way as to change the direction of taper and the boulder then should change its direction of motion (Fig. 22). If this path brings the course of the boulder into opposition to the local slope direction, the boulder will prematurely come to rest perched at the distal end of a debris train. In other cases, a debris train may end when the boulder has disintegrated to some small unit-size controlled by a limiting terminal rotational velocity and the tensile strength of the rock. This size may be small enough that the boulder is no longer recognizable amid its other fragments.

Applicability of the Bingham model to other rock avalanches

The Sherman Glacier rock avalanche possesses all of the features (Shreve, 1966, 1968a, 1968b, Table 2) supposedly characteristic of large rock avalanches, that according to Shreve, are best typified by the Blackhawk slide. Shreve (1959, 1966, and 1968b) proposed that the "type" Blackhawk slide, slid on a cushion of compressed air trapped in an initial fall. Nevertheless, the dispersal of debris and the features developed within the Sherman avalanche can
well be accounted for by expected kinetic interactions of materials known to have been present in the avalanche and by a simple rheological model of deformation as a Bingham plastic. Introduction of a nearly perfect lubricant such as an air layer into the model is not only unnecessary but makes the system too well lubricated.

The apparent lubrication of many rock avalanches can be studied through the calculation of an average coefficient of friction (see p. 57). The many compilations of apparent average coefficients of friction of rock avalanches (Heim, 1932; Shreve, 1966, and 1968b; Scheidegger, 1973; and Hsu, 1975; Table 5) are, however, not directly comparable to the various values estimated above for the Sherman Glacier rock avalanche, because I have chosen to make a more theoretically sound estimate from the displacement of the center of mass of the avalanche (as Hsu, 1975, p. 132, has also suggested), rather than from the maximum values of height of fall and distance travelled, and I have also calculated values for specific segments of path during the avalanche.

Among the many rock avalanches that have been documented, two are presented in such detail that similar estimates can readily be made from them.

Shreve's (1968b, Plate 1) map of the Blackhawk slide is sufficiently detailed that the displacement of the center of mass of that avalanche can be estimated if the source region is reconstructed. If the $280 \times 10^6$ m$^3$ of debris in the Blackhawk slide underwent a
30 percent dilation during that avalanche, it can be fitted back into the scar on its source, Blackhawk Mountain. I estimate that its center of mass dropped about 900 m in 5.5 km, which gives an effective average coefficient of kinetic friction of 0.166. This is greater than Shreve's estimate of 0.15, largely because my estimate involves a consideration of the shape of the original block on the mountain. If this block slid as a single block for the first 600 m until it reached a major change in slope at the base of the mountain where flexure would have caused disintegration (as was demonstrated experimentally by Rengers and Müller, 1970) with a coefficient of basal sliding friction of 0.600, then the coefficient of basal friction over the remainder of the 5.5 km of displacement must have been 0.111, in order that the distance-averaged value be 0.166.

A coefficient of kinetic friction for a part of the Elm slide in Switzerland (Heim, 1882, and 1932) can be estimated from Heim's estimate of the minimum speed (30 m sec\(^{-1}\)) for the debris to rise up an opposite slope from its source, and the distance travelled to the distal margin. In this case, the debris slowed from 30 m sec\(^{-1}\) to 0 m sec\(^{-1}\) in about 1500 m while traversing a slope of 3° 54'\(^{\prime}\). This provides for a minimum effective coefficient of friction of 0.099. If the initial speed had been 40 m sec\(^{-1}\), the coefficient would have been 0.12.

The fact that coefficients for three avalanches moving across diverse substrates are essentially identical, strongly suggests that they behaved in a similar manner, and that all similar lobate rock
avalanches should behave in a like manner; that is, essentially as I have determined for the Sherman Glacier rock avalanche. An effective coefficient of basal friction of about 0.11 may be characteristic of the latter part of large rock avalanches; a manifestation of some basic property or similarity of properties of aggregates of loose heterogeneous rock debris in rapid motion.

Before proceeding to examine this in greater detail, it is useful to examine the first and most basic assumption in Shreve's air-layer model, the mechanism of trapping air, to determine why this model is inappropriate to rock avalanches, and to examine why Shreve saw a need for a frictionless lubricant.

Shreve (1968b, p. 41) suggested that the Blackhawk landslide added a lubricating layer of air after it was launched into the air by a low ridge across its path, and fell upon and trapped the air beneath the arching parabolic canopy of the avalanche debris. This unsupported assertion was not substantiated, and in fact this behavior is not a characteristic of massive materials in motion. In flight, the debris sheet would have been slowed only by air resistance as the avalanche displaced air from its path. Because the ratio of density of air to that of debris is about 1:2500, air resistance could cause only negligible loss of momentum. A parabolic trajectory is the normal path described by a projectile with negligible air resistance, and a projectile in flight exerts no influence on the air beneath it. The air enclosed beneath the parabolic canopy of
debris should therefore have been at atmospheric pressure, and essentially unaffected by passage of the avalanche overhead just as the air behind a waterfall is similarly unaffected. The air mass should have remained in place while the avalanche passed the ridge. Unless a high vacuum was left in the lee of the ridge, it would have been impossible for the debris sheet to have slid off on a "cushion" of air. Thus Shreve's air-layer model deviates from reality at his first assumption.

If the avalanche was to remain airborne, it would have to glide or fly. Through Voellmy's (1955) equation (p. 52) and using an avalanche density of 2500 kg m$^{-3}$ and a thickness of 50 m at launching, the velocity required to keep the avalanche in flight by pressure at the base would have to have been about 1560 m sec$^{-1}$ (Mach 4.7!) Flight of objects can also be assisted by lift at the upper surface, but loose debris has no tensile strength and surface lift cannot be a possibility for keeping an avalanche air-borne. It can, however, be a factor in lifting clasts from the upper surface of the avalanche to form the clouds of dust frequently observed, although the turbulent flow of air may be a sufficient cause of dust clouds.

A further major defect in the air-layer model is introduced by the low shear strength of avalanching debris (0.02 bar for the Sherman Glacier avalanche, and 0.2 bar for the Blackhawk slide). The Blackhawk debris could not sustain a basal air pressure of
12 bar, as required by the air layer model, without failing at the slightest weakness to let the air rise rapidly to the surface in great high-pressure bubbles.

Because air-layer lubrication is clearly not involved in rock avalanches, and because they apparently slide directly on their substrates (p. 64), it is useful to examine if other processes could occur at the base of rock avalanches besides simple sliding to provide such low coefficients of friction. Rolling friction, for example, can be as much as two or three orders of magnitude less than sliding friction. Thus, if clasts at the base of an avalanche move by a combination of sliding and rolling, as apparently was the case in the Sherman Glacier rock avalanche (Fig. 9), an effective coefficient of kinetic friction of between about 0.6 and 0.006 could be expected.

At the base of the Blackhawk slide, the shear stress was about 0.07 bar near the end of the run, while the expected frictional force at the base (for $\mu = 0.6$) was 1.8 bar. If deformation principally was at the base of the debris, as apparently was the case for a material of viscosity of the order of $10^7$ poise (Fig. 19) the basal clasts were subject to a torque of 2250 Newton m$^{-1}$ for a mean size of 0.025 m (the present model size of clasts at the surface, Shreve, 1968b, p. 28). In the Sherman Glacier rock avalanche the torque at the base at the end of the run was only 200 Newton m$^{-1}$ (for clasts of 0.075 m diameter, Marangunic, 1968, p. 91). Thus, rolling of clasts should have been more dominant in the Blackhawk slide than
in the Sherman Glacier avalanche, but rolling still occurred in the latter avalanche (viz. the transverse mode in basal clast fabrics, Fig. 9). For this reason, the apparent reduction in basal friction, in comparison to that expected for the substrates involved, should have been greater in the Blackhawk slide than in the Sherman Glacier rock avalanche; but this does not appear to be the case, both the soft snow-covered surface of Sherman Glacier, and the hard, probably dry alluvium beneath the Blackhawk slide were traversed with the same apparent coefficient of friction as were the rain-softened fields at Elm. If they had all slid on air, this would present no problem, except that a coefficient of friction of 0.11 is too high by at least an order of magnitude for air (p. 64), and a means of trapping and maintaining the air is lacking.

A part of the estimated coefficient of friction is derived from the frictional forces within the flowing debris, but it is evident from Fig. 19 that these can contribute but a small part to the total braking force. Most of the friction is at the base, and most of the estimated coefficient of friction is from basal friction.

Both Heim (1882, and 1932) and Hsü (1975) discuss, at great length, the fluid-like flow of rock avalanches (or sturzstroms, as they prefer), and this fluid flow was also evident to Kent (1966), and to Müller (1964, and 1968). In his attempts to demonstrate that flow rather than sliding is the characteristic of rock avalanches, Hsü neglects some of the evidence that he himself presents (from
Heim's descriptions): that in its traverse of the Unterthal, the Elm sturzstrom scoured the valley bottom and carved parallel furrows (Hsu, 1975, p. 131).

Shreve (1966, and 1968b) on the other hand, was so much more impressed by the apparent sliding in rock avalanches that he failed to treat, but obviously was not unaware of the flow required to convert a thick rigid block into a thin flexible sheet. It was to account for this sliding at high speed as a flexible sheet that Shreve proposed the lubricating air layer.

As I have demonstrated for the Sherman Glacier rock avalanche, the true situation involves both the sliding and flow, with sliding being dominant. The thin flexible sheet is an accurate description of a rock avalanche in flow.

In his discussion of the preservation of original stratigraphic order in rock avalanches, that so impressed Heim and Shreve, Hsu (1975, p. 133) reasons that Shreve's objection to viscous flow can be overcome simply by postulating that the flow was non-viscous. It is obvious that both Shreve and Hsu thought in terms of Newtonian viscosity, and that the Bingham viscosity that I have suggested is a "non-viscous" flow in Hsu's usage. What is less obvious is that both must have considered only subcritical fluid flow, the most commonly treated mode of fluid flow, whereas the feature that they are attempting to explain is a consequence of supercritical flow.

A model of supercritical laminar flow of debris as a Bingham plastic provides an excellent model for the rheology of the rapidly
flowing loose aggregate of heterogeneous debris that makes up a rock avalanche.

In his further analysis of flow in rock avalanches, Hsü (1975, p. 134) applies the concept of flow of cohesionless grains (Bagnold, 1954, 1956), but he is apparently puzzled by the need for an "interstitial fluid" (Hsü, 1975, p. 135) that was a part of Bagnold's analysis. To account for the fluidity of lunar avalanches, that lack water and air for a dispersing medium between cohesionless grains, Hsu suggests that the abundant fine component might act as a dispersing medium in most rock avalanches and thus account for the apparent reduction in internal friction. Trawinski (1953) had earlier noted that addition of a fine component to gas-fluidized grains brought about a reduction in apparent viscosity, and suggested that the finer grains might act as "ball bearings". Hsü, however, apparently regards the role of fines as increasing the density of the dispersing medium, whereas Botterill and Van der Kold (1971) suggest that they clog pore space and reduce permeability.

Hsü's density analysis (Hsü, 1975, p. 136) might be applicable to terrestrial avalanches where air is present, but it cannot be applied to a lunar avalanche, because a vacuum has no resistance to compression and can not provide a reduction in "effective normal pressure", no matter how full of dust it may be.

Much of Bagnold's analysis is independent of a dispersing fluid, it requires only that the cohesionless grains be dispersed, in order that they may move about past one another, just as the molecules of
a fluid are dispersed. The same forces that keeps the molecules of a fluid dispersed is available to rock avalanche debris in motion. The dispersive force in a fluid stems from the kinetic energy of the molecules, their thermodynamic temperature, that controls their relative velocities and spacing, or mean free path (which relates to pressure). In rock avalanches, the kinetic energy of the clasts keeps them separated during countless elastic impacts. This separation constitutes a dilation of the debris and an increase in porosity.

Because the frequency of collisions is a factor in dispersion, the addition of a fine component which increases the number of particles and thus the frequency of collisions, must also increase the dispersion of the larger clasts so that they may more readily pass one another - a requirement in deformation of a granular substance. This greater dispersion of some controlling clast-size thus constitutes an increase in ability to deform and a reduction in Bingham viscosity.

A granular substance that is dispersed by the vibrating motion of its clasts must also be highly compressible, and that compression, that decreases the dilation, must bring about an increase in Bingham viscosity. That is, the Bingham viscosity of a rapidly flowing granular fluid is strongly dependent on the normal stress, as was suggested for one of the flow units in the Sherman Glacier rock avalanche (Unit 1, Fig. 17).

The Vaiont slide described by Müller (1964, and 1968) apparently did not undergo as extensive a dispersion as is apparent in many
other rock avalanches. Müller (1964, p. 210) recognized a "quasi-
plastic" flow, but noted that the internal displacements were very
small, and that the block was essentially translated without massive
deformation. Evidently this slide did not fall far enough for
individual clasts to gain sufficient kinetic energy to adequately
disperse the debris. It had to deform with an extremely high
viscosity, possibly orders of magnitude higher again than the highest
in the Sherman Glacier rock avalanche, and it thus behaved more as a
thick flexible sheet, than as a fluid or plastic flow. In connec-
tion with the Vaiont slide, Broili's (1967, p. 79) use of the terms
"friction sand" and "rounded bun-like structures" in his description
of the cataclastic zone of shear in the slide, clearly implies a
rolling of clasts within the zone, just as Marangunic' and Bull
(1968, p. 392) found in the Sherman Glacier rock avalanche.

At the opposite extreme must be the extremely mobile Huascaran
avalanche (Hsü, 1975, p. 138), which through its tremendous height
of fall, should have had ample kinetic energy for dispersal of
clasts, and should therefore have exhibited a very low viscosity,
but in this case, an interstitial fluid of mud was undoubtedly
present.

In smaller rockslides and rockfalls, two different situations
can arise: (1) the height of fall may be small, so that no signifi-
cant dispersal of clasts takes place and the single block is merely
translated; or (2) the height of fall may be so large that the gain
in kinetic energy in the fall is sufficient to "totally" disperse
the rock, so that the comparatively few small clasts can not interact with one another to the extent that they are able to in the massive rockfalls and rockslides that become rock avalanches. In analogy to fluids, the rock mass boils, and becomes a gas that disperses by diffusion rather than flow. A simple Bingham plastic flow model is obviously no longer applicable.

The mechanical-fluidization mechanism that I have proposed as the means by which the internal friction of moving rock debris is reduced to the often seemingly impossible values displayed in rock avalanches is no more than the cause of a true thixotropy, whereby a formerly rigid aggregate looses its coherent structure to become a more readily flowing plastic as its kinetic energy rises. Müller (1964, p. 208) was thus fully justified in applying the term thixotropic to the behavior of the Vaiont slide. He had no reason, however, to justify his later qualification of "mass-thixotropy" (Müller, 1958, p. 84-85); the phenomenon that he describes is a true thixotropy - the gradual or spontaneous loss of structure and consequent increase in ability to flow that occurs when the kinetic energy of a thixotropic substance is increased beyond some critical value.

Rock avalanche debris appears to behave as a complex Bingham plastic with a variable normal-stress dependent viscosity, and a yield stress. The viscosity and yield stress apparently are functions of the coefficient of sliding friction of rock against rock in an environment where many elastic impacts significantly reduce the number of rocks in contact with one another.
Differences between one rock avalanche and another result from such parameters as: the strengths and elastic properties of the parent rocks; the initial states of fragmentation of the parent bodies; the sizes of the parent bodies, particularly their thickness; the various parameters of the avalanche paths such as: height of fall; surface slope; obstructing, and diverting obstacles; changes of slope that cause flexure and fragmentation; and the softness and compressibility of the substrate; and lastly the various parameters of the avalanche debris itself such as: porosity and permeability; and the mode of disintegration of larger clasts.

Prediction of the reach of rock avalanches

The purpose of developing an understanding of the mechanics of rock avalanches is to predict their occurrence and the area of their destructive influence to save lives and property. My analysis has been directed towards understanding the flow of avalanches, and is of little use in predicting their occurrence, but it can be applied to determining an area of destruction.

Without a full understanding of the behavior of rock avalanches, the approach to prediction of their reach must be empirical. This approach has been taken by Scheidegger (1973) and by Hsu (1975). A significant correlation between the logarithm of avalanche volume V and the logarithm of apparent coefficient of friction μ was determined by Scheidegger (1973, p. 235) from data for 33 avalanches.
He found that

$$\log_{10} \mu = -0.1567 \log_{10} V + 0.6242$$

with a correlation coefficient of 0.82. Through this empirical relationship, the reach of a potential avalanche can be predicted from the estimated volume and possible height of fall (\( \mu \) is the ratio of fall to reach).

A further refinement, of model experiments with bentonite suspensions, was added by Hsü (1975) to essentially the same data used by Scheidegger. Through his models, Hsü (1975, p. 138) was able to suggest that apparent coefficient of friction might not be the best indicator of mobility. A new parameter was defined, "excessive travel distance", the extra distance travelled beyond that expected of a simple frictional slide with a "normal" coefficient of friction (\( \mu = 0.62 \)). In his model experiments, Hsü found that this distance depended only on volume and mobility of his bentonite suspensions and not on height of fall; he thus deduced that "excessive travel distance" is determined solely by the volume and mobility of the avalanche, and postulated a semilog function to relate excessive travel distance to the logarithm of volume. Although Hsü was unable to predict the mobility of a given avalanche, his Fig. 8 can be used to predict a maximum excessive travel distance if maximum mobility is assumed. He suggested that mobility might be related to height loss, or confinement, and in the case of the Huascaran avalanche, to interstitial mud.
When a rock avalanche spreads over the terrain below its source, work is done in overcoming friction. The source of energy to do this work is the loss of potential energy in the fall. For this reason, Hsu's suggestion that "excessive travel distance" is independent of height of fall is theoretically unsound for an empirical approach to prediction because it implies that the work done is independent of the energy used. The energy for dispersal of debris comes from the loss in potential energy, so that the soundest empirical approach must be to correlate distance travelled to the potential energy lost in the fall. From data provided in Hsu's table 1, I have calculated maximum horizontal distance travelled and potential energy lost (Table 5). A linear relationship between the logarithm of the potential energy lost and the logarithm of maximum distance travelled is apparent (Fig. 24). The Huascaran avalanche may have behaved as a mud flow, and the Vaiont slide was exceedingly rigid.

This empirical relationship has the theoretical advantage that, when no potential energy is lost, the avalanche does not move. A notable feature of the relationship is the fact that as avalanche volume increases, there apparently is a very marked decrease in the rate of increase in travel distance, so that in terms of maximum travel distance, smaller avalanches are more efficient than larger ones, so that apparent coefficient of friction is clearly no measure of efficiency.

A sounder basis for prediction, however, is one based on the rheology of rock debris. The Sherman Glacier rock avalanche flowed
Fig. 24. Correlation between travel parameter and work parameter for rock avalanches. Open circles: maximum distance travelled versus "maximum" energy loss (calculated as a loss of potential energy, but using the drop in height from source to distal margin). Dots: square root of area covered versus actual potential energy lost (calculated from the estimated drop of the center of mass). The numbers correspond to the numbered rock avalanches in Table 5.
Table 5. Relationship between the travel parameters of travel distance and area covered and the work done through the loss of energy in rock avalanches.

<table>
<thead>
<tr>
<th>No.</th>
<th>Locality of event</th>
<th>Volume ($\times 10^{5}$m$^3$)</th>
<th>Equivalent coefficient of friction$^a$</th>
<th>Area (Km$^2$)</th>
<th>Maximum travel distance$^b$ (Km)</th>
<th>&quot;Maximum&quot; energy loss$^c$ (log Joules)</th>
<th>Potential energy loss$^d$ (log Joules)</th>
</tr>
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<tbody>
<tr>
<td>1</td>
<td>Airolo</td>
<td>0.5</td>
<td>0.64</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>Schachental</td>
<td>0.5</td>
<td>0.58</td>
<td>2.6</td>
<td>13.72</td>
<td>13.18</td>
<td>-</td>
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<tr>
<td>3</td>
<td>Val Lagone</td>
<td>0.5-0.8</td>
<td>0.44</td>
<td>2.4</td>
<td>13.69</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>4</td>
<td>Mombiel</td>
<td>0.8</td>
<td>0.47</td>
<td>0.6</td>
<td>13.18</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>5</td>
<td>Iacacaran (Peru)</td>
<td>2</td>
<td>0.30</td>
<td>23.1</td>
<td>14.59</td>
<td>14.342</td>
<td>-</td>
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<td>6</td>
<td>Mengen (1)</td>
<td>2-3</td>
<td>0.45</td>
<td>1.0</td>
<td>13.91</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>7</td>
<td>Mengen (2)</td>
<td>5-6</td>
<td>0.42</td>
<td>1.4</td>
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<td>-</td>
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<td>8</td>
<td>Elm</td>
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<td>0.58</td>
<td>14.68</td>
<td>13.537</td>
<td>-</td>
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<td>9</td>
<td>Sherman (U.S.A.)</td>
<td>10.1</td>
<td>0.21 (0.19)</td>
<td>8.25</td>
<td>14.96</td>
<td>14.204</td>
<td>-</td>
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<tr>
<td>10</td>
<td>Little Tahoma Fk. (U.S.A.)</td>
<td>11</td>
<td>0.29</td>
<td>5.18</td>
<td>14.65</td>
<td>14.653</td>
<td>-</td>
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<tr>
<td>11</td>
<td>Disentis</td>
<td>10-20</td>
<td>0.36</td>
<td>2.2</td>
<td>14.91</td>
<td>-</td>
<td>-</td>
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<tr>
<td>12</td>
<td>Corna di Dosde</td>
<td>20</td>
<td>0.32</td>
<td>3.8</td>
<td>15.23</td>
<td>-</td>
<td>-</td>
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<td>Madison (U.S.A.)</td>
<td>29</td>
<td>0.27</td>
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<td>13.93</td>
<td>14.083</td>
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<td>14</td>
<td>Vorarlsee</td>
<td>30</td>
<td>0.33</td>
<td>3.3</td>
<td>15.36</td>
<td>-</td>
<td>-</td>
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<tr>
<td>15</td>
<td>Frank (Canada)</td>
<td>30</td>
<td>0.25</td>
<td>4.0</td>
<td>15.26</td>
<td>14.559</td>
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<tr>
<td>16</td>
<td>Goldau</td>
<td>30-40</td>
<td>0.21</td>
<td>1.9</td>
<td>15.2</td>
<td>-</td>
<td>-</td>
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<tr>
<td>17</td>
<td>Grand Ventre (U.S.A.)</td>
<td>38</td>
<td>0.17</td>
<td>2.65</td>
<td>14.950</td>
<td>-</td>
<td>-</td>
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<tr>
<td>18</td>
<td>Hops (Canada)</td>
<td>47</td>
<td>0.37</td>
<td>5.6</td>
<td>15.83</td>
<td>-</td>
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<tr>
<td>19</td>
<td>Diazelba</td>
<td>50</td>
<td>0.34</td>
<td>5.4</td>
<td>15.91</td>
<td>-</td>
<td>-</td>
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<tr>
<td>20</td>
<td>Scima da Saoseo</td>
<td>80</td>
<td>0.27</td>
<td>5.0</td>
<td>16.15</td>
<td>-</td>
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<td>21</td>
<td>Obersee</td>
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<td>0.36</td>
<td>10.0</td>
<td>16.28</td>
<td>-</td>
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<tr>
<td>22</td>
<td>Kanderstal</td>
<td>140</td>
<td>0.19</td>
<td>4.1</td>
<td>16.20</td>
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<tr>
<td>23</td>
<td>Poschiavo</td>
<td>150</td>
<td>0.36</td>
<td>10.0</td>
<td>16.28</td>
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<tr>
<td>24</td>
<td>Apollo 17 (Moon)</td>
<td>200</td>
<td>0.20</td>
<td>7.5</td>
<td>17.0</td>
<td>-</td>
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<td>25</td>
<td>Silver Reef</td>
<td>220</td>
<td>0.13</td>
<td>15.5</td>
<td>16.97</td>
<td>-</td>
<td>-</td>
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<td>26</td>
<td>Vaiont (Italy)</td>
<td>250</td>
<td>0.34 (0.25)</td>
<td>2</td>
<td>15.97</td>
<td>15.217</td>
<td>-</td>
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<tr>
<td>27</td>
<td>Blackhawk (U.S.A.)</td>
<td>250</td>
<td>0.13 (0.17)</td>
<td>9.6</td>
<td>15.877</td>
<td>-</td>
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<tr>
<td>28</td>
<td>Dojen</td>
<td>600</td>
<td>0.11</td>
<td>6.6</td>
<td>16.48</td>
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<td>0.25</td>
<td>7.5</td>
<td>17.0</td>
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<td>30</td>
<td>Fernpass</td>
<td>1,000</td>
<td>0.09</td>
<td>15.5</td>
<td>16.97</td>
<td>-</td>
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<td>31</td>
<td>Siders</td>
<td>1,000-2,000</td>
<td>0.14</td>
<td>17.4</td>
<td>17.41</td>
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<td>32</td>
<td>Tamins</td>
<td>1,300</td>
<td>0.095</td>
<td>13.5</td>
<td>17.08</td>
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<td>Pamir</td>
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<td>0.24</td>
<td>6.3</td>
<td>17.34</td>
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<td>34</td>
<td>Engelberg</td>
<td>2,500 to 3,000</td>
<td>0.22</td>
<td>6.4</td>
<td>17.31</td>
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<td>35</td>
<td>Flims</td>
<td>12,000</td>
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<td>15.9</td>
<td>18.23</td>
<td>-</td>
<td>-</td>
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<tr>
<td>36</td>
<td>Sai Darreh</td>
<td>20,000</td>
<td>0.08</td>
<td>270</td>
<td>18.99</td>
<td>17.594</td>
<td>-</td>
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<td>37</td>
<td>Tsilokosky</td>
<td>1,200,000</td>
<td>0.06</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

Principally from Hsu (1975) and Scheideler (1973) with corrections and additions.

$^a$ value in parenthesis calculated from the displacement of the center of mass.
$^b$ recalculated from Hsu, 1975, Table 1 in most cases.
$^c$ calculated from total height of fall, avalanche volume, the gravitational acceleration, and an assumed debris density of 2000 kg m$^{-3}$ in most cases.
$^d$ calculated using estimates of the vertical displacements of the center of mass.
until it had depleted its kinetic energy, but it had far from run out of potential energy. That avalanche stopped because it had thinned until the force causing flow had dropped below the yield strength of the plastic debris. That is, the properties of an avalanche that control its spread are volume and yield strength, because the latter governs the final thickness of the deposit, and the former then governs the area covered. The shear stress at the base of an avalanche is not merely dependent on the debris thickness, but also on the surface slope. For avalanches that cover very large areas, the surface slope conforms closely to the basal slope, and the topography of the area over which the debris disperses therefore has an influence on debris thickness, and of course, it controls to a large extent the direction of motion that in turn controls the shape of the area covered, and through that, the maximum distance travelled.

The reasons for much of the scatter in the relationship expressed in Fig. 24 now become apparent. In calculation of potential energy, the vertical displacement of the centers of mass should have been used, rather than the sometimes very different measure from the top of the surviving scar to the furthest and lowest margin. Those avalanches that flowed upslope over parts of their paths (as most do, to some extent, in mountainous regions), had some energy restored to potential energy. This was particularly significant in the Vaiont slide (Müller, 1968, p. 15-16). And the shape of the area covered by each of the avalanches is determined by the local and regional
topography, and shape was not considered in "distance travelled". This topographic problem can be removed if the dispersal parameter of area is used, instead of maximum distance travelled (Fig. 24).

If rock avalanche debris truly behaves as a Bingham plastic, then a direct relationship must exist between the area covered by an avalanche, and the actual potential energy loss in the fall. This loss of potential energy is the physical measure of the actual work done in spreading the avalanche: it is not the cause of the spreading; that is a result of flow.

It remains now to predict the shear strength of avalanche debris. The Sherman Glacier rock avalanche apparently had a shear strength of 0.02 bar, and this value appears to be suitable for other much smaller avalanches in its vicinity as well. The Blackhawks slide had a yield stress of 0.2 bar, this value also appears suitable for the much smaller Elm slide, and these left deposits that are nearly an order of magnitude thicker than that of the Sherman Glacier rock avalanche. With an average thickness of about 100 m (Watson and Wright, 1969, p. 121), the Saidmarreh avalanche is nearly an order of magnitude thicker again and had a yield stress of 2 bar. The cause of these differences in yield stress is not immediately obvious; it may not be a simple function of mass because the Elm slide was smaller than the avalanche at Sherman Glacier. A compilation of avalanche thicknesses, and studies of variation in thickness within each avalanche may reveal a number of favored modes of yield stresses that can be utilized by flowing rock debris. This would
permit an accurate prediction of the reach of a given potential rock avalanche, if its yield stress and volume could be predicted.

In terms of prediction of avalanche hazard, it is the yield stress of avalanches that is important, and not the Bingham viscosity that is utilized during flow, because the latter only controls the rate of flow, and the time before the yield stress is reached, while the former determines where the avalanche will come to rest.

A mechanical reconstruction of the Sherman Glacier rock avalanche

The Sherman Glacier rock avalanche of 27 March 1964 was triggered by continued strong shaking during the Great Alaska Earthquake. The $10.1 \times 10^6$ m$^3$ of well jointed and fractured graywacke and argillite, and about $2 \times 10^6$ m$^3$ of snow and ice, was perhaps at least partially fluidized in the violent shaking the first dislodged snow and a small cirque glacier from Shattered Peak, and this was almost immediately followed by a flood of rock rubble that flowed and slid down all three sides of the peak.

The initially great thickness (about 100 m) and low permeability (perhaps about 10 - 100 millidarcy) so restricted the initial growth of a significant porosity that is a necessary part of strain in granular materials, that basal clasts were unable to roll. This maintained a high basal friction during the fall from Shattered Peak to Andres Glacier and resulted in a loss of about half of the total available energy to friction in the first 30 seconds of the 3.5 minutes of avalanching. When it hit Andres Glacier, the avalanche
was travelling at about 67 m s\(^{-1}\). Most of the debris was then deflected by an impact with a 150-m high spur and slid down the local slope of the snow-covered Andres Glacier before sliding across and down the regional slope of Sherman Glacier. About a quarter of the mass swept over the spur before reaching Sherman Glacier.

Throughout the avalanche, the debris behaved as a complex, perhaps dilatant, Bingham plastic with a yield stress of about 0.02 bar and two apparent Bingham viscosities. At thicknesses greater than about 2.5 m, the avalanche deformed with a Bingham viscosity of \(4 \times 10^7\) poise perhaps because thickness and permeability (now about 1 - 10 darcy) continued to limit the internal strain-rate. When the avalanche had thinned to about 2.5 m, the rate of flow of air to fill the growing void space was no longer a limiting factor, and only the low friction between violently jostling elastic clasts controlled the apparent Bingham viscosity (about \(1 \times 10^6\) poise) of the deforming avalanche. These very high apparent viscosities prevented the development of turbulence and kept deformation in the avalanche largely confined to shear at the base of the debris, so that for the most part, the avalanche slid as a thin flexible sheet, with solid rather than fluid behavior being the more visible. This was particularly so because of the additional effect of a rigid surface crust where the shear stress was low, and because the fluid flow was supercritical.

Along most of its path, the avalanche slid on snow with a coefficient of basal kinetic friction of 0.11, which suggests
that this avalanche may have quickly gathered a basal plaster of ice and compact snow and essentially slid with ice sliding on wet snow. Air, snow, and water forced out of the underlying snow (density initially about 450 kg m$^{-3}$) by the weight of the avalanche and by the reduced pressure within the avalanche, may have speeded the dilation of the debris, and allowed it to spread more thinly than it might otherwise have spread for debris of its permeability.

Collapse of pore space during settling of the debris near the end of the avalanche caused explosion of some of the air inhaled during dilation of the debris. The rapid transition from dilated debris with sliding and rolling kinetic friction to collapsed pore space and static friction between clasts in the avalanche brought the debris to a very sudden stop at the end of its run.

With the exception of a possible role played by abundant snow and ice in the Sherman Glacier rock avalanche, this mechanical reconstruction is applicable to all other major rock avalanches. These avalanches can not have been lubricated by sliding on discrete layers of compressed air because they lack a mechanism to trap an air layer and also lack the strength to maintain one if they could trap it. Instead, rock avalanches deform in supercritical laminar flow as a Bingham plastic with deformation largely confined to shear at the base, so that they slide without turbulence, essentially as thin flexible sheets, in direct contact with the ground, with little friction by rolling and sliding on clasts at the base of the avalanche.
CHAPTER II

EFFECTS OF A DEBRIS COVER UPON ABLATION AT SHERMAN GLACIER

2.1 Introduction

The most immediately obvious effect of the thick debris cover upon Sherman Glacier has been the substantial reduction in ablation beneath the cover, as compared with adjacent debris-free ice. This has caused the avalanche deposit to apparently rise above the surrounding glacier on a pedestal of ice that ranged between 10 m and 60 m high in 1969.

Early work at Sherman Glacier ignored ablation of ice and snow beneath and within the avalanche deposit (e.g. Shreve, 1966). When glaciological work was initiated, it was believed that 5 cm was a reliable estimate of the ablation (Bull and Marangunic, 1968, p. 312). Direct measurement of melting of 11.6±4 cm of ice beneath debris over 35 days in 1967 led Marangunic (1972, p. 97) to revise the estimate of annual ablation to 200 kg m\(^{-2}\) a\(^{-1}\). In 1970 I was able to obtain a direct measure of melting at one site of 420 kg m\(^{-2}\) a\(^{-1}\) (Section 2.3).

The very wide range in estimates of ablation suggested a need to examine the factors influencing ablation beneath debris, and to assess the areal variation in ablation beneath the avalanche deposit at Sherman Glacier. This chapter attempts to satisfy these needs with the very
limited amount of available data.

2.2 Previous work in other regions

Effects of thin and very-thin debris covers on melting of ice have long been studied (see studies by Sharp, 1949; Østrem, 1965; and McKenzie, 1969). Layers of debris less than a mean particle-diameter thick, or less than a few millimeters thick in the case of a sand or silt cover, cause increased melting over that of adjacent debris-free ice (Østrem, 1959, p. 228). This augmented ablation results from increased absorption of incident radiation. From sub-surface temperature measurements Østrem (op. cit.) estimated a debris-surface temperature of 3.6°C above local air temperature due to this cause.

Østrem (1965, p. 23), and many others, have noted that thicker debris-covers significantly retard ablation, but few have made useful measurements. A layer of sand and gravel 0.4 m thick reduced melting by 80 percent at a test area on Isfjellsåtna (Østrem, 1965, p. 24), but near the same locality Østrem (1965, p. 31) inferred from temperature profiles that some melting still takes place beneath 2.5 m of debris. At Grasubreen, Østrem (1965, p. 24) found no melting taking place in 1961 and 1962 beneath 15 m of debris of an ice-cored moraine. He (1965, p. 29) reasoned that buried ice will not melt if the cover is thick enough; i.e. thicker than the depth of summer thaw and the ice effectively lies in permafrost. This depth is related to local snow-cover conditions, and to winter and summer temperatures at the spot.

Sharp (1949, p. 298) found that insulating qualities of debris
also depend on grain-size and moisture content; a 3-cm cover of moist sand provides the same protection as 15 to 30 cm of coarse debris. Sharp (1949, p. 299) considered that the inferiority of coarse materials as insulators may be due in part to the ease with which air and water circulate through large openings, and gain access to the underlying ice. He stated that trickling rainwater causes considerable melting of ice even though the debris mantle may be thick, but he had made no determination of such melting, and no analysis of heat transfer.

McKenzie (1969, p. 422) estimated melting of ice beneath thick debris in a collapsing kame terrace where stagnant glacier-ice was buried by 3 to 6 m of sandy gravel from an estimate of rate and duration of conductive heat-transfer through debris. For a location where the debris was 3 m thick, McKenzie (1969, p. 423) computed an annual melt of 0.24 m, using a temperature gradient estimated from temperatures taken to a depth of 1.3 m, and assuming a conductivity for the debris of 1.5 W m$^{-1}$ deg$^{-1}$ and an ablation season of 5 months. He noted (op. cit., p. 422) that heat may also be transferred by percolation of rainwater, change of state, radiation, and convection within the gravel, but he did not estimate the amounts of heat transferred by any of these processes or the depth in the gravel to which the transfer might take place. He noted however (op. cit., p. 420), that rainwater percolation is an important mechanism of heat transfer from the heated gravel surface to the buried ice. Measured temperature changes at depth within gravel during periods of prolonged heavy rainfall led him to this deduction. My investigations of melting at Sherman Glacier suggest that
McKenzie may have greatly underestimated the rate of ablation of ice by underestimating both thermal conductivity and thermal gradient. Because many nonconductive heat transfer mechanisms operate near the soil surface, the thermal gradient that is necessary for calculation of melting is the gradient near the interface with the ice where conduction is the dominant process in heat flow (Section 2.6.1).

2.3 Melting of ice beneath debris at Sherman Glacier

At Sherman Glacier no reliable representative direct measurements of ice melt beneath the avalanche cover have been made. Buli and Marangunic (1968, p. 312) considered 5 cm a\(^{-1}\) was a reliable estimate based on conduction of heat through dry debris. However, Marangunic (1972, p. 97) measured melting of 11.6±4 cm (106 kg m\(^{-2}\)) beneath an artificial debris mound on Sherman Glacier over a period of 35 days during July and August 1967, by regarding as insignificant down-melting of very short aluminum ablation-poles. For another locality, near Marker 16 (Fig. 2), Marangunic (1972, p. 97) estimated melting of 200 kg m\(^{-2}\)a\(^{-1}\) by heat flow under wet and dry conditions and length of ablation season. He assumed heat flow to occur 7 hours a day for two months, but heat flow for 24 hours a day, perhaps for as much as 4 months each year in the area of his test site is more probable. This ablation interval however varies widely over the debris, depending chiefly on elevation and local snow-drifting.

In 1970, I used another method to assess ablation beneath the debris cover. A trench was excavated into the ice-pedestal that has
developed beneath debris, at the eastern side of the debris cover (Fig. 2), within a few meters of where a similar trench had been dug in 1967. The trench exposed winter snow from the 1963-1964 balance year, preserved as horizontally layered firn on top of vertically foliated glacier-ice by the overlying rock-avalanche debris. It had decreased in thickness by 1.4 m since 1967 (Fig. 25). Compaction of snow during metamorphism could be ignored because it had reached glacier-ice density by 1967. The decrease in firn thickness is an average loss of ice of 420 kg m\(^{-2}\)a\(^{-1}\). Local variation in thickness of 1963-1964 snow cover can not be adequately assessed and has been ignored.

### 2.4 Heat flow

In general, a flow of heat takes place whenever a difference of temperature exists between different parts of a body. This flow of heat is related to temperature difference by the law:

\[
h_i = -k_{ij} \frac{\partial T}{\partial x_j}
\]

(Equation 2.1)

where \(h_i\), components of the rate-of-heat-flow vector, measure the quantity of heat passing, in unit time, unit areas normal to three axes; \(\frac{\partial T}{\partial x_j}\) are components of the temperature gradient vector; and \(k_{ij}\) are coefficients relating these two vectors. \(k_{ij}\) thus form a second rank tensor and this is called thermal conductivity (where \(i\) and \(j\) are integers ranging from 1 to 3, denoting the three axes of the system).

Heat flow in debris can be calculated using a simplified law, but
Fig. 25. Cross-section of the eastern edge of the rock avalanche and underlying pedestal of snow and ice near marker 51 (1967 data from Marangunić, 1972, Fig. 30).
it must be remembered that this is then an approximation. The debris-cover approximates a thin planar sheet in which thermal gradient is normal to the plane of the sheet. Furthermore, thermal conductivity of a uniform aggregate is isotropic, even though individual components may be anisotropic in thermal properties. The rate-of-heat-flow vector must thus parallel the thermal gradient and be normal to the plane of the cover. The relationship between temperature gradient \( \frac{dT}{dx} \) with \( x \) vertical and heat flow \( h \) becomes

\[
h = -k \frac{dT}{dx}
\]

(Equation 2.2)

where \( k \) is the thermal conductivity of the debris.

The rate of flow of heat within a debris cover may be calculated if both thermal conductivity and temperature gradient in the debris are known. Ablation of ice beneath the debris cover is by heat flow through the base, so it is the thermal conductivity and temperature gradient at the base that is important to this work. When ablation-rate has been calculated, the added knowledge of time during which heat flows from debris to ice enables the annual ice loss to be estimated.

2.5 Thermal conductivity

Thermal conductivity of soil is difficult to determine with accuracy. Kersten (1966) found it to vary with grain-size, moisture content, density, composition, temperature, and state of moisture (solid, liquid, or vapor). Wechsler (1966) had difficulty in duplicat-
ing results, particularly at different heating rates and with different
designs of equipment. He found that laboratory determinations produce
more consistent results (within ± 8 percent for dry materials, and
uncalibrated equipment, and within ± 3 percent for calibrated equipment),
but soil conductivity is critically dependent on soil structure, such as
variations in void space, grain size, and moisture content with depth,
and is more reliably determined in situ. Wechsler found thermal conduc-
tivity of moist soils very difficult to determine, because of moisture
migration along the thermal gradient; a factor that gave erroneously
high conductivity from the measured rate of heat flow.

No in situ, or laboratory determinations of thermal conductivity
of rock avalanche debris from Sherman Glacier have been attempted. In
his assessment of ablation, Marangunic (1972, p. 95) utilized de Vries'
(1963) determination of "Fairbanks sand" under wet and dry conditions
(2.5 and 2.3 W m⁻¹deg⁻¹). Kersten (1966) also assessed thermal
conductivity of Fairbanks sand; he obtained a value of 1.48 W m⁻¹deg⁻¹
for 5 percent moisture by weight (11 percent by volume). Kersten's
value is consistent with Wechsler's (1966) and his own determinations
of other similar materials. McKenzie (1969, p. 423) selected
1.5 W m⁻¹deg⁻¹ as a reasonable estimate for gravel in a kame terrace in
Glacier Bay on the basis of Wechsler's and Kersten's analyses.

Values of in situ thermal conductivity of water-saturated thawed
permafrost (K. Everett, personal communication) vary from 0.7 to 2.28
W m⁻¹deg⁻¹. Thermal conductivity of these saturated soils varies
inversely with moisture content expressed as a percentage of dry weight.
Materials such as peat, with moisture contents exceeding 100 percent of dry weight, have low conductivities, asymptotically approaching that of pure water (0.56 W m\(^{-1}\) deg\(^{-1}\)), while less porous soils, with less water, have higher conductivities, approaching that of a non-porous rock made up of the soil’s minerals.

This inverse relationship becomes linear when water content is expressed as a proportion of total wet volume. This is not unexpected, because thermal conductivity of a uniform mixture is the mean of the constituents’ conductivities, weighted in accordance with the volumes of constituents in the mixture, provided their thermal properties are actually or statistically isotropic.

2.5.1 Thermal conductivity and porosity

Consider a mixture of crushed greywacke, water, and air. Water has a thermal conductivity of about 0.6 W m\(^{-1}\) deg\(^{-1}\), and air, approximately 0.1 W m\(^{-1}\) deg\(^{-1}\). Thermal conductivity of greywacke has a range of possible values, but 2.3 W m\(^{-1}\) deg\(^{-1}\) (a reasonable estimate for material of granitic composition, Clark, 1966, p. 461) is probably representative.

The contribution of rock to soil conductivity varies linearly with the volume of rock present in the soil, and hence varies linearly with soil porosity. For grains of uniform lithology, Fraser (1935, p. 917) found soil porosity to depend on grain-size, and grain-size distribution, grain shape, method of deposition, and the many changes that follow deposition. He found porosity to decrease as grain-size increases (for natural materials) and to increase as sorting increases.
He also found porosity of uniform sized crushed materials to be greater than that of similar rounded materials. Porosity of freshly poured, loose, dry sand is greater than that of sand that has been compacted by vibration (Fraser, 1935, p. 936). Some of Fraser's data is presented in Table 6.

Thermal conductivity of soil varies more with variation in porosity than with water content (Fig. 26), but the range of possible conductivity values, for realistic values of porosity and water content, is small. For a soil composed of crushed greywacke, a thermal conductivity in the range of 1.5 to 2.0 W m$^{-1}$deg$^{-1}$ can be expected: the lower limit in very porous, well drained soils, and the upper limit in less porous, and poorly drained soils.

Porosity, permeability, and water saturation of the debris cover on Sherman Glacier have not been determined, but porosity must exceed 20 percent because Marangunić (1972, p. 95) recorded a water content of 20 percent by volume in one sample. A porosity of 30 percent would be reasonable for a material of such poor sorting, but it may vary spatially, perhaps by as much as 20 percentage points.

Porosity has changed with time as finer particles have been flushed by rain and melting snow, from the surface to concentrate at depth. Thus, thermal conductivity at the base has probably increased slightly with time. Variation in water content at the base is unimportant because the debris is in contact with melting ice and must be constantly at or near saturation.
<table>
<thead>
<tr>
<th>Material</th>
<th>Porosity</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>loose</td>
</tr>
<tr>
<td>mixed spheres (3 sizes)</td>
<td>30-38</td>
</tr>
<tr>
<td>standard sand</td>
<td>39</td>
</tr>
<tr>
<td>uniform spheres</td>
<td>40-43</td>
</tr>
<tr>
<td>uniform crushed quartz</td>
<td>48</td>
</tr>
</tbody>
</table>

Table 6. Porosity (percent) of various aggregates for loose and compacted conditions (data from Fraser, 1935)
Fig. 26. Variation of thermal conductivity of an ideal soil with variation in porosity and water content (percent of saturation).
A thermal conductivity of $1.8 \text{ W m}^{-1}\text{deg}^{-1}$ has been chosen as a suitable value for the basal debris on Sherman Glacier. This value is unfortunately critically dependent on the choice of $2.3 \text{ W m}^{-1}\text{deg}^{-1}$ as a suitable conductivity for the parent rock.

2.5.2 Thermal conductivity and permeability

Thermal conductivity of soil is unaffected by permeability. However, water flow is controlled by permeability, and advection of heat by flowing water is an efficient heat-transfer mechanism. As eluviated fine debris has clogged pore passages, permeability has decreased, and advection of heat through the base has declined over much of the cover. Some advection of heat must still persist in the basal debris because a hydraulic gradient is present in the saturated debris and "ground water" flow exists. This ground water is drained by a sub-debris pseudo-karst drainage system within the glacier. Ablation by advective heat flow through ground water movement is concentrated within this drainage and, while ablation by this mechanism can not be seen, its effects are plain to see in local collapse of the debris surface, and in cavern systems exposed in crevasse sections.

Estimation of advective heat flow through the basal debris (Section 2.9) shows that conduction is the dominating heat transfer mechanism at the base.

2.5.3 Glacier flow and conductivity

Glacier flow can change thermal conductivity of a loose debris
cover upon the glacier surface through mechanical properties of loose aggregates. An uncompacted aggregate of loose clasts is compressible. If debris overlies an area of glacier ice in compressive flow, pore space will diminish, while debris over slightly extending flow will have increasing pore space. If the extensive flow is sufficient to produce crevasses, the cover is completely disrupted. Lateral shear at the ice surface also shears the debris, and dilation through shear of one clast past another must maintain a high porosity within shearing debris.

The gradient of the curve of soil conductivity versus porosity is very low \((-2.2 \times 10^{-4}\) in Fig. 26) and expected changes in porosity are small, thus the effect of spatial changes in glacier flow on ablation generally is too small to be considered in an assessment of mass balance, but it may be important in long-term mass wastage and eventual production of ice-disintegration landforms.

Crevasses in the glacier beneath the debris, on the other hand, have a pronounced effect upon heat flow and ablation. The avalanche fell in late March 1964 and swept over a surface where crevasses were snow-filled or snow-covered. Flow of ice beneath the debris has opened new crevasses, and old snow-bridges have collapsed. Debris continues to cascade into crevasses that are maintained as shallow V-shaped channels 1 to 15 m deep, floored by boulders large enough to span the crevasse widths. The ice of the crevasse walls is only thinly covered, but only the deepest crevasses have ice exposed in them. Reduced ablation under the thick debris on the glacier surface has combined with augmented ablation under the thin debris on crevasse walls to produce a distinct-
tive, hummocky topography in areas of intense crevassing, such as the southwestern sector of the debris (Fig. 29). The growth of crevasse etching from 1964 to 1973 has been one of the more noticeable changes in the landscape of the avalanche surface (Fig. 28).

The volume of ice lost from crevasse walls has not been assessed. The rate of loss varies from crevasse to crevasse as crevasse orientation and thickness of debris cover vary, and as debris cascades down their walls. It has significantly increased in magnitude with time, as crevasses have grown wider, and as new ones have formed. The volume of ice lost from crevasse walls in 1964 may have been negligible, but, by 1971, in most crevassed areas, more ice was being lost from crevasses than from inter-crevasse surfaces, for by that time, inter-crevasse surfaces often amounted to less than a quarter of the local area.

2.6 Thermal gradients in the debris cover

In 1967, C. Marangunić measured several temperature profiles within the debris on Sherman Glacier. At one trench (Fig. 25), temperature was measured at four points in the lower 50 cm of the 2 m-thick debris. A line fitted to these data (Fig. 29b) gives a thermal gradient of -8.8 deg m⁻¹ at the ice-debris interface. Another temperature profile (Fig. 29a and 30), in 1.1 m of debris near marker 16 (Fig. 2), yields an even higher thermal gradient of -18.9 deg m⁻¹ at the base.

2.6.1 Factors affecting thermal gradients

A more open, well drained meshwork of debris near the surface
Fig. 27. Hummocky topography developed in an area of intense crevassing in the southwestern sector of the debris.
Fig. 28. Growth of crevasse etching through the debris cover, 1964 to 1973. (aerial photography by Austin S. Post)
Fig. 29. Basal temperature profiles: A) from a pit near marker 16, 15 August 1967 (air temperature - 16.7°C) and B) from a trench near marker 51, 26 August 1967 (air temperature - 4°C). Data by Č. Marangunic.
should have a lower conductivity than clogged, saturated debris at the base. Hence, if conduction alone is the heat transfer mechanism within all of the debris, the temperature profile of Fig. 30 can not be an equilibrium one: more heat leaves the base than passes through the mid-section.

At the trench site, temperature 1.5 m below the debris surface was half a degree warmer than surface air temperature. A more complete temperature profile at this site could not be obtained without serious errors induced by local topography. A complete profile was obtained near marker 16 (Fig. 2) that has a very high gradient of 40 deg m⁻¹ in the upper 15 cm that is a result of an unusually high surface temperature (13.7°C). Meteorological records (Marangunic, 1972, p. 35-62) show no direct sunshine on the day of measurement, and maximum air temperature for that day, at a nearby standard meteorological screen at the same elevation, was only 10°C. The surface must have been heated nearly 4°C by diffuse radiation emitted by, and transmitted through cloud cover. Bull and Marangunic (1968, p. 312) found heating of as much as 10 degrees on days of sunshine (Fig. 34). Hardly more than one hour of sunshine was recorded in 21 days preceding the measurements, shown in Fig. 30, so direct beam solar heating can not be a factor in maintaining the upper 50 cm of the profile above mean air temperature for that 21-day period; this heating too, must be by absorption of diffuse radiation from cloud cover.

During the 21 days preceding measurement of the marker 16 profile on 15 August, 1967, 210 mm of rain fell at the base camp, almost all of
Fig. 3C. Temperature profile in pit west of marker 16, on 15 August 1967 - total cloud cover, no sunshine, no rain (data by Č. Marangunid).
this came in only 8 of these days, but traces fell on all other days. In the two days preceding measurement of the trench profile on 26 August, about 140 mm of rain fell, and at least a trace of rain had fallen on all but two of the preceding 31 days. These same 31 days brought almost 16 hours of sunshine to Sherman Glacier. This weather is typical for summers at Sherman Glacier, thus, with respect to all meteorological parameters, these two measured temperature profiles should be typical of normal late summer conditions within the Sherman Glacier rock-avalanche debris, not only for 1967, but for all years of study.

Advection of heat by infiltrating rainfall is an obvious addition to heat transfer by conduction. Take, for example, 140 mm of rain falling at a mean air temperature of 7°C on a surface heated 4 degrees above air temperature. As this rainwater percolates through the debris, it carries enough heat to melt almost 20 kg m⁻² of ice, and this can occur in only two days! In 1970 rainfall averaged 33 mm per day over 45 days! Because temperatures were similar in both 1967 and 1970, advected heat could have melted some 200 kg m⁻² of ice in 45 days in 1970.

Heat is also advected by air currents within the upper debris which is permeable enough at the surface that a perceptible flow of air within it was felt at several localities, but this was never closely investigated. Almost-constant wind, and innumerable protuberances on the surface to promote extreme turbulence and locally strong pressure gradients, ensure that air flow is significant in heat transfer in the
upper debris in most areas.

Marangunić (1972, p. 49) noted that supersaturation occurs often in air over debris on Sherman Glacier and is a characteristic of the eastern edge of the debris cover, where chilled air from the ice surface mixes with warm air over the debris surface. Where air circulates within the debris, saturation and supersaturation of air also occur, as warm saturated air from the wet debris surface is carried into the cooler interior of the debris cover. Condensation from this air would contribute greatly to heat transfer within the debris.

All of these factors contribute to advective heat flow and are important in the upper part of the debris, but near the base, the high thermal gradient forces most heat to flow by conduction, and low permeability inhibits advection.

There is insufficient data to analyse the important effects of variation in temperature with time, from day to day during the year, and from year to year.

2.6.2 Areal variation in thermal gradient

The debris cover, below the equilibrium line on Sherman Glacier, ranges 250 m in elevation, from 200 to 450 m above sea level. At a standard adiabatic lapse rate (-0.01 deg m⁻¹), a difference of 2.5 degrees might be expected in temperatures at the upper and lower extremes. Bull and Marangunić (1968, p. 312) however found a progressive 10 degree warming of air passing over the debris, in excess of such
adiabatic heating (Fig. 34), under unusual conditions of full sun and slight breeze. At least half of this amount of heating can occur under cloud cover (Section 2.6.1). Perhaps, then, a surface temperature difference of 7.5 degrees (the $2.5^\circ$ difference from the lapse rate plus half of the $10^\circ$ warming) could be expected over the debris.

The two measured profiles suggest that thickness of debris plays a role in variation of basal thermal gradient; i.e. the gradient at the base of 2 m of debris is one half that below 1 m of debris, which is closer to the average thickness of the avalanche cover in this region.

Profile A (Fig. 29) might be considered as representative of the temperature profile with depth at the eastern, upper end of the debris cover. The other profile (B) is probably representative only of the extreme margin of the avalanche cover, the pock-marked raised rim along the upper margin (Fig. 1), where debris thickness was doubled by overriding or "bulldozing" during the avalanche.

The temperature profile at the lower distal end of the avalanche cover can only be crudely estimated, because no measurements of it were made. The same processes operate to transfer heat in the upper layers there and a similar, though slightly steeper thermal gradient, might be expected because of the higher surface temperature. If this is so, then the basal thermal gradient will be far higher at the lower, distal end of the avalanche cover than at the proximal end (Fig. 31, curve B).
Fig. 31. Hypothetical temperature profiles in one meter of debris
A) at 400 m elevation, and B) at 200 m elevation.
2.7 Heat flow and ablation rate beneath debris at Sherman Glacier

Heat flow by conduction, and, hence, ablation rate vary linearly as thermal conductivity or thermal gradient separately vary (Equation 2.2 - Section 2.4). Variation in heat flow and ablation rate with varying thermal gradient for several values of thermal conductivity are shown in Fig. 32. From this graph, a conductivity of 1.8 W m⁻¹ deg⁻¹ provides a heat flow of 15.8 W m⁻² at the trench site, and 34 W m⁻² at the pit near marker 16 (Fig. 2), the site chosen as most representative of the upper area of debris. At the lower end of the cover, heat flow might be as high as 65 W m⁻². Daily ablation rates for these heat flows respectively are 4.1, 3.8, and 16.8 mm (water equivalent of ice) (heat of fusion of ice is 334 J g⁻¹).

2.8 Total ablation by conduction beneath debris at Sherman Glacier

The ice loss of 420 kg m⁻² a⁻¹ at the trench site (Section 2.3) could be accomplished in about 100 days at the calculated rate of 4.1 mm per day. The onset of ablation is not an abrupt change from zero to half a gram of water per day, and cessation is not abrupt either; both occur with regional seasonal temperature changes. Thus, ablation season must be greater than 100 days; however, climatic records from nearby Cordova Airport (Cordova FAA) indicate that it is less than 150 days (U.S. Weather Bureau, 1963).

If the measured 420 kg m⁻² a⁻¹ ice loss occurs from an effective heat flow of 15.8 W m⁻² over 100 days, then 880 kg m⁻² a⁻¹ might be lost to a heat flow of 34 W m⁻² at 400 m elevation in a similar period. At
Fig. 32. Heat flow and ablation rate for varying values of thermal conductivity, $k$, and thermal gradient.
200 m elevation at least 1.7 Mg m\(^{-2}\) a\(^{-1}\) of ice might be melted by conductive heat flow, although at this lower elevation the ablation season would be longer.

The average rate of loss of ice beneath avalanche debris cover on Sherman Glacier is probably about 1 Mg m\(^{-2}\) a\(^{-1}\). This is ice lost through melting by conducted heat alone, assuming a uniform conductivity at the base of the debris of 1.8 W m\(^{-1}\) deg\(^{-1}\).

2.9 Ablation by advection beneath debris at Sherman Glacier

Ablation by advected heat carried by the "ground-water" flow within the debris may locally be important. How much ice is lost by this process cannot be fully assessed: neither the temperature nor amount of flow are known with any precision, but some estimates can be made.

The amount of ice ablated by flowing water depends upon the amount and temperature of water flowing from the debris. Water flow comes from direct rainfall and from melting of winter snow on the debris surface, but heat carried by snow-melt is small and can be neglected. Rainfall averages close to 500 mm a month during summer on Sherman Glacier, so, in 4 months, runoff of 2 Mg m\(^{-2}\) can be expected. If the bottom 15 cm of debris is saturated and water flow into the glacier is tapped from the top of this layer, then a water temperature of 2\(^\circ\)C can be expected near the upper end of the avalanche cover (from Fig. 29). This could melt 50 kg m\(^{-2}\) a\(^{-1}\) of ice. Although this is insignificant when compared with melting by conducted heat over the same interval, it is not spread over the entire area of debris but is confined to where
water penetrates to glacial ice. Over an area of 8.25 km² of debris cover, this ablation rate amounts to more than half a million cubic meters of ice lost each year from channels, fissures, and tunnels within the glacier.

2.10 Additional melting below the debris cover

Melting of ice within the glacier also occurs through frictional heat generated by the loss of potential energy of water flowing in the glacier and by ice flow (Table 7).

All of the melt below the surface is hidden from view in tunnels within the glacier, but it amounts to 16 percent of the actual mass loss from the glacier near the upper edge of the cover. Before the avalanche, when net balance was about 4,000 kg m⁻² a⁻¹ at the surface in this region, this internal melting was only 4 percent of the total mass loss. At that time, a 28 kg m⁻² a⁻¹ increase in melting from the increased runoff from the surface was more than offset by an absence of advective heat transfer from the warmed, low albedo, rock debris surface which now removes 50 kg m⁻² a⁻¹.

2.11 Effect of large clasts on heat flow and ablation

An unknown percentage of the cover on Sherman Glacier is made up of clasts whose diameters exceed the estimated mean thickness of debris (1.65 m), so the model of a uniform thin sheet of constant thermal conductivity at its base, upon which all of the calculations are based, is not fully correct.
Runoff from the debris area

- Summer melt by conduction through debris: 880
- Summer melt by advection through debris: 50
- Summer melt of winter snow: 2000
- Summer rainfall: 2000
- Total runoff: 4930
- Height of fall (m): 300
- Potential energy loss ($J \text{ m}^{-2} \text{a}^{-1}$): $1.45 \times 10^7$
- Ice melt by frictional heating in the energy loss: 43

Runoff from above the debris area

- Estimated summer melt and rainfall: 3000
- Area of catchment ($\text{km}^2$): 43.5
- Drop in height of the subglacial stream (m): 100
- Potential energy loss beneath the debris ($J \text{ a}^{-1}$): $1.28 \times 10^{14}$
- Ice melt by frictional heating in the energy loss: 46.3

Ice flow

- Average velocity through the glacier ($\text{m a}^{-1}$): 50
- Shear stress at the base ($\text{Newton m}^2$): $1.20 \times 10^5$
- Work done ($J \text{ m}^{-1} \text{a}^{-1}$): $6 \times 10^3$
- Ice melted by friction of ice flow: 18

Other melting

- By geothermal heat flow: 5
- By advection (Section 2.9): 50

Total ice melt below surface: 163

Total ice melt above surface (Section 2.8): 880

Table 7. Melting within the glacier beneath the debris cover (units are kg m$^{-2}$ a$^{-1}$ unless indicated).
The parent rock of the avalanche is nearly impermeable to water and air, so heat flow through large clasts is by conduction alone. For a clast 1.65 m thick, and surrounded by finer debris, with a surface temperature of 9°C, a thermal gradient of ≥-5.5 deg m⁻¹ and a heat flow of ≈ 13.0 W m⁻² would exist; this would melt ≈ 3.4 mm of ice per day (Fig. 32), and ≈ 340 kg m⁻² a⁻¹ in a hundred day ablation season. This is less than half of that beneath surrounding loose debris (880 kg m⁻² a⁻¹, Section 2.8).

That a material of higher conductivity can be a more effective insulator than materials of lower conductivity, is confirmed by observation that, with time, large boulders do appear to rise from the avalanche cover, as ice beneath surrounding finer debris is ablated more rapidly than beneath the larger clasts. For those even larger than mean thickness of the debris, and for large ones atop of the debris, this difference in ablation is even more marked. Heat flow through them is lower, but they also intercept more rainfall than do flat sections of debris that do not protrude into the near-horizontal rain carried by a near-constant down-glacier wind. Runoff from this rain pours off their edges and carries more heat to the base of the surrounding debris.

At the largest boulder (visible in Fig. 1) on the glacier surface, two additional effects cause higher ablation beneath debris surrounding it. First, because it is 17 m high, it always protrudes through the winter snow cover, and acts like any area of bare rock in absorbing solar radiation and reflecting and radiating heat to cause accelerated melting of surrounding snow. Thus, early in summer, a circular trough
is developed about it, forming a wind-free radiation trap that warms the debris surface. Second, the clast interfered with debris flow during the avalanche, resulting in slight thinning of debris immediately adjacent to it.

2.12 Other aspects of ablation in debris-covered areas

Melting beneath the thickened rim at the upper end of the rock-avalanche cover is only half that beneath most debris near that area (Section 2.8). This differential ablation is about half a meter of ice each year. A raised rim (Fig. 3), that initially was little more than 1 m higher than the interior debris surface in 1964 progressively grew by differential ablation to stand some 5 m above the surrounding debris in 1971. This ridge also accumulates wind-drifted snow, and snowdrifts now persist into late summer on its leeward western side. This persistent cold blanket also reduces ablation rate beneath the cover, and serves to accentuate growth of the western limb of the ridge and hence of the ridge itself. The locally steep, ice-surface gradients that have developed through differential ablation also serve to control the movement of "groundwater" that can carry heat from high points to low points and thus serve to accentuate the relief.

2.13 An independent check of ablation rate

Č. Marangunić measured firn stratigraphy (Fig. 33) at two pits excavated in avalanche debris and underlying snow on Sherman Glacier in an area west of marker 16 (Fig. 2). Although these were measured a year apart, on 15 August 1966 and on 15 August 1967, he did not compare
Fig. 33. Stratigraphic sections from pits west and northwest of marker 16. (data by Č. Marangunić)
If snow-cover in the region of marker 16 was uniform in the winter of 1963-1964 (this assumption is examined later), melting beneath debris in this area can be assessed directly through the change in firn thickness. Marangunić measured 2.27 m of firn in 1966, and 1.21 m in 1967: a loss of 1.06 m. He recorded firn densities of between 840 and 900 kg m\(^{-3}\) in 1966 but, in 1967, using .5 liter density tubes, he obtained densities ranging from 700 to 760 kg m\(^{-3}\). A ten-centimeter-sided cube of ice cut in 1967 gave a density of 800 kg m\(^{-3}\), and this probably is the more reliable of the 1967 values, but the discrepancy between 1966 and 1967 is perplexing. The difference, if real, is not due to snow metamorphism because that should increase the density.

If the 1.06 m of ice lost had a density of 800 kg m\(^{-3}\), then 850 kg m\(^{-2}\)a\(^{-1}\) was lost; on the other hand the loss would be 910 kg m\(^{-2}\)a\(^{-1}\) if the density was 860 kg m\(^{-3}\) (the mean for 1966). The mean of these values is 880 kg m\(^{-2}\)a\(^{-1}\) which is the same as the calculated value (Section 2.8). The equality of these values is fortuitous.

The possibility that snow-cover in the vicinity of marker 16 was non-uniform in the winter of 1963-1964 can be assessed from Marangunić's snow stratigraphy for the two pits (Fig. 33). If "icy crusts" have lateral continuity, and persist in time, stratigraphies from the two pits can be correlated, as has been done in the figure. Such a correlation as I have made suggests that no uniformity of snow-cover existed, and implies melting that does not exceed 200 kg m\(^{-2}\)a\(^{-1}\) for
density 800 kg m\(^{-3}\) (250 kg m\(^{-2}\) a\(^{-1}\) for density of 860 kg m\(^{-3}\)). Marangunic's estimate was 200 kg m\(^{-2}\) a\(^{-1}\) for this locality, but he used incorrect data in his estimate (Section 2.3), and it is also less than half of my direct measurement at a site where the basal thermal gradient was only half that at this site. Marangunic did not use his measured sections to estimate ice loss.

My estimates of ablation rate suggest that by 1974 firnified snow from the 1963-1964 winter was only present beneath the eastern rim of the debris cover, where it has been able to survive ablation since 1964, and locally where it may have been exceptionally thick, perhaps as a crevasse filling. Marangunic's estimate, on the other hand, suggests that the firn was still widespread by 1974 because of the low ablation. Of course, my analysis is inapplicable to those few small areas of the debris that often retain a snow-cover on the debris surface from one year to the next, and which can have only negligible ablation beneath.

Without a re-examination of the location within the next few years, the problem probably will never be resolved, because the reference horizon provided by a summer surface beneath firn will be gone.

2.14 Ablation at the terminus

Bull and Marangunic (1968, p. 312) attributed the difference between their measured net ablation at the glacier terminus of 11 m, and Field's (1965, p. 330) estimated value of 6.1 to 7.9 m for net ablation down-glacier of the rock avalanche cover, to a secondary effect of the cover. They found that absorption of solar radiation by
Debris produces a 10 degree increase in near-surface air temperature on clear days (Fig. 34) and thought that heated air would produce much more than normal melting as it travelled over the narrow, debris-free zone at the glacier terminus. However, Field's estimate was based on rates of retreat of the terminus, surface slope, and assumed ice velocities, before any determination of ice velocity had been made at Sherman Glacier.

Air temperature 40 cm above marker 36 is not noticeably different from that to be expected if debris was not present on the glacier (Fig. 34). Ablation rate at this point can not be significantly different now from what it would have been had the avalanche not occurred. If net ablation rate at markers 36 and 52 is unaffected by the debris cover, as the evidence presented suggests (Fig. 35), then that at marker 35 must similarly be unaffected, since it lies on a linear interpolation between them. A direct relationship also exists between net ablation and elevation along the center flow-line of the glacier (Fig. 39, Section 3.1.3). A least-squares straight line fitted to all of the data is not significantly different from a line fitted to data for only the area above the debris. Thus there is no justification for concluding that the data from below the debris zone belong to a distinct data set.

Rock-avalanche debris on Sherman Glacier has no noticeable effect on ablation down-glacier from the debris. Either the heated air layer is thin and the quantity of heat that warms the air is small, or little or none of the heated air reaches the ice. The latter may be a more
Fig. 34. Profile of temperature differences of air 0.4 m above the surface of Sherman Glacier, after correction for daily variations and for adiabatic cooling, on clear days with slight breeze, in the summer of 1966. Temperature of the terminus has been used as a reference temperature. (adapted from Bull and Marangunic, 1968, Figure 4).
Fig. 35. Net balance along the center line of flow on Sherman Glacier, August 1965 to August 1966. (Adapted from Bull and Marangunic, 1968, Fig. 2)
important factor, but the former is probably also true.

2.15 Discussion and summary

In many areas it is not feasible or even possible to directly assess the rate of melting of ice beneath a debris cover, yet this rate is essential in calculation of glacier mass balance (Chapter III) and in determining the cause of changes in surface elevation of the debris (Section 4.3.8e). I have shown that from simple measurements of temperature profiles within the debris, and of porosity and water content of the basal debris, estimates of melting can be made that compare favorably with direct measurements where they are available.

Calculated heat flow through debris shows that melting beneath debris at Sherman Glacier has been greatly underestimated in the past. Direct measurement at one atypical site gave an annual ablation of 420 kg m\(^{-2}\), twice the previous highest estimate of 200 kg m\(^{-2}\) (Marangunic, 1972, p. 97). Analysis of factors affecting thermal conductivity and temperature gradients within the debris, suggests that this direct measurement is less than half the ablation rate beneath debris of average thickness near the measurement site (880 kg m\(^{-2}\) a\(^{-1}\)), and that, at lower elevations, it is yet double again (1.7 Mg m\(^{-2}\) a\(^{-1}\)).

Direct measurement at ablation poles up-glacier of the debris cover gives a net balance of around -4 Mg m\(^{-2}\) a\(^{-1}\) (Fig. 37). Thus debris cover has inhibited ablation beneath it by about 80 percent at 400 m elevation. Down-glacier of the debris, net balance at about 200 m is around -8 Mg m\(^{-2}\) a\(^{-1}\) (Fig. 36); again ablation beneath the
debris has been inhibited by about 80 percent.

Percolating rainwater and even air-flow also carry heat from the surface into the debris, but they are not important at the base. Heat advected from the top of the "groundwater" table, by drainage of water from the debris into the glacier, may melt about $0.5 \times 10^6 \, m^3 \cdot a^{-1}$ from tunnels, fissures, and channels, but this is only 5 percent of the melting due to conduction.

Differential melting of ice beneath boulders, as compared with melting beneath adjacent fine debris, causes the boulders to rise through the debris by perhaps as much as $0.5 \, m \cdot a^{-1}$. This is entirely due to the absence of advective heat flow in the large boulders. Other effects of differential melting can be seen where debris locally is thicker, such as at some margins, and where locally it is very thin, such as on crevasse walls.

The debris apparently does not increase melting beyond its margins. Where the margins of the cover have been extended by debris cascading down the ice pedestal upon which it now stands, accelerated melting takes place, but no measurements have been made. Elsewhere melting is still the same as it would be if the avalanche had not occurred.

The 80 percent reduction in melting over this large area of Sherman Glacier has significantly altered the glacier mass balance.
CHAPTER III

MASS BALANCE AT SHERMAN GLACIER

3.1 Direct assessment of balance

Sherman Glacier was not chosen for study because of its suitability for mass-balance work, but any study of response of a glacier to change in balance regime requires some knowledge of mass balance, both present and immediate past. To this end, measurements of some balance parameters were made at Sherman Glacier from 1965 to 1971, and these have been related to climatological data for nearby Cordova Airport to obtain an estimate of balance trend over the last 30 years.

Mass-balance terminology recommended to the International Commission of Snow and Ice of the International Association of Scientific Hydrology (UNESCO/IASH, 1970) has been adhered to whenever possible. Available data, however, are incomplete and inadequate for a standard assessment of mass balance.

Determination of mass balance at Sherman Glacier has presented a number of special problems that are seldom discussed in descriptions of mass-balance measurements, yet these problems can not be unique to this glacier.
3.1.1 Some problems and solutions

Size of glacier has been perhaps the largest problem to be dealt with. With an area of close to 55 km², Sherman Glacier exceeds the recommended upper limit of size for mass-balance studies (Østrem and Stanley, 1969, p. 1) by 40 km². Although, for the most part, mass-balance measurements have been carried out by only one or two people, size of glacier has not prevented an adequate coverage of accessible areas. Other problems placed more stringent limitations on the work.

Access presented a major problem; the area of positive net-balance on Sherman Glacier is a series of largely inaccessible tributaries, and this has prevented precise mass-balance work.

The problem of access was insurmountable in most years and had to be overcome by a technique that has not been checked. In 1967, a helicopter was available for a very limited time for transport into higher tributaries, and net-snow-accumulation within them, near the end of the 1966-1967 measurement year, was estimated by Marangunić (1968, p.109) from snow cores and rod soundings. In later years, and in reassessing earlier years, I have assumed that the reasonable pattern of balance, established by Marangunić for 1967, has remained constant, and that balance for all tributaries can be inferred from balance measured in two readily accessible tributaries that present little hazard to the lone investigator. Because this assumption has remained untested, no attempt at standard error analysis has been made for the balance work. Instead, climatological records have been analysed to see if measured
balances are in accord with balances estimated from the climatic record.

Direct mass-balance studies at Sherman Glacier have been limited to assessment of annual and net balances because the field seasons have covered only unknown portions (perhaps less than half) of the total summer melt periods so that individual summer and winter balances are unknown.

Periods when studies have been carried out at Sherman Glacier (Table 8) are only loosely related to balance. August is the only interval common to balance and academic years that permits ablation poles to be visible above winter snow, and an investigator to be present to observe them. Times, in the balance year, of balance maximum and balance minimum, are unknown, and neither winter nor summer balance may be determined directly for any year.

In standard mass-balance determinations, two time systems of measurement may be used: 1) a stratigraphic system; and 2) a fixed-date system (UNESCO/IASH, 1970, p. 18). The two systems can not be combined without, in general, introducing large errors, for the two systems can give equivalent results only if balance is monitored continuously through a year; a rare situation in which choice of system is left to the whim of the observer.

Circumstances can arise, however, that make separation of systems impossible. In each year of study at Sherman Glacier, separation was prevented by a variety of factors. Ablation at ablation poles all too often exceeded the depths to which poles were drilled, while accumulation
<table>
<thead>
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</tr>
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<td>1969</td>
<td>5 July</td>
<td>21 August</td>
<td>McSaveney</td>
</tr>
<tr>
<td>1970</td>
<td>1 July</td>
<td>24 August</td>
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</tr>
<tr>
<td>1971</td>
<td>31 July</td>
<td>24 August</td>
<td>McSaveney</td>
</tr>
</tbody>
</table>

Table 8. Glaciological studies at Sherman Glacier
almost always exceeded the height of accumulation poles.

In general, mass balances of different parts of Sherman Glacier have been assessed in slightly differing time systems. In all, four, and perhaps five systems have been used, and mass-balance work on the glacier has been far from standard. The four systems have been applied to four distinct areas of the glacier, each characterized by a distinct balance regime.

The smallest area, that between avalanche cover and glacier terminus, is characterized by very high ablation and a progressively shrinking area. An estimate of annual balance was made for this area by Marangunić (1968, p.109) for the time period 22 July, 1966 to 15 August, 1967, from 6 poles, none of which remained standing at the end of the measurement interval. Poles in the area, reset in 1967 for ablation known to exceed 10 meters of ice each year, were never recovered. My assessments of annual balance for the glacier have made use of Marangunić’s estimates for this region, with due allowance for effect of climatic change. Ice velocity and the change in profile of the terminus from 1967 to 1969 (Fig. 104) suggest an ablation rate of 11 m of ice each year so that Marangunić’s estimates apparently are reliable.

It is unfortunate that, although this area is only some 3 percent of the glacier area, it contributes 20 to 25 percent of the annual net loss of mass below the equilibrium-line. In time, as the area is greatly reduced, its contribution to mass balance will decline significantly, but this has not yet occurred.
A modified fixed-date system has been applied to the area beneath avalanche cover. Balance beneath the cover has been discussed in the previous chapter. Ice-loss in this area is determined from measurements of the distance between the summer surface of 1963 and the ice-debris interface at known dates. It has been assumed to be constant from year to year.

Although there is some uncertainty in balance beneath the avalanche cover, the rate of loss of ice is sufficiently slow, that, despite a large area of cover (15 percent of total glacier area), balance for this area is only about 15 percent of total annual balance for the ablation zone.

Annual balance in the remainder of the ablation area has been determined in a true fixed-date system, with 31 August marking the change from one year to the next. Ablation poles have been maintained in this area in all years, and measured ablation rates have been used to extrapolate ablation from varying measurements to a fixed date. In addition, net balances for specific poles in this area have been assessed in a stratigraphic system.

Balance for the accumulation area of Sherman Glacier has been most difficult to determine. Because of excessive snowfall a fixed-date system could not be used in this area, and a stratigraphic system has been utilized. In most years this has been only an estimated stratigraphic system. Measurements of net snow accumulation above a previous summer surface have been made by probe and pit studies during most
field seasons. Measurements have been corrected to an estimated balance year by extrapolation of rates of ice loss. Extrapolation was made to a balance year, rather than to a fixed date, because it required extrapolation of data at only one end of the balance period, and because in one year (1967-1968) balance in the accumulation area was assessed entirely by pit studies in which enclosing summer surfaces were observed.

The problem of inferring balance patterns in later years from the pattern of balance established in 1967 may perhaps be considered to introduce a fifth system for the inaccessible tributaries.

In some few years ice-loss from the ablation area could have been assessed in a stratigraphic system, but this was not possible for all years. Therefore, through necessity, annual balance in the ablation area was maintained in a fixed-date system, while net balance of the accumulation area was kept stratigraphic (Table 9).

Because the position of the equilibrium line fluctuated widely from year to year, balance of some areas has been assessed by different methods at different times. For this reason, and for more compelling reasons presented later, detailed yearly balance-altitude curves have not been prepared.

3.1.2 Total annual balance for Sherman Glacier

Total annual balance for Sherman Glacier (Table 9) has been determined by planimetric measurement of contoured net-balance maps (for
<table>
<thead>
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</tr>
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<tbody>
<tr>
<td></td>
<td>In accumulation area</td>
</tr>
<tr>
<td></td>
<td>T&lt;sub&gt;g&lt;/sub&gt;†</td>
</tr>
<tr>
<td>Average</td>
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</tr>
<tr>
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</tr>
<tr>
<td>1962-1963&lt;sup&gt;*&lt;/sup&gt;</td>
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</tr>
<tr>
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<tr>
<td>1965-1966</td>
<td>50</td>
</tr>
<tr>
<td>1966-1967</td>
<td>24</td>
</tr>
<tr>
<td>1967-1968</td>
<td>45</td>
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<td>46</td>
</tr>
<tr>
<td>Total since 1963</td>
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</tr>
</tbody>
</table>

* Balance based solely on position of firn line at end of measurement year

† T<sub>g</sub> = 10<sup>12</sup> gm

‡ Based on an area of 54.1 km<sup>2</sup>

Estimates of standard error in these estimates could not take into account inadequacies in method and have not been made.

Table 9 . Mass balance of Sherman Glacier, 1962 to 1971
Fig. 36. Map of mass balance for the 1966-1967 measurement year on Sherman Glacier (from Marangunić, 1972, Fig. 32).
discussion of method, see Østrem and Stanley, 1966, p.18). The art of
contouring areas of widely scattered data is best termed subjective,
especially since it necessarily requires a preconception of variation
between points. For Sherman Glacier, the contouring of large areas
that lack data requires even greater artistic skill. An annual-balance
map for the measurement year ending 31 August 1967 (Fig. 36 from
Marangunic, 1968, p.110) was utilized to define the pattern of balance
in other years, supplemented by only a few additional data (detailed
measurements for one tributary, Andres Glacier, and isolated observa-
tions from crevasses in several others), and a degree of intuition.

Position of the firn line on Sherman Glacier at the end of a
measurement year apparently bears a consistent relationship to balance
measurements. This relationship permitted estimates of balance to be
deduced from firn line positions, recorded on aerial photographs taken
around the end of the measurement year, for years in which no balance
measurements were made. Such estimates are possible because balance
variations often result largely through uniform increases or decreases
in balance at all elevations, without change in gradient of balance
(see Fig. 37 and 57). Similar techniques for balance estimation
have been used by Meier and Post (1962, p. 70).

3.1.3 Balance gradients

One technique that has been utilized to portray and determine
balance information is the integration of balance-altitude and area-
altitude curves (Ahlmann and Thorarinsson, 1938; Meier and Tangborn,
Fig. 37. Net balance versus elevation for the ablation area above the avalanche-debris cover for six balance years.
1965; and Orheim, 1972). This technique has not proved useful on Sherman Glacier because a single balance-altitude relationship does not exist for this glacier. Instead, different balance-altitude curves can be prepared for different areas of the glacier. These different curves result from differing climatic and heat-balance regimes, and to combine them in a single curve is to mask information and to introduce a source of significant error because only a small part of the glacier system has been sampled.

Meier (1961, p. B-207) considers the activity index (the gradient of the balance-altitude curve at the equilibrium line) a useful glaciological parameter that characterizes a glacier. Orheim (1972, p. 50) suggests that a mean activity index (the gradient of a minimum-variance line fitted to data in the vicinity of the equilibrium line) is a more usable parameter. Activity indices that correspond closely to both definitions have been determined for Andres Glacier and for the central trunk of Sherman Glacier. The activity index for Sherman Glacier (determined along the center-line) is 21.3 kg m\(^{-3}\) a\(^{-1}\); that of a typical maritime glacier; but that of Andres Glacier (6.7 kg m\(^{-3}\) a\(^{-1}\)) is only one third of this.

The line fitted to the data in Fig. 39 passes through \(-11\) m net balance at 105 m, the elevation of the glacier terminus, indicating that ablation in the glacier terminal region has not been significantly altered by the presence of the avalanche debris.

But net balance on Sherman Glacier varies systematically with
Fig. 38. Balance gradient on Andres Glacier (6.7 kg m$^{-3}$ a$^{-1}$).
Fig. 39. Variation in net balance with elevation along the center line of Sherman Glacier (gradient is 21.3 kg m\(^{-3}\) a\(^{-1}\)).
parameters other than altitude. Across a transverse profile at about 450 m elevation, a net-balance gradient of about 60 kg m$^{-3}$a$^{-1}$ is present over a distance of 2.5 km (Fig. 40). A satisfactory explanation for this gradient has not been formulated. It appears to vary in time as a function of amount of net balance (see data for 1967, 1968, and 1970 in Fig. 40) and hence varies largely as a function of winter balance (see p.228). Steepest gradient occurs in a year of smallest winter balance, and lowest gradient occurs in a year of greatest winter balance. Such a relationship does not follow from a simple explanation of areal variation in radiation balance, as might be expected in a high-latitude east-west trending valley.

A possible explanation of this balance gradient, and its variation from year to year, might be an unequal distribution of wind velocity at the glacier surface, with highest mean velocity on the northern side of the glacier. Marangunic (1968, p. 81) showed that much ice is ablated by turbulent heat transfer and this is a function of wind speed. In addition, wind also affects snow-drifting and settling, and hence distribution of snow accumulation. Higher wind speeds might be expected on the northern side of the glacier because mean wind-direction is slightly oblique to the orientation of the valley of Sherman Glacier, hence the southern side of the glacier may lie in a partial wind shadow, while the northern side is more exposed. This explanation suggests that asymmetry of net balance should also be reflected in asymmetry of ablation rate. This asymmetry is recorded in measurements of summer-ablation rate (Fig. 41). However, the gradient of summer ablation rate
Fig. 40. Net balance at markers along a transverse profile across Sherman Glacier at about 430 m elevation.
Fig. 41. Daily melting rates along a transverse profile across Sherman Glacier at about 430 m elevation in August 1969.
\( \frac{d\hat{s}}{dx} \), integrated over a full summer, appears to account for the total observed gradient of net ablation \( (a_n) \), without recourse to a gradient in winter precipitation \( (c_w) \), which should also occur.

3.1.4 **Effect of uplift on mass balance of Sherman Glacier**

Sherman Glacier lies close to the region of maximum uplift in the 1964 Alaska Earthquake, and was raised about 2 m (Plafker, 1965, Fig. 2). Two mass-balance gradients have been measured on the glacier: one of 6.7 kg m\(^{-3}\) a\(^{-1}\) on Andres Glacier; and another of 21.3 kg m\(^{-3}\) a\(^{-1}\) on the trunk portion of Sherman Glacier. These values may not be simple functions of elevation alone, but may also be functions of other factors in valley geometry, but, in any case, the true variation in balance with elevation may lie between these values. Other values ranging from 6 kg m\(^{-3}\) a\(^{-1}\) to 13 kg m\(^{-3}\) a\(^{-1}\) that should be independent of valley geometry can be estimated from a summer balance-summer temperature curve (Fig. 54, Section 3.2.13).

Over a total glacier area of 54.1 km\(^2\), the change in balance for a 2 m rise is 0.725 to 2.305 Tg over the possible range of balance gradient. Sherman Glacier is about 1.8 km wide at its terminus, and net mass balance at the region of eventual mass gain is about 6.5 Mg m\(^{-2}\) a\(^{-1}\). Thus, if the glacier was previously in equilibrium, it would have to gain additional surface area of 1.115 x 10\(^5\) m\(^2\) to 3.546 x 10\(^5\) m\(^2\) by increasing its length by 62 to 197 m to remain in equilibrium. Post (1967, p. D-34) computed that a glacier subjected to such a small balance change would ultimately thicken by about 1 m. This, of course,
would add a further 0.362 to 1.152 Tg and increase the expected change in length to between 93 and 296 m.

A mere 2 m rise in elevation of the land can have a significant effect on glaciers in a region of uplift.

3.2 Estimation of balance from climatological records

The relationship between glacier mass balance and local climate is complex and not easily evaluated, but, if great accuracy is not essential, a number of simplifications can be made, and a simple relationship established that utilizes only major climatic parameters that most directly effect mass changes. Balance estimates obtained through this relationship may not be as accurate as direct measurements, but if it can be adequately established, and a long enough climatological record is available, balance trends can probably be reliably assessed. Assessment of balance trends is the object of this exercise.

3.2.1 Factors effecting mass changes

Three climatic parameters most directly effect mass changes on glaciers: 1) winter snowfall; 2) duration of summer melt period; and 3) summer temperature. The latter two parameters are often combined as a single parameter, degree-days, and the first is often ignored. principally because of lack of data in the record (see, for example, Orheim, 1972a, p. 52, 55, and 56; and Liestøl, 1967, Table III). If climatic parameters are not determined year-round at the glacier, they must be estimated indirectly through climatological records from remote
stations; this requires relating climate of the remote stations to local glacier climate.

3.2.2 Previous estimates of balance from climatological records

A number of attempts have been made to use climatological records for determining glacier mass balance over longer intervals than that of measured balance. Liestøl (1967) developed a 150-year record for Storbreen utilizing only summer temperature (1 May to 31 October) and winter precipitation (1 October to 31 May) recorded at Bergen. Khodakov (in Troitsky and others, 1966) developed a similar length record for IGAN utilizing summer temperature (June to August) and winter temperature (October to May) recorded at B. Khadat (listed as Syktyvkar in Grosval’d and Kotlyakov, 1969). Chizhov (in Chizhov and others, 1968) utilized the same parameters to estimate mass balance of Novaya Zemlya Ice Sheet for about 60 years. Several of these records are discussed later (Section 3.5.1), following derivation of a mass balance record for Sherman Glacier.

3.2.3 Climatological records available for Sherman Glacier

Few climatic parameters have been determined at Sherman Glacier, and these only for the summer field season, but data are collected year-round at Cordova Airport, 13 km distant and well exposed to regional climate with few local topographic influences. Although monthly summaries of daily weather are available for this station - Cordova F.A.A. (N.O.A.A. and U.S. Weather Bureau, 1942 to present), the most readily accessible and usable information are monthly mean temperature
and monthly total precipitation. Winter snowfall at Cordova is recorded, but is not a useful parameter in this study since: 1) it is recorded as depth of snowfall, regardless of density; and 2) winter rain is a frequent occurrence at this low-elevation maritime site while the near absence of icy layers in winter snow attests to the non-significance of winter rain on Sherman Glacier. My balance assessments have utilized only monthly temperature and precipitation data.

3.2.4 Duration of winter

Estimation of winter snowfall from records of monthly total precipitation requires an estimate of duration of winter, the period when precipitation falls dominantly as snow, and hence, the period between balance minimum and balance maximum. Duration of winter at Sherman Glacier was arbitrarily designated as the interval when the curve of monthly mean temperature at Cordova Airport is below an equivalent freezing point temperature of the local cloud base at the glacier, assuming an adiabatic lapse rate of $-0.0075 \text{ deg m}^{-1}$ (Section 3.2.5) (see, for example, Fig. 42). To speed computation, only the months of April, May, and June, and September, October, and November were plotted for any year, and estimates of duration of winter were made from smooth curves fitted to these data. No significant long-term trend of change in length of winter at Sherman Glacier is apparent (Fig. 43).

3.2.5 Adiabatic lapse rate at Sherman Glacier

In most field seasons, air temperature on Sherman Glacier was
Fig. 42. Mean monthly temperatures for Cordova F.A.A. showing how the duration of winter was estimated for Sherman Glacier, assuming a local cloud base of 600 m and an adiabatic lapse rate of -0.0075 deg m⁻¹. (Data from U. S. Weather Bureau)
Fig. 43. Variation of estimated duration of winter at Sherman Glacier.
measured within the atmospheric inversion layer over the ice surface and cannot be used in estimating a lapse rate. In 1967, however, Marangunić maintained a meteorological station at about 550 m elevation on the south wall of the valley of Sherman Glacier (Marangunić, 1972, p. 35), well above any inversion layer. Mean temperature at this site during August 1967 was 9.2°C (op. cit., p. 38) while at Cordova F.A.A., at about 15 m elevation, mean temperature for that month was 13.2°C (N.O.A.A., 1967). Thus, the apparent adiabatic lapse rate was -0.0075 deg m⁻¹. This is essentially the same as the expected lapse rate for the average relative humidity of this region (90 percent for August, U.S. Weather Bureau, 1963). At this humidity, and a mean temperature of 13.2°C at 15 m elevation, the saturation elevation would be reached at about 177 m elevation. Thus, the dry lapse rate of -0.0099 deg m⁻¹ is applicable for the first 162 m, while the saturated lapse rate of -0.0066 deg m⁻¹ is applicable for the remaining 373 m, for an effective lapse rate of -0.0076 deg m⁻¹. The value of -0.0075 deg m⁻¹ has been used as the best estimate for this region. Although it probably is inappropriate to use this same value in winter, no attempt was made to calculate a winter value.

3.2.6 Winter precipitation at Cordova F.A.A.

Winter precipitation at Cordova F.A.A. is defined as precipitation that falls during winter at Sherman Glacier as defined by a curve of monthly temperatures. This assessment of winter and winter precipitation is based on a model of continuous uniform precipitation in any month and of smoothly sinusoidal temperature change, month by month.
through a year. The reliability of the winter precipitation estimates can not be assessed, however, they lead to balance estimates that are consistent with observations of balance at Sherman Glacier.

3.2.7 Winter precipitation at Sherman Glacier

Total winter precipitation has never directly been determined for any year on Sherman Glacier. Marangunić made several attempts at measurement, but was unsuccessful in maintaining a totalizing precipitation gauge. The precipitation is quite large and highly variable from year to year and place to place, and in all probability it often far exceeds the capacity of most gauges.

Winter precipitation at Sherman Glacier can only be evaluated from its relationship to winter precipitation recorded at Cordova Airport, but no record of winter snowfall exists for Sherman Glacier. This can be overcome by assuming that winter precipitation follows the same relationships as does summer rainfall. This is an assumption that can not directly be tested with available data. Several factors suggest that it might not be valid: 1) snowfall is difficult to measure because of drifted snow so that there may be a systematic variation in reliability of summer and winter precipitation records for Cordova Airport; 2) distribution of monthly totals of precipitation for Cordova Airport is non-uniform through the year, suggesting seasonal changes in precipitation regime; and 3) summer precipitation may not be reliably determined for Sherman Glacier where rainfall almost invariably is accompanied by strong to gale force wind and rain gauges were not
shielded. In addition, there is only a small amount of data on summer rainfall for Sherman Glacier and a distinct summer relationship can be only poorly defined.

3.2.8 Summer rainfall at Cordova F.A.A. and at Sherman Glacier

Daily weather observations are taken at 12:00 hr (Alaska Central Time) at Cordova F.A.A. This is an inconvenient observational time on Sherman Glacier and was not duplicated. Thus daily records of summer rainfall are not directly comparable, instead, data for Cordova F.A.A. and for Sherman Glacier were grouped into distinct storms and compared as total rainfall from each storm (storm is merely a convenient term for a period of rain). As a consequence of this grouping, systematic change in rainfall distribution during any one storm can not be determined; such change is not apparent in daily data, where hour of record is a dominant control.

Thirty-nine distinct storms can be recognized in six field-seasons of record (no rain fell during the six days of work in 1968); these storms make up two thirds of total field time, although the proportion varied greatly from year to year. Data for 37 storms has been utilized (Fig. 44): one storm in 1971 had an anomalously high ratio of rainfall at Sherman Glacier to that at Cordova F.A.A. of 24.5 and was deleted from further analysis because of its influence on most statistics; and for one day during a storm in 1966, Marangunić's notes record "rainfall not taken" - a gale blew continuously that day while 112 mm of rain fell at Cordova Airport, the highest daily total during any field season.
Fig. 44. Correlation between rainfall at Sherman Glacier and at Cordova F.A.A. for individual rainstorms, by year.
In general rain occurs simultaneously at Cordova F.A.A. and at Sherman Glacier, so that any line fitted to the data of Figure should pass through the zero intercept. A frequency histogram of distribution of ratios of rainfall at Sherman Glacier to that at Cordova Airport (Fig. 45) shows that the distribution is both non-uniform and non-gaussian (it is not log-normal either), making standard statistical analysis meaningless. An unweighted mean ratio of 2.32 is indicated (standard deviation of the mean is 0.31, indicating that, for a normal distribution, the mean ratio is distinctly different from 1). A mean ratio weighted in accordance to amount of rain has a value of 1.88, which is not significantly different from the unweighted mean.

Although the modal summer storm in the Cordova region brings near uniform rainfall to the two localities (Fig. 45), most of the summer rainfall on Sherman Glacier comes in the less frequent storms that show a pronounced orographic effect. The net result of all summer storms is a rainfall at Sherman Glacier that is about twice that at Cordova Airport. The data are insufficient to better define the distribution and to more precisely determine the relationship. If this same relationship holds for winter precipitation, snowfall on Sherman Glacier might be twice the winter precipitation at Cordova Airport.

Grouping of storm data by field season and observer (Table 10) shows some systematic trend toward higher ratio with time, but no clear change with observer. Weather records for Cordova Airport show a suggestion of change in rainfall regime in this period but it is not well defined.
Fig. 45. Frequency histogram of summer-rainfall-ratios for storms (ratio of rainfall at Sherman Glacier to that at Cordova F.A.A.).
<table>
<thead>
<tr>
<th>Summer</th>
<th>Frequency within .25 above the stated ratio of 0.50 0.75 1.25 1.75 2.25 2.75 3.25 3.75 4.25 4.75 5.25 5.75 6.25 6.75 7.25 7.75</th>
</tr>
</thead>
<tbody>
<tr>
<td>1965</td>
<td>2  2  -  -  -  1  -  1  -  -  -  -  -  -  -  -  -</td>
</tr>
<tr>
<td>1966</td>
<td>1  1  3  -  -  -  -  -  -  -  -  -  -  -  -  -  -  -</td>
</tr>
<tr>
<td>1967</td>
<td>-  2  2  1  2  -  -  -  -  -  -  -  -  -  -  -  -</td>
</tr>
<tr>
<td>1969</td>
<td>-  4  1  1  -  -  -  -  -  -  -  -  -  1  -  -  -  -</td>
</tr>
<tr>
<td>1970</td>
<td>-  -  2  -  1  1  -  -  1  -  -  1  -  -  -  -</td>
</tr>
<tr>
<td>1971</td>
<td>-  2  -  1  -  -  -  -  -  1  -  1  -  -  -  1</td>
</tr>
<tr>
<td>Total</td>
<td>3  11  6  5  2  2  1  1  0  2  1  1  1  0  0  1</td>
</tr>
</tbody>
</table>

Table 10. Distribution of storms by ratio of rain at Sherman Glacier to that at Cordova Airport.
3.2.9 Winter snowfall at Sherman Glacier

Total precipitation for winter at Cordova Airport is quite variable from year to year (Figs. 46 and 47); a probability diagram (Fig. 47) suggests that it can be expected to vary some 54 percent about a mean of 1.107 m over a 100 year period. Over the last decade there has been a trend toward increasing winter precipitation (Fig. 46). The previously derived factor of 1.88 for summer rainfall was used to estimate snowfall at the base camp on Sherman Glacier over the period of record for Cordova Airport (this estimate is also shown in Fig. 46).

Reliability of these estimates can not be directly assessed, although earlier analysis of summer rainfall suggests that they could be quite unreliable. Reliability can, however, be subjectively assessed through the consistency of derived balance assessments. Reliability might also be assessed through the distribution of winter precipitation totals for the 30 years of record (Fig. 47). Steps in the distribution can only result from sparcity of data. That the data cluster in well-defined linear segments is an indication that they are at least internally consistent, and closely follow a simple normal distribution.

3.2.10 Accumulation and winter balance

Orheim (1972, p. 3) believes that precipitation data involving snow are incomplete and unreliable and can not be used to determine accumulation because: 1) he found a lack of correlation between measured monthly precipitation at two stations 6 km apart on Deception Island, South Shetland Islands, and thought it due to imprecise measure-
Fig. 46. Winter precipitation at Cordova F.A.A. and at 430 m on Sherman Glacier (smoothed by a 5-year moving binomial function).
Fig. 47. Probability diagram for winter precipitation at Cordova F.A.A.
ments and local topography; and 2) Swithinbank (1957, p. 68) found at Maudheim on the Antarctic ice sheet and adjacent ice shelf that an unshielded precipitation gauge provides no direct information about either accumulation or atmospheric precipitation.

Although he could possibly have justified ignoring precipitation data for Deception Island on other grounds, Orheim fails to prove his point since: 1) Deception suffers a polar maritime climate that differs in multitudinous ways from the polar continental climate experienced by Maudheim; and 2) a lack of correlation between precipitation at two localities does not imply a lack of correlation between precipitation and accumulation at either locality. In addition, precipitation is a statistical phenomenon, occurring as discrete units scattered in time and space. Small samples of a statistically uniform population can appear non-uniform; perhaps monthly data represent a too-small sample. The danger of rejecting data simply on the basis of a lack of correlation is well illustrated by data from Cordova F.A.A. and Sitka Magnetic, two meteorological stations 450 km apart in coastal Alaska (Section 3.2.18 and Figs. 59, 60, and 61). Data from these stations do not correlate from year to year, but do correlate when the data is averaged over five years.

Swithinbank was careful to restrict his conclusion of the unreliability of precipitation gauges to Antarctica, where amount of drifted snow is large in proportion to atmospheric precipitation. He did find gauges quite accurate during calm-weather snowfall.
In maritime climates drifting of snow might be expected to be less significant for a number of reasons: 1) snowfall is large; 2) snow is drifted dominantly during snowfall; and 3) warm snow is cohesive and warm snow surfaces generally soft and these factors inhibit drifting once snow has settled. Rapid metamorphism, and input of solar radiation between snowfalls, restricts massive snowdrifting to times of snowfall. A relationship might also be expected between amount of drifting snow and snowfall in such circumstances, and although the amount of snow collected by a gauge may not equal actual snowfall, on average it may be related to it.

Swithinbank (1957, p. 73) concluded that accumulation is an immediate consequence of atmospheric precipitation and that, in large uniform areas of small slope, drift of snow appears to become a constant, so that total accumulation becomes a measure of true precipitation. There is no reason to suppose that the reverse is invalid and that precipitation would not become a measure of accumulation. The problem lies in measuring precipitation.

3.2.11 Accumulation and winter balance at Sherman Glacier

Winter weather at Cordova Airport probably reduces the problem of drifting snow. Precipitation is large, and temperature of falling snow is high. A significant amount of winter precipitation at Cordova Airport also falls as rain and wet snow. Under these circumstances, it might be expected that winter precipitation recorded at Cordova Airport bears some relationship to accumulation on Sherman Glacier.
Winter ablation at Sherman Glacier is likely to be small because surface melt refreezes in cold snow, and evaporative loss is small at low temperatures, especially when high humidity prevails. Thus, a direct relationship might be expected between accumulation and balance. Hence, it should be possible to assess winter balance at Sherman Glacier through winter precipitation at Cordova Airport.

To determine a relationship between precipitation measured at Cordova Airport and winter balance at Sherman Glacier an iterative procedure is necessary because only one data pair is available for initial comparison of the two parameters. This data pair is available through the accident of the rock avalanche that, on 27 March 1964, covered and preserved a layer of winter snow beneath avalanche snow and rock debris. A total of 1.056 Mg m\(^{-2}\) of winter precipitation had fallen to that date at Cordova Airport, while 1.76 Mg m\(^{-2}\) of snow (2 m depth, of density 880 kg m\(^{-3}\) in 1967) was preserved beneath the avalanche cover on Sherman Glacier near the center line of flow at the upper edge of the avalanche at about 425 m elevation (Fig. 25).

No relationship can be adequately determined from one data pair, but some estimate can be obtained if assumptions are made. One hypothesis can be put forward that, when no snow has fallen at Cordova Airport, winter balance at Sherman is zero. This hypothesis (hypothesis 1 on Fig. 48) implies that any difference between estimated snowfall at Sherman Glacier and winter balance is directly proportional to the balance. In such a case the difference must be caused by processes involving snowfall and must be attributed to losses through drifted
Fig. 48. Two hypotheses for estimating winter balance at Sherman Glacier from winter precipitation data for Cordova F.A.A.
snow. Another hypothesis (hypothesis 2, Fig. 48) assumes that the 0.183 Mg m\(^{-2}\) difference between estimated snowfall for the period October 1963 through March 1964 and the measured balance for that interval represents a uniform mass loss that occurs continuously through the winter and amounts to some 0.267 Mg m\(^{-2}\) over a whole winter. Of course, the entire difference and more might be fully accounted for in the uncertainty of estimating snowfall at Sherman Glacier.

The range of precipitation values for 30 winters at Cordova Airport is between 0.6 and 1.7 m. Over this range the difference between the two hypotheses is 'small'. Alternative hypotheses could be suggested, such as: 1) a hypothesis lying somewhere between the previous two; or 2) a hypothesis that mass loss is inversely proportional to snowfall. Neither of these can be significantly different from the other two considering the range of precipitation and the uncertainties involved.

If higher snowfall is due to more frequent snow storms, mass loss might decrease with increasing snowfall; if more intense storms are the cause, mass loss might be constant or it might increase with increased drifting of snow. In all cases, over the small range of precipitation values for winter at Cordova the difference between any of these hypotheses is 'small'.

A more significant hypothesis relating winter ablation to winter temperature is discussed and used later (Section 3.2.13).
3.2.12 Summer temperature

In a simplified climatic model of uniform sinusoidal change in temperature with time, duration of summer and of winter are closely related to one another, and mean summer temperature is related to duration of summer and hence to the number of degree-days and to amount of melt or summer balance. For use in this somewhat oversimplified climatic model I have defined summer temperature as the mean of monthly mean temperatures for June, July, August, and September (without weighting for number of days in each month).

In 30 years of record at Cordova Airport, summer temperature follows mean annual temperature only in periodicity but not in long term trend; while summers have warmed, winters have become even colder, for a net decline in annual temperature (Fig. 49).

The greater amplitude of winter changes is a function of inherent capability for greater temperature extremes in arctic winters than in arctic summers, and the poor method of obtaining daily means (one half the sum of daily temperature extremes).

This climatic trend toward warmer summers and cooler winters could be caused by a decrease in size or a southward shift of the "Aleutian low" that dominates weather in this region. Such changes would increase the influence of high-pressure systems that draw air from the Alaskan interior. Climatological data for Sitka, southern Alaska, which has the longest temperature and precipitation record in Alaska, are consistent with these hypotheses (Fig. 61a), but here the decrease in size or
Fig. 49. Summer, winter, and balance year temperatures for Cordova F.A.A., 1943 to 1972 (smoothed by a 5-year moving binomial function).
southward shift has brought pronounced summer cooling because there is no equivalent of the warm Alaskan interior from which to draw heated air. A temperature-correlated trend of winter ablation discussed later (Section 3.2.13), that may be related to cloudiness, is also consistent with this interpretation of differential interaction of air masses between interior Alaska and the Gulf of Alaska.

3.2.13 Net balance at 430 m from climatological records

Net balance at a point \( b_n \) is related to winter balance \( b_w \) and summer balance \( b_s \) by the relation

\[
b_n = b_w + b_s
\]

and hence for a point in the ablation zone

\[
b_s = b_n + (-b_w)
\]

where \( b_n \) is loss of glacier ice and \(-b_w\) is loss of winter snow. For a given summer temperature \( T_s \), snow and ice might have different ablation rates (\( k_s \) and \( k_i \) respectively), related by the expression

\[
k_s = A k_i
\]

where \( A \) is a constant. If no snow falls in summer

\[
b_n = k_i T_s t_i
\]

where \( t_i \) is duration of melt of glacial ice. This is a simplification justified by two factors: 1) the limited range of summer temperature (Fig. 49) and 2) the small amount of data available (however, it is
later shown that summer melting varies exponentially with temperature. If duration of melt of winter snow is $t_s$ and

$$t_s + t_i = 1 \text{ summer}$$

it follows that

$$b_n = k_i T_s (1 - t_s)$$

and since

$$-b_w = A k_i T_s t_s$$

$$b_n = k_i T_s - (-b_w)/A$$

Net balance ($b_n$) at marker 4 on Sherman Glacier has been measured for six years and can be corrected to a fixed point through an estimate of balance gradient (Fig. 39) and changes in surface elevation (Fig. 79); estimates of $T_s$ and $b_w$ are also available for those years. Thus a solution for $k_i$ and $A$ can be made.

Although a minimum-variance solution is possible, it was found more profitable to solve by simple elimination, because in the solution for $A$ two values were obtained that differed greatly from the other 13 solutions. These deviant values resulted from division of numbers formed by subtraction of nearly equal quantities and have been deleted from computation of a mean $A$. A mean value for $A$ of 0.938 with standard deviation of the mean of 0.069 was obtained. This value of $A$ is not significantly different from 1.0; a gratifying result since it suggests that the heat needed to melt ice is the same as that needed to melt an
equivalent mass of snow. It also suggests that there is no significant albedo difference between ice and snow in this area, although the value of \( A \) of 0.938 is consistent with expected albedo differences. This value of \( A \) is computed on a second approximation to winter balance \( b_w \) involving a temperature-dependent winter ablation \( a_w \) that is discussed later. Winter balance estimates by the two earlier hypotheses also give \( A \) values that statistically do not differ from unity.

If \( A \) is of unit value

\[
 b_s = k_i T_s = b_n - b_w
\]

When the difference between measured net balance, corrected for change in elevation, and winter balance estimated through hypothesis 2 (Section 3.2.11) is plotted against summer temperature for balance years ending 1965 through 1970 (Fig. 50), an ablation rate for ice and snow \( k_i = -0.88 \text{ Mg deg}^{-1}\text{summer}^{-1} \) is obtained that yields estimated net balance \( b_n \) values that differ consistently from observed net balance. When the deviations are added to the estimated winter ablation \( a_w \) and plotted against estimated accumulation (Fig. 51) no relationship is apparent, yet relationships between snowfall and winter ablation were assumed in hypotheses 1 and 2 (Section 3.2.11).

When the same deviations plus estimated winter ablation were instead plotted against winter temperature, a linear relationship became apparent. Rather than use this relationship, however, it is more useful to adjust the gradient of this relationship to provide minimum variance
Fig. 50. Variation of estimated summer balance at 430 m with variation in summer temperature at Cordova F.A.A., assuming a constant winter balance correction of 0.267 m.
Fig. 51. Correlation of estimated winter ablation with winter accumulation (constant winter ablation was assumed in Fig. 50).
in the summer balance - summer temperature relationship (Fig. 53).

The relationship

\[ a_w = 0.125 \ T_w - 0.033 \]

(Fig. 52) gave the smallest apparent variance from a linear relationship between summer balance and summer temperature (note, however, that the relationship then becomes nonlinear, Fig. 53.)

The apparent absurdity of decreasing winter ablation with increasing winter temperatures is perhaps readily resolved. Winter temperatures given are those for Cordova Airport at 15 m elevation. At 430 m and higher, winter temperatures are far below freezing so mean winter temperature cannot be a direct cause of ice loss. Air moving off the Gulf of Alaska is relatively warm and moist: if cyclonic activity is prevalent, winter temperature and cloudiness will be greater than normal, air will be moist, and direct influx of solar radiation will be lower than normal, hence sublimation of ice will also be less than average and could even contribute to accumulation as hoar. If anticyclonic activity is prevalent, cold air from the Alaskan interior will flow over the Chugach Mountains. This air is warmed adiabatically and although it is still very cold it is quite dry; cloudiness will be below normal and direct input of solar energy above normal, hence sublimation will be greater than average.

Alternatively, the differing climatic regimes that bring warm or cold winters might bring differing ratios of winter precipitation to Cordova Airport and Sherman Glacier, or the deviations might be related
Fig. 52. Winter balance correction (plotted as winter ablation) as a linear function of winter temperature at Cordova F.A.A. The open circle is the correction for the winter of 1963-64. The gradient is chosen to provide minimum variance in summer balance estimates (Fig. 53).
to errors in estimating winter precipitation and duration of winter.

The first hypothesis gives a subjectively more satisfying explanation of the high correlation between "winter ablation" and winter temperature than do the latter two, but there is no objective means of discriminating between them.

If summer balance is now estimated through the relationship

\[ b_s = b_n - c_w - a_w \]

where \( c_w \) is winter accumulation, and plotted against summer temperature, a nonlinear relationship is obtained (Fig. 53) from which summer balance can be estimated for years for which summer temperature is available. A curve was fitted to the data rather than a simple straight line because several factors that contribute to ice melt, such as radiation, vary as higher powers of temperature than 1. The accuracy of data may not warrant such a curve. Only 4 of 30 years were warm enough for their estimated balances to be greatly increased by the difference between the curve and a straight line.

The nonlinear relationship between summer balance and mean summer temperature (Fig. 53) very closely approximates an exponential relationship. To examine this relationship, the natural logarithm of the "observed" summer balance (corrected measured net balance minus calculated winter balance) was plotted against the reciprocal of the estimated mean summer temperature at 430 m on Sherman Glacier (calculated as mean summer temperature at Cordova F.A.A., minus 3°C
Fig. 53. Variation in estimated summer balance at 430 m on Sherman Glacier with summer temperature at Cordova F.A.A. assuming a variable winter balance correction determined from winter temperature (Fig. 52).
from an adiabatic lapse rate of 0.0075 deg m\(^{-1}\) expressed in degrees Kelvin. The linear relationship (Fig. 54) is as good a fit to the data as is the curve of Figure 53 (see Table 11). In this relationship, the number of degrees of freedom (3) is better defined than the curve of Figure 53. Two degrees of freedom are used to define the straight line of Figure 54, while another is used to define the gradient of the winter ablation line. On the other hand, in the curve of Figure 53, probably at least three degrees of freedom are used to define the curve, and this leaves only two degrees of freedom to determine the standard deviation.

The relationship between summer balance (\(b_s\)) and mean summer temperature at 430 m on Sherman Glacier (\(T^0\)) from Figure 54 is

\[
b_s = -1.041 \times 10^{21} \exp (-1.093 \times 10^5/[RT])
\]

where \(R\) is the gas constant (8.308 J deg\(^{-1}\) mole\(^{-1}\)).

The numerator in the exponent is analogous to an activation energy (\(q\)), the energy needed to melt one mole of ice

\[
q = 1.093 \times 10^5 \text{ J mole}^{-1}
\]

but the heat used to melt ice at 0\(^\circ\)C (\(q_m\)) is only 6.006 \times 10^3 J mole\(^{-1}\). In addition, ice at 273.15\(^\circ\)K radiates heat (\(q_r\)) according to the Stefan-Boltzmann law

\[
\frac{dq}{dt} r = E \sigma T^4
\]

where \(\sigma\) is Stefan's constant (5.667 \times 10^{-8} \text{ J m}^{-2} \text{ deg}^{-4} \text{ sec}^{-1}), \text{ and } E \text{ is}
Fig. 54. Relationship between the natural logarithm of observed summer balance and the reciprocal of mean summer temperature at 430 m on Sherman Glacier. Mean summer temperature at Sherman Glacier was calculated from mean summer temperature at Cordova F.A.A. minus 2.99°C from an adiabatic lapse rate of -0.0072 deg m⁻¹.
<table>
<thead>
<tr>
<th>Balance year</th>
<th>$T_s^{1}$</th>
<th>$b_n^{2}$</th>
<th>$b_w^{3}$</th>
<th>$b_s^{4}$</th>
<th>$\ln(b_s)$</th>
<th>Residual to $b_s^{5}$</th>
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</thead>
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<td>9.38</td>
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<td>2.072</td>
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<td>1.3159</td>
<td>-0.0001</td>
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<td>1.037</td>
<td>-4.695</td>
<td>1.5465</td>
<td>0.0076</td>
</tr>
<tr>
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</tr>
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<td>2.766</td>
<td>-5.309</td>
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<td>-0.0649</td>
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<tr>
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<td>-1.420</td>
<td>3.186</td>
<td>-4.606</td>
<td>1.5274</td>
<td>0.0041</td>
</tr>
</tbody>
</table>

Number of degrees of freedom = 3

Standard deviation = $\pm 0.054 \text{ Mg m}^{-2} \text{ a}^{-1}$

Probable error = $\pm 0.088 \text{ Mg m}^{-2} \text{ a}^{-1}$

Probable error from fig. 53 = $\pm 0.090 \text{ Mg m}^{-2} \text{ a}^{-1}$

1 Summer temperature
2 Net balance (measured)
3 Winter balance (calculated)
4 "Observed" summer balance $(b_n - b_w)$
5 Residuals between estimated summer balance (from fig. 54) and observed summer balance (above)

Table 11. Analysis of residuals between estimated and observed summer balance for the period 1965 to 1970.
the emissivity of ice (0.986 for ice and snow). The time required to
melt one mole of ice (18 g) from a square meter of ice surface where
mean summer balance is \(-4.8167\ \text{Mg m}^{-2}\) for a summer of approximately
120 days is 38.8 seconds, thus

\[
q_r = 1.207 \times 10^4 \text{ J mole}^{-1}
\]

and the sum of radiation and melting is \(1.807 \times 10^4 \text{ J mole}^{-1}\), which is
only 16.5 percent of the estimated activation energy. This ratio of
energies is probably a measure of the efficiency of the heat transfer
process from the air to the ice under average summer conditions. Thus,
heat transfer is about 16.5 percent efficient, although the melting
itself has an efficiency of only 5.5 percent. This "efficiency" of heat
transfer includes such factors as the averaged albedo of the ice and
snow over a summer, and the rate of heat flow through the inversion
layer over the ice.

The coefficient of the balance relationship (\(-1.041 \times 10^{21}
\text{ Mg m}^{-2}\text{a}^{-1}\)) is a melting-rate factor that combines the effects of many
variables such as duration of summer, the heat-transfer rates of
different mechanisms of heat flow (convection, conduction, condensation,
and radiation) and the relative proportions of these mechanisms. In the
time frame of the observations in this study, the melting-rate factor
apparently has been remarkably constant so that at least the various
effects have varied so as to keep the rate constant, or they themselves
have not varied significantly.

If summer is assumed to be about 120 days long on Sherman Glacier,
the daily melting rate can be predicted from measured air temperature and the above relationship between temperature and summer balance.

On 7 August, 1967 temperature, about 2 m above the ice surface at 485 m elevation, averaged about 6.4°C so that about 31.6 kg m\(^{-2}\) of ice should have ablated that day, but ablation of 53.5 kg m\(^{-2}\) was recorded at a nearby ablatograph (data collected by Č. Marangunić). This temperature, however, was recorded from within the inversion layer over the glacier surface. Above the inversion layer at a meteorological station on a hillside at about 500 m elevation, a mean temperature of 8.1°C was recorded and for an adiabatic lapse rate of -0.0075 deg m\(^{-1}\) an air temperature of 8.3°C would be expected at the elevation of the glacier surface (note that the temperature used in establishing the balance relationship did not include the temperature inversion); hence, a melting rate of 52 kg m\(^{-2}\) day\(^{-1}\) is the better estimate of the ablation, and this is in excellent agreement with the measured value (53.5 kg m\(^{-2}\)).

August 7 was a day of heavy rain and strong wind, with no sunshine. The agreement between measured and predicted ablation is also good on days of sunshine (estimated 18 kg m\(^{-2}\) for 2°C and measured at 12 and 17 kg m\(^{-2}\) on two days of that average temperature). However, on a day of cloud cover, but no rain, when mean temperature was 4°C, the estimated ablation was more than twice the measured ablation (37 versus 14 kg m\(^{-2}\)). Because the relationship is established for averaged summer conditions it is surprising that it is applicable to any individual days.

Relationships have now been determined that permit net balance at a point \(b_n\) to be estimated from the three "measured" climatological
parameters: winter precipitation at Cordova Airport ($P_w$); winter temperature ($T_w$); and summer temperature ($T_s$)

\[ b_n = 1.88 P_w + 0.125 T_w - 0.033 + b_s \]

where

\[ b_s = -1.041 \times 10^{21} \exp \left( \frac{-1.093 \times 10^5}{R(T_s + 270.16)} \right) \]

Estimates of balance parameters for Sherman Glacier at 430 m for the last 30 years are provided in Table 12 and Figure 55. Analysis of residual differences between estimated and corrected observed net balances (Table 11) suggests that actual net balance lies within 89 kg m$^{-2}$a$^{-1}$ of the estimated net balance for all years (at the 90 percent confidence level). This implies a precision of estimation of net balance that is within ±3 percent, but observed net balance is not known to that precision in some years, and winter balance is not an accurately estimated parameter either (see also Section 3.2.11).

The short period for which balance calibration is available coincides with an interval of extremes in climatic parameters; maximum and minimum values of most observed parameters over the 30 years of record occurred between 1965 and 1971 - a fortunate coincidence that enables balance to be estimated for other years by interpolation rather than by extrapolation. Interpolation is the more reliable means of estimating values, but unusual climatic conditions must have prevailed to produce all extremes and the period may not be entirely suited to the purpose of calibration.
<table>
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Table 12. Meteorological parameters and estimates of the components of the mass balance of Sherman Glacier at 430 m elevation for the period 1943 to 1972 (in Mg m$^{-2}$ a$^{-1}$ unless indicated).
Fig. 55. Net balance at a point on Sherman Glacier: lower curve ( ) $b_n$ at 430 m elevation, and upper curve ( ) $b_n$ at 630 m elevation. Smoothed with a 5 year moving binomial function.
A smoothed balance curve (Fig. 55) for the 30 years of record shows that trends toward warmer summers and colder winters (Fig. 49) have more than offset a trend toward increased winter precipitation (Fig. 46); net balance has become increasingly negative with time over these 30 years at 430 m elevation on Sherman Glacier by about 0.5 Mg m$^{-2}$a$^{-1}$ - a 20 percent change.

The distribution of net balances departs very little from a true gaussian distribution (Fig. 56). Net balance can be expected to vary some 75 percent about a mean of -2.785 Mg m$^{-2}$a$^{-1}$ over a 100 year period. Winter precipitation has a probable variation of some 54 percent over the same interval and this could account for a 40 percent variation in net balance. Because winter precipitation and temperature, and summer temperature vary in phase with time (compare Figs. 46 and 49), variation in winter precipitation, and hence in winter balance, is the most significant factor in variation of net balance with time at 430 m, but it is not greatly so (40 percent vs. 35 percent).

The conclusion that variations in winter weather are more important to net balance than are summer variations on Sherman Glacier is quite the opposite of the conclusions of Orheim for balance variations at Deception Island (Orheim, 1972, p. 55 and 57) where climate is also very strongly maritime. But Orheim was only able to show that net balance was correlated to summer temperature, not that it was the principal cause of net balance variation. Visual comparisons of curves of summer and winter temperatures and winter precipitation for Sherman Glacier show that significant correlations exist between these parameters and
Fig. 56. Probability distribution of net balance $b_n$ at 430 m for 34 years of estimation ($b_n$ is mean, $\sigma$ is standard deviation).
hence a correlation should be expected between summer degree-days and net balance. This demonstrates only a common cause, not cause and effect. However, this does not negate the validity of estimating summer degree-days from net-balance measurements when a correlation can be established; this was Orheim's objective. If summer degree-days can be correlated with winter weather at Déception Island his climatic reconstruction would be more valuable, otherwise it probably ignores half of the possible cause of variation.

3.2.14 Balance at other elevations from climatological records

Balance at 430 m elevation on Sherman Glacier was calculated from the relationship

\[ b_n = R P_w + a_w + b_s \]

where \( R \) was an orographic effect equal to 1.88 at 430 m. If winter ablation \( (a_w) \) is assumed to be controlled largely by cloudiness and humidity, and is not principally an artifact of method of computation, it cannot be strongly influenced by change in elevation and can be considered constant for all areas of the glacier that experience winters of similar duration to that at 430 m. Inasmuch as winter precipitation at Cordova \( (P_w) \) is partially determined by duration of winter, it is influenced by change in elevation, but not greatly so.

Both \( R \) and \( b_s \), however, are strongly influenced by change in elevation. In calculating net balance at 430 m, summer balance \( (b_s) \) was assumed to be a simple function of summer temperature and the
balance data support this over the small range of available temperature
data (Fig. 54). A summer-balance-elevation function can be established
from a balance-temperature function through an adiabatic lapse rate
(approximately 0.01 deg m\(^{-1}\)). Thus the balance-temperature curve
already developed (Fig. 54) can be used to estimate summer balance for
elevations that do not differ more than a few hundred meters from 430 m.

Variations in the orographic factor are less readily determined.
Rainfall was measured at 465 m elevation as well as at 425 m in the
summer of 1969 and a possible small orographic effect was detected:
P\(_{465}\) : P\(_{425}\) was 1.08, but for the same summer the orographic factor
between Cordova Airport and 425 m was also negligible. The orographic
factor must be determined another way.

Net balance (b\(_n\)) at 630 m was measured as + 0.17 Mg m\(^{-2}\)a\(^{-1}\) for
the 1965-66 balance year, and estimated as - 0.70 Mg m\(^{-2}\)a\(^{-1}\) for 1966-67.
Since

\[
R = \frac{b_n - a_w - b_s}{P_w}
\]

R may be determined from the measured and estimated parameters. R was
found to be 3.8 (precipitation at 630 m is about twice that at 430 m).
Balances at 630 m (Table 13, and Fig. 55), were estimated through the
relationship

\[
b_n = 3.8 P_w + a_w + b_s
\]
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<th>Winter pptn. at 630 m</th>
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<th>Winter balance</th>
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<td>1969</td>
<td>0.934</td>
<td>3.549</td>
<td>-0.543</td>
<td>-4.123</td>
<td>-3.580</td>
<td>-0.574</td>
</tr>
<tr>
<td>1970</td>
<td>1.864</td>
<td>7.083</td>
<td>0.050</td>
<td>-3.270</td>
<td>-3.320</td>
<td>3.813</td>
</tr>
<tr>
<td>1971</td>
<td>1.488</td>
<td>5.654</td>
<td>-0.433</td>
<td>-4.233</td>
<td>-3.800</td>
<td>1.421</td>
</tr>
<tr>
<td>1972</td>
<td>1.371</td>
<td>5.210</td>
<td>-0.758</td>
<td>-4.058</td>
<td>-3.300</td>
<td>1.152</td>
</tr>
</tbody>
</table>

Mean 1.318 5.010 -0.326 -3.642 -3.317 1.380

Table 13. Estimates of the components of mass balance at 630 m elevation on Sherman Glacier for the period 1943 to 1972 (in Mg m⁻² a⁻¹).
The balance estimates predict a 1969-70 winter balance at 630 m of about 7 Mg m\(^{-2}\) a\(^{-1}\) and a net balance for the year of 3.8 Mg m\(^{-2}\) a\(^{-1}\). By the end of July, when about 3/5 of the summer ablation should have occurred, a balance of 4.54 Mg m\(^{-2}\) a\(^{-1}\) water equivalent of 1969-70 snow should remain: this would have a depth of 7.6 m at density 0.6 Mg m\(^{-3}\). At the end of July 1970 8.5 m of snow of density 0.6 Mg m\(^{-3}\) was measured at this elevation.

3.2.15 Estimated-net-balance gradients

Net-balance gradients computed from estimated net balances at the two elevations (430 and 630 m) are plotted for all balance years in Figure 57. The simple hypothesis, that most change in balance from year to year is accomplished solely by a uniform shift of the balance-altitude curve, without change in gradient, is shown to be a generally quite reliable first approximation. Variation in proportion of winter snowfall to summer balance is the cause of departures from this hypothesis.

Over the last 15 years there has been a pronounced trend toward a steeper balance-gradient; net balance at 630 m has changed more than net balance at 430 m (compare curves in Fig. 55). This is principally a reflection of the increasing contribution of precipitation in balance at increasing elevations.

3.2.16 Net balance of Sherman Glacier from climatological records

Correlation of measured net balances at a point at 430 m with areal net balances for the whole of Sherman Glacier should enable estimates
Fig. 57. Net-balance gradients between 430 and 630 meters on Sherman Glacier from 1943 to 1972.
of areal net balance to be made for the entire length of climatological record, but no relationship is apparent in a plot of the two parameters (Fig. 58). Because areal net balance of the glacier is an integration of net balances at all points on the glacier and since net balances at different points can be related through balance-altitude curves, a correlation should exist between actual net balance at 430 m and actual areal net balance for the glacier. Net balance at 430 m is obtained largely by direct measurement, with only small corrections for pole shift: calculated areal net balances for Sherman Glacier appear to be greatly in error.

More than one pair of summer temperature and winter precipitation values can produce the same areal net balance \( B_n \), just as more than one pair can produce the same point net balance \( b_n \). Thus, expected correlation between \( b_n \) and \( B_n \) cannot be perfect, nevertheless, given the limited range of the two parameters \( T_s \) and \( P_w \) and their own cross-correlation, some significant correlation should be apparent.

Areal net balance for 1965-66, 1966-67, and 1967-68, are probably determined to within \( \pm 20 \) teragrams. The 1968-69 balance year was probably two months longer than the measurement year and some 20 Tg of that balance should be transferred to 1969-70. The 1964-65 measurement year was assigned the same areal balance above the equilibrium line as was the 1965-66 year (Marangunić, 1968, p.111) but climatic evidence suggests that this was not so, because the earlier winter was snowier and warmer and the earlier summer colder. The net balance above the equilibrium line may have been 27 Tg more than indicated. Readjust-
Fig. 58. Correlation of areal net balance of Sherman Glacier with measured net mass balance at 430 m (arrows signify suggested corrections to balances).
ment of all data to a balance year for all parts of the glacier would move all points on Figure 58 toward an improved correlation.

If the line on Figure 58 represents the correlation between net balance at 430 m and areal net balance for Sherman Glacier, the glacier would be in mass balance with a net balance of about -3 m at 430 m elevation. Since 1964 net balance at this elevation has been about -3 m water equivalent of ice, and the data suggest that Sherman Glacier is approximately in balance. This, however, is inconsistent with measured thinning of the glacier (Fig. 79) that results from continued mass loss from a continued negative net mass balance.

3.2.17 Effect of debris cover upon mass balance of Sherman Glacier

Insulation by avalanche debris has reduced ablation in the region of avalanche cover by 80 percent (Section 2.15). Without the debris cover, areal net balance would be some 35 Tg more negative and the glacier would not be in balance.

3.2.18 A long-term net balance record for Sherman Glacier

Although climatic parameters for Cordova F.A.A. and for Sitka Magnetic (450 km southeast of Cordova) correlate only poorly, or not at all from year to year over the last 30 years (Figs. 59 and 60), the cyclic trends of climate at the two localities have been in phase and of similar amplitude (Figs. 61a and b). The general trends of climate have not been entirely dissimilar either. This long-term correlation of climate is expected because weather at the two localities is influenced
Fig. 59. Correlation of summer temperature at Cordova F.A.A. with that at Sitka Magnetic, for the period 1943 to 1972 (from N.O.A.A. 1970; N.O.A.A., 1957 to present; and U.S. Weather Bureau, 1963).
Fig. 60. Correlation of winter precipitation at Cordova F.A.A. with that at Sitka Magnetic for the period 1943 to 1972 (from N.O.A.A., 1970; N.O.A.A., 1957 to present; and U.S. Weather Bureau, 1963).
Fig. 61. Meteorological parameters for Sitka Magnetic with data for Cordova F.A.A. (since 1942) for comparison: smoothed summer temperature and smoothed winter precipitation (smoothed by a 5-year moving binomial function) (from Clayton, 1927, 1934 1947; and U.S. Weather Bureau).
dominant by the same weather system, the "Aleutian low". Thus, a hundred-year-long smoothed balance record for Sherman Glacier can be calculated from the record of weather at Sitka despite the apparent lack of correlation from year to year.

Several alternative methods can be used to estimate the balance record: two have been used to show that different approaches yield similar records. In one method, the data for summer temperature and winter precipitation for Sitka Magnetic were converted to equivalent temperature and precipitation values at Cordova from relationships estimated from Figures 59 and 60 (assumed to be \( T_C = T_S - 1 \), and \( P_C = P_S - 0.3 \) for this purpose, despite the apparent lack of correlation). These values were then used to estimate net balance from the sum of summer balance and winter snowfall at 430 m on Sherman Glacier estimated by the relationships previously established (Fig. 53 and Section 3.2.13). A winter ablation correction was not made. The calculated net balances were then smoothed by a 5-year equal weighted moving average. The resulting balances were consistently about 0.75 Mg m\(^{-2}\)a\(^{-1}\) more positive than the previous estimates from Cordova data, and a - 0.75 Mg m\(^{-2}\)a\(^{-1}\) correction was added to obtain net balance at 430 m on Sherman Glacier (Fig. 62). This correction partially compensates for the variable winter ablation correction used in the earlier estimate (Section 3.2.13), but its magnitude is about three times the average ablation correction for the last 30 years. The correction may, in part, arise from the choice of unit gradients in the precipitation and temperature relationships above. These gradients were guessed in the absence of good
Fig. 62. Net balance at 430 m on Sherman Glacier, estimated from climatological data for Sitka Magnetic. Dashed line is the net balance at 430 m needed for an equilibrium areal net balance of Sherman Glacier. Line connecting dots is net balance at 430 m on Sherman Glacier estimated from climatological data for Cordova F.A.A. Both estimates have been smoothed with a 5-year equal weighted moving average.
correlations.

The other approach taken to estimate net balance makes use of the information that summer balance and winter precipitation contribute about equally to variation in net balance (Section 3.2.13). The ordinate scales of the smoothed curves of summer temperature and winter precipitation for Sitka (Figs. 61a and b) were normalized to give the curves equal graphical variance, and then summer temperature was subtracted from winter precipitation. The resultant curve (Fig. 63) thus expresses the smoothed variation in net balance that can be expected for the known variations in summer temperature and winter precipitation. The ordinate scale of this curve can be calibrated by comparison with the curve of estimated net balance at 430 m on Sherman Glacier, smoothed over the same time interval (5 years) (Fig. 55). With suitable adjustment of scale, this 100-year-long record of balance is applicable to each glacier that meets the following conditions: 1) its weather is influenced dominantly by the Aleutian low; and 2) winter precipitation and summer balance contribute about equally to its balance variations.

3.2.19 Trends in net balance at Sherman Glacier

Balance curves calculated through the two methods (Fig. 62) show a long interval of low balance in the first three decades of this century. A rising trend in balance since about 1925 apparently culminated in the 1940's. This culmination is more apparent in the balance curve derived from the Cordova data (Fig. 55, and also plotted on Fig. 62), but it is also present in the other two curves. Since the 1940's, net balance has
shown a slight drop.

The balance curves suggest that net balance at 430 m on Sherman Glacier has not been more than about $-2.5 \text{ Mg m}^{-2}\text{a}^{-1}$ at any time in the past 100 years (when averaged over 5 years). Without the debris cover, Sherman Glacier would be approximately in balance with a net balance at 430 m of about $-2 \text{ Mg m}^{-2}\text{a}^{-1}$, so that Sherman Glacier has apparently been thinning by at least half a meter a year for the last century.

The relationship between net balance at 430 m on Sherman Glacier and areal net balance for the entire glacier (Fig. 58) was established for a glacier that does not differ significantly in size and shape from that of Sherman Glacier in 1968 (that is, the glacier did not change much between 1965 and 1971). But, over periods longer than the 7 years of study, changes in size and surface elevation of the glacier are significant.

At the culmination of a glacier advance about 1910 (Section 3.8) the trunk portion of Sherman Glacier was about 80 m thicker than in 1968 as indicated by stranded lateral moraines. If the balance gradient of $21.3 \text{ kg m}^{-3}\text{a}^{-1}$ established earlier (Fig. 39) has not changed significantly with time, and this is supported by the estimated balance gradients over the last 30 years (Fig. 57), the net balance of the glacier would have been 1700 $\text{ kg m}^{-2}\text{a}^{-1}$ more positive at any one locality in 1910 and the net balance for the entire glacier upglacier from the present terminus would have been about 34 Tg more positive.

This 34 Tg correction almost compensates for the -35 Tg correction
Fig. 63. A century long record of net balance variation at a point on glaciers near Cordova and Sitka, Alaska. Net balance scale must be calibrated for a particular glacier by correlation with measured values.
that must be applied because of the avalanche. Thus, at about the turn of the century, Sherman Glacier would have been in equilibrium with a balance at 430 m of about \(-3 \text{ Mg m}^{-2} \text{a}^{-1}\) (from Fig. 58). The smoothed balance curves show that this equilibrium balance was exceeded during the mid-1880's, so that the areal net balance of Sherman Glacier was positive at that time. Estimated net balance at 430 m between 1882 and 1887 exceeded \(-3 \text{ Mg m}^{-2} \text{a}^{-1}\) in all but one year, 1886 (Table 14).

The twelve-year gap in the climatological records for Sitka Magnetic between 1887 and 1899 probably includes a period of negative balance for Sherman Glacier because the mid-1880's glacier-mass-balance maximum that is apparent in many balance estimates (Fig. 65) was followed in the 1890's by a pronounced balance minimum.

Net balance at 430 m next exceeded \(-3 \text{ Mg m}^{-2} \text{a}^{-1}\) in the early 1930's but, by then, the glacier probably had thinned by 40 m. This would lead to a 17 Tg balance correction that would compensate for only half of the 35 Tg avalanche correction. Thus the glacier would have been in equilibrium around 1930 if net balance at 430 m had been about \(-2.6 \text{ Mg m}^{-2} \text{a}^{-1}\) (from Fig. 58). Net balance at that time peaked at only \(-2.7 \text{ Mg m}^{-2} \text{a}^{-1}\) (Fig. 62).

Net balance next peaked around 1947 (Fig. 62) (1945 by the Cordova data, Fig. 55), at about \(-2.3 \text{ Mg m}^{-2} \text{a}^{-1}\). This probably brought Sherman Glacier to near equilibrium at that time.

The dashed curve in Figure 62, showing the 430-m net balance needed for an equilibrium areal net balance of Sherman Glacier, has been
<table>
<thead>
<tr>
<th>Balance year</th>
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<td>-2.44</td>
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<tr>
<td>1885</td>
<td>-2.76</td>
</tr>
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<td>1886</td>
<td>-3.37</td>
</tr>
<tr>
<td>1887</td>
<td>-1.89</td>
</tr>
</tbody>
</table>

Table 14. Net balances at 430 m on Sherman Glacier, estimated from climatological records for Sitka Magnetic, from 1882 to 1887. There is a 5 year gap in the records prior to 1882, and a 12 year gap after 1887. The data are not smoothed.
estimated from rates of surface lowering and the estimated net balance gradient. Rates of surface lowering have been estimated from the calculated rate of retreat of the terminus over the last 60 years, but this is known only in an averaged form (Section 3.3). The rate of surface lowering can be determined, with the precision needed for this study, only if both the areal net balance and the glacier response to this balance are known. Determination of the former is the immediate object of the exercise, and evaluation of the latter is the ultimate aim of this text.

3.3 Measured balance variations around the world

Records of measured glacier mass balance that span intervals approaching 30 years are few in number (Fig. 64) and are confined to the northern hemisphere (specifically Europe). Mass balance of the Grosser Aletschgletscher (Switzerland) has been measured since 1940 (Kasser, 1970) and of Storgläciären (N. Sweden) since 1945 (Schytt, 1959; and 1963 to present). Other quite long records are available for Storbreen (Norway) (Liestøl, 1967; 1968 to present), for Glacier de Sarennes (Switzerland) (Kasser, 1967), and for the Hintereisfjerner (Austria) (Hoinkes, 1970). Few long records are available for glaciers in the Americas; mass balance of South Cascade Glacier has been measured since 1952 (Meier and Tangborn, 1965; World Data Center, personal communication) and that of Blue Glacier since 1955 (La Chapelle, 1965; World Data Center, personal communication).
430 meters on Sherman Glacier
- 630 meters on Sherman Glacier
- South Cascade Glacier (Meier and Tangborn, 1965, World Data Center)
- Blue Glacier (La Chapelle, 1965, World Data Center)
- Storbreen (Liestøl, 1967-present)
- Hintereisferner (Hoinkes, 1970)
- Storglaciären (Schytt, 1959-present)
- Aletschgletscher (Kasser, 1970)
- Fissure on Deception Island (revised from Orheim, 1972a)
- Glacier de Sarenes (Kasser, 1967)

Fig. 64. Measured balance variations around the world (smoothed by a 5-year moving binomial function).
3.4 Comparison of balance records

Some differences are apparent among the curves of mass balance (Fig. 64), but there are also notable similarities, especially among the North American records. The curve of smoothed measured mass balance for South Cascade Glacier is identical in form to that of smoothed estimated balance at 630 m on Sherman Glacier, while that for Blue Glacier is more similar to those at 430 m.

To some degree the trends, but more certainly the periodicity of balance changes at all glaciers are quite similar to one-another. There is little reason to suppose that the estimated mass-balance records for Sherman Glacier are not realistic estimates. While the variations from record to record are not always perfectly in phase, there is no suggestion of any form of antiphase relationship over the span of these records.

A mass-balance record derived from snow, ice, and firn stratigraphy exposed in a volcanic fissure on Deception Island (Orheim, 1972b) is included in Figure 64 to demonstrate that these generalities about northern-hemisphere glacier mass-balance trends apparently are global in applicability over this span of time. This stratigraphic record probably lacks the absolute accuracy of time scale that is inherent in historical records because its time scale is an interpretation of a stratigraphy that may not have recorded some years if they were years of negative balance. This record is my reinterpretation of the stratigraphy presented by Orheim (1972a and 1972b). Orheim reached the
opposite conclusion as to global variation in balance from his interpretation of this stratigraphy. My interpretation, however, faithfully adheres to the criteria set forth by him for recognition of annual layers in his stratigraphy and Orheim's interpretation apparently does not. The record of stratigraphy itself appears to be reliable in that a record of balance can be inferred from it that compares favorably with measured records from elsewhere on this planet (Fig. 65) and this appears to be the best criterion for assessing reliability.

3.5 Long estimated-balance records from around the world

Although there are no continuous measured mass-balance records that span more than the last 30 years, there are some few estimated records that span considerable intervals. A net-balance record based on run-off and precipitation data for the Grosser Aletschgletscher (Kasser and Muller, 1960; Kasser, 1970) spans 70 years (Fig. 65). Liestøl (1967) developed a mass-balance record for Storbreen (Fig. 65) utilizing correlation of measured mass balance and meteorological records from two stations; Luster, some 37 km from Storbreen (1900 to 1963), and Bergen, some 200 km distant (1816 to 1963). Khodakov (in Troitskiy and others, 1966; also reported in Grosval'd and Kotlyakov, 1969) developed a balance record for IGAN (Fig. 65) and Obrucheva Glaciers utilizing correlation of measured mass balances and meteorological records from B. Khadat (given as Syktyvkar in Grosval'd and Kotlyakov, 1969) that were initiated in 1818.

Liestøl compiled his record using summer temperature and winter
Fig. 65. Long estimated-balance records from around the world (10-year running means plotted at approximately equal variance, see original reference for balance scale): Storbreen, Norway (Liestøl, 1967, 1968-present); IGAN, Polar Urals (Troitskiy and others, 1966); Deception Island, South Atlantic (my interpretation from Orheim, 1972a); Aletschgletscher, Switzerland (Kasser and Müller, 1960; Kasser, 1970). Also plotted are 10-year running means of summer temperature in central England plotted upside down (from Manley, 1970) and the unsmoothed sunspot Wolf number curve (Athay and Warwick, 1961; and Lincoln, 1960-present). Every second sunspot cycle has been plotted as negative to show the "22-year" periodicity of polarity.
precipitation while Khodakov used winter temperature as a substitute for precipitation after establishing the existence of a correlation between winter balance and winter temperature.

3.5.1 Quality of the long estimated records

In assessing the quality of the Storbreen and IGAN records, Orheim (1972a, p. 70-71) suggested that the Storbreen record is much more likely to accurately reflect past mass-balance variations than is the IGAN record. Only one of his five reasons for this suggestion is valid: the Storbreen record is in good agreement with known volume changes and variations of the terminus of Storbreen since 1900 whereas no such tests can be made of the IGAN record. There is, however, sufficient agreement between the two balance records that the IGAN record can be tested against the changes at Storbreen. Khodakov omitted four years from the IGAN record because, in a span of four years, he was unable to compile a summer or winter temperature in one year and was able to compile only one or other of them in each of the other three. Orheim (1972a, p. 71) noted that Grosval'd and Kotlyakov (1969) "misleadingly represent" the completeness of the IGAN record by suggesting that fewer years are missing, but their figure 2 clearly shows a dashed line over the 14 years of record of 10-year running means where four years of balance are missing and the words "No data available" clearly interrupts this dashed line over the portion of the curve where there is no data. Orheim misrepresents the IGAN data by omitting this interruption from his own presentation of the record. In his criticism of the Russian work Orheim notes that: 1) the Storbreen record is based on a continuous
series of observations since 1816 (Orheim, 1972a, p. 71); and 2) the
Luster meteorological data is used to estimate the Storbreen record
since 1900, but Liestøl (1967) states in his Table III, p. 28, and on
his figure 27 that no observations of precipitation at Bergen
Observatory were recorded until 1861 and he used the 1861 to 1962 mean
value for all winters prior to 1861. There is a fifty-year gap in half
of the variation in the Storbreen record (balances for this period,
1816 to 1861, have been omitted from my Figure 65). Orheim (1972a,
p. 72) noted this shortcoming of the Storbreen record but apparently
considered that it was less significant than the four year gap in the
IGAN record. Liestøl (1967, p. 41) also stated that precipitation
records for Bergen were used through the Storbreen record - from 1816 to
1949, and it is not clear if his Table III contains any data from Luster.
The implication of Liestøl's sentence "The correlation (of temperature
and ablation) between this last station (Bergen) and Storbreen is not
so good as for Luster, but, nevertheless, it gives a very good basis
for calculations of ablation on Storbreen" is that his Table III
contains no Luster data at all and the estimated balance record is based
solely on Bergen records and their correlation with balance at Storbreen.
However, his figures on the preceding page use the name Luster in
connection with this estimated balance.

Orheim (1972a, p. 71) also suggests that the longer series of
comparisons of mass balance and meteorology for Storbreen (16 years)
than for IGAN (11 years) makes the Storbreen relationship "sounder and
more significant". Significance can only be determined through a
correlation coefficient that is directly proportional to the sum of the products of the deviations of the parameters for each year from their mean values and inversely proportional to the number of observations; "sounder" has no apparent meaning in this context. Liestøl and Khodakov did not determine correlation coefficients, but both present plots of data pairs (Liestøl, 1967, Figs. 25 and 26; Troitskiy and others, 1966, Fig. 97). Liestøl's diagrams show acceptable correlations, but they are far from perfect. Khodakov's diagram shows less scatter, but his plot contains only six data pairs, not eleven as indicated by Grosval'd and Kotlyakov (1969) (perhaps later balance measurements have supported the established relationship).

Braithwaite (1972, p. 156) and Orheim (1972a, p. 71) criticize the theoretical value of the correlation of winter balance with winter temperature, but Khodakov (Troitskiy and others, 1966, Figs. 97 and 98) shows a strong empirical correlation between snow accumulation and winter temperature for 15 data pairs from two stations. The implication of Khodakov's correlation is that variations in balance and temperature apparently stem from a common cause: no cause-and-effect was ever implied.

Braithwaite (1972, p. 156) inferred that the IGAN record was an extrapolation, but balance estimates were made largely by interpolation on their figure 97 (Troitskiy and others, 1966). Khodakov had to extrapolate his established relationship for only about 20 percent of his data; the same is true for Liestøl's work.
Because the Storbreen record fails to include compensation for variation in winter balance for the period 1816 to 1862, the IGAN record is probably the better record over that time span, but both are in need of substantiation of the relationships upon which they are based. Statistical analysis would help in assessing their reliability.

Inasmuch as net balance is the difference between the mass that falls and the mass that leaves, the Aletschgletscher record is founded on a much more direct theoretical basis than the other two records, but for this record, problems of measurement of run-off and precipitation are critical, because balance is determined from their difference with minor corrections. In all, the errors in measurement of these quantities are probably no worse than the errors often involved in direct assessment of glacier mass balance. The Aletschgletscher record is probably the most reliable record of long-term mass-balance variation of these three records, as Orheim (1972a, p. 71) also concluded.

3.6 Long mass-balance records from stratigraphy

Although many mass-balance determinations are made with the aid of stratigraphic studies of shallow pits, only a few long records have been obtained by measurement of stratigraphic sections in glacier ice or firn.

Apparently only six long stratigraphic-balance records that span more than a century have been published (Fig. 66): all but one of these are from the Antarctic continent - Little America V and Byrd Station (Gow, 1968), Wilkes S2 (Cameron, 1959 and 1964), South Pole Station (Giovinetto, 1960; and Giovinetto and Schwerdtfeger, 1964),
Fig. 66. Long balance records from stratigraphy (10-year running means plotted at approximately equal variance, see original reference for balance scale): Little America and Byrd Station (Gow, 1968); Wilkes Station (Cameron, 1964); Plateau Station (Koerner, 1971); South Pole Station (Giovinetto and Schwerdtfeger, 1964); Deception Island (1 - my interpretation of stratigraphy in Orheim 1972a, 2 - Orheim's interpretation).
and Plateau Station (Koerner, 1971) - and even the other one is from an
island only 100 km north of the Antarctic Peninsula - Deception Island
(Orheim, 1972a and 1972b; this stratigraphy has also been reinterpreted
to accord with the criteria presented by Orheim, 1972b, p. 7). These
are not the only stratigraphic records: Bull (1971, Table 7, p. 382),
Giovinetto (1964, Table 8, p. 146), and Barkov and Petrov (1966,
Table 8, p. 200 ) present listings of some additional shorter records;
and Bader and others (1955) prepared another from a core taken at Site 2
in Greenland, but these are all comparatively short records.

Ice cores spanning very long time intervals have been obtained
from Antarctica, Greenland, and Devon Island, but only disconnected
parts of their records of net balance have yet been analysed and there
is little prospect of continuous records ever being compiled from them
because they would be of limited use and the cores probably do not
contain such a record.

Stratigraphic records in glacier ice are not restricted to the
polar regions as the above records might suggest. In fact, the clearest
annual stratigraphy is not formed in the polar environment where
analysis of firn stratigraphy is described as an art (Gow, 1968, p. 6)
but on so-called temperate glaciers, where large annual increments of
ice are often separated by sharply defined summer surfaces marked by
wind-blown dust and organic matter. But no long stratigraphic record
has been deciphered for a temperate glacier.
3.7 A stratigraphic record from Sherman Glacier

Sherman Glacier is one of these glaciers where a clear annual layering is marked by dark summer horizons (Fig. 67). Relative widths of the dark-light couplets have been measured from enhanced enlargements of this photograph (two examples are shown in Figs. 68 and 69) and corrected for photographic distortion and scale with the aid of an orthophotographic map. However, this record is quite useless at present for a variety of reasons. Measured flow rates and analysis of aerial photographs shows that most of the stratigraphy shown in Figure 70 was deposited in the general vicinity of the '630-m' location for which a 30-year estimated balance record was developed earlier in this chapter. That record suggests that eight of the last thirty years (Table 13) are recorded as hiatuses in the stratigraphic record. Not only would mean net balance estimated from such a stratigraphy be 25 percent too large, but the period and amplitude of balance variations would also be greatly in error. Although some movement markers were placed in this region of Sherman Glacier, there were far too few to adequately define the surface strain-rate field to correct measured outcrop lengths of layers for the areally rapidly varying strain rate. In addition, both ends of the stratigraphy are covered: the lower end is covered by rock avalanche debris (it is recorded on pre-avalanche aerial photographs taken in 1963 and 1950 although the quality of density contrast in these negatives has not been evaluated); but, more importantly, the upper end of the record may be covered by firn and this leaves the chronology free-floating in time. Table 13 shows a net balance for 1964-1965 that
Fig. 67. Stratigraphy of ice and dirt layers on Sherman Glacier.
(Aerial photograph taken 28 August 1967)
Fig. 68. Solarized enlargement of the upper end of the ice and dirt stratigraphy. The black surface is the summer surface of 1964-1965.
Fig. 69. Enlarged negative print of a mid-portion of the ice and dirt stratigraphy.
Fig. 70. Outcrop length of layers on Sherman Glacier (4 year moving average).
was nearly twice as large as balance for the preceding two years. This suggests that 1964-1965 firn masks the firn line of at least the previous year and possibly many more. Finally, some years of very small positive balance may not have been recognized in analysis because of the finite resolution of the photographs and the same is possible for years whose winds brought too little debris to the glacier surface to produce a visible density contrast on the photographic print. It was this latter factor, accentuated by photographic prints that were inadequately processed to reveal the full range of densities of emulsion of the negatives, that led Marangunić (1972, p. 117) to suggest a quadrennial origin for the observed layering.

3.8 Discussion of mass-balance variation at Sherman Glacier

The century-long record of glacier mass balance at Sherman Glacier, developed from the climatic record for Sitka (Fig. 62) suggests that the last significant interval of positive mass balance at Sherman Glacier was probably in the 1880's. Considerable variations of negative balance have occurred in later decades, with periods of near-equilibrium balance centered around 1918, 1933, 1946-1947, and 1954 during times of cool summers and moist winters. The 1946-1947 balance maximum, which has been the largest since the mid-1880's, may even have been slightly positive.

Consideration of kinematic-wave velocity (≈ 250 m a⁻¹, Section 4.4.3) and an average flow path for such a wave (about 5 km at present, and about 6 km at the turn of the century when the glacier was longer)
suggests that the positive balance event should have initiated an advance about 1910 and that there should have been marked decreases in the rate of retreat, or perhaps even brief "still stands", around 1940, 1953, 1967, and 1974.

Tuthill and others (1968, p. 325) estimate that one advance of Sherman Glacier perhaps culminated around 1910. Field and others (undated manuscript) note a quite minor advance about 1930 (more correctly, a period of moraine development that may only have been a still stand). These advances have been dated by growth-ring counts on trees, and estimations of the time taken to establish trees on the moraines; the advances were not observed. Tuthill and others also note that the average rate of recession of the glacier during the 1950's was only 15 m a\(^{-1}\) compared with 20 to 25 m a\(^{-1}\) for the previous half century, and 46 m a\(^{-1}\) between 1959 and 1964. They had estimates of the position of the glacier terminus in 1910, 1950, 1959, and 1964 from which to derive these estimates. Marangunic' (1972, p. 113) noted an up-to-20 m advance of parts of the glacier terminus between 1966 and 1967, but much of it merely remained stationary, and, by 1969, a general slow recession was in progress (Section 4.3.9). The kinematic wave generated by the balance maximum of 1954 should arrive at the terminus in 1974. In 1966 the wave should have been 2 km from the terminus, in the vicinity of marker 28 (Fig. 2). The velocity profile for 1965-1966 (Fig. 81) shows a large velocity jump at that location (8 km along the X axis), but where this jump went in later years is difficult to determine. This history of glacier response is in
remarkable agreement with the long estimated record of mass balance.

Kinematic waves travelling through the trunk portion of Sherman Glacier (see Fig. 81) have a wavelength that suggests their generation by mass balance variation every 3 to 5 years (Section 4.4.3) which is in accord with the unsmoothed balance record estimated from the data from Cordova F.A.A. (Fig. 55).

It therefore appears that the century-long record of balance reliably represents the variations in mass balance at Sherman Glacier that are due to changes in climate, and this permits the observed behavior of the glacier to be checked against the expected behavior in response to the climatic change. From this, the response of the glacier to the rock avalanche can be determined.
CHAPTER IV

RESPONSE OF SHERMAN GLACIER

TO CHANGE IN MASS BALANCE

4.1 Introduction

Study of the response of Sherman Glacier to change in mass balance largely has been confined to observation of changes that were manifest at the surface of the glacier while detailed study was in progress between July 1965 and August 1971. These changes in form and behavior of the surface were monitored by terrestrial triangulation surveys of movement markers, by measurements of strain networks, by aerial photographic interpretation, and by visual observation.

4.2 Behavior prior to the rock avalanche

Virtually nothing is known of the behavior of Sherman Glacier prior to the detailed studies that began after the March 1964 earthquake and rock avalanche. Of the seven observations of the glacier to March 1964 recorded by Tuthill and others (1968, p. 320), two were mere sightings and the other five were aerial photography. No glaciological work was undertaken until July 1965, although subsequently, Tuthill and others (1968, p. 322) determined surface lowering and recession of Sherman Glacier from glacial geology, aerial photography, and the
Cordova C-3 and C-4 sheets of the United States Geological Survey topographic map series.

The known history of Sherman Glacier to 1964 is one of continued rapid net recession. For the first half of this century net recession was around 20 to 25 m each year; this dropped to 15 m each year during the next decade, but rose to 46 m a year for the period 1959 to 1964. Total recession from a moraine estimated to have formed around 1910 has been about 1200 m (Tuthill and others, 1968).

Of particular value to a study of glacier response to the 1964 event is a series of vertical aerial photographs taken 27 August 1963 by Austin S. Post, U. S. Geological Survey. These photographs define a "datum state" of Sherman Glacier against which changes in surface morphology with time since the avalanche have been compared.

4.3 Behavior subsequent to the rock avalanche

4.3.1 Methods of study

a) Triangulation surveys for displacement of the glacier surface from year to year

Positions of movement markers on the glacier were determined by reduction of conventional triangulation surveys, made each year from 1965 to 1971 with a Wild T2 or Kern DKM2 theodolite from survey stations established by Marangunić (1968, p. 8-19; 1972, p. 9-20). Markers were aluminum stakes drilled into the ice each year, or supported by rock cairns on the area of debris cover. Many markers were followed through
the entire six years of study; others, lost through various causes, were replaced in early phases of the study or abandoned in later phases.

Marker coordinates were determined by solving for the point of intersection, in the local grid system, of equations of the horizontal lines of sight between two survey stations and each marker (see Fig. 71). The vertical coordinate was determined from the horizontal distance between one station and the marker and the angle of depression or elevation from station to marker. No more sophisticated analysis was warranted, either by the data, or by the results.

Marangunić (1972, p. 20) gives horizontal errors of ± 0.2 m and a vertical error of ± 1 m for the location of any survey station within the local grid system. Because surveys of any one marker were, in general, repeated from the same pair of stations every year, and changes in position and changes in annual displacement are of such magnitude, these errors are insignificant to the study.

Positions of movement markers at the beginning and end of the study are shown in Figure 72. Error ellipses on these positions are smaller than the plotted points at this scale.

b) Triangulation surveys for displacement from week to week

Positions and velocities of selected markers were also determined by repeated triangulation surveys at weekly intervals during July and August 1969. These surveys were made to high precision in order to determine whether variations in velocity could be related to climatic
\[
x = \frac{a_1 x_1 - y_1 - (a_2 x_2 - y_2)}{a_1 - a_2}
\]

\[
y = a_1 (x - x_1) + y_1
\]

\[
z = z_1 + \left[ (x - x_1)^2 + (y - y_1)^2 \right]^{\frac{1}{2}} \tan \theta
\]

where \( \theta \) is the vertical angle from \( A \) to \( P \).

**Fig. 71.** Method of solution for marker coordinates \( P (x, y, z) \) given measured angles to the marker from two known stations \( A \) and \( B \).
parameters, such as rainfall, more reliably than has been done in the past.

Directions to markers on two selected transverse lines across the glacier were measured with a Kern DKM2 theodolite eight times each over a period of one hour from each of two survey stations. The line of markers from 1 to 7 was observed from stations M14 and B3 (Fig. 72). The line of markers including marker 28 and marker M13 (Fig. 72) was observed from stations B2 and M17.

Compensation for marker motion during the three-hour-long surveys was by linear extrapolation of each of the one-hour trends of angular change to determine directions to markers at a common epoch. Coordinates of markers were obtained by standard coordinate geometry in a local grid system utilizing the line of sight between the two observing stations as a baseline of known length and orientation.

In addition to obtaining marker displacement from week to week, estimates of marker motion during each three-hour survey were made from the estimated marker positions and from rates of angular change by the method outlined below.

In the geometrical nomenclature of Figure 73 the coordinates of marker M are given by

\[ X = C \frac{\sin \beta}{\sin (\beta - \alpha)} \cos \alpha \]
Fig. 73. Simple triangulation model where $\alpha$ and $\beta$ are observed changing directions to a moving marker $M$. 
and

\[ Y = C \frac{\sin \beta}{\sin (\beta - \alpha)} \sin \alpha \]

where

\[ \alpha = f_1(t) \]
\[ \beta = f_2(t) \]

and C is a known distance between the two observing stations. By differentiating \( X \) and \( Y \) with respect to time and substituting for \( X \) and \( Y \)

\[
\frac{dx}{dt} = \frac{x}{\tan \beta} \frac{d\beta}{dt} - y \frac{d\alpha}{dt} - \frac{X}{\tan (\beta - \alpha)} \left( \frac{d\beta}{dt} - \frac{d\alpha}{dt} \right)
\]

\[
\frac{dy}{dt} = \frac{y}{\tan \beta} \frac{d\beta}{dt} - x \frac{d\alpha}{dt} - \frac{X}{\tan (\beta - \alpha)} \left( \frac{d\beta}{dt} - \frac{d\alpha}{dt} \right)
\]

c) Estimation of surface strain at strain nets

Deformation of two strain nets, one of four markers on the debris cover (utilizing M13, 28, 44, and 48, Fig. 2) and another of five markers centered on marker 4, were determined each year from 1965. The net at marker 4 was abandoned beneath snow in 1970, but the other was followed each summer season until August 1971.

At marker 4, all distances within the net were flat-taped with steel tape at 4.5 kg tension; but on the debris cover, only one side of that net was measured by taping, the remainder of its geometry was determined from theodolite surveys at each of the four markers.
Marangunic (1972, p. 106-112) and Bull and Marangunic (1968, p. 314-315) analysed strain at these nets according to the method of Nye (1959), but this method is somewhat unsatisfactory for a study of changes in strain-rate and a more realistic approach was taken for the two nets. Their analyses also contain an error in sign convention and arithmetic errors.

4.3.2 Analysis of surface strain

Analysis of deformation at the strain net on the debris cover proceeded on the premise that seasonal variations in surface strain-rate were immaterial to this study of long-term glacier response. Thus, the variations in length of the measured side and in magnitude of determined angles could be expressed as smooth simple functions of time on a time scale of seven years. Measured values were plotted against time and smooth curves sketched through the data (Figs. 75 and 76). Significant deviations of the data from these curves were disregarded; being probable blunders in measurement.

Figure 74 provides the geometrical nomenclature utilized in this analysis of deformation. Strain-rate at half-yearly intervals was determined directly for each of the two triangles shown, from the curves fitted to the changing geometrical parameters in Figures 75 and 76 through the following method.

Side \( c \) was the measured side and all labelled angles (designated by letters capped by \(^\circ\)) were observed. Lengths of the other sides of the two triangles are given by
Fig. 74. Geometrical nomenclature utilized in analysis of the debris strain net.
Fig. 75. Change in length of the measured side of the strain net on the debris, 1965 to 1971.
Fig. 76. Change in measured angles in the strain net on the debris, 1965 to 1971.
\[ a = c \frac{\sin \hat{A}_1}{\sin \hat{C}_1} \]

\[ d = c \frac{\sin \hat{D}}{\sin \hat{C}_1} \]

\[ b = c \frac{\sin \hat{D} \sin \hat{C}_2}{\sin \hat{C}_1 \sin \hat{B}} \]

\[ e = c \frac{\sin \hat{D} \sin \hat{A}_2}{\sin \hat{C}_1 \sin \hat{B}} \]

Strain-rate (\( \dot{\epsilon}_\theta \)) along line \( c \) is

\[ \dot{\epsilon}_{\theta c} = \frac{1}{c} \frac{dc}{dt} \]

where subscripted \( \theta \) designates direction of the strain. Similarly

\[ \dot{\epsilon}_{\theta a} = \frac{1}{a} \frac{da}{dt} \]

\[ \frac{da}{dt} = \frac{\partial c}{\partial t} \frac{\sin \hat{A}_1}{\sin \hat{C}_1} + \frac{\cos \hat{A}_1}{\sin \hat{C}_1} \frac{\partial \hat{A}_1}{\partial t} - \frac{\sin \hat{A}_1 \cos \hat{C}_1}{\sin^2 \hat{C}_1} \frac{\partial \hat{C}_1}{\partial t} \]

hence

\[ \dot{\epsilon}_{\theta a} = \dot{\epsilon}_{\theta c} + \frac{1}{\tan \hat{A}_1} \frac{d\hat{A}_1}{dt} - \frac{1}{\tan \hat{C}_1} \frac{d\hat{C}_1}{dt} \]
Similarly

\[ \dot{\epsilon}_{\theta_d} = \dot{\epsilon}_{\theta_c} + \frac{1}{\tan D} \frac{d\hat{B}}{dt} - \frac{1}{\tan C_1} \frac{d\hat{C}_1}{dt} \]

\[ \dot{\epsilon}_{\theta_b} = \dot{\epsilon}_{\theta_d} - \frac{1}{\tan B} \frac{d\hat{B}}{dt} + \frac{1}{\tan C_2} \frac{d\hat{C}_2}{dt} \]

and

\[ \dot{\epsilon}_{\theta_e} = \dot{\epsilon}_{\theta_d} - \frac{1}{\tan B} \frac{d\hat{B}}{dt} + \frac{1}{\tan A_2} \frac{d\hat{A}_2}{dt} \]

Principal components of the strain-rate tensor and their directions for each triangle can then be determined by, for example, the Mohr circle method.

In this analysis the direction of side e was utilized as an intermediate axis of reference. Results were referred to a fixed axial system through calculated rotation of line e (Fig. 77) (between markers 44 and 28) obtained through repeated triangulation surveys.

The network of poles about marker 4 was treated similarly, although in this case, with all distances measured, rates of change in lengths of sides of individual triangles were read directly from Figure 88.

4.3.3 Results of yearly triangulation surveys

a) Total displacement of the glacier surface

Horizontal coordinates of movement markers at the beginning and
Fig. 27. Computed rotation of the reference axis of the strain net on the debris.
end of the survey period are presented as vectors of net displacement in Figure 72. For some markers this period is shorter than six years and the vectors are not a representation of relative displacement between markers. In detail, horizontal displacements are not along straight lines as indicated in this figure, but follow curves that may be irregular or smooth, as shown for the particular markers that constitute a strain net on the debris cover (Fig. 78).

The line of markers from 21 to 70 approximate the center flow-line of the glacier. Changes in elevation of this line (Fig. 79) reveal changes in longitudinal profile of the glacier with time. Detail of vertical change at the debris strain net is shown in Figure 80.

No analysis of errors in these marker coordinates has been attempted. Whether markers have followed smooth or irregular paths can not be resolved from the calculated displacements; the capabilities of the theodolites used far exceeded those of the users in an often hazardous and inclement environment, and an error analysis of these non-redundant surveys uses only the instrument’s capabilities. Successive changes in marker position are of significant magnitude and most changes are in the expected direction. The data appear to be free of significant blunders.

b) Annual displacement

By applying small corrections for differing dates of survey from year to year to distances between successive yearly marker positions, annual displacements of each marker (Table 15) were calculated.
Fig. 78. Displacement of the debris strain net, 1965 to 1971, with magnitudes and directions of the principal stresses within each triangle of the net.
Fig. 79. Positions of markers along the center line of Sherman Glacier, 1965 to 1971.
<table>
<thead>
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<th>Marker</th>
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<td>21</td>
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<tr>
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<td>7</td>
<td>14.36</td>
</tr>
<tr>
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<td>-</td>
</tr>
<tr>
<td>31</td>
<td>-</td>
</tr>
<tr>
<td>46</td>
<td>-</td>
</tr>
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<td>49</td>
<td>27.75</td>
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<tr>
<td>18</td>
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<tr>
<td>8</td>
<td>37.18</td>
</tr>
<tr>
<td>74</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 15. Annual displacements of markers (in meters). See Fig. 2 or 72 for locations of markers. Values between years are means for two or more years.
Annual displacements of markers along the center flow-line are shown in Figure 81. Although some variance among these longitudinal surface-velocity profiles may be due to small errors in surveying from year to year, a major part of the variance must represent real flow variation with time.

4.3.4 Results of weekly triangulation surveys

a) Position and flow vectors

Two transverse lines of markers were surveyed for horizontal position at approximately weekly intervals, as weather permitted, during the later part of July and early part of August 1969. One line, immediately upglacier of the debris cover, was observed four times: on 23 July, 31 July, 7 August, and 19 August. Another line, on the debris cover, was observed three times: on 1 August, 9 August, and 19 August. Positions of the markers on the two lines are shown with successive velocity vectors in Figures 82 and 83.

An error analysis through a minimum-variance solution of these positions suggests that all markers are located to less than ±0.005 m in X position and ±0.02 m in Y position, at the 90 percent confidence level. For most markers precision of location is better than ±0.01 m (90 percent level). This is better presented in terms of the diameter of the marker poles themselves: each calculated marker position most probably lies within the actual diameter of the marker pole at each occasion of survey.
Fig. 81. Velocity profiles along the center line of flow of Sherman Glacier, showing possible crest of kinematic wave (numbered 1 to 5).
Fig. 82. Positions and velocity vectors of the upper transverse line of markers, July and August 1969 (vector scale $x 10^{-3} \text{ m day}^{-1}$).
Fig. 83. Positions and velocity vectors of the lower transverse line of markers, July and August 1969 (velocity scale $10^{-3}$ m day$^{-1}$).
This precision of location is better than the coupling between markers and the glacier surface. Markers on the cover are well coupled to the debris - seldom to worse than one or two pole diameters, but markers on the debris-free ice stand in ablated cones of depression that can be as much as 0.15 m in base diameter. A part of this coupling problem was reduced by ensuring that markers were held upright in the center of their ablated holes by crude tripod arrangements of poles before surveying began each week. Whether they remained in this position during each survey is not known; no collapse was ever observed.

This obvious coupling problem was not recognized at the start of surveying the upper line of markers. Its effect is seen in the apparent erratic motion of markers 1, 7, and 72 (Fig. 82) that is entirely due to straightening of poles for the second survey. Displacements of all markers in this line between the first and second surveys have been ignored.

Magnitudes of velocity vectors for the two lines are shown in Figures 84 and 85. The radical change in surface velocity profile at the upper line from its state before 7 August to its state after 7 August can not be explained by lack of coupling.

b) Instantaneous velocities

Instantaneous velocities (Section 4.3.1b) for the lower line are shown in Figure 86. Significant retrograde motion is impossible according to currently known theories of glacier motion. This aspect of the surface motion study must be dismissed as an interesting but
Fig. 84. Magnitudes of short-term velocity vectors for 8 markers in a transverse line upglacier from the debris cover (error bars at 90 percent confidence level).
Fig. 85. Magnitudes of short-term velocity vectors for 7 markers in a transverse line on the debris cover (error bars at 90 percent confidence level).
Fig. 86. "Instantaneous" velocity profile of a transverse line on the debris cover.
fruitless exercise.

One set of observed changes in apparent directions to markers coincided with clearly recorded simultaneous changes in meteorological parameters, such as wind-speed, air temperature, and humidity, and thus apparently was caused by horizontal refraction of ray paths through a varying inversion layer of cooled air flowing over the glacier surface. The sets of continuous-recording meteorological equipment maintained at several stations on the glacier surface, however, were inadequate to assess the meteorological parameters at the precision of time-scale of the observed changes in direction, so that this aspect can not be investigated with available data (the instruments responded too slowly and could not keep synchronous time).

This general technique, for determining instantaneous velocities, would be useful if horizontal refraction could be overcome by making simultaneous observations at two different monochromatic wavelengths. Laser distancers would be particularly useful, because the substitution of measured distances instead of measured directions greatly simplifies the solution.

4.3.5 Results from surface strain-net measurements

a) Upglacier from the debris cover

The symmetrical arrangement of five markers in the strain-net at marker 4, upglacier from the debris cover, was divided into four triangles numbered anticlockwise 1 through 4 (Fig. 87). The strain-
Fig. 87. The strain net at marker 4: position, orientation, and magnitude of principal horizontal strain rates.
rate tensor for each triangle was calculated for each of three points in time: July 1, 1966; January 1, 1968; and July 1, 1969, using lengths and their rate of change with time (Section 4.3.2) read from curves fitted to the measured parameters (Fig. 88). Magnitudes of the principal components of this tensor and their changes with time for each triangle are shown in Figure 89. Figure 87 shows the deformation and approximate displacement of the net, along with magnitudes and directions of the principal horizontal strain-rates at the beginning and end of study.

A small negative rotation of the network evident in Figure 87 is a pseudo-rotation stemming from use of the direction of line 4-S and 4'-S' as a local axis of reference. Absolute rotation of the set of triangles is unknown but small; curvature of a medial moraine that cuts through the net (Fig. 67) suggests that rotation in this area is small and negative; evidence from the measured velocity profile across the glacier at this net is ambiguous.

b) Surface strain-rate on the debris cover

Calculated strain-rates along the sides and one diagonal of the strain net on the debris cover are presented in Figure 90. Those along the other diagonal were not calculated because it was convenient to treat the net as two triangles divided by the one diagonal whose direction was observed. Strain-rates were calculated for half-yearly points in time, from 1965 to 1972, using the method previously outlined (Section 4.3.2).
Fig. 88. Change in length of the measured sides and diagonals of the strain net at marker 4. Lengths in meters.
Fig. 89. Variation in time of the magnitudes and directions of the principal strain rates in the 4 triangles of the strain net at marker 4, 1966-1970.
Fig. 90. Variation in time of the magnitude of the strain-rate along the four sides and one diagonal of the strain net on the debris, 1965 - 1972.
In this analysis, as in the last, true values of individual strain rates are relatively unimportant; it is the trend of change in strain rate that is of importance to this study. An inflexion of the curve of measured length of one side with time (Fig. 75) appears to dominate the shape of the strain rate curves. If this inflexion were unreal, an artifact of errors in measurement, a discussion of the significance of the computed changes would be useless. This inflexion, however, does not appear in computed lengths of some other sides of the net (Fig. 91). Although measurement of the one side was not repeated in all years, in those years in which it was repeated, agreement to within 0.2 m was obtained. The inflexion around 1969 would be insignificant only if the error in taping approached 0.5 m. This is too large even for taping over the rough crevassed debris surface and it must be concluded that the inflexion is significant.

Principal horizontal strain rates and their directions for the two triangles of the strain net are shown in Figure 92. Vertical strain rates calculated through conservation of volume are shown in Figure 93.

Rotational velocity about a vertical axis (Fig. 94) was calculated for each of the two triangles as the mean of the rates of change of directions of the three sides.

4.3.6 State of stress at the two surface strain nets

Stresses for each triangle of the two strain nets (Figs. 95 and 96) were calculated from the estimated principal strain rates through a simple flow law put forward by Nye (1957) that relates effective strain-
Fig. 91. Variation in time of the computed lengths of the sides and diagonals of the strain net on the debris.
Fig. 92. Variation in time of the magnitude and directions of the principal horizontal strain rates in the strain net on the debris, 1965 to 1971.
Fig. 93. Variation in vertical strain-rate (computed as $\dot{\varepsilon}_2 = - (\dot{\varepsilon}_1 + \dot{\varepsilon}_3)$) from 1965 to 1972.

A) for the triangle M13 - 28 - 44
B) for the triangle M13 - 44 - 48
Fig. 94. Rotation of the two triangles of the strain net on the debris, 1965 to 1972.
Fig. 95. Principal horizontal stresses in the strain net at marker 4, 1966 - 1970.
Fig. 96. Principal horizontal stresses in the strain net on the debris, 1965-1972.
rate ($\dot{\varepsilon}$) to effective shear-stress ($\tau$) by the relationship

$$\dot{\varepsilon} = A \tau^n$$

where $A$ and $n$ are constants. This simple law and Glen's (1958) original values of the constants ($A = 0.148 \text{ a}^{-1} \text{bar}^{-4.2}$, and $n = 4.2$) were used because the true values of the stresses within the nets were not essential to this study.

4.3.7 Vertical change at the strain nets

From surface balance at each marker, absolute elevations of markers on the glacier surface and their change with time, and absolute elevation of the glacier bed and its variation in space, vertical motion of a point of ice can be determined with respect to the glacier bed. Data on the first three have been presented earlier, the latter two were calculated by Marangunic (1972, p. 21-34), from a gravity survey made in 1966.

If average vertical strain-rate through a column of ice is known, then rate of change in height of the column, due to internal strain, can be simply calculated as a product of strain-rate and original thickness. Vertical strain-rate averaged through the ice thickness at each strain net is unknown, but, as a first assumption, vertical strain-rate determined at the surface from conservation of volume and horizontal strain-rates can be utilized by assuming constant vertical strain-rate with depth.

Data on absolute change in thickness at each strain net and
expected change in thickness due to internal strain are presented in Tables 16 and 17.

An adequate error analysis of these data can not be undertaken. A large source of error is in determination of change in elevation of the glacier base which has been determined by interpolation between estimates calculated from a surface gravity survey. Computation of ice thickness from gravity data provides only a highly smoothed picture of bottom topography, smoothed over areas that are large in comparison to the square of ice thickness (which is about 250 to 350 m), while the critical gradient required for analysis of ice thickness change is the gradient over one annual displacement (about 50 m) in the direction of displacement.

An error of ±2.5 m vertical change in 55 m of horizontal movement in estimation of gradient of the bed at marker 4 would account for most of the observed extra thickness change at that net, but the difference in computed change between the two halves of the net probably can not be accounted for in this fashion.

The remarkable constancy of difference between the two halves of the strain net (Triangles 1 and 4, and 2 and 3, Table 16) can not be attributed to differences in ice thickness because this would require about a 50 percent contrast in ice depth. A part of the total discrepancy may be due to differences in basal melting or freezing; in this temperate glacier, however, basal melting is very much more probable, and a basal freezing rate of 2 to 3 m each year is highly
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<td>429.6</td>
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<tr>
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<td>Bed elevation $B$ (m)</td>
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<td>$\frac{dB}{dt}$ (m a$^{-1}$)</td>
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<tr>
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<td>Thickness change (m a$^{-1}$)</td>
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**Vertical strain rate**

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<th>3</th>
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</thead>
<tbody>
<tr>
<td>($\times 10^{-3}$ a$^{-1}$)</td>
<td>1.2</td>
<td>1.0</td>
<td>2.7</td>
<td>2.8</td>
</tr>
</tbody>
</table>

**Thickness changes due to strain (m a$^{-1}$)**

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>4</th>
<th>2</th>
<th>3</th>
</tr>
</thead>
<tbody>
<tr>
<td>($\times 10^{-3}$ a$^{-1}$)</td>
<td>0.4</td>
<td>0.3</td>
<td>0.9</td>
<td>0.9</td>
</tr>
</tbody>
</table>

**Thickness change not due to strain (m a$^{-1}$)**

<table>
<thead>
<tr>
<th></th>
<th>1</th>
<th>4</th>
<th>2</th>
<th>3</th>
</tr>
</thead>
<tbody>
<tr>
<td>($\times 10^{-3}$ a$^{-1}$)</td>
<td>2.6</td>
<td>2.6</td>
<td>2.1</td>
<td>2.1</td>
</tr>
</tbody>
</table>

**Difference between two halves**

<table>
<thead>
<tr>
<th></th>
<th>0.5</th>
<th>0.5</th>
<th>0.5</th>
</tr>
</thead>
<tbody>
<tr>
<td>of the net (m a$^{-1}$)</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Treated as bed moving relative to ice.

Table 16. Parameters for change in thickness at marker 4.
<table>
<thead>
<tr>
<th></th>
<th>1966</th>
<th>1969</th>
<th>1971</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Mean surface elevation (m)</strong></td>
<td>354</td>
<td>347</td>
<td>335</td>
</tr>
<tr>
<td>(\frac{dS}{dt}) (m a(^{-1}))</td>
<td>-1.6</td>
<td>-4.2</td>
<td>-4.1</td>
</tr>
<tr>
<td><strong>Bed elevation B (m)</strong></td>
<td>41.9</td>
<td>35.3</td>
<td>28.5</td>
</tr>
<tr>
<td>(\frac{dB}{dt}) (m a(^{-1}))</td>
<td>-2.6</td>
<td>-2.6</td>
<td>-2.6</td>
</tr>
<tr>
<td><strong>Surface balance (b_n) (m a(^{-1}))</strong></td>
<td>-1.5</td>
<td>-1.5</td>
<td>-1.5</td>
</tr>
<tr>
<td><strong>Ice thickness (m)</strong></td>
<td>312</td>
<td>312</td>
<td>306</td>
</tr>
<tr>
<td><strong>Absolute ice change (m a(^{-1}))</strong></td>
<td>-0.1</td>
<td>-2.7</td>
<td>-2.6</td>
</tr>
<tr>
<td><strong>Thickness change (m a(^{-1}))</strong></td>
<td>2.7</td>
<td>-0.1</td>
<td>0.0</td>
</tr>
<tr>
<td><strong>Vertical strain rate</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>((x10^{-3} a^{-1})) Triangle 1</td>
<td>-14.5</td>
<td>-11.5</td>
<td>-0.3</td>
</tr>
<tr>
<td>Triangle 2</td>
<td>-18.5</td>
<td>-20.5</td>
<td>-8.5</td>
</tr>
<tr>
<td><strong>Thickness change due to</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>strain (m a(^{-1})) Triangle 1</strong></td>
<td>-4.5</td>
<td>-3.6</td>
<td>-0.1</td>
</tr>
<tr>
<td>Triangle 2</td>
<td>-5.8</td>
<td>-6.4</td>
<td>-2.6</td>
</tr>
<tr>
<td><strong>Thickness change not due to</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>strain (m a(^{-1})) Triangle 1</strong></td>
<td>7.2</td>
<td>3.5</td>
<td>0.1</td>
</tr>
<tr>
<td>Triangle 2</td>
<td>8.5</td>
<td>6.3</td>
<td>2.6</td>
</tr>
</tbody>
</table>

* Treated as bed moving relative to ice

**Table 17.** Parameters for change in thickness at the debris strain net.
improbable if not impossible.

The simplest explanation of the discrepancy is that the initial assumption of uniform strain-rate through the ice column is invalid and there is horizontal advection of ice at depth that is unaccounted for in surface strain-rates. A depth-averaged vertical strain-rate of about $9 \times 10^{-3} \text{ a}^{-1}$ would account for all the observed thickness change at marker 4; this is from three to ten times calculated vertical strain-rates at the surface. Vertical strain-rate at the surface net itself varies through a factor of nearly three over the four triangles.

At the strain net on the debris, the discrepancy between actual thickness change and estimated change due to internal strain is even more remarkable. A basal freezing rate of $8.5 \text{ m a}^{-1}$, which would explain the discrepancy, is impossible.

One simple explanation of the discrepancy is that basal gradient has been wrongly estimated from gravity data. In Table 18 basal gradient has been estimated from the other parameters of Table 17. While these estimated gradients themselves are quite acceptable as local values, enough data are available to compute a continuous basal gradient function. Such a function, when integrated between observed gravity stations, would probably give ice thicknesses that deviate beyond the limits of precision of thickness estimates from gravity values, although subglacial topography of $10 \text{ m}$ amplitude, which would be consistent with the data of Table 18, could not be detected with observations of gravity. Advection of ice at depth probably must again be invoked. Depth-
<table>
<thead>
<tr>
<th></th>
<th>1966</th>
<th>1969</th>
<th>1971</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Change in surface height</strong></td>
<td>-1.6</td>
<td>-4.2</td>
<td>-4.1</td>
</tr>
<tr>
<td><strong>Surface balance</strong></td>
<td>-1.5</td>
<td>-1.5</td>
<td>-1.5</td>
</tr>
<tr>
<td><strong>Absolute ice change</strong></td>
<td>-0.1</td>
<td>-2.7</td>
<td>-2.6</td>
</tr>
<tr>
<td><strong>Thickness change due to strain</strong></td>
<td>Triangle 1</td>
<td>-4.5</td>
<td>-3.6</td>
</tr>
<tr>
<td></td>
<td>Triangle 2</td>
<td>-5.8</td>
<td>-6.4</td>
</tr>
<tr>
<td><strong>Assumed thickness change not due to strain for both triangles</strong></td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td><strong>Basal gradient (meters per annual displacement)</strong></td>
<td>Triangle 1</td>
<td>4.4</td>
<td>0.9</td>
</tr>
<tr>
<td></td>
<td>Triangle 2</td>
<td>5.8</td>
<td>3.7</td>
</tr>
<tr>
<td><strong>Gradient from gravity survey</strong></td>
<td>-2.6</td>
<td>-2.6</td>
<td>-2.6</td>
</tr>
</tbody>
</table>

Table 18. Estimation of basal gradient beneath the debris strain net (units are m a⁻¹).
averaged vertical strain-rates of $+8.7 \times 10^{-3} \text{a}^{-1}$, $-3.2 \times 10^{-3} \text{a}^{-1}$, and $0.0 \text{a}^{-1}$ for the three points in time would be necessary to account for the observed change.

Large-scale "extrusion flow" of basal ice around irregular bedrock topography may account for necessary strain thickening at depth.

The strain thickening may, alternatively, be a necessary consequence of erosion of $90 \text{kg m}^{-2} \text{a}^{-1}$ of ice over the entire area of the debris cover by subglacial streams (Table 7). This $0.83 \times 10^6 \text{m}^3 \text{a}^{-1}$ of ice is lost from walls of tunnels within the ice, largely from a few, or perhaps one major tunnel at the base of the glacier.

Consider a loss of $0.83 \times 10^6 \text{m}^3 \text{a}^{-1}$ from the wall of a half-cylinder of 15 m radius and 2 km length, at the base of the glacier. This amounts to a loss of $8.75 \text{m a}^{-1}$ from the walls and roof of the tunnel, and is compensated by tunnel closure due to ice flow that causes horizontal compressive strain in the ice above the tunnel. If the $17.5 \text{m a}^{-1}$ horizontal closure is spread over an average width of 300 m through the glacier, then the 300 m thick glacier should thicken by $17.5 \text{m a}^{-1}$ but $8.75 \text{m a}^{-1}$ is lost from the tunnel roof, leaving a net glacier thickening of $8.75 \text{m a}^{-1}$. The observed strain thickening at the debris strain net is clearly of this magnitude. This raises the possibility that maps of surface strain-rate could be used to locate the position of large subglacial streams and their changing position with time.
4.3.8 Other observations

a) Crevasse orientation at the debris strain net

Bull and Marangunic (1968, p. 314) reported "highly anomalous" surface tensile stresses and a large rotation in direction of stress at the strain net at marker M13 on the debris cover, and this was discussed further by Marangunic (1968, p. 124, or 1972, p. 110-111). Their analyses suggested that by 1969, or shortly thereafter, principal tensile stress should be parallel to the axis of the glacier and a significant advance of the glacier might be expected.

The strain net is in an area of longitudinal crevassing so tensile stress should never have been considered anomalous, but significant rotation of the stresses should have been accompanied by change in crevassing direction or change in deformation at existing crevasses. To study this change, orientations of all crevasses around the circumference of the net at marker M13 were recorded in 1969 (Fig. 97). It became apparent during this measurement that no large change in stress orientation could have occurred and this has been confirmed in subsequent strain analysis. The larger mode of crevasse orientation (Fig. 97) is approximately at right-angles to the principal tensile stress as expected, but more significantly, the departure from a right-angle relationship can be accounted for by the known small rotation of the direction of principal tensile stress and by expected rotation of the crevasses themselves, as determined from measured rotation of the strain net (these rotations have been added to the rose diagram of
Fig. 97. Rose diagram of crevasse orientations around the strain net on the debris cover, in 1969. Direction of rotation of crevasses and directions of principal tensile stress also shown.
Fig. 97).

A very much smaller, secondary mode of crevasse orientation (Fig. 97) parallel to the principal tensile stress was made up entirely of small fresh cracks in the debris cover. These were attributed to the change in stress direction described by Bull and Marangunic' (1968, p. 314) until it became apparent that the magnitude of the cracks was inadequate for the known tensile strain-rate. Subsequent rigorous strain analysis (Section 4.3.5) showed that these cracks formed as the other principal strain became mildly tensile for a few years around 1969 (Fig. 92).

b) Crevasse pattern over the debris cover

The rock avalanche of 1964 flowed over many crevasses that were buried by winter snow. In the years since April 1964, most or all of the snow cover has melted, and debris has collapsed into the underlying crevasses. Some new crevasses may also have formed. The pattern of crevasses that has been revealed by this collapse (Fig. 98), for the most part, is identical to that recorded on aerial photographs taken in August 1963: with one notable exception (Fig. 99) in a region immediately to the east of the upglacier edge of the cover, north of markers 51 and 52 (Fig. 2).

The contrast in crevasse development in Figure 99a and b, of a more chaotic pattern replaced by a more orderly pattern, suggests that disruption of flow by a large reigel at the confluence of Ana and Sherman Glaciers (Fig. 2) has diminished with time. This can not be
Fig. 98. Crevasse pattern beneath the debris cover on Sherman Glacier (traced by the author from an orthophotographic map compiled by H. Brecher from 1967 aerial photography). The approximate positions of the strain nets are also shown.
Fig. 99. Change in crevassing over a reigel upglacier from the debris cover: A) 1963 B) 1969. (photography by Austin S. Post)
through thickening of the glacier or smoothing of the bedrock: the glacier has thinned significantly in this region (Fig. 11). Decreased velocity of flow over the reigil is a most probable cause of this changing crevasse pattern. Deceleration has been recorded for all markers upglacier from the debris cover (Fig. 81) and this probably is in response to the observed thinning that is due to a long-continued negative mass balance. There is no reason to expect that this change in crevasse pattern upglacier from the debris is a consequence of the rock avalanche.

The sequence of photographs of crevasses appearing through the debris cover from 1964 to 1973 (Fig. 100) appears to show motion of the debris cover and underlying ice over a bed irregularity in the vicinity of marker 70 (Fig. 2) at the distal edge of the rock avalanche and terminus of the glacier. There is no sign that the presence of the debris cover has in any way altered an existing crevasse pattern in this region. The same crevasse pattern can be seen in this area on aerial photographs taken in 1950.

c) Growth of an ice pedestal

The different rates of melting of ice beneath debris and of adjacent debris-free ice has resulted in growth of a pedestal of ice beneath the debris. Between 1967 and 1970, it grew 4.2 m at one site at the eastern edge of the avalanche debris. Measured loss of ice beneath debris at this site was 1.4 m (Fig. 25) while on adjacent debris-free ice, at marker 52 (Fig. 2), it was about 7.7 m (Fig. 37). At the same time, ice thinned by 1.5 m beneath the debris, and by 9 m in the
Fig. 100. Growth of crevasses at the western edge of the debris cover, 1964 to 1973 (photography by Austin S. Post, U.S. Geol. Survey).
debris-free area (Fig. 79).

The directly measured average rate of growth of the ice pedestal is 1.4 m a\(^{-1}\); from the measured rate of thinning of the ice, the expected rate of growth is 2.5 m a\(^{-1}\); whereas the expected rate from differential ablation at the site is 2.1 m a\(^{-1}\). The ice pedestal which was 10 m high in 1967 and 14.2 m high in 1970, might be expected to have kept the vertical strain-rate more compressive in the ice column beneath it than it might otherwise have become, but this is not apparent in the data of the previous paragraph. From that data, it can be calculated that the absolute vertical velocity of a point of ice in the debris-free area is 0.43 m a\(^{-1}\) downward, whereas beneath the debris it is only 0.03 m a\(^{-1}\) downward (calculated from observed thinning minus net balance). At this site, the presence of debris appears to have made the vertical strain-rate less compressive, and this is the reverse of the expected result. Curvature of the glacier bed should also be used in assessing these changes, but it is probably not significant in this region (Fig. 112). The rates of thinning are calculated from change in elevation of markers that are several hundreds of meters from the pedestal edge (markers 51 and 52, Fig. 2), so little note should be taken of the discrepancy between this data. The measured ablation on the debris-free ice also comes from several hundred meters away, marker 52, and the -0.7 m a\(^{-1}\) difference between measured pedestal growth and suggested differential ablation is consistent with expected differences in ablation between ice at the marker and ice at the pedestal foot. Accumulation of drifted snow against the pedestal foot should significantly reduce net ablation in this area, despite an
increased summer melt due to heat radiated from the warm dark debris.

Thus, growth of an ice pedestal at the upglacier edge of the debris cover has had no observed effect on glacier flow.

4.3.9 Changes down-glacier from the rock avalanche

a) Change in area

A significant decrease in area of ice down-glacier from the debris cover is illustrated in the sequence of photographs in Figure 101. Decrease in area stems from two causes: passive motion of the debris atop moving ice (60 to 80 m a\(^{-1}\) at marker 70) that produced a reduction in area of about 10 percent each year between 1964 and 1970; and an active motion of debris down the slopes of the ice pedestal that has developed through differential ablation of ice beneath the debris (see Fig. 105). This gravity sliding of debris produces only a minor decrease in area, but it has blurred the finely digitate distal margin of the avalanche deposit (compare Fig. 101 b and f).

Change in position of the terminus has not appreciably altered the area of glacier down-glacier from the debris.

b) Change in position of the terminus

Marangunic\(^{1}\) (1972, p. 113) noted small changes in position of the terminus of Sherman Glacier between 1965 and 1967. Between 1965 and 1966 the glacier receded 5 to 10 m. By the time of the 1967 survey, the northern part of the terminus had advanced up to 20 m but much of
Fig. 101. Decrease in area of ice down-glacier from the debris, 1964 to 1969 (photography by Austin S. Post).
the terminus remained stationary and some sections even receded. This response continued to 1969 (Fig. 102). In 1970 some sectors that had shown slight advance by 1967, showed a small (2 m) retreat from the 1969 position. At the end of the 1970 summer field season some 25 m of a more active portion of the terminus was seen from the air to be collapsed and partially floated away (Fig. 103). It had been undermined by one of several river outlets when 0.55 m of rain fell in a four-day period. By the summer of 1971, a dark coating of debris spilling from the rock avalanche cover was causing slight retreat (5 m) of the southern side of the terminus.

This vacillating behavior of the terminus of Sherman Glacier over the last eight years stands in marked contrast to the continuous rapid retreat recorded by Tuthill and others (1968) for the previous fifty years. They record an average annual retreat of 46 m between 1959 and 1964, but note that by 1965 annual recession had dropped to zero. This is not in accord with available aerial photographs; these show that between 24 August 1964 and 25 August 1965 a moraine formed about 20 m in front of the 1965 terminus. Aerial photographs taken 27 August 1963 show a pair of parallel annual moraines 20 to 25 m apart at the 1963 terminus. Thus, at least half of the observed change in rate of retreat of the glacier terminus is unrelated to the rock avalanche, and must be considered as lying within the normal range of annual retreat that must fluctuate in response to arrival at the terminus of kinematic waves of climatic origin.

Virtually the entire response of the terminus is due to arrival of
Fig. 102. Map of the changing position of the glacier terminus. 1965 data by W. O. Field, 1966 and 1967 data by Č. Marangunić, 1969 data by the author.
Fig. 103. Collapse of a portion of the terminus in the summer of 1970. The terminus previously had been at the small moraine ridges.
kinematic waves, because debris has hardly begun to have a direct
effect, but the fraction of change due to a kinematic wave generated by
the rock avalanche can not be evaluated (see Section 4.4.6).

The coincidence of change in rate of retreat of the terminus of
Sherman Glacier with deposition of the rock avalanche may be just that -
a coincidence. The only change at the terminus that with any certainty
can be attributed to the rock avalanche is the recent initiation of
retreat through accelerated melting caused by the thin debris cover
that is sliding and being washed from the avalanche cover.

c) Change in profile of the terminus

As debris has been carried forward atop of the glacier the average
slope of the ice surface at the terminus has steepened (Fig. 104).
Some of this steepening is a passive response as the distance between
debris and terminus lessens and some is a passive response to a local
longitudinal compressive strain, but some part may also be a response
to change in thickness and accompanying increase in compressive strain
rate. The latter two show as an increasing convexity of the slopes of
the debris-covered ice pedestal. The ice pedestal here has grown at
about 5 m a⁻¹.

d) Moraine formation

During most of the years of observation of the terminus of Sherman
Glacier, substantial moraines have developed along most of the ice
front, the largest being on the northern side. Four types of moraine
Fig. 104. Change in profile of the glacier terminus, 1967 to 1969 (1967: from a 1:10,000 contour map by H. Brecher; 1969: from a terrestrial triangulation survey).
have been distinguished; although all are essentially annual moraines that have been obliterated in succeeding years by variations in the outwash streams. Three of these moraine types are ice-cored push-moraines made entirely, or dominantly, of a thin layer of outwash gravel over re-activated buried ice.

The one moraine type (Fig. 105) for which no evidence of pushing was observed, consists of a low ridge 1.5 m higher than the outwash plain and slightly lower than the active glacier-ice level during the ablation season. Except during times of heavy ablation, ice overhangs the crest of the ridge and a gap usually exists between ice and ridge crest. Debris, melting out of the base of the ice, cascades down the outward-facing front of the ridge whose slope is at the angle of repose of fresh wet bouldery till. This ridge, probably composed entirely of till, rests on outwash and appears to contain no ice-core, although ice is undoubtedly present below outwash beneath the till ridge, as it is for some distance down the outwash plain (this is evidenced by continued development of pitted outwash). This ridge grew continuously but not significantly over two years from 1968 to 1970 (compare Fig. 106 a and b).

The several types of push-moraine are distinguished more on the basis of content than on mode of formation or morphology. Two types are composed entirely of outwash gravel and buried ice. One of these (Fig. 107) consists of a thin layer (1 cm to 1 m) of outwash gravel over a ridge of re-activated buried ice, while in the other (Fig. 110) no ice is found in the ridge at a depth of more than 1 m. The third
Fig. 105. Schematic cross section of moraine formed by deposition of till at the terminus of Sherman Glacier (for scale see Fig. 106 a and b).
Fig. 106. Moraine formed by deposition of till: A) 1969  B) 1970. Note the advance of the debris-protected ice pedestal.
Fig. 107. Thin-cover push moraine at the terminus of Sherman Glacier in 1970, and schematic cross-section of thin-cover push moraine.
Fig. 108. Ice-flanked push moraine at the terminus of Sherman Glacier and schematic cross-section of ice-flanked push moraine.
Fig. 109. Till between ice flank and moraine ridge.
Fig. 110. Thick-cover push moraine at the terminus of Sherman Glacier and schematic cross-section of thick-cover push moraine.
Fig. 111. Stepped outer face of the thick-cover push moraine.
of the push-moraine types (Fig. 108) is faced on the glacier side by till (Fig. 109), and by a thin slab of active glacier ice containing some englacial debris. All three types of push-moraine have numerous small rounded step-like formations in the outwash in front of them (Fig. 110), and, where their fronts are not steepened to the angle of repose of damp outwash (45° for outwash at Sherman Glacier), the outward facing slope is also stepped (Fig. 111). These steps are discrete dislocations of the gravel. Some of these "faults" had lateral as well as vertical displacement, where glacier ice pushed obliquely against moraine.

No direct observation of pushing has been made. An attempt was made in 1969 to measure the forward motion of a small push-moraine in front of Sherman Glacier. By the summer of 1970, ice had receded 2 m at that point and the moraine had remained stationary.

At ablation rates that have been estimated as as much as 11 m of ice each year, it could be expected that the ice ridge flanking the moraine in Figure 108 would waste away within one or two months if it were not continually replenished by glacier motion. Measurements made over a one-month period in July and August 1969 showed a decrease in height of only 50 cm for the highest point on one ice ridge (J.-R. Kläy, personal communication). This indicates a continuous upward component of motion of the ice. Over the same period the relative height of the ridge above the ice-floored trough beside it did not change as far as could be determined. Thus bending of the thin tongue of ice is not a major mechanism in the formation of the unusual feature, if it occurs at all. If it were a major factor, then the relative height would be
expected to grow because the vertical velocity component of ice below the trough would be less than that in the ice ridge, while both are subject to about the same ablation.

It is thought that the form of the ice flank, which was first present in 1968 as a small feature, and which had grown to be nearly 6 m high by the summer of 1970, grows during the winter and early spring through the combined effects of differential ablation at low angles of the sun and of differential snow accumulation about the moraine ridges. Aerial photographs taken in early spring of 1964 show that snow accumulates in drifts at the base of the ice tongue and that the accumulation pattern is also controlled by valley configuration. Deep snow occurs on the south side of the valley where it persists in the shadow of the mountains into late spring. The transition from snow-covered to snow-free terminus in early spring is exactly the location of ice flank formation. This may be fortuitous, but snow reduces ice ablation and it may also alter the temperature distribution within and beneath the ice.

The ice flank is maintained in the summer probably by slight albedo differences caused by a concentration of dirt along the base of the trough. This is insufficient to cause the form to grow, but seems able to offset differences in ice-loss that would be expected between adjacent concave and convex surfaces. No strain-rate determinations have been made in this area but differences in strain-rate could also contribute to the stability of the ice form.
This interesting feature was destroyed by flooding and undermining by "Sherman River" near the close of the 1970 field season.

The ice-cored push-moraines are formed primarily by a re-activation of stagnant ice, buried beneath outwash, by seasonal advances of the glacier. Some of the outwash may not have been subjected to much horizontal pushing but may merely have been lifted passively on the re-activated ice. The ridge-forms are still, in a sense, push moraines because they owe their origin directly to pushing by the advancing ice. Because it is not known to what extent the outwash gravel is frozen during winter, it is not possible to discuss the role this might play in moraine formation; nothing was ever observed during a summer field season that suggested that the gravel would need to have been frozen to have deformed in the manner observed.

Several of the moraine types described above represent no significant concentration of debris and can not be expected to leave clearly identifiable moraine ridges after all ice has ablated, and, in fact, none of these ridges have survived.

4.4 Discussion

4.4.1 Surface displacement

In the total displacement of markers on the surface of Sherman Glacier (Figs. 72, 79, and 112), only the vertical component (Fig. 79) shows clear evidence of influence of the debris cover. This is as expected: ablation is reduced by the debris and this directly alters
Fig. 112. Profiles of the center line of Sherman Glacier, 1965 and 1970.
(Bottom profile from Marangunic, 1968)
Fig. 113. Annual transverse velocity profile upglacier from the debris cover.
vertical displacement of the surface. Any effect of debris on horizontal displacement must take place indirectly through change in strain-rate due to the changing load that in turn is due to the change in ice thickness. Ice thickness in the region of debris cover (Fig. 112) has remained for the most part unchanged and, therefore, any change in ice thickness that has occurred can have only a very minor influence on horizontal displacement.

It should be noted, however, that ice thickness changed significantly in the debris-free regions, by as much as - 2 m a\(^{-1}\) (about 0.5 percent each year) upglacier from the debris (Fig. 112), and by - 3 m a\(^{-1}\) (up to 100 percent) down-glacier from the debris (Fig. 104). Glaciological theory predicts that internal deformation of ice and glacier basal sliding vary as high powers of ice thickness, hence ice-flow in debris-free areas should have decreased dramatically. A large decrease in flow has been detected upglacier from the debris (Fig. 113) but later discussion shows that this decrease may not be due to change in thickness (Section 4.4.2).

Thus the significant influence of the debris cover upon surface displacement is not the change that it has produced, but the change it has inhibited. In many respects the debris-covered portion of the glacier is now in equilibrium, being able to maintain a constant profile from year to year (Fig. 112).

4.4.2 Velocities

Annual displacements (essentially horizontal velocity components)
of markers along the center line of flow of Sherman Glacier (Fig. 81 and Table 15) show no obvious change that can be attributed to presence of a debris cover. Changes in surface velocity with time, at any fixed point on the glacier, show no significant or consistent trend at the upper limit of the debris; indeed, the changes at all points along a flow-line appear to be remarkably uniform. Thickness upglacier from the debris has decreased some 2.3 percent over the span of observation while within the debris-covered area there has been little change. Thus thickness change does not appear to appreciably alter surface velocity despite the fact that observed velocity change at a transverse profile upglacier from the debris (Fig. 113) is consistent with observed thinning.

This same velocity change is also consistent with known trends in summer rainfall (Table 19) but probably not with available water. But the data can not be used to support a hypothesis of velocity variation due to variation in basal water supply. Total displacement for the year 1968-1969 was at a faster rate than displacement during the period 7 August to 19 August 1969 when available water was above the yearly average.

The cause of most of the observed velocity variation is unknown.

4.4.3 Kinematic waves

The velocity profiles in Figure 81 show other changes with time that are more readily explained. The velocity function along the center line is not a simple function of distance and time. A large
<table>
<thead>
<tr>
<th>Year</th>
<th>Rainfall (m)</th>
<th>Net balance (Mg m(^{-2}) a(^{-1}))</th>
<th>Summer balance (Mg m(^{-2}) a(^{-1}))</th>
<th>Velocity (m a(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1966</td>
<td>1.22</td>
<td>-3.6</td>
<td>-4.6</td>
<td>57.6</td>
</tr>
<tr>
<td>1967</td>
<td>1.20</td>
<td>-4.4</td>
<td>-5.7</td>
<td>55.9</td>
</tr>
<tr>
<td>1968</td>
<td>0.61</td>
<td>-2.6</td>
<td>-5.3</td>
<td>54.4</td>
</tr>
<tr>
<td>1969</td>
<td>0.44</td>
<td>-4.0</td>
<td>-5.0</td>
<td>48.1</td>
</tr>
</tbody>
</table>

Table 19. Comparison between surface velocity and balance parameters at marker 4 on Sherman Glacier and summer rainfall at Cordova FAA (June-September) for the years 1966 to 1969.
part of the change in velocity with distance is a function of configuration of the glacier bed, but there are perturbations in the velocity curves that move with time (Fig. 81), and thereby fit the definition of kinematic waves (see, for example, Paterson, 1969, p. 198). These perturbations have poorly defined and variable wavelengths of between 750 and 1250 m (Fig. 81) and appear to move about 250 m each year, about four times the annual motion of the ice, but this is also the spacing of markers and thus the limit of resolution of the data. The spacing and velocity of the waves suggest that they form every three to five years. The mass balance curves (Figs. 55 and 62) developed in Chapter III suggest that if the kinematic waves are generated by balance changes the glacier is apparently only able to damp out the annual balance variation from summer to winter; the year-to-year variations appear as kinematic waves.

4.4.4 Surface strain at marker 4

A feature of strain-rates at marker 4 is the lack of symmetry across the net, parallel to flow (Fig. 87). The probable explanation of this lack of symmetry is that the net straddles regions of different ice structure and fabric (Figs. 67 and 114). To the north of marker 4, flow is probably dominated by shear in a near horizontal plane induced by an ice fabric inherited from passage of this ice through the base of a large icefall, a region of very high longitudinal stress. South of marker 4, shear in a vertical plane is evident, both in ice fabric and gross structure and in strain measurement (Fig. 87) where simple shear dominates the observed strain. This fabric is inherited from a flow
Fig. 114. Map and cross-section of foliation at the strain net at marker 4.
regime initiated many kilometers upglacier from marker 4 in a region of very strong lateral compression.

Difference in ice fabric does not appear to influence the transverse velocity profile on a time scale of years (Fig. 113) but may have a variable influence from week to week (Fig. 84). Marangunic (1972, p. 107) calculated that less than a quarter of observed surface velocity in this region is due to internal deformation. Slight changes in deformation rate due to differences in strength and orientation of ice fabric may be undetectable at the scale of measurement in this study.

4.4.5 Surface strain at the debris strain net

In contrast to strain-rates at marker 4, surface strain-rates at the debris strain net show little change over the area of the net, but large change in time. A large part of this latter change surely is due to motion of the net from an area of longitudinal crevassing towards a crevasse-free region and is thus caused by bedrock configuration, but another part of the change moves faster than ice velocity and must be unrelated to the bed.

A delay of four to five months occurs between the maximum of longitudinal strain-rate in the upper triangle of the net and the maximum in the lower triangle (Fig. 92). The centroids of these triangles are about 100 m apart, resolved along a flow line. Thus, the apparent delay represents a wave of extending flow traveling at approximately 250 m a⁻¹. A wave of extending flow is identically a kinematic wave: the propagation of some flow property at faster-than-ice velocity.
The seven years of observation were apparently too short to detect the expected co-existent wave of compressive flow.

Data for this analysis were collected at yearly intervals and the strain net is little more than 250 m across in the direction of flow. It is problematical whether the technique of data analysis is capable of revealing the phenomenon that is apparent in the results. The results are, however, consistent with results from data on ice velocities (Section 4.4.3) and crevasse formation (Section 4.3.8a).

4.4.6 Kinematic waves generated by the debris cover

Kinematic waves are generated by positive changes in mass balance (\( \Delta b_n / \Delta t > 0 \)). The Sherman Glacier rock avalanche increased the mass balance of Sherman Glacier and hence generated a train of kinematic waves. The wave of greatest magnitude should be generated where the greatest increase occurs and this is at the western distal margin of the avalanche cover. A wave generated at this margin and travelling at four times ice velocity would be expected to arrive at the glacier terminus five years after initiation if mean ice velocity is 60 m a\(^{-1}\), and some response of the terminus should have been evident in 1969. The greatest change at the terminus, however, was recorded between 1966 and 1967. The thinning of ice towards the terminus may have accelerated passage of the kinematic wave, but this would not be the expected result. The decrease in ice velocity towards the terminus may be such that a kinematic wave travelling at four times ice velocity may not have reached the terminus during the entire period of observation. If mean
velocity through the debris-free region down-glacier from the cover is only $40 \text{ m a}^{-1}$ (an average of $60 \text{ m a}^{-1}$ at the avalanche margin and $20 \text{ m a}^{-1}$ for the extreme terminal edge of the ice), the wave should have arrived at the terminus in 1972, the year after field study ceased, and all observed changes at the terminus are due to balance changes of climatic origin.

4.5 Responses of other glaciers to thick debris covers

Responses of other glaciers that have had thick debris covers dumped on their surfaces are very poorly documented. Many glaciers, throughout the world, have significant portions of their ablation areas blanketed by thick debris, but no inventory of them has been made. These covers result from a multitude of processes: some are catastrophic in origin, through processes such as rock avalanching and volcanic-ash falls; whereas others are more uniformitarian, and form by slow accretion of debris through many repeated small rockfalls and landslides, or through slow ablation of ice that contains debris.

Response of a glacier should be the same no matter what the origin of debris, but in the absence of documented observation, this must remain a supposition. A part of the lack of documentation must result from the problem of studying ice through thick debris, another part from the lack of attention generally given uniformitarian events which take place so slowly as to be over before they are noticed, but a large part of the lack must result from the lack of attention paid by glaciologists to rubble on glaciers. Such features on glaciers are
most usually investigated by glacial geologists who pursue the problem of effect of ice upon debris.

4.5.1 Debris covers emplaced in the St. Elias Range in 1899

Tarr (1909) and Tarr and Martin (1912 and 1914) believed, after witnessing the dramatic advances of a number of glaciers in southern Alaska, that avalanching was an important cause of glacier advance: that a series of large magnitude earthquakes in the St. Elias Range in 1899 had shaken such a quantity of rock and snow from the mountainsides that the mass balances of many glaciers became temporarily perturbed and spectacular advances resulted. Although some thick debris covers formed at that time, Tarr and Martin seem not to have been concerned by any other property of the cover than its mere mass, and did not differentiate rock avalanching from snow avalanching. The validity of this "Earthquake Advance Theory" of added mass through avalanching was questioned by Post (1967, p. D37) who showed that avalanching of snow was insignificant in the 1964 Alaskan earthquake because few large glaciers in Alaska have areas nearby that are topographically suited to large scale snow avalanching.

Tarr and Martin (1914, p. 179) dismissed tectonic uplift as a likely cause of the advances they observed after 1899. Post (1967, p. D34) estimated that the 2 m uplift of the Chugach Mountains in 1964 in the vicinity of Cordova would cause about a 1 m increase in thickness of glaciers and that this probably could not be detected. Earlier, I showed that this 2 m uplift could cause an advance of Sherman Glacier
of between about 100 and 300 m if the glacier had formerly been in equilibrium. Some glaciers in the St. Elias Range were uplifted about 14 m in 1899 (Tarr and Martin, 1912, Plate 14) and this uplift may have triggered the advances that they observed, and not avalanching as supposed.

My earlier analysis of climate and mass-balance relationships in southern Alaska (Section 3.3.19) suggests that the St. Elias Range may have experienced a brief period of positive balance about twenty years before Tarr and Martin visited the area. This might also be expected to have triggered advances of glaciers about the time of their visits.

4.5.2 Debris on Netland Glacier

Post (1967, Fig. 24) presents a view of Netland Glacier, Yukon Territory, that shows a large rock avalanche deposit that may have been in place as long as 12 years. The source of the avalanche and displacement of the debris since falling are not evident in the figure. The photograph, taken in 1964, shows Netland Glacier abutting into a vegetated digitate terminal moraine ridge while, a few kilometers upglacier, bare lateral moraine ridges stand above the level of the ice. Some unusual form of advance of the glacier had occurred before 1964, and the debris cover is apparently its cause. The glacier has not been re-photographed since 1964 (Austin Post, personal communication).

The upglacier margin of the rock avalanche is oddly curved and apparently ridged, and the glacier immediately upflow from the avalanche debris appears excessively crevassed, but causes of these features are
not evident in the photograph.

4.5.3 Old debris covers on Slide Glacier

If Tuthill's (1966) interpretation of the origin of superglacial drift on Slide Glacier (formerly "Sioux Glacier") in the Martin River valley system, south-central Alaska, is correct, he shows one of the longer sequences of frequent observations of a thick avalanche cover on a glacier. However, it is most probable that the appearance of an avalanche deposit on the 1938 photograph (Tuthill, 1966, Fig. 2, by B. Washburn) is an optical illusion, and this latter interpretation better fits with Tuthill's later sequence of photographs that show only a medial moraine that widens towards the glacier terminus, and a glacier that has changed very little in 20 years.

4.5.4 New debris cover on Slide Glacier

Reid (1969) undertook repeated surveys of markers on the thick debris cover that resulted from the 1964 avalanche on Slide Glacier, to determine its effects, but was only able to repeat his surveys once. He calculated displacements of markers along one transverse and one longitudinal line of markers (Fig. 115), ten markers in all, and made leveling traverses along both lines in each year (Figs. 115 and 116). His displacement data show an expected decrease in velocity down-glacier and an expected rise of the glacier surface over much of the debris cover during the year of study (Fig. 115).

The maze of crevasses that appeared around Reid's marker 10
between 1965 and 1966 probably resulted from collapse into existing crevasses of snow bridges preserved beneath the debris since the winter of 1963-1964 as at Sherman Glacier, and not from any dramatic change in glacier behavior as he suggests.

His longitudinal profile of Slide Glacier shows changes in elevation that are larger than those recorded for Sherman Glacier (Fig. 79), but not greatly so. The increase in thickness, at the lower end of the profile recorded by Reid, results from three causes: reduction in ablation, longitudinal compression, and movement of ice into locations where ice was previously thinner. Reid's transverse profile (Fig. 116) shows a change from 1963 to 1965 that is probably not "a good approximation" (Reid, 1969, p. 363). If marker 5 was raised 28 m between 1963 and 1965, as Reid inferred, the debris cover had a remarkable effect. If the debris was as much as 3 m thick and reduced ablation by as much as 6 m of ice each year at marker 5, then 13 m of snow probably lay under marker 5 when the avalanche fell.

Comparison of his Figures 3 and 5 suggests that it is more probable that marker 5 lies atop of a short length of medial moraine now buried by avalanche debris. Reid also observed that even areas such as medial moraines that have been debris-covered at least since 1938, and which were not covered by the 1964 avalanche, have also responded to the new loading, suggesting that the glacier responds more to regional loading than to the local loading of a raised medial moraine.

The response of Slide Glacier to emplacement of its debris cover has thus been very similar to that of Sherman Glacier; that is,
Fig. 115. Longitudinal profile of Slide Glacier (from Reid, 1969, Figure 9).
Fig. 116. Transverse profiles of Slide Glacier (from Reid, 1969, Figure 8).
different parts of the debris-covered Slide Glacier have thickened, thinned, or stayed the same thickness, dependent on local ablation, internal flow, and bedrock topography. Any further response went undetected or did not occur during the brief study.

4.5.5 A prehistoric response of Chelingletscher, Swiss Alps

Röthlisberger (1969, and also Zoller and others, 1966) suggested a rock-avalanche-induced surge of Chelingletscher to explain a part of the glacial history of the Göscheren Valley. He presents evidence that about $1.5 \times 10^6 \text{ m}^3$ of debris was transported 4 to 5 km along slopes that average $4.8^\circ$ along an upper section and $1^\circ$ along the lower kilometer. The debris is now 4 to 10 m thick, with a hummocky upper surface on which hummocks arch across valley with convex side down valley. He also mentions lateral ridges on the valley wall that stand 30 to 50 m above the valley floor. He suggested a glacier surge mechanism to explain this occurrence but his description suggests to me that the event was a typical large rock avalanche of the type described in Chapter I and by Shreve (1968b). A glacier, if present at that time, appears to have provided only a passive substrate beneath the avalanche, and played no active role in movement of the debris. In contrast, Sherman, Slide, and Bronva (Section 4.5.8) Glaciers have played active roles in debris motion since emplacement of their avalanches.

4.5.6 Rock avalanches at the mouth of Ötz Valley; Inn Valley, Austria

Heuberger (1966) described two rock avalanches at the junction of Ötz and Inn Valleys below Tschirgant. He proposed that the smaller
Tschirgant landslide at Haiming fell upon the Ötz Glacier at about the time of the Steinach maximum (Older Dryas, about 10,000 years B.C.) whereas the larger Tschirgant landslide fell on the same glacier but immediately before the later Gschnitz maximum (Younger Dryas, about 8,800 years B.C.). He proposed a series of events (Table 20) in the history of the Ötz Valley glacier around the times of these maxima but this series was primarily based on studies in the region of the two avalanches. The sequence bears a striking similarity to an expected sequence resulting from two rock avalanches that each dammed the Inn Valley on separate occasions and glaciers may not have been involved except in the oversteepening of the mountain slopes. Shreve (1968b) has shown that rock avalanche deposits can be internally stratified, that they can erode and deposit materials along their paths, and that they can travel over soil and vegetation at times without noticably disturbing them. Recognition of these features, and other properties of large rock avalanches, and knowledge that lateral moraine deposits from the Ötz Valley glacier still occur to an elevation of several thousand meters on the walls of Tschirgant and would have lain in the avalanche paths, suggest that Heuberger's interpretation of the series of events is not the only one possible, and may not be the most probable one.

Debris-covered glaciers ultimately become stagnant and eventually leave large high lateral and terminal moraine ridges. If the two avalanches at the Ötz Valley mouth had fallen on ice, some remnants of these large moraine ridges should survive, but they do not. These
<table>
<thead>
<tr>
<th>Heuberger interpretation</th>
<th>McSaveney interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oldest Dryas</td>
<td>Same</td>
</tr>
<tr>
<td>Separation of Ötz and Inn Glaciers. Ice-dammed lake at Imst, Inn Valley, dammed by Ötz Glacier, lake drained repeatedly.</td>
<td>Same</td>
</tr>
<tr>
<td>Bolling</td>
<td>Same</td>
</tr>
<tr>
<td>(10,750-10,350 BC)</td>
<td>Same</td>
</tr>
<tr>
<td>Retreat of Ötz Glacier, disappearance of Imst lake, first vegetation in valley bottom, but no extensive forest.</td>
<td>Same</td>
</tr>
<tr>
<td>Steinach (Older Dryas)</td>
<td>Rock avalanche fell from Tschirgant and swept over Haiming soil to dam Inn Valley, lake formed.</td>
</tr>
<tr>
<td>Advance of Ötz Glacier over frozen till and soil to end at Haiming. Imst lake formed again. Landslide of Haiming fell from Tschirgant onto Ötz Glacier.</td>
<td>Inn River incises through debris and lake drains.</td>
</tr>
<tr>
<td>Allerod (10,000-8,800 BC)</td>
<td>Inn River incises through debris and lake drains.</td>
</tr>
<tr>
<td>Retreat of Ötz Glacier leaving youngest till. Soil formed at Karres. Last outbreak of Imst lake extensively destroying Steinach ground and end moraine. Extensive forests formed in the region.</td>
<td>Inn River incises through debris and lake drains.</td>
</tr>
<tr>
<td>Gschnitz (Younger Dryas)</td>
<td>Large rock avalanche sweeps from Tschirgant and again dams Inn Valley, lake forms.</td>
</tr>
<tr>
<td>(ca. 8,800 BC)</td>
<td>Large rock avalanche sweeps from Tschirgant and again dams Inn Valley, lake forms.</td>
</tr>
<tr>
<td>Advance of Ötz Glacier to the foot of Tschirgant. Just before maximum advance the great Tschirgant landslide fell. Loess formation, solifluction vigorously erodes and destroys Steinach moraine and Allerod soil and loess.</td>
<td>Large rock avalanche sweeps from Tschirgant and again dams Inn Valley, lake forms.</td>
</tr>
<tr>
<td>Post-glacial</td>
<td>Lake drained.</td>
</tr>
<tr>
<td>Inn and Ötz rivers cut into the debris left at the Ötz Valley mouth.</td>
<td>Lake drained.</td>
</tr>
</tbody>
</table>

**Table 20.** Chronology of Late Glacial events at the junction of the Ötz and Inn Valleys (from Heuberger, 1966).
ridges would be large and blocky and probably would have survived the brief period of intense solifluction that Heuberger suggests destroyed the expected moraines. This area deserves close reexamination.

4.5.7 Debris-covered glaciers in the Mt. Cook region of New Zealand

Thick debris blankets are conspicuous features of lower reaches of all of the larger glaciers in the Mt. Cook region of New Zealand. So far as is known, the major part of these debris covers has accreted through slow processes of ablation of debris-laden ice and accumulation of multitudinous small rockfalls and debris-filled snow avalanches, and has been in place with little change during a century of observation. Major glaciers on the western side of the Southern Alps adjacent to this region lack these continuous debris blankets.

Glacial histories of moraine development over the last 350 years have been evaluated recently for a number of glaciers in both of these regions: the Mt. Cook region by Burrows (1973), and the Westland National Park region by Wardle (1973). Similarities between their chronologies are apparent in Table 21.

Presence of a debris blanket on a glacier's ablation zone apparently does not prevent it from responding directly to climatic change, nor is there any spurious advance that might be attributed to a catastrophic emplacement of debris cover. Relative sizes of each of the various advances vary more among the debris-covered Mt. Cook glaciers than among the debris-free Westland ones, but this is probably due more to vagaries of climate and glacier response than to debris covers.
<table>
<thead>
<tr>
<th>Westland</th>
<th>Fox</th>
<th>Mueller</th>
<th>Tasman</th>
</tr>
</thead>
<tbody>
<tr>
<td>Advance to 1968</td>
<td>Advance to 1970</td>
<td>Shrinkage to 1970</td>
<td>Shrinkage</td>
</tr>
<tr>
<td>Advance to 1954</td>
<td>Thickening 1948-53</td>
<td>Advance to 1930</td>
<td>Shrinkage, then stillstand to 1913</td>
</tr>
<tr>
<td>Advance to 1934</td>
<td>Advancing in 1934</td>
<td>Shrinkage 1916-27</td>
<td>Large advance to 1890</td>
</tr>
<tr>
<td>Advance to 1909</td>
<td>Advancing in 1892</td>
<td>Shrinkage, then stillstand to 1913</td>
<td>?</td>
</tr>
<tr>
<td>Small advance about 1880</td>
<td></td>
<td>Advance to 1890</td>
<td>Advance to 1850</td>
</tr>
<tr>
<td>Advance before 1840</td>
<td>Advance before 1840</td>
<td>Shrinkage 1860-80</td>
<td>Advance to about 1820</td>
</tr>
<tr>
<td>Advance before 1790</td>
<td>Advance before 1790</td>
<td>Advance to about 1810</td>
<td>Shrinkage to about 1780</td>
</tr>
<tr>
<td>Advance before 1620</td>
<td>Advance before 1620</td>
<td>Shrinkage to about 1780</td>
<td>Large advance to 1750</td>
</tr>
<tr>
<td>Advance before 4730 B.P.</td>
<td></td>
<td>Large advance to 1750</td>
<td>Two advances to 1680</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Advances 1580-1650</td>
<td>Advance to 1550</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Advance to 1550</td>
<td>Series of 4 advances last 2 at about 1250 and 1150</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Series of 7 undated advances</td>
<td></td>
</tr>
</tbody>
</table>

Table 21. Comparison of fluctuations of Westland and Mount Cook glaciers (from Wardle, 1973; and Burrows, 1973)
Fig. 117. Moraine spacing at four New Zealand glaciers, showing the contrast in spacing of moraines between the debris-free Westland glaciers and the debris-covered Mount Cook glaciers (from Wardle, 1973; and Burrows, 1973)
A more striking difference between glacier response on the two sides of the mountain range is the spacing of moraine systems (Fig. 117). At Mueller, Hooker, and Tasman Glaciers in the Mt. Cook region, terminal moraines of more than four centuries of glacier fluctuation are compressed into single moraine ridge complexes that are scarcely more than half a kilometer wide, (Burrows, 1973, Figs. 12, 13, 14, and 16). In contrast, on the western side of the mountains, Fox and Franz Josef Glaciers, which are similar in size to Mueller and Hooker Glaciers, have well-separated moraine ridges dating from the last 200 years spread over 3 km of valley floor (Wardle, 1973, Figs. 13 and 16). Size of glacier apparently exercises little control over this spacing: the exceptionally long Tasman Glacier does not differ in response from its shorter neighbors, while Fox and Franz Josef Glaciers exhibit responses that seem typical of the western glaciers despite their atypical size.

While it is possible that differences other than presence or absence of a debris cover are causes of these differences in moraine spacing, the persistence of a continuous thick ice mass beneath a debris cover for long periods must change a glacier's response to climatic change. Debris-covered glaciers can substitute down-wasting and thickening for the more usual glacier retreat and advance, and this must tend to maintain their termini at near constant locations, with consequent build-up with time of enormous moraine ridge complexes.

4.5.8 A historic rock avalanche on Brenva Glacier, Italy

On 14 November 1920, the tongue of Brenva Glacier on the Italian
side of Mont Blanc was buried under a blanket of detritus (Fig. 118) during a landslide that fell from near the summit (Capello, 1941, p. 127). Observation of the position of the terminus of Brenva Glacier has been made almost annually since about 1909 (Fig. 119). Brenva Glacier began advancing about 1914 along with many glaciers in the Alps (Cerutti, 1971, p. 262), but all others had ended their advances by about 1925, while Brenva Glacier continued until 1941, in a total advance of 500 m that brought it close to its maximum extent of 1818 (Capello, 1941, p. 133). The cause of this abnormal advance was the subject of some divergence of opinion among Italian glaciologists (Capello, 1941, p. 127); some believed it a result of increased precipitation, while others, including Capello, believed that the continued advance after 1925 was a result of the debris cover. Capello (1941, p. 135) was not sure whether the advance was due to the mass of debris itself or to reduction of ablation beneath the cover, but was certain that it was independent of variation in snowfall.

After 1941, the tongue became stagnant for all practical purposes (Cerutti, 1971, p. 259) and slow down-wasting began. It was almost totally separated from its source of accumulation in 1959 (Capello, 1971, Fig. 2).

In the summer of 1968, in response to favorable climate, Brenva Glacier began a resurgence of activity marked by rapid thickening and an eventual advance that has continued to the present (Lesca, 1972, p. 93). Between 1959 and 1971, occasions when maps were made, the glacier thickened 20 to 30 m and gained some \(14 \times 10^6\) m\(^3\) of ice, mostly
Fig. 118. The tongue of Brenva Glacier, Italian Alps, nine years after a major landslide from Mont Blanc, 14 November 1920 (from Figure 1, Capello, 1941).
Fig. 119. Comparison of behavior of the termini of Italian glaciers - 1880 to 1940. Note that no sudden change in behavior of Brenva Glacier occurred after 1920 (from Capello, 1941, Figure 4).
Fig. 120. The terminus of the advancing debris-covered Brenva Glacier, Italian Alp, 1930 (from a photograph in Capello, 1941, Figure 5).
between 1965 and 1971 (Lesca, 1972, p. 96-97). Between 1970 and 1971 the reactivated debris-covered glacier advanced 20 m while surface velocity 500 m upglacier was 50 m a$^{-1}$ (Lesca, 1972, p. 95).

Thus, the response of a glacier to emplacement of a rock avalanche had been well studied and documented as early as 1941. Yet the same question still arose in the case of Sherman Glacier. Was the observed behavior of Sherman Glacier due to climate or due to the avalanche cover? No one appears to have profited by the past experience and several erroneous inferences and predictions were made about the behavior of Sherman Glacier that were inconsistent with the well-documented record for Brenva Glacier. The Brenva Glacier study is also valuable in that it enables the future behavior of Sherman Glacier to be predicted with greater certainty.

By 1971, Sherman Glacier had not yet developed a more-or-less vertically cliffed terminus, such as that present in 1941 at Brenva Glacier (Fig. 120), and the large lateral and terminal moraines such as were present at Brenva (Figs. 118 and 120), had not yet formed.

4.6 Future behavior of Sherman Glacier

Between 1964 and 1971, the rock avalanche on Sherman Glacier wrought few significant changes in flow of the glacier beneath it. Yet, there are changes that can reasonably be expected. These are: 1) an advance of the terminus, 2) a change in flow of ice beneath the debris, 3) the development of a terminal moraine, and 4) a change in the future response of the glacier. It is presumed that these changes will
take place.

4.6.1 Advance of the terminus

Position of the terminus of any glacier is determined by the flow of ice into the terminal region and the net balance there. By causing the glacier to thicken near the terminus (by as much as 10 m a\(^{-1}\) between 1967 and 1970, Fig. 104), the veneer of debris has increased the mass flux into the terminal region of Sherman Glacier, while, through the passive transport of the thick cover towards the terminus, net balance in the region is being significantly increased. Thus, the pre-avalanche rapid retreat of the glacier can not be maintained, regardless of whether or not the present cessation of retreat is independent of the avalanche. At some time in the future the reduction in area of debris-free ice will so increase the areal net balance of that region that inflow of ice will more than compensate the loss through melting, and an advance will occur.

Marangunić (1972, p. 126-127) suggested that the debris-covered ice could be considered as an isolated mass that would seek an equilibrium profile similar to an ice cap. Using Orowan's (1949) simplest expression for this profile he suggested that the glacier ultimately would entirely fill the outwash area between Sherman and Sheridan Glaciers, an advance of 3.3 km.

Rather than seek a theoretical solution, I have sought to extrapolate from current data to estimate the magnitude of the expected advance. It is certain that no future advance can be extrapolated from
the current standstill and retreat, but the rate of lowering of the 
surface of the advancing debris front can be extrapolated to predict 
when the debris might reach the level of the outwash plain, when, 
because the ice will then be gone from beneath the debris front, the 
advance should cease, or be very much impaired.

Surface lowering of the leading edge of the debris front is $5 \text{ m a}^{-1}$ 
(Fig. 104) and the front is moving forwards at $65 \text{ m a}^{-1}$ at the southern 
side of the terminus (Fig. 104). In 1967 the debris stood 100 m above 
the outwash plain (Fig. 104). Simple extrapolation of the rate of 
lowering predicts that in 20 years the debris will be level with the 
outwash. At this time, ice velocity will be zero (i.e. no ice). A 
mean forward speed of $32.5 \text{ m a}^{-1}$ maintained for 20 years will produce 
an advance of 650 m for the leading edge of the debris, which would be 
a net glacier advance on the southern side of only 300 m from the 1967 
position (1967 is used because the form of the glacier is precisely 
known in that year through mapping from aerial photography).

Along the center-line of flow at marker 70, surface lowering is 
about 1 m each 10 m of forward displacement (Fig. 79). From the 1967 
elevation of marker 70 (300 m), the debris should lower to the 100 m 
elevation of the outwash plain after a lateral displacement of about 
2 km, which is a net glacier advance of about 630 m from the 1967 
position of the terminus (Fig. 6). This advance could be completed in 
50 years if marker 70 maintains an average speed of $40 \text{ m a}^{-1}$.

In 50 years, at present rates of thinning of $2 \text{ m a}^{-1}$, ice thickness
near the present upglacier edge of the debris will be some 100 m less thick, a one third reduction in thickness of the glacier at this point (Fig. 112). If this thinning is associated with a balance regime that would cause a 45 m a\(^{-1}\) retreat of the terminus of a debris-free glacier (46 m a\(^{-1}\) was the average rate of retreat from 1959 to 1964, Section 4.2), the glacier terminus should then lie 2.25 km east of the 1967 terminus, and should also be some 0.75 km east of the upper edge of the debris cover (which has a forward speed of about 60 m a\(^{-1}\)). That is, the debris-covered ice should be separated from Sherman Glacier in less than 50 years unless there is a significant deterioration in climate and increase in mass balance. This suggested retreat is well in accord with an expected terminal position of a glacier that is only 200 m thick at the present upglacier edge of the debris.

A 630 m advance in 50 years provides for an average rate of advance of less than 13 m a\(^{-1}\), but in the first ten years none of this advance has occurred, and none should be expected at this central portion of the terminus in the first twenty years until the foot of the ice pedestal reaches it. Thus, the glacier should advance at an average rate of about 20 m a\(^{-1}\) for the thirty years after 1984. The distance of parts of the irregular debris front from the glacier terminus varies across the glacier and makes these estimates somewhat imprecise, but the rate is remarkably similar to the current rate of advance of debris-covered Bremner Glacier for similar ice velocities (20 m a\(^{-1}\) for ice flow of 50 m a\(^{-1}\), Lesca, 1972, p. 95).
4.6.2 Change in flow of ice beneath the debris

There are several reasons for predicting that the 630 m advance is probably a maximum estimate. One reason is the observed increase in the rate of surface lowering at marker 70 (Fig. 79) which, if continued, will run the debris into the ground sooner than predicted by a linear extrapolation of the present rate. Another reason is the prediction that the debris-covered section will be isolated from Sherman Glacier before the advance is completed. A third reason is that the rate of flow of the glacier will significantly decrease with time. This has been considered in a simple way in the decrease in flow at the terminus, but major slowing of flow will also become apparent over the entire glacier, as the glacier thins through flow and ablation, and particularly as the surface gradient decreases as the glacier both thins and extends itself. The surface gradient change may become particularly significant as Sherman Glacier begins to separate from the debris-covered mass. Before this event occurs, surface gradient will decline to zero, then reverse itself over the upglacier area of the debris. The accompanying reversal in flow will retard the predicted separation, and may also reduce the predicted advance.

The predicted separation also will be retarded by an increased mass flux into Sherman Glacier from Eliza, Andres, and Fidalgo Glaciers (Fig. 2). Mass balances of these three tributary glaciers have been significantly increased by debris that blankets their entire ablation zones (Fig. 3). The increased contribution of ice from these glaciers will greatly alter the pattern of strain within the glacier, but flow
from these glaciers is slow, and the size of their ablation areas small, and no change is yet apparent. The effect of debris on flow from these glaciers will be quite long-lived because thick layers of debris persist under snow and firn into their accumulation zones and the surface covers of the ablation zones can be replenished from those sources for some time to come.

4.6.3 Development of a terminal moraine

Each winter, at least since the winter of 1962-1963, Sherman Glacier has pushed up a small annual moraine at its terminus, but none has been very large, and all rapidly have been swept away by summer meltwater streams, usually swelled by abundant runoff from rain. The arrival of avalanche debris at the terminus will increase the size of moraines that are built each year, but not by any significant amount, and increase in size alone will do little to ensure their survival. When some of the many large boulders in the debris, however, reach the outwash plain, they will not be eroded by the streams which probably lack the competency to move them, and a bouldery moraine should develop at the glacier terminus.

The presence of a large moraine ridge will greatly alter the sediment regime of "Sherman River". At present, base level at Sherman Glacier is apparently controlled by outlet levels of lakes dammed by Sheridan Glacier. Unless the volume of fine gravel brought to the terminus is sufficient to bury the boulders, a series of rapids will form at the bouldery moraine ridge, and a set of matching terraces will
form below the moraine, as it takes over the control of local base level. As the moraine grows, it will probably also stabilize the outlets from beneath the glacier, and this too, should help preserve the moraine.

Lateral moraines will also form, from debris that falls off the sides of the glacier. These lateral moraines will be unusual in that they will not define a former synchronous maximum extent of the glacier,, but will be time transgressive and unsuitable for use in reconstructing a former ice profile. The upglacier ends of these lateral moraines were formed when the avalanche fell in 1964, but the terminal ends may not form for another 50 years. By then, the ice will be 100 m thinner at the upper end of the lateral moraines.

4.6.4 Change in future response

If Sherman Glacier follows the behavior of other debris-covered glaciers, the long persistence of ice beneath cover will change the glacier's response to future climatic deterioration, and the glacier will continue to utilize the same terminal position for many years, despite its varying mass balance, with periods of down-wasting and virtual stagnation during balance minima, followed by phases of rapid passage of kinematic waves and thickening. Sherman Glacier, however, differs from most other debris-covered glaciers in that its cover is a result of a unique event, not likely to recur at this glacier for many centuries (consider the statistics of the thousands of glaciers in Alaska and the low frequency of major rock avalanches of about one every two years). Thus, with time, Sherman Glacier will rid itself of
its debris cover and revert to its former behavior.

4.7 Glacial history and the Sherman Glacier rock avalanche

Before concluding an examination of the effects of a rock avalanche on Sherman Glacier, it is useful to reflect on what aspects will be preserved for the geologic record and how they might be interpreted.

The rock-avalanche origin of superglacial debris on Sherman Glacier can readily be recognized through the preservation of a flow pattern and through a lithological continuity from boulder to boulder in trains across the debris. Both of these characteristics are absent from the debris cover on Tasman Glacier in New Zealand, which is not due to a single avalanche, but they are not characteristics that can be expected to survive through deposition, and they certainly would be difficult to recognize on the floor of a moss-enshrouded Alaskan rain-forest. If the geology of the region had not been so homogeneous, and if the source area had been a distinctive rock type, the origin of the debris might readily have been apparent, and the debris that travelled almost directly across to the opposite side of Sherman Glacier would be a clear indicator of a rock avalanche origin. Instead, the only clear indicator of an unusual origin is the debris over soil on the overtopped spur opposite Shattered Peak in Andres Glacier, and this area is so remote from the rest of the glacial geology that it almost certainly would be overlooked in any study.

A part of the glacier's response to the avalanche will be recorded in the profile of lateral moraines that will be left by the debris. As
has been suggested earlier (Section 4.6.3), these lateral moraines will not record a former ice profile, but a composite one that bears no relation to any former shape of the glacier (Fig. 121). If the time-transgressive nature of this moraine is not recognized, a former mass-balance regime with greatly increased mass turnover and a greatly lowered equilibrium line could be inferred from such a profile.

Other peculiarities in glacial history of Sherman Glacier make it extremely unlikely that the timing of the avalanche could be determined from the deposits that will be left. Wood is abundant in the present margins of Sherman Glacier, but it is not modern wood; it is fresh unrotted logs that are the remnants of a forest destroyed by a glacial advance some \( \approx 2,000 \) years ago. In any sampling of material for \(^{14}C\) dating, this old wood and not modern wood would most probably be taken, and even if many samples were dated, the very rare late 20th-century dates could readily be discarded as contaminated. It is also doubtful that the glacier will advance into any region where much modern wood could be incorporated, so that the true age and time-transgressive character of the moraine will not readily be recognizable. Thus, in any usual investigation of glacial geology, the deposits left by the avalanche might be interpreted as glacial deposits dating from an early phase of Neoglacialization when the glacier equilibrium line was low (about 200 m lower than today's) but the balance gradient was very steep. It is improbable that enough evidence of the actual origin will be preserved to contradict this erroneous interpretation, and it is even more certain that not enough of this evidence could be found.
Fig. 121. Possible profile of lateral moraine formed by rock avalanche debris during the 50 years following the avalanche.
If the avalanche origin of the deposits at Sherman Glacier will not be recognizable in the distant future, it seems to be of questionable value to interpret other deposits left by glaciers as being the results of rock avalanches unless some clear indicator is present, such as an unexpectedly wide distribution of some unique restricted rock type whose source is known. This is a disquieting conclusion when one considers that perhaps 5,000 Alaskan glaciers have endured large rock avalanches in the last 10,000 years, and they are probably not uncommon elsewhere either.
SUMMARY

The Sherman Glacier rock avalanche

The glacially oversteepened mountainsides of Alaska have long been a source of landslides. As in previous earthquakes, a major earthquake in the Chugach and Kenai Mountains in 1964 triggered thousands of slides. Whole ridge crests and peaks fell away in huge rock avalanches to spread widely over glaciers in the valleys below. Fifty-one large rock-avalanche deposits were noted and several attracted attention because of possible effects on the glaciers beneath them and because of their remarkable ability to disperse across gentle slopes. One such avalanche occurred when $12.1 \times 10^6$ $m^3$ of rock, snow, and ice fell from a peak where bedding, joints, and faults were inclined towards a steep face, to cover $8.25$ $km^2$ of Sherman Glacier to a depth of $1.65$ m.

The Sherman Glacier rock avalanche lost half of its available energy to high basal friction ($\mu$ averaging 0.4) in the initial fall from Shattered Peak before porosity, and permeability had developed sufficiently to permit significant internal strain and rolling of basal clasts or, alternatively, before the kinetic energy of the clasts built up to a sufficient level to cause a loss of internal cohesion of the grains and mechanical fluidization of the mass of the avalanche. A maximum speed of $84$ $m$ sec$^{-1}$ was reached at the foot of this peak. The avalanche then slid on snow with an average speed of $26$ $m$ sec$^{-1}$. 330
with a sole of avalanche debris perhaps smoothed by a plaster of snow and ice. The coefficient of basal sliding friction was 0.11. During the 3.5 minutes that the avalanche was in motion, it deformed at an average rate of 0.04 sec\(^{-1}\) as a complex Bingham substance with a yield stress of about 0.02 bar and two apparent viscosities. At shear stresses between 0.02 and 0.1 bar the debris flowed with an apparent viscosity of about 1 \(\times 10^6\) poise that was probably largely a function of friction between adjacent moving clasts. At higher shear stresses (> 0.1 bar) the debris flowed with an apparent viscosity of 4 \(\times 10^7\) poise when the strain-rate was limited by the thickness and permeability of the avalanche which limited the flow of air within the debris. This flow of air is a necessary part of the dilation and collapse of pore space in the debris that occurs with internal strain of granular masses. The yield stress of 0.02 bar apparently is the result of a transition from kinetic to static friction between clasts when their relative motion is very small (< 0.001 m sec\(^{-1}\)).

These high apparent viscosities permitted only slow internal strain so that deformation during the avalanche was largely confined to the base and the debris slid as a thin flexible sheet without turbulence. The finite yield stress gave the avalanche an apparently rigid crust that failed in longitudinal and lateral tension to form transverse fissures and longitudinal grooves and ridges in the debris surface. The grooves served as zones of stress concentration within the surface crust, and lateral shear stress, from the gradient in lateral shear within the flowing debris, was transmitted through the crust to these
grooves, which were also the sites of zones of lateral shear.

The Sherman Glacier rock avalanche behaved in ways previously thought to require a basal lubricant of air, yet it slid on snow. Lubrication by a basal layer of compressed air is not a viable mechanism for any rock avalanche: 1) because it probably is too efficient a lubricant to permit an avalanche to stop within its known confines; 2) because rock avalanches lack a viable mechanism for first trapping an air layer because they can neither glide nor fly, nor trap any air over which they might be launched in their travels; and 3) because rock avalanche debris lacks the strength to maintain a discrete basal air layer, particularly if the air is under pressure enough to overcome the limiting effect of permeability and thickness of the debris. Instead, the low basal friction that permits wide dispersal of rock avalanches is the kinetic friction of clasts rolling and sliding at the base.

Effect of the avalanche or melting of the glacier

Calculation of heat flow through the avalanche cover suggests that melting of ice beneath debris has been underestimated in the past. Measurement at one site gave ablation of 420 kg m$^{-2}$ a$^{-1}$, twice the previous highest estimate. Analysis of factors affecting thermal conductivity and temperature gradients within the debris suggest that this direct measurement is only half the ablation (880 kg m$^{-2}$ a$^{-1}$) beneath debris of near average thickness near the measurement site, and that at lower elevations, it is yet double again (1.7 Mg m$^{-2}$ a$^{-1}$).
The debris cover has inhibited ablation by about 80 percent.

Advection of heat by percolating rainwater may melt an additional 50 kg m\(^{-2}\) a\(^{-1}\), but this 0.5 \(\times\) 10\(^6\) m\(^3\) a\(^{-1}\) loss of ice beneath the entire cover is localized in channels and tunnels in the ice. Advection of heat also causes greater heat flow through the permeable debris than through large impermeable boulders, despite their greater thermal conductivity. Thus, differential ablation beneath the deposit causes large boulders to be thrust through the debris. Differences in debris thickness also cause differential ablation. Opening of crevasses has greatly reduced the continuity of cover in some regions and resulted in accelerated melting.

The debris has had no noticeable effect on ablation of adjacent debris-free ice despite an observed 10\(^\circ\)C warming of air passing over the cover.

**Mass balance of Sherman Glacier**

Mass balance of Sherman Glacier was measured from 1965 to 1971, and although measurement was beset with problems that severely limit the value of the assessments, areal net balance appears to have been positive since the winter of 1965-1964, largely as a result of three years of very positive balance. Prior to 1964 the glacier probably had an average annual deficit of 25 Tg.

Uplift of 2 m during the earthquake may have increased mass balance by 1 to 3.5 Tg which would cause a 100 to 300 m advance if the
glacier had been in equilibrium, but the increase is less than 14 percent of the average deficit of 25 Tg before the avalanche, and its effect should be well masked by the other balance changes.

From accurate determinations over six years of stratigraphic mass balance at a point at about 430 m elevation on the glacier, and climatological records from nearby Cordova F.A.A., Alaska, a relationship was established between net balance and those climatic parameters that most directly affect mass changes such as winter snowfall, summer temperature, and duration of summer melt. Winter temperature was also found to be important for an unknown reason perhaps related to winter ablation. This relationship enabled mass balances at 430 m and at 630 m elevation to be determined for the last 30 years, apparently with some precision. These estimates suggest that the direct assessments of areal net balance are poorly determined, although of the correct sign and magnitude. In most years, areal net balance may have been determined to about ± 20 Tg.

Comparison of the estimated record with measured records from other areas around the world suggest that the Sherman Glacier record may be quite reliable, and that balance at Sherman Glacier has varied in phase with those of other glaciers during the last 30 years. A record of long-term variations of net balance at Sherman Glacier over the last century derived from climatological records for Sitka, Alaska, also appears to have followed the gross trends of estimated records for Storbreen in Norway, and for Lednik IGAN in the Polar Urals, both of which seem to be reliable records. This long record of balance for
Sherman Glacier is also in excellent agreement with the known history of the glacier's behavior.

A stratigraphy of clean and dirty ice exposed at the surface of Sherman Glacier was examined, but the record is unsuitable for development of a balance record, because 1) the stratigraphy only records years of positive net balance; 2) an unknown portion of the record is covered; and 3) the velocity field in the region of development and deformation of the stratigraphy is largely unknown.

Response of Sherman Glacier to the changes in balance

Surface motion and strain on Sherman Glacier show no changes that can be interpreted as responses to the avalanche during the period of observation from 1965 to 1971. Changes in the surface profile of the glacier suggest that this may be because the glacier maintained a near-constant profile beneath the cover and the lack of change may have been the response.

Surface motion and strain changed with time as kinematic waves of climatic origin travelled through the glacier at speeds of \( \sim 250 \) m a\(^{-1}\). An advance of about 20 m of a sector of the glacier terminus between 1966 and 1967 was initiated by arrival at the terminus of one such wave probably generated during a period of near equilibrium balance in the early to mid 1940's and apparently was not a response to the rock avalanche. A kinematic wave generated by the balance change at the leading edge of the debris may not have arrived at the terminus until 1972.
Arrival of this wave should initiate an advance of about 630 m that will take \( \sim 50 \) years. By then, the debris-covered ice will have separated from Sherman Glacier. The advance will form a significant terminal moraine that will be protected from erosion by a coating of very large boulders from the avalanche. Lateral moraines are also developing, but the upglacier ends of these moraines formed in 1964, and the terminal ends may not form for another 50 years, so that the lateral moraine will not define a synchronous glacier profile. The evidence for this probably will not be preserved in a readily recoverable form because of a near lack of modern wood that could be incorporated in a moraine below a late-19th-century trim-line, and because of an abundance of ancient wood beside the glacier from a glacial advance that destroyed a mature spruce forest perhaps 2,000 years ago, which will be incorporated in the moraine.

It is probable that the only clear evidence of the avalanche origin of this future advance will be found on the crest of a spur overtopped by the avalanche, but this is so isolated from the future glacial geology that it probably would be overlooked in a glacial geological study.

**Suggestions for future work**

**Rock avalanching**

In this study I have paid particular attention to aspects of the avalanche deposit that record physical properties of the avalanche itself, such as basal friction, internal strain, velocity, shear
strength, porosity and permeability. Past studies of rock avalanches have tended to ignore these and have concentrated on simple descriptions of the deposits and mechanisms of transport, with little apparent attempt to connect the two, with the consequent development of a mechanism of transport that is invalid.

Further study of the process of rock avalanching should continue in the direction of evaluating physical properties of avalanching media, and in particular, to studying the variation of properties during the avalanche. There is a need for early observation of detail before weathering, such as by rainwash, obscures detail. Thickness and surface slope at a point reveal the basal shear strength of the avalanche as it came to rest, but no systematic study of variation in thickness and surface slope has been made. Internal and basal fabric reveal the mode of deformation by rolling and sliding, but too few observations have been made to develop a quantitative assessment of the role of either. A theory of the role of permeability in rock avalanches has been developed by Shreve (1968c) to investigate the permeability necessary to maintain an air layer, and a new theory that considers the role of permeability in internal strain should be investigated. If the parameters controlling the yield strength of the debris could be determined then the area of devastation of a potential rock avalanche could be simply determined from its potential volume, because the strength of the debris controls the thickness of the deposit.
Ablation beneath debris

Study of ablation of ice beneath debris has been hampered by a lack of direct measurements over a range of different debris thicknesses and compositions. Probably the simplest measurements could be made by mechanically drilling through debris covers and placing ablation poles deep into the underlying ice to provide a stable reference. Accurate knowledge of thermal conductivity and spatial and temporal variations in thermal gradient are also needed.

Very few studies have more than mentioned the possible role of rainfall in ablation of ice. At Sherman Glacier, as at other maritime glaciers, a very high summer rainfall makes this a probably important role.

Glaciers and climatic change

Sherman Glacier is unsuited to precise areal mass balance studies so that further balance study there can not be recommended. Reliable long records of balance can, however, be estimated from point measurements of balance and climatological records, provided all important parameters contributing to balance are considered. All published long records from stratigraphy need careful reevaluation, but this is impossible for many of the records because most authors fail to provide an account of the stratigraphy from which an independent assessment can be made. One problem in developing an accurate record of balance from stratigraphy or climatology is the recognition of accuracy. Correlation with several or more independent records is the best criterion for this.
The problem of missing years in stratigraphic records is not easily overcome, but with time and the development of many records, it may be possible to recognize where years of negative balance have occurred in any one record - this time is not near at hand. Correlation of stratigraphic records with records of directly observed annual phenomena provides a means of assessing "missing years" but some circular reasoning begins to enter when a correlation is forced by adding years of zero balance.

Orheim's (1972a and b) record from Deception Island appears to have the potential of providing an exceptionally valuable record of balance, because a clear stratigraphy has been published and can be independently assessed. Because my interpretation negates all of Orheim's interpretations, correlations, and conclusions, there is an urgent need for another independent assessment of this record, or alternatively, a need for an independent assessment of ages of many parts of the stratigraphy, and of the correlation of his fissure and crater stratigraphies.

Response of Sherman Glacier to the rock avalanche

There can be little particular value in following the future response of Sherman Glacier for its own sake. Other glaciers have responded to similar events in the past, and their response is already known. Continuing studies, such as Austin Post's annual aerial photography of the glacier that was initiated before the avalanche, are an excellent means of monitoring glacier behavior and should be continued.
As this will automatically record the response of the glacier, further detailed study should be deferred for about 40 years.

**General glaciology of Sherman Glacier**

Two aspects of the debris cover on Sherman Glacier can be of great value to glaciology: 1) it has enabled a part of the glacier to maintain a constant profile over a prolonged period, a unique situation for which most glaciological theory has been developed, but which rarely occurs in nature; and 2) it contains many unique objects that can be recognized and followed from year to year on aerial photographs. The surface strain-rate field of the debris-covered portion of Sherman Glacier from 1964 to 1973 is recorded with great accuracy as the changing positions of boulder images from year to year on the aerial photographs, and this can be resolved with some precision by photogrammetry utilizing known ground control.

A pilot study to determine the feasibility of determining surface strain-rate from existing photographs of Sherman Glacier has been initiated by Dr. Peter J. Morgan and myself but the results are not yet available. My analysis of results from two surface strain-nets suggest that a map of surface strain and its variation in time will be of considerable value in assessing the roles of ice fabric, basal topography, kinematic waves, and subglacial channels in flow of a real glacier.
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