TILL DEFORMATION BENEATH BLACK RAPIDS GLACIER, ALASKA, AND ITS IMPLICATION ON GLACIER MOTION

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By

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By

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Abstract

The motion of a glacier is largely determined by the nature of its bed. The basal morphology and its reaction to the overlying ice mass have been subject to much speculation. because the glacier bed is usually difficult to access, and good field data are sparse.

In spring 1997 a commercial wireline drill rig was set up on Black Rapids Glacier, Alaska, to extract cores of basal ice, subglacial till, and underlying bedrock. One of the boreholes was equipped with three tiltmeters to monitor till deformation, and a piezometer to record pore water pressure. The surface velocity and ice deformation in a borehole were also measured.

The drill successfully reached bedrock twice after penetrating a till layer, some 5 to 7 m in thickness, confirming an earlier seismic interpretation. The till consisted of a sandy matrix containing clasts up to boulder size. Bedrock and till lithology indicated that all the drill holes were located to the north of the Denali Fault, a major tectonic boundary along which the glacier flows.

The mean annual surface velocity of the glacier was 60 $\mathrm{m\,a^{-1}}$, of which 20 to 30 $\mathrm{m\,a^{-1}}$ were ice deformation, leaving 30 to 40 $\mathrm{m\,a^{-1}}$ of basal motion. The majority of this basal motion occurred at a depth of more than 2 m in the till, contradicting previously held ideas about till deformation. Basal motion could occur as sliding of till over the underlying bedrock, or on a series of shear layers within the till. This finding has implications for the interpretation of the geologic record of former ice sheets, for geomorphology, and for glacier dynamics.

The effect of a thick till layer on ice flow and on quantities observable at the glacier surface was calculated. These include velocity changes on secular, seasonal, and shorter

time scales. A mechanism for uplift events and dye tracing responses was suggested. An easy surface observation that could serve to clearly distinguish a glacier underlain by till from the more traditional view of a glacier underlain by bedrock could not be identified.

Contents

List of Figures . . . viii
List of Tables . . . x
Acknowledgements . . . xi

1 **Introduction** **1**

2 **Subglacial Drilling at Black Rapids Glacier, Alaska, USA: Drilling Method and Sample Descriptions** **2**

2.1 INTRODUCTION . . . 3
2.1.1 Black Rapids Glacier . . . 3
2.1.2 Geologic setting . . . 5
2.1.3 Drill site location . . . 6
2.1.4 Sampling of subglacial till . . . 7
2.2 METHODS . . . 8
2.2.1 Drilling . . . 8
2.2.2 Logistics . . . 12
2.3 THE DRILL HOLES . . . 12
2.4 RESULTS . . . 14
2.4.1 Basal ice . . . 14
2.4.2 Subglacial till . . . 18
2.4.3 Bedrock . . . 21
2.5 DISCUSSION . . . 22
2.5.1 Ice content of the till samples . . . 22
2.5.2 Stability of an ice-water-till conglomerate . . . 23

2.5.3 Formation of ice in the till 23
2.5.4 Erosional and depositional processes 25
2.5.5 Debris layers in basal ice 27
2.5.6 Implications for surge behavior 27
2.6 SUMMARY 28
2.7 ACKNOWLEDGEMENTS 29

3 Glacier motion dominated by processes deep in underlying till 30
3.1 INTRODUCTION 31
3.2 SETTING 32
3.3 METHODS 33
3.3.1 Surface motion and deformation of the ice 33
3.3.2 Measurements within the till 34
3.4 RESULTS 36
3.4.1 Surface displacement and ice deformation 36
3.4.2 Measurements in the till 38
3.4.3 Engineering properties of a till sample 41
3.5 DISCUSSION 42
3.5.1 Alternative interpretations of the data 42
3.5.2 Short term fluctuations 43
3.5.3 Motion deep in a till layer: other observations 46
3.5.4 Is till deformation at a depth of over 2 m physically plausible? . . . 47
3.5.5 Erosion and till mass balance 49
3.5.6 Implications for the surge behavior 49
3.6 CONCLUSION 50
3.7 ACKNOWLEDGEMENTS 52

4 Implications of till deformation on glacier dynamics 53
4.1 INTRODUCTION 54
4.1.1 Motivation 54
4.1.2 Previous work 54
4.1.3 Black Rapids Glacier 55

4.2 FLOW THROUGH A GLACIER CROSS SECTION 55
4.2.1 Boundary conditions 56
4.2.2 The finite element method 59
4.2.3 Calculated flow through a cross section 60
4.2.4 Sensitivity 62
4.3 VELOCITY VARIATIONS 64
4.3.1 Secular variations 65
4.3.2 Seasonal variations 67
4.3.3 Speed-up events 69
4.4 TILL VOLUME CHANGES 70
4.4.1 Uplift events 71
4.4.2 Dye tracing 74
4.5 DISCUSSION 74
4.5.1 Transfer of basal shear stress 74
4.5.2 Till versus bedrock 77
4.6 CONCLUSION 78
4.7 ACKNOWLEDGEMENTS 79

Bibliography **80**

List of Figures

2.1 Outline of Black Rapids Glacier 4
2.2 Transverse profile with borehole locations 6
2.3 Longitudinal glacier profile 7
2.4 Picture of the drill rig 10
2.5 Schematic drawing of the core barrel 11
2.6 Drill log of N2 hole 13
2.7 Drill log of N1 hole 15
2.8 Drill log of N1A hole 16
2.9 Drill log of Center hole 17
2.10 Grain size analysis of recovered till samples 19

3.1 Postion of the instruments 35
3.2 Dual axis tilt measurement 36
3.3 Velocity, water pressure and tilt 37
3.4 Total tilt and water pressure 39
3.5 Stereo plots of tilt 40
3.6 Inferred distribution of glacier motion 40
3.7 Postion of instruments if sliding is dominant 44
3.8 Modeled pore pressure and tilt 45
3.9 Maximum effective pressure as a function of depth 48

4.1 Water pressure 58
4.2 Cross section of Black Rapids Glacier used in the FE study 58
4.3 Finite element grid 60

4.4 Velocity contour plot 61
4.5 Calculated ice deformation at the N1 location 61
4.6 Transverse velocity profile 63
4.7 Measured scaled transverse velocity profiles 63
4.8 Secular velocity changes 65
4.9 Velocity vs. depth of water table 66
4.10 Calculated basal motion 67
4.11 Lake drainage event 69
4.12 Expansion of the till layer 72
4.13 Spring speed-up 73
4.14 Diagram of stress re-distribution 75
4.15 Stress re-distribution 76

List of Tables

3.1 Engineering properties of a till sample from borehole N1A 42

4.1 Sensitivity of modeled velocity 64

Acknowledgements

It has been stated that the acknowledgements are the most-read part of any thesis (Heavner, 1999). I will therefore avoid being brief. Some of the acknowledgements specific to one part of the thesis can be found at the end of each chapter/paper.

Herr Professor Doktor Obersturmbannführer William D. Harrison, my thesis advisor, is no newcomer to the field of glaciology. I greatly benefited from his experience as a field worker and his careful and precise way of analyzing data. And he also taught me to never split an infinitive. Keith Echelmeyer has made many good suggestions to improve this work. His airplane and his piloting skills have often been very handy. I have enjoyed a lot of discussions with him about a variety of subjects in these past years. Climbing mountains and skiing was great fun too. My other two committee members, Craig Lingle and Lou Shapiro, made many useful comments and suggestions. Roman Motyka was the main link to the world of geology, and he provided me with a wonderful opportunity for a boat trip in Southeast Alaska.

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Obwohl däheimu wyt äwäg isch, isch mär das niä ä so vorcho. Telefon, email und Flugzyg machund d'Verbindigä hitzutag churz. Äs härzlichs Merci an mini Eltru, Schweschträ (mit Nachwux), und an alli Kollegä wa mi no nit vergässu hent. Merci vill mal öi ani Bsüächär.

And last but not least: Thank you Alaska for the great mountains, glaciers, and aurora!

Chapter 1

Introduction

In spring 1997 a wire line drill rig was set up on Black Rapids Glacier in the central Alaska Range. There were two major goals for this drilling effort: to sample subglacial material, including basal ice, subglacial till, and underlying bedrock ('see what's there') and to instrument the boreholes, so the distribution of glacier motion could be described. These are the subjects of chapters 2 and 3, respectively. Chapter 4 is a modeling study that explores the consequences of having till beneath the glacier and checks whether our understanding of till physics is sufficient to explain various surface observations.

Each of these three parts is written in the form of a paper. Chapters 2 and 3 have been accepted for publication in the Journal of Glaciology. Chapter 4 is in preperation to be submitted to the same journal. This format explains some of the peculiarities: Each chapter contains its own abstract and acknowledgements. The bibliography, however, has been combined at the end of the thesis. Because the chapters are written as stand-alone papers, some repetition is inevitable, especially in the introductory part of each chapter. References across chapters are made as Truffer and others, 1999 (Chapter 2) and Truffer and others, in press (Chapter 3). The contributing authors are listed in the beginning of each chapter. The formatting of the papers was changed to be in accordance with UAF thesis requirements.

Chapter 2

Subglacial Drilling at Black Rapids Glacier, Alaska, USA: Drilling Method and Sample Descriptions

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ABSTRACT[1]. We employed a commercial wireline drill rig to investigate the subglacial conditions of Black Rapids Glacier, a well studied surge-type glacier in the central Alaska Range. The four main goals were: to assess the capabilities of the commercial drilling industry for sampling subglacial material, to investigate the basal morphology, to determine the subglacial geology, and to emplace borehole instruments. The drilling was done in an area where seasonal and secular variations in speed are large, and where seismic studies

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suggested the presence of a till layer. Four holes were drilled at three locations to a maximum depth of 620 m. Three holes yielded samples of basal ice and till, although recovery of the latter was generally poor. Bedrock was sampled in one or possibly two of the holes. In the area sampled the glacier is underlain by a till layer some 4 to 7 m in thickness, confirming the seismic interpretation. It consists of a sandy matrix containing a sizable volume, at least 20 - 30 %, of larger clasts. Limited samples of the matrix indicate that near the top of the till its porosity is 40%, and that some of the pore water is frozen. Geologic studies suggest that the drilling area lies to the north of the Denali Fault, a major tectonic boundary followed by the glacier, and that most of the till is locally derived.

2.1 INTRODUCTION

2.1.1 Black Rapids Glacier

Black Rapids Glacier is a 40 km long surge-type glacier in the central Alaska Range. It first caught public attention during its last surge in 1936/37, when the surge front was observed from the Black Rapids Roadhouse, situated along the Valdez to Fairbanks trail (Fig. 2.1). There the astounded residents radioed the news of the approaching glacier to the world (Hance, 1937). An attempt to document the previous surge history has been made by Reger and others (1993).

The surge behavior and the accessibility of Black Rapids Glacier have made it an object of intensive scientific study. In 1970 the U.S. Geological Survey initiated a surveying program, measuring annual velocities and mass balance. This program was later continued by University of Alaska Fairbanks (Heinrichs and others, 1996). Harrison and others (1975) measured near surface temperatures and showed that Black Rapids is a temperate glacier with a cold, about 20 m thick, surface layer in the ablation area. Starting in 1982 cameras were placed at up to four locations along the glacier to measure the velocity of survey poles. In addition, vertical strain and passive seismicity were recorded. In 1993, a more detailed program was started, with radio echo soundings (Gades, 1998), hydrological studies (Raymond and others, 1995; Cochran, 1995) and seismic detection of subglacial changes (Nolan and Echelmeyer, 1999a and b). In 1996 we hot water drilled seven holes through the ice to obtain accurate ice depths and to prepare for the subglacial wireline drilling carried out in

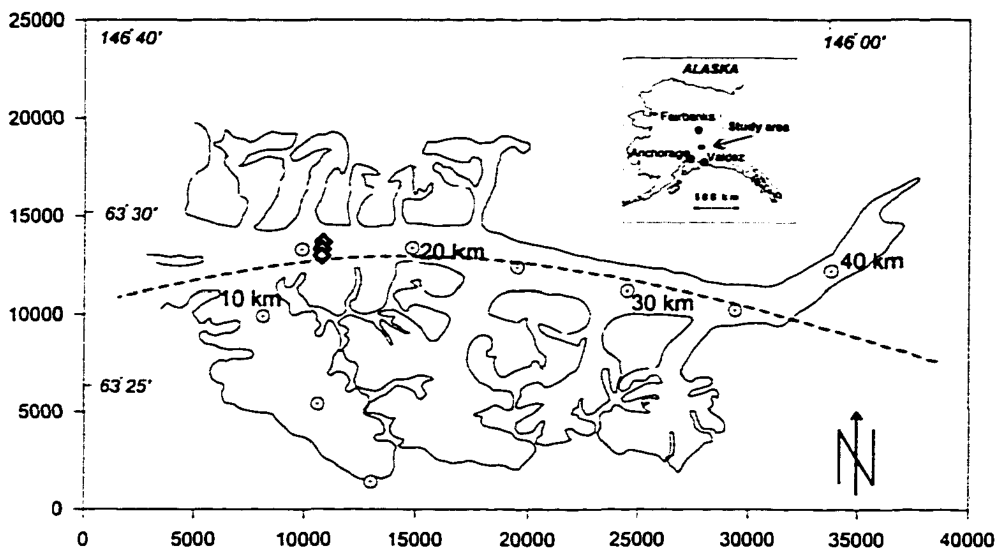

Figure 2.1: Outline of Black Rapids Glacier. The diamonds show, from top down, the location of holes N2, N1, and Center. The open circles are located every 5 km along the centerline from the head of the glacier and a local coordinate system is shown (in meters). The approximate location of the Denali Fault is indicated by the dashed line. The glacier is situated just east of the Richardson Highway and the Trans Alaska Pipeline.

spring 1997. The waterlevels in these boreholes were recorded during the summers of 1996 and 1997.

2.1.2 Geologic setting

Black Rapids Glacier lies in the central Alaska Range, which is characterized by glacierized peaks of over 4,000 m in elevation. The range lies along the Denali Fault, a major tectonic boundary that extends from southeast Alaska through the Alaska Range to the Bering Sea. The fault underlies Black Rapids Glacier in the vicinity of the drilling site. Plafker and others (1994) consider the fault to be active. Probable rates of dextral Holocene movement average about 15 $\mathrm{mm\,a^{-1}}$. Estimates of Quaternary offset on the fault range from 1 to 6.5 km, with an estimated total displacement of 300 to 400 km since Late Cretaceous (Nokleberg and others. 1994). In the vicinity of Black Rapids Glacier, the fault separates rocks belonging to the 'Aurora Peak' terrane on the north from the 'MacLaren' terrane to the south (Nokleberg and others. 1992). The former include Cretaceous gabbros, metagabbros, metadiorites, amphibolites, and Silurian schists. The MacLaren terrane is composed of Late-Cretaceous gneissic granitic rocks, chiefly granodiorite, quartz diorite, and minor granite of the 'East Susitna Batholith', and of Late-Cretaceous and older schists and amphibolites. The glacier's accumulation area is situated in the 'MacLaren' terrane from where it flows north and then curves to the east into a valley defined by the fault. The main part of the glacier starts flowing into the fault valley 14 km from its head. and some 2 km upstream from the drill site (Fig. 2.1). A major tributary enters at about 23 km, which is below the drill site.

We attribute significant importance to the geologic setting of the glacier because we believe that it is somehow related to the surge behavior. Post (1969) recognized that surging glaciers in Alaska are not randomly distributed, and that most of the surging glaciers in the Alaska Range directly overlie the Denali Fault (his Figure 1). Wilbur (1988) did not find a one-to-one correlation between glacier geometry and surge behavior and concluded that surge behavior was also related to regional mountain geometry and ultimately to geologic conditions. A similar conclusion was reached in a study on glaciers in Svalbard (Hamilton and Dowdeswell, 1996). Work at Findelengletscher, Switzerland, has also shown a strikingly different velocity behavior on each side of the glacier (Iken and Truffer, 1997), which may relate to differences in bedrock geology (Bearth, 1953).

2.1.3 Drill site location

The drilling was done in the area between 16 and 17 km from the glacier head (Fig. 2.1). This is where the seasonal and longer term variations in speed are largest, and basal motion is thought to be large (Heinrichs and others, 1996). It is about 4 to 5 km downstream of the present day equilibrium line, and lies downglacier from marginal and supraglacial lakes (potholes) that frequently drain in early summer (Sturm and Cosgrove, 1990) and cause speedup events. The glacier undergoes a spring speed up, typically in early June. Nolan and Echelmeyer (1999 a and b) inferred the existence of a subglacial till layer — at least 5 m in thickness — from a seismic study. This layer became seismically transparent during speed up events. The subglacial seismic changes were observed at the N1 site (Fig. 2.2), where two of the boreholes were drilled. Figure 2.3 shows a longitudinal radio echo sounding profile (Gades, 1998 and unpublished data). The boreholes were drilled in a subglacial hollow, directly upstream of a riegel. The ice thickness is decreasing and the glacier is slowing down between 14 km and 20 km (Heinrichs and others, 1996).

It should be kept in mind that all the results presented here describe a relatively small part of the glacier, albeit an interesting one.

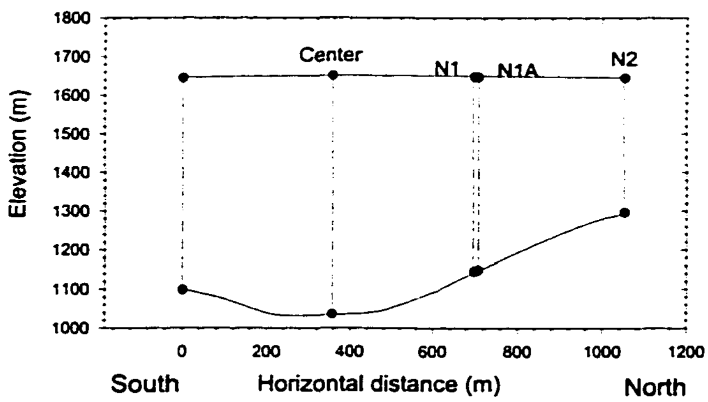

Figure 2.2: Transverse profile with borehole locations. The southern-most hole is a hot water hole where no subglacial samples were taken. The ordinate axis shows elevation above sea level.

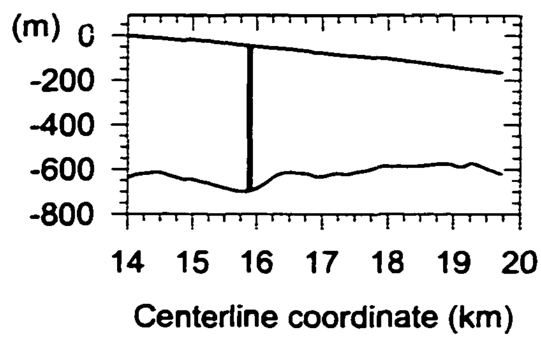

Figure 2.3: Longitudinal glacier profile. Profile of ice surface and glacier bed from the 14 km to the 20 km index site (Fig. 2.1). The vertical line shows the location of the boreholes. Data are from Gades (1998 and unpublished).

2.1.4 Sampling of subglacial till

Basal glacial processes are important to understand glacier dynamics. This has led several investigators to take a closer look at subglacial sediments, sampling them and making in-situ measurements. For the remainder of this paper we will refer to any unlithified subglacial material as till (Paterson, 1994). Early observations of subglacial till were made at Blue Glacier by Harrison and Kamb (1973) and Engelhardt and others (1978). Samples of fine grained till with a small clast content have successfully been retrieved from beneath Ice Stream B with a piston corer (Engelhardt and others, 1990; Kamb, 1991). Boulton and Hindmarsh (1987) studied the till beneath Breidamerkurjökull through a tunnel at the glacier terminus. Extensive borehole studies of in-situ subglacial till properties have been undertaken on Trapridge Glacier (e.g. Blake and others 1992, 1994; Fischer and Clarke, 1994), Storglaciären (e.g. Iverson and others, 1994; Hooke and others, 1997) and Columbia Glacier (Humphrey and others, 1993).

With simple borehole instruments, it is only possible to penetrate subglacial till to a depth of about 40 cm or less. It is often difficult to assess the exact depth of an instrument in relation to the ice-till interface — assuming a sharp interface exists — because the hot water drill stirs up the till and can actually drill through fine grained tills. On the other hand, debris rich ice can stop the hot water drill above the actual glacier bed.

It is extremely difficult to sample subglacial till. Unlike the marine sediments underlying the West Antarctic ice streams, most glacial till contains numerous clasts of many sizes. Such clasts will invariably stop or jam piston corers. Sampling attempts have thus often been met with only little success. A split spoon sampler was used on Storglaciären yielding disturbed samples (Iverson, pers. comm.). Blake and Clarke (1991) describe a sediment sampler that works by sucking sand and pebbles up into the sampler that was used on Trapridge Glacier. This sampler is size-selective, however, and does not yield representative samples (Clarke, pers. comm.).

The idea behind our project was to exploit existing drilling technology in an attempt to retrieve complete core samples of basal ice, subglacial till and the underlying bedrock. This would allow us to study the basal morphology and the bedrock geology. At the same time we hoped to use some of the boreholes to install instruments for in-situ measurements of borehole tilt and pore water pressure. We feel that obtaining undisturbed samples of subglacial till, although very difficult to do, is of primary importance. Although there is a large research effort in modeling till deformation, there are only a limited number of in-situ measurements and actual samples.

This paper describes the drilling technology used and an analysis of the recovered samples. The results from the borehole instruments will be published separately.

2.2 METHODS

2.2.1 Drilling

Drilling into the basal ice, till and bedrock, and the sampling, were done with a commercially contracted wireline drill system. The size of the drill rig was determined by the diameter of the core sample needed and the portability of the drill rig. A Longyear Super 38 (Fig. 2.4) was selected allowing us to drill to at least 650 m and obtain a 60 mm diameter core. This drill is helicopter portable, breaking down into 5 pieces with the largest piece weighing approximately 450 kg. Drill rod of HQ size (78 mm inner diameter) was used in the shallower holes and NQ size (60 mm inner diameter) in the deepest hole. This was done for safety reasons because an HQ string of more than 600 m was deemed too heavy for the drill rig; 6866 kg for HQ, as opposed to 4544 kg for NQ. In some cases the narrower drill rod can be used inside the larger, with the larger as casing for the hole. Sections of drill rod, each 10 ft. (3.05 m) long, were added as drilling proceeded.

The bottom section of the drill string is called a core barrel (Fig. 2.5). It carries a shoe with the drill bit at its end. It contains an inner tube that can be pulled out through the entire rod by a wire cable; hence the name wireline drilling. The inner tube is equipped with a core retainer. This can be either a spring retainer that works through wedging action or a plastic basket, depending on the expected sample material. The advantage of this kind of drill is that the inner tube with the core can be retrieved through the drill rod. A sample of ten feet (3.05 m) or less can thus be pulled out without taking the entire drill string out of the hole. The drill rod effectively cases the hole during sampling. The core barrel is equipped with a latching mechanism for the inner tube. Failure of this latching mechanism was a chronic problem that we were unable to diagnose. Sometimes the problem could be fixed by raising the entire drill string a few meters and rotating it rapidly. This is not an ideal solution, because it allows the lowest part of the borehole to cave in and necessitates some redrilling.

Two drill bits were used: a carbide bit to drill through ice and soft till and an impregnated diamond core bit to drill through till and bedrock. The carbide bit wears out quickly when used in hard rock, and drilling through ice with the diamond bit was as slow as 0.5 $\mathrm{m\,h^{-1}}$, as opposed to 2-5 $\mathrm{m\,h^{-1}}$ with the carbide bit. A surface-set diamond core bit might be a better choice for drilling in ice. Typical drilling speeds in till were about 2 $\mathrm{m\,h^{-1}}$ and in bedrock $\sim 1 - 1.5 \mathrm{m\,h^{-1}}$. Pulling out the core barrel through 600 m of drill string takes about half an hour, provided everything goes well.

Ideally, softer material is sampled with a punch corer. Some core barrel designs combine a retractable punch corer with a drill bit. This makes it possible to push the punch corer ahead through softer sediments with the bit engaging any harder material. In our case, this setup was not often used because the subglacial till contained cobbles of the size of the diameter of the punch corer, and these could easily damage or plug the corer and, hence, require pulling the entire drill string.

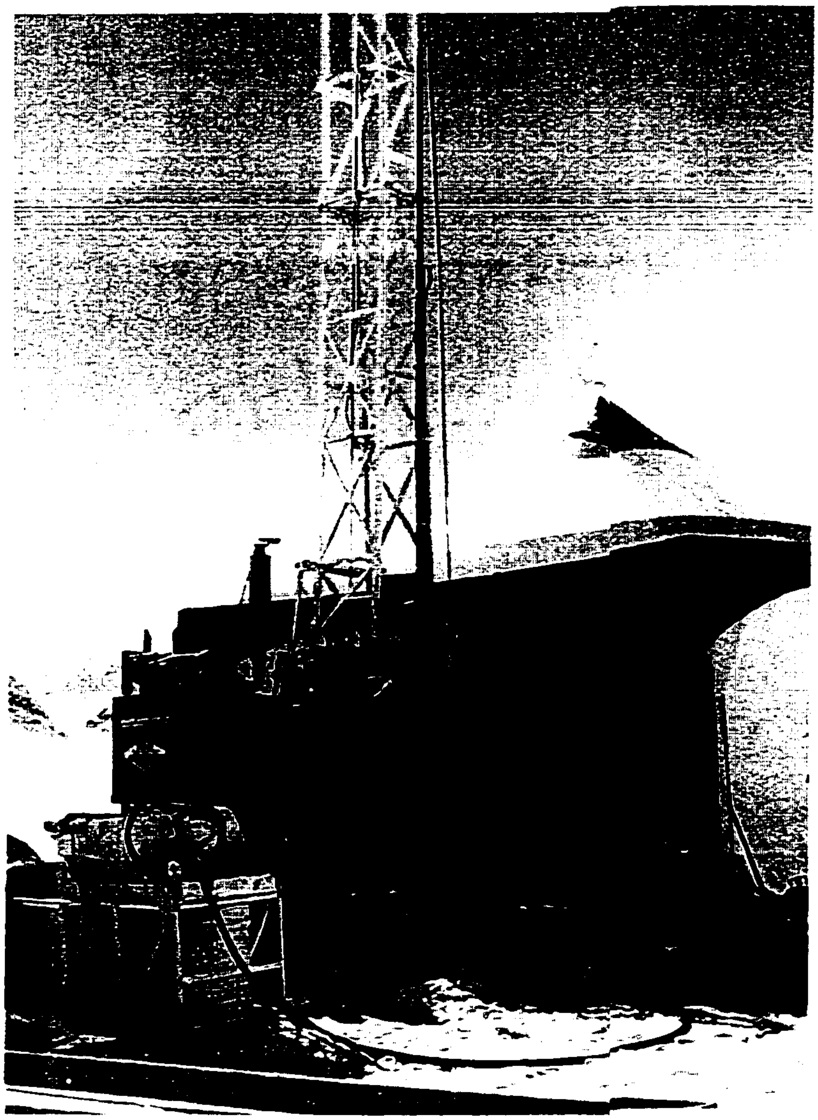

Figure 2.4: Picture of the drill rig. As drilling proceeds, ten foot sections of drill rod are lifted up on the tower and then connected to the drill string.

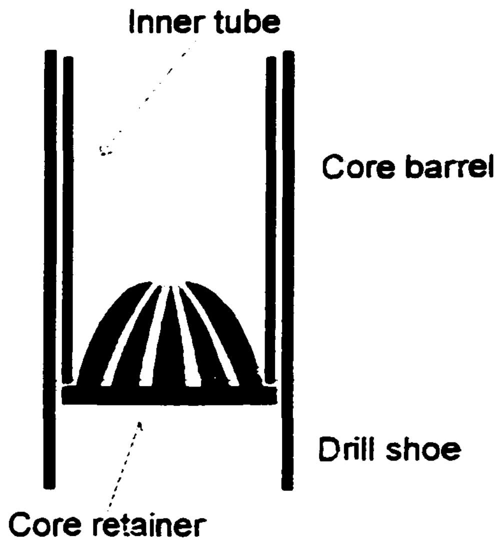

Figure 2.5: Schematic drawing of the core barrel. It is the lowermost section of the drill string. The drill bit is attached to the core barrel's bottom. The inner tube is retrievable through the drill string. It carries a core retainer at the bottom. A plastic basket core retainer that was used to sample till material is shown.

Because we were only interested in basal material, we predrilled holes into the ice using a conventional hot water drill (e.g. Iken and others, 1988), which is much faster and cheaper. The water was recovered from the borehole for reuse using a submersible pump. The rotary drill also required recovered water for cooling and flushing the drill bit. Ice depths were known from seismic and radar measurements and hot water drilling done the previous year. In some holes we stopped the hot water drill a few meters short of the bed, so that basal ice could be sampled as well. The actual ice-till interface was within 2 m of what we expected from hot water drilling at N1 and Center. At N2 we expected to reach the bed at 341 m, but we had not reached it when we abandoned the hole at 348 m.

During normal operation the drill rod cases the hole and keeps till from collapsing and closing the hole. If the drill string has to be pulled out, a special polymer ('drill mud') is pumped into the hole before pulling. This helps prevent the hole from collapsing for a short period of time. For reasons not entirely clear to us, a smaller concentration of the polymer was also used during normal drilling operation. If the hole stays open after removal of the drill string, borehole instruments may be emplaced once the drill string is removed. Ideally, instruments could be introduced through the drill string, which would then be pulled out over the instrument cable. Unfortunately our instruments were too large to fit through the drill bit. We installed a string of three tiltmeters and one piezometer at the base of one of the boreholes.

2.2.2 Logistics

A drilling operation of this kind adds a whole new dimension to field work on valley glaciers. Including camp and fuel, a total of over 40,000 kg of material was flown to the glacier. This was done by a combination of helicopter and fixed-wing aircraft. The fixed-wing aircraft was much less expensive, but loading and unloading was more efficient with a helicopter and the helicopter was also needed to assemble the drill rig on the glacier.

When assembled the drill rig was mounted on a large steel toboggan to distribute its weight. It could be moved on the glacier using large snow anchors ('dead men') dug into the snow. The rig was then pulled using its own wireline winch system.

Our drilling was performed in late April and early May, before any significant melting occurred, thus avoiding most problems with transport on a wet snow surface. Also, we expected the drilling to be easier before any accelerated basal motion occurred.

Due to the relatively high standby cost of this drill rig, it proved advantageous to work in two twelve hour shifts. Special arrangements were needed to keep the crew out of the wind and the low temperatures, which at that time of the year can still reach -30°C. Directly adjacent to the drill rig a large heated tent was set up, allowing most of the work to be done in somewhat less harsh conditions (Fig. 2.4).

2.3 THE DRILL HOLES

Four holes were drilled at three sites on a transverse section: N2, N1 and Center (Fig. 2.2). At N2, wireline drilling commenced at a depth of 336 m and continued through 12 m of ice (Fig. 2.6) before the hole was abandoned because of problems with moving the core barrel up and down the string. Subsequent inclinometry showed that the hot water drilled hole was over 4° out of plumb at some places. We suspect that this was caused by debris rich ice or by rocks on the surface that fell into the hole. The proximity of a medial moraine and the slow hot water drilling rate lend support to this inference.

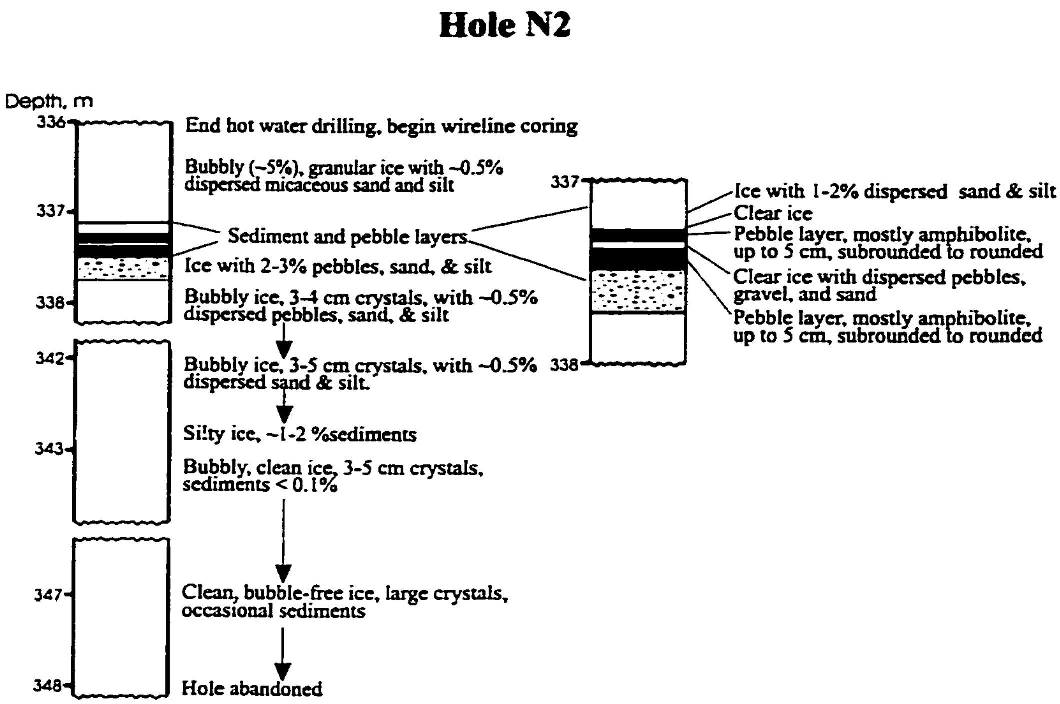

Figure 2.6: Drill log of N2 hole. A combination of radio echo sounding and seismic measurements yielded a depth of 360 m.

Wireline core drilling at N1 commenced at 488.5 m and eventually reached 510.1 m in depth (Fig. 2.7). Hole N1 provides the most complete picture of the ice-till-bedrock structure of the four holes. Drilling through the till presented major challenges. Equipment failure and other problems required pulling of the entire drill string three times before N1 was completed, a tedious and labour-intensive undertaking. Each time the drill string was removed, till collapsed into the uncased hole, requiring redrilling. Because of this we cased the hole with the wider HQ rod and drilled inside it with NQ rod. In this manner we eventually reached bedrock or a large boulder. The hole had to be abandoned when the lowest 6 m of the string was twisted off. This could have been caused by borehole deformation and/or basal motion.

A second hole at this site, N1A, was drilled to obtain till samples and to place a string of three tiltmeters and one piezometer into the till layer. To accelerate the drilling process and avoid time consuming wireline coring through ice, hot water drilling was used to penetrate all the way to the ice-till interface, located at ~498.5 m (Fig. 2.8). Thus no ice samples were obtained but till was successfully recovered in the first 3 m section of core below the interface. Instruments were installed successfully in the till after the drill string was pulled.

The Center hole was the deepest (about 610 m, Fig. 2.9). NQ rod was used to reduce the weight of the drill string. Wireline coring of ice began at 602.0 m but was abandoned in favor of hot water drilling at 607.5 m. The hot water drill was lowered inside the drill rod. thus avoiding pulling of the entire drill string. Coring recommenced at 614.8 m and continued to 621.2 m after penetrating bedrock at 619.5 m.

2.4 RESULTS

2.4.1 Basal ice

Ice samples were obtained from near the bottom of three of four holes drilled. However, only at N1 was ice retrieved from the ice-till interface. The coring process commonly fractured the ice into short centimeter-scale disks.

At N2 wireline drilling stopped several meters above the bottom. At 337–338 m, coring intersected two distinctive, several centimeter thick layers of subrounded to rounded pebbles, primarily black amphibolite (Fig. 2.6). The pebble layers contained no ice when recovered

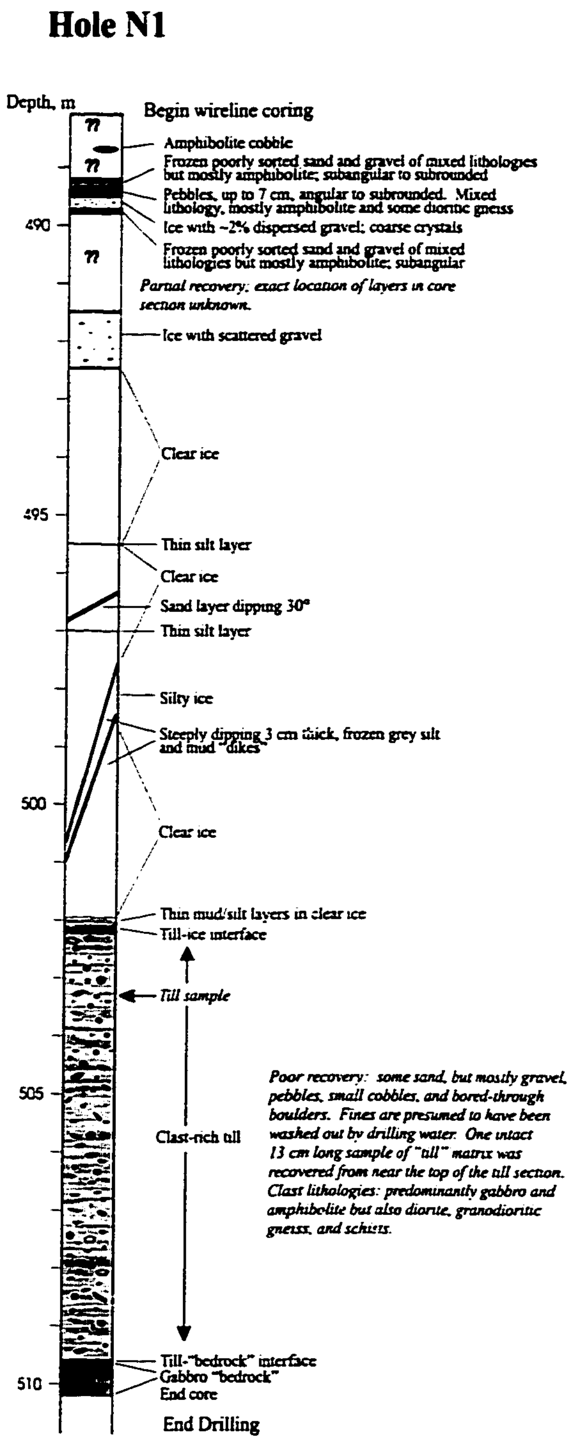

Figure 2.7: Drill log of N1 hole. A combination of radio echo sounding and seismic measurements yielded a depth of 500 m.

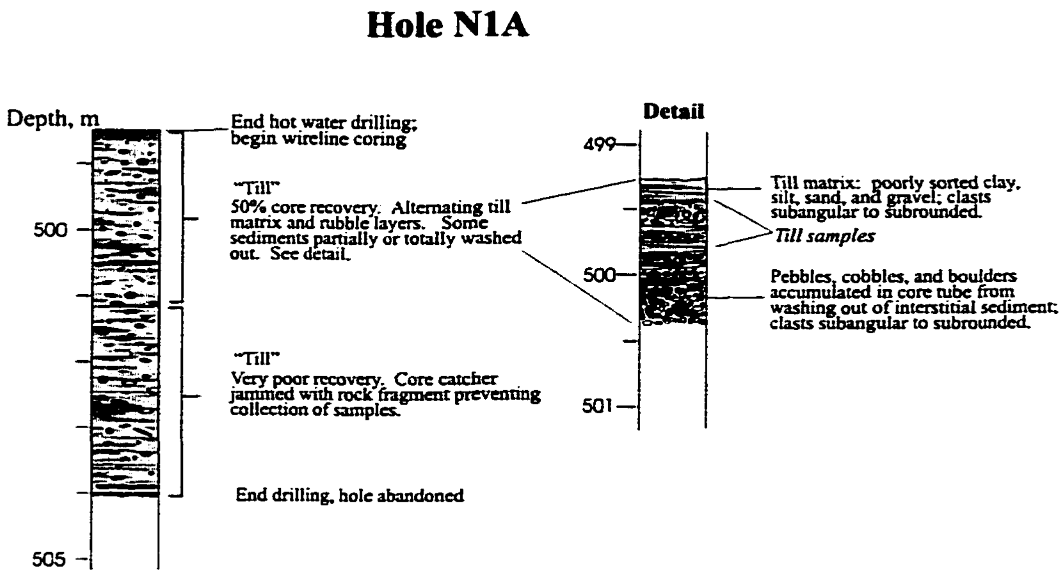

Figure 2.8: Drill log of N1A hole.

and lacked finer grained sediment but were separated by a thin layer of clear ice containing dispersed finer sediments. They were bracketed by ice containing entrained pebbles and finer sediments. The ice generally decreased in bubble content and entrained sediments with depth below the layers and crystal sizes measured from 3 to 5 cm.

The first section of core at N1 was only partially recovered but it contained distinctive layers of frozen sediment and a layer of pebbles (see Fig. 2.7). Sediments finer than coarse sand were absent. Pebbles were more angular than those at N2 and included dioritic gneiss in addition to amphibolite. A thin ice layer separated two of the sediment layers like at N2. No other ice was recovered in this section of the core. It is possible that the sediments and structure are artifacts of the hot water drilling that preceded the wireline coring as it causes most of the sediments it drills to accumulate at the bottom of the hole. Below this section, we recovered a nearly continuous core of ice from ~491.5 m to the till interface at ~502 m (Fig. 2.7), consisting mostly of clear, relatively bubble free ice, with occasional thin silt layers. A 30° dipping sand layer was encountered at 496.5 m. Between 497.5 m and 501 m, two close-set, sub-parallel, steeply dipping silt-mud layers of variable thickness (averaging 3 cm) were encountered. They have the appearance of mud-silt dikes filling ice fractures. Several thin, millimeter scale, mud-silt layers occur just above the till interface.

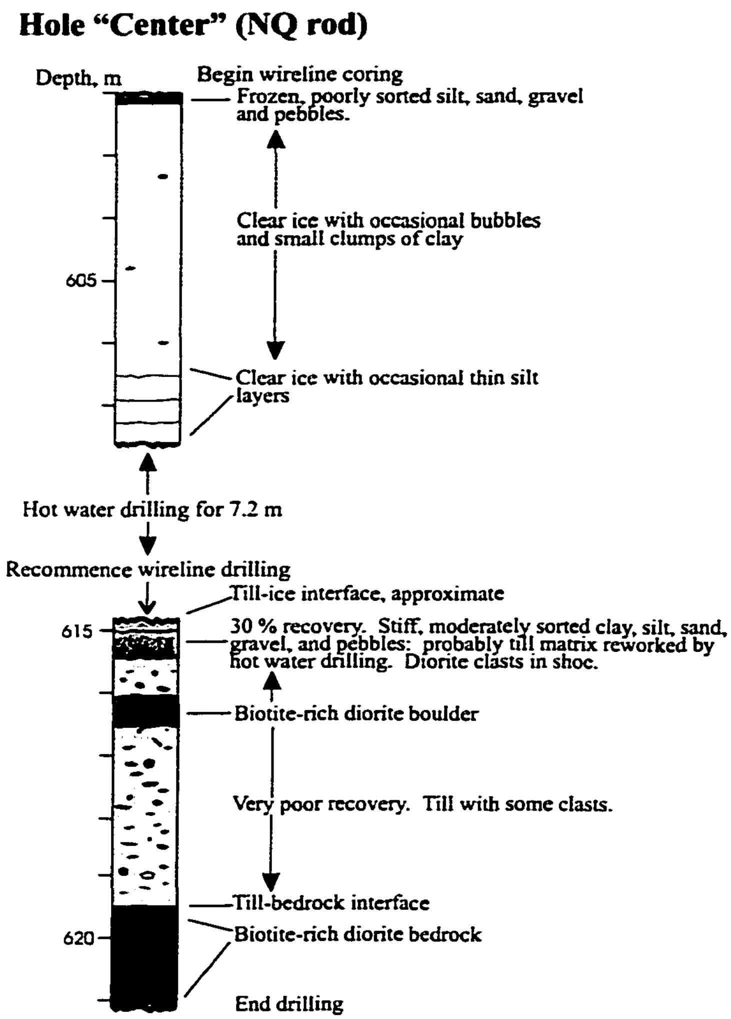

Figure 2.9: Drill log of Center hole. A combination of radio echo sounding and seismic measurements yielded a depth of 620 m.

Analysis showed that over 60% of these layers consist of grains smaller than 44 μm; silts and clays. Details of structure immediately at the interface were not preserved and some sediments and ice have been lost in the drilling process.

About 7.5 m of ice was recovered from the Center hole. A 6 cm layer of frozen, poorly sorted sediments was found near the top of the retrieved sample, a layer which may have stopped the initial hot water drill from progressing (Fig. 2.9). The remainder of the core consists of clear ice with occasional bubbles, clumps of clay, and thin silt layers.

2.4.2 Subglacial till

Till was reached in three of the four holes. Recovery of till proved to be extremely difficult because of two factors. The first factor involved the use of pumped water to cool and flush the drill bit in order to prevent the bit from melting, as did happen on one occasion. The stream of water caused the fine sediments, sand, and even gravel to be washed away from the core barrel, preventing sampling. The other factor involved the clogging of the entrance of the core barrel with clasts and fragments of rocks which then prevented entry of any further material. When using the spring retainer most of the material slipped out, while the basket retainers (Fig. 2.5) tended to get clogged. The problem of washing out can be reduced by using a minimal amount of water to cool the drill bit, and perhaps by using a combination of punch- coring and rotary drilling, but this was not done for the reasons described earlier.

Despite these obstacles, retrieval of small samples of till matrix was achieved on two occasions through judicious drilling. In addition, a variety of rock fragments were recovered, giving some indication of the nature and size of clasts contained in the till. The clast shape classification used in the following sections is entirely subjective. We divided shapes into four categories: rounded, subrounded, subangular and angular clast shapes.

N1 hole

The first sample of intact till matrix was recovered in hole N1. Prior to sampling the drill string was removed from the hole, and then reinserted about 60 hours later. Either material had fallen in from the sides and we recored that or we lost the lower part of the hole and drilled a new one. This is apparently not uncommon. While recoring, we sampled a section of till about 100 ml in volume. It was located within the first 1.0 m below the ice-till interface. The fragment was mechanically coherent when it came out, but it decayed into a slurry after exposure to above melting temperatures for less than two hours. The water content of the sample, being defined to be the ratio of the weight of water to that of dry soil (Lambe and Whitman, 1979), was 21%. The results of a grain size analysis are shown in Figure 2.10a; the sample contained mostly sand and has a similar grain size distribution as other basal tills (e.g. Jiao and others, 1989). Only 6% of the sample is silt and clay as compared to 30% to 40% clays in Ice Stream B samples (Tulaczyk and others, 1998). Grain sizes greater than 2 mm were mostly lithic fragments. The larger grain sizes were subangular to subrounded with some degree of roundness observed in grain sizes as small as 1 mm. Sediment below this size was much more angular and was predominately composed of individual mineral grains.

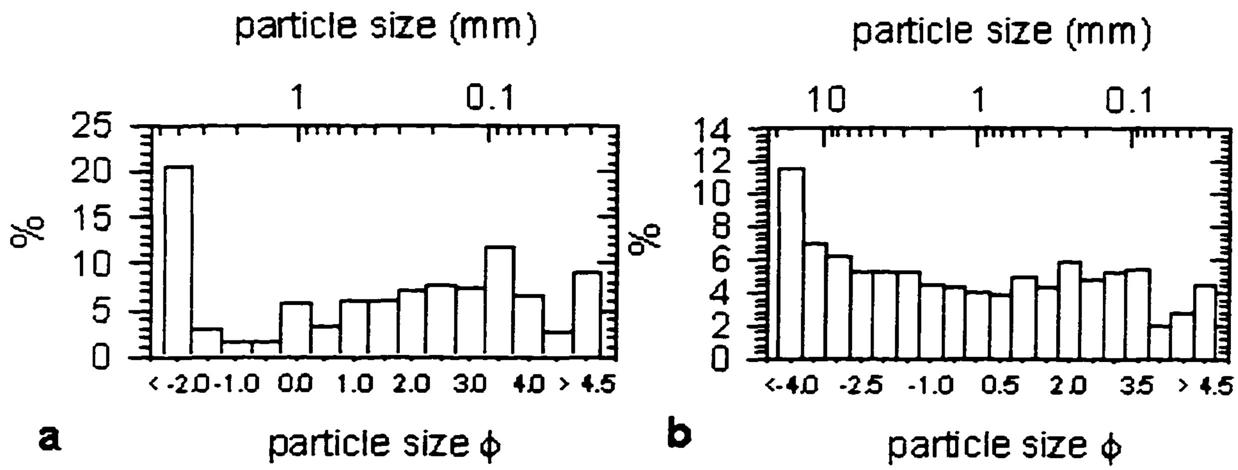

Figure 2.10: Grain size analysis of recovered till samples. They were recovered from the N1 borehole (a) and the N1A borehole (b).

An undisturbed cylindrical sample, 11 cm by 5 cm, was recovered in hole N1A. It was located at the top of the first core below the ice-till interface. Only 1.2 m out of 2.4 m were sampled leaving an uncertainty of 1.2 m for the exact location of the sample. The sample was mechanically coherent when it came out and we stored it at subfreezing temperatures. Its water content was 29%, and the porosity was 40%. When the sample was thawed, it turned into a slurry, much like the N1 sample described above. Laboratory measurements showed that the water content at the Atterberg liquid limit (Bowles, 1992) was 10%, less than half the observed water content. Only 4.4 % of the sample were in the silt and clay fraction (Fig. 2.10b).

Other, smaller, pieces of till were closely monitored when retrieved. They could easily be broken apart. There were some ice crystals on the samples' surfaces. We believe that we saw interstitial liquid water, however, when breaking them apart. The ice could have formed because of the pressure release or because of the generally low air temperatures.

We do not think that it would have been possible to retrieve these samples if all the water had been in liquid form. We therefore conclude that some of the water was in the form of ice, albeit microscopic in size as ice crystals were not observed inside the specimens, even with a hand lens. The fact that the samples could easily be broken apart leads us to believe that they were not frozen solid in situ. When drilling through debris layers in basal ice at N2 and N1, the ice was not melted by the drilling. Judging from that experience we do not think that the drilling could have melted a substantial fraction of interstitial ice. Also, if ice had been melted by the drilling, we would expect a melted outer layer and a frozen inner core. This was not observed.

Roughly 165 clasts greater than 1 cm were recovered from till in holes N1 and N1A, including cores and fragments of several 7 to 13 cm thick cobbles and boulders. The clasts consisted of 42% (by number) gabbro similar to underlying 'bedrock' (see below). 25% amphibolite, 14% gneissic granite, 14% light-colored micaceous schist, and 5% dioritic rocks similar to bedrock at Center (see below). Most of the clasts were oblong; the majority were subangular to subrounded although angular and rounded clasts were also found.

Because of the poor recovery, the volume concentration of clasts greater than pebble size is difficult to determine. A minimum can be estimated from the first sample extracted from N1A. The thickness of till recovered is about half the 2.4 m drilled. If we assume the missing fraction is due to washing away of gravel-size and smaller sediments between the intact till pieces, then rocks pebble size or larger would compromise at least 20-30 percent of the till at N1A. Presuming bedrock was reached, we drilled through 7.5 m of till at N1.

Center hole

Till from the top of the Center hole appears to have been reworked by hot water drilling judging from the moderate sorting the sediments displayed upon recovery. Far fewer clasts

(~25) were recovered than at N1 and N1A, perhaps because of the narrower diameter of the drill core. Alternatively, till at Center may not be as clast rich as N1 and N1A. A large, 50 cm thick, boulder of diorite composition was drilled through. The remaining clasts were predominately biotite rich diorite, similar to underlying bedrock, plus quartz pebbles, micaceous schist, and granodiorite. The till layer at the Center hole was 4.5 m thick, 3 m thinner than at N1.

2.4.3 Bedrock

We recovered rock cores from the bottom of two of the drill sites, N1 and Center (Fig. 2.2). The rock core retrieved from N1 was 0.6 m in length. Geochemistry and mineralogy identify the rock as a two- pyroxene, metal sulfide bearing gabbro. The lithology of the basal rocks most closely resembles the 'gabbro of Mt. Moffit' which outcrops at the base of Mt. Moffit to the north of the drill site (Nokleberg and others, 1992). Although fractured and fragmented by drilling, the core appears to be continuous. However, the core displayed significant dissimilarities in pyroxene characteristics between top and bottom suggesting the core may not be continuous bedrock, and instead may be boulders overlying bedrock of similar lithology. Alternately, the dissimilarities may indicate small scale layering in the intrusive.

Fractured but continuous rock core, 1.5 m in length, was recovered from the bottom of the Center hole (Fig. 2.9). Grinding between the different pieces of bedrock had occurred during the drilling. Geochemistry and mineralogy of rock samples taken from the top, bottom, and middle of the core are essentially identical — a two-pyroxene, metal sulfide-bearing, biotite-rich diorite with minor quartz and potassium feldspar but no hornblende. The identical chemistry and length of core argue that the entire sample is bedrock although it could also be a single very large boulder. Thus, although Center and N1 are only 300 m apart, the lithologies of 'bedrock' samples from the two drill sites appear to be distinctly different. The closest possible nearby relative to Center bedrock is metadiorite of somewhat similar composition that is exposed 6 to 10 km west of the field site in the Aurora terrane on the north side of the fault (Nokleberg and others, 1992). However, Center bedrock is distinctly plutonic, not metamorphic. No diorite occurs to the south or upstream of the drill sites. The closest possible relatives in MacLaren terrane rocks are the gneissic granodiorite of the 'East Susitna Batholith', south of the glacier (Nokleberg and others. 1992). Thus, not only do the 'bedrock' lithologies differ, they also appear distinctly different from mapped outcrops upstream of the sites. Given the dextral offset along the Denali fault, the bedrock encountered in the drill holes could be slivers that have been displaced long distances from their point of origin, and may have no relationship to rocks surrounding the glacier. Megascale faults such as the Denali usually have broad shear zones, ranging up to 1 km or more in width and our drill sites may very well lie over such a shear zone. Alternatively, the gabbroic batholith exposed at the base of Mount Moffit to the northeast of the drill site, may extend under the glacier and could be the same gabbro encountered at N1. The latter hypothesis could be tested with further analysis of each gabbro and Ar-dating.

2.5 DISCUSSION

The results presented above lead us to the following questions:

(i) How much ice is contained in the till?

(ii) Is the till-water-ice mixture stable?

(iii) How was this ice formed?

(iv) What can we infer about erosional and depositional processes?

(v) What is the origin of the debris layers in the basal ice?

(vi) What are the implications for the surge behavior?

2.5.1 Ice content of the till samples

In the above section we argued that some of the till samples recovered in the uppermost meter of the till layer at the N1 and N1A boreholes contained ice. The samples were not frozen solid, since they could easily be broken apart. If the ice was simply a part of the solid soil structure, about half of the water (10 to 15% of the total sample's weight) would have to be in solid form in order to keep the water content below the liquid limit. We think this is an upper limit, because ice also has a bonding effect that can increase the stability of the whole matrix. We are not able to give a lower limit. Also, it is not clear how the ice is distributed in the till. This is an important question and relates to the stability of the till-water-ice mixture.

2.5.2 Stability of an ice-water-till conglomerate

When either one of the till samples was dried and reconstituted with a water content of more than 11%, it essentially turned into a slurry. This means that the sample is supersaturated at zero effective pressure (difference between overburden and pore water pressure). That leaves us with two possibilities:

a) The till consists of a fragile skeleton of high porosity (higher than a sample consolidated in the lab) that is stabilized by the bonding action of the ice. There would be grain to grain contact and the soil skeleton would fully carry the overburden pressure.

b) Some of the ice helps bear the overburden pressure. If the interstitial water is connected to a subglacial hydraulic system it would presumably be at a lower pressure and therefore have a higher melting point. Heat flow from the water to the ice would be established and the ice would melt.

Therefore, we sampled till that was potentially unstable in the encountered configuration. This unstability could be temporal — ice in till was formed at some earlier time in a different flow regime and is now melting — or spatial, where ice is being formed upstream and melting out at the drilling site. If such a hypothetical instability is more than a local effect, it could relate to observed velocity variations on different time scales (Heinrichs and others, 1996).

2.5.3 Formation of ice in the till

Two processes allow the formation of ice in till: regelation and freeze-on of water and till particles onto the basal ice. Regelation of basal ice into a till layer was described by Iverson and Semmens (1995) who also explored this idea experimentally. They concluded that — under conditions similar to those at Black Rapids — one would not expect regelation to proceed much further than 1 m into the till layer. At this depth, the downward migration of the ice front would be roughly balanced by melting due to the geothermal heat flux and strain heating, but we did not sample any till matrix below 1.2 m from the ice-till interface. Thus regelation can possibly explain ice in the till.

A more detailed treatment of regelation into a till layer shows that a regelation front would be very uneven because ice would not regelate into small pore spaces due to surface tension effects. This was treated by Everett (1961) and applied to subglacial tills by Tulaczyk (1998). The driving force for regelation is the effective pressure (difference between ice overburden and pore water pressure). A simple calculation shows that effective pressures of 100 kPa — typically observed beneath Black Rapids Glacier — would not drive regelation through pore spaces smaller than $\sim 10^{-6}$ m. Borehole measurements show effective pressures that are one order of magnitude higher during short periods in summer. During such times regelation could proceed through passages larger than $\sim 10^{-7}$ m. The recovered till was rather coarse however (Fig. 2.10) to use this mechanism to explain the high water contents. A knowledge of the size of pore spaces and passage ways would be required to settle the question ultimately.

Ice in a water saturated till matrix can also be formed as a consequence of cold patches, where the local temperature is below the pressure melting point. In general there are more heat sources than heat sinks beneath a glacier (geothermal heat, strain heating). Drake and Shreve (1973) proposed a heat pump effect: water that is melted in high pressure areas flows away and does not refreeze at the immediately adjacent low pressure area. This leaves cold patches that advect downglacier. Robin (1976) estimates that a temperature anomaly of as much as 0.5°C below the local melting point could extend about 2 m into the basal ice. Such a cold layer would only be enough of a heat sink to freeze a few centimeters of interstitial water at the top of the till layer, which is too little to account for the sample recovered from near the top of the layer.

Nolan and Echelmeyer (1999b) proposed large variations in the local overburden pressure due to hydraulic jacking. This would happen at times of high water discharges, such as lake drainages, when the subglacial channels are overpressured. A reduction of the overburden pressure would raise the melting point and some water would freeze. One should expect this ice to melt again as the ice relaxes and the overburden is effective once again.

Röthlisberger (1968) suggested that water derived from an area where the glacier is thicker (and the melting point lower) could flow downstream and freeze there, because it would be below the local melting point. The drilling site is just downstream of a maximum in ice thickness. Between the location of maximum thickness and the drilling site there is a decrease in ice thickness of 11 m and a drop in bedrock elevation of 3 m (Gades, unpublished data). The 11 m change in thickness causes a change in the pressure melting point of about 0.0073 K. This is just slightly more than the warming of the meltwater that would be expected due to the loss of gravitational energy. However, these numbers are all well within the margin of errors of the radio echo sounding — about 10 m — and we cannot make any definite conclusions in this regard.

In summary, regelation and freeze-on of cold water could both potentially lead to a till-water-ice mixture. The occurrence of water as well as ice can be explained by inhibited regelation due to small passages for the former mechanism and differences in the local pressure melting point for the latter.

2.5.4 Erosional and depositional processes

The till thicknesses of 4.5 and 7.5 m suggest a depositional environment, unless erosion is occurring underneath the till layer. Erosion of bedrock by a till layer has been modeled by Hindmarsh (1996) and Cuffey and Alley (1996), who find that erosion under a till layer will not happen unless basal motion is very high or the substrate very weak. Tulaczyk (1998) treats subglacial till as a plastic material containing clasts. He predicts the existence of discrete shear zones. These shear zones can change position due to water pressure fluctuations that diffuse into the till layer. He concludes that a diurnal water pressure variation of 10 kPa amplitude would affect a till with a hydraulic diffusivity of about $10^{-6} \mathrm{m}^2 \mathrm{s}^{-1}$ to a depth of about 0.3 m. Consolidation tests on Black Rapids till show a hydraulic diffusivity of about $10^{-5} \mathrm{m}^2 \mathrm{s}^{-1}$. If seasonal variations are taken into account, then the pressure variation and therefore the weak zone can diffuse to the bottom of a 7.5 m thick layer. It is therefore plausible that the whole layer is active and that erosion is happening, intermittently, at the base of the active till beneath Black Rapids Glacier. This conclusion holds true even if the hydraulic diffusivity were almost a magnitude lower. A lower diffusivity should be expected in the uppermost ice bearing till, because the above mentioned tests were done on unfrozen samples.

If there is no erosion beneath the till layer, then the drill site must be situated in a depositional environment. It is possible, however, that deposition is a sporadic process, as discussed by Iverson and Semmens (1995). They show that changes in the effective pressure and in the sliding speed can change subglacial conditions from regelation into basal till to melt out of basal debris. Regelation and melt out (deposition) are competitive effects. Because melt out is a function of the basal shear stress and basal motion, and regelation is a function of the effective pressure, it is possible that deposition only happens at times of low effective pressures and high basal speeds (high melt rate), such as in early summer or under surge conditions. If massive melt out were to take place during a surge one would expect debris to accumulate in basal ice during quiescent conditions. The basal ice is relatively clean, however. This argues against such a cycle and the scenario of erosion occurring underneath a till layer would therefore seem more likely. Note also that the presence of ice near the top of the till layer indicates that melt out and deposition are not taking place, at least at the time of the sampling.

The lithology of the till clasts indicates that the erosional source is somewhere within 2 km upstream of the drilling site, above which point the glacier bends to the south and overlies different bedrock, south of the Denali Fault. Alternatively, the till could have been supplied from one of the northern tributaries with transport distances of up to 12 km. Only 14% of the recovered clasts were derived from that area. These could have been derived supraglacially, although some rounding was observed, suggesting contact with a basal traction zone. The vast majority of the clasts is therefore of local origin and was eroded from the bed since the 2 km distance would not be enough to transport surficial debris to the bed. The moderate rounding of clasts and the relatively coarse till matrix also suggest that the transport distance of the till is not very long (Haldorsen, 1981; Hooke and Iverson, 1995). Clasts of boulder size were encountered in the till. They indicate that plucking is occurring upstream of the drill site (Röthlisberger and Iken, 1981; Hallet, 1996). The lithological diversity of the recovered clasts indicate that the bed upstream of the drill sites must be relatively inhomogeneous, reflecting the diversity of the bedrock around the glacier.

2.5.5 Debris layers in basal ice

Basal debris layers, separated by clean ice, occurred as high as 12 m above the bed. Lawson and Kulla (1978) and Strasser and others (1996) encountered a similar situation at Matanuska Glacier. Their isotopic study suggests that the ice was refrozen melt water, at least partially surface derived. Boulton (1979) suggested that basal debris layers can form as a result of the three dimensional flow around bedrock bumps. However, we drilled into an area of massive till and do not believe that bedrock was exposed to the ice within a short distance of the drilling site. The debris encountered in our cores does not look surficial, because it is of similar lithology as the underlying bedrock and because it shows signs of abrasion and crushing. Some of the layers closer to the ice-till interface are steeply inclined (up to 60°) mud layers that could have formed by filling in basal cracks. Such layers would rotate to a dip of 30° dipping layer in about three years under the strain rates expected. If the basal ice is deforming under simple shear, these layers must be fairly fresh. Alternatively, thrust faults could explain the debris layers (e.g. Rabus and Echelmeyer, 1997).

2.5.6 Implications for surge behavior

The possible instability of a till-water-ice mixture discussed above could relate to the surge behavior. However we have sampled such a till at only one location and it could be a very local phenomena.

The most striking result of the drilling is the thickness of the till layer encountered, suggesting large erosion rates, either now or in the past. Humphrey and Raymond (1994) observed erosion rates of 0.3 m of bedrock in 20 years on Variegated Glacier. About two thirds of the sediment evacuation occurred in the two years of the surge and the bulk in probably only two months. Black Rapids Glacier last surged in 1936/37 when a significant amount of erosion could have happened in a similar manner. In both cases, however, it is unclear whether the high erosion rates are a prerequisite rather than a consequence of the surge behavior.

2.6 SUMMARY

We have shown that a commercial wireline drill rig is capable of drilling in a subglacial environment under an ice thickness of up to 600 meters. The problem of basal motion can be overcome by drilling early in the season and by reducing the drilling time to a minimum. Early season drilling also ensures a more solid working surface, when we could move the drill rig across the glacier. There are substantial difficulties in sampling subglacial till. A combined punch corer – rotary drill would probably be an ideal combination for this task, but cobbles the size of the core diameter make the use of such a system difficult. Water for cooling and flushing has to be used judiciously; too much of it will wash out all the fine material, and too little will eventually cause melting of the drill bit. Successfully latching the inner tube to the core barrel presented a problem never quite solved. As our experiences have shown, drilling and sampling subglacial till is as much an art as it is a science, even for commercial drillers.

Bedrock was recovered at two sites. N1 lies within the Denali Fault zone, while Center may be a tectonic sliver that is not exposed on either side of the glacier. The tectonic shear zone is at least 300 m wide, possibly as much as 1 km. The peculiar setting of the glacier, which crosses the fault about 2 km upstream of the drilling site, allows us to make some conclusions about the origin of the till.

A till layer of 7.5 m at N1 and 4.5 m at Center was encountered, confirming the predictions of Nolan and Echelmeyer (1999a and b). The till is of local (north of the Denali Fault) origin, with only 14% of the clasts originating south of the fault. The degree of rounding indicates a certain amount of travel in the basal zone. Ice-bearing till was sampled near the top of the till layer ($\leq$1.2 m below ice-till interface). This ice could have been formed by regelation or freezing of cold water. The till contains a minimum of 20% of clasts that are pebble size and larger. The till matrix is mostly sandy with a very low clay and silt fraction. The water content of samples just below the interface is 20–30%, much higher than the liquid limit, suggesting the presence of ice. Such an till-water-ice mixture is potentially unstable.

Debris layers in basal ice were encountered as far as 12 m above the ice-till interface. Immediately above the interface at N1, steeply inclined mud dikes were sampled. They either represent flow along shear planes or recently filled-in basal cracks.

2.7 ACKNOWLEDGEMENTS

The fieldwork described in this paper was an order of magnitude larger than that which we normally encounter on valley glaciers. It would not have been possible without the help of many field assistants and helpers. We want to thank Anne Harrison for a wonderful job as a cook and morale upkeeper, the drillers and drill helpers of Elgin Exploration Co., Calgary, Canada, and Dale Pomraning, Jesse Collins, Laurence Sombardier, Robin Wilson, Olaf Eisen, Hermann Engelhardt, Ian Willis, Matt Nolan and Dan Elsberg. The Polar Ice Coring Office did the drill contracting. Two reviewers provided valuable comments. This work was supported by the NSF grant OPP 9423 477.

Chapter 3

Glacier motion dominated by processes deep in underlying till

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ABSTRACT[1]. Black Rapids Glacier is a 40 km long, surge-type glacier in the central Alaska Range. In spring 1997 a wire line drill rig was set up at a location where the measured surface velocities are high and seasonal and annual velocity variations are large. The drilling revealed a layer of subglacial 'till', up to 7 m thick, that is believed to be water saturated. At one location a string of instruments, containing three dual-axis tiltmeters and one piezometer, was successfully introduced into the till. The tiltmeters monitored the inclination of the borehole at the ice-till interface, 1 m into the till, and 2 m into the till for 410 days. They showed that no significant deformation occurred in the upper two meters of the till layer, and no significant amount of the basal motion was due to sliding of the ice over the till. The measured surface velocity at the drill site is about 60 $\mathrm{m\,a^{-1}}$, of which 20-30 $\mathrm{m\,a^{-1}}$ can be accounted for by ice deformation. Almost the entire amount of basal motion, 30-40 $\mathrm{m\,a^{-1}}$, was taken up at a depth in the till greater than 2 m, possibly in discrete shear layers, or as sliding of till over the underlying bedrock. We propose that

[1]in press: Journal of Glaciology

the large-scale mobilization of such till layers is a key factor in initiating glacier surges.

3.1 INTRODUCTION

Early models of basal motion commonly assumed that glaciers were underlain by clean bedrock. Some borehole studies, however, showed the presence of a layer of sediments beneath the ice (Engelhardt and others, 1978). The importance of such a layer for glacier dynamics was demonstrated by the work of Boulton and Hindmarsh (1987) who measured the positions of strain markers and pore water pressures in a sediment layer beneath the terminus of Breidamerkurjökull in Iceland, and by Echelmeyer and Zhongxiang (1987), who measured the distribution of basal motion in a tunnel at the terminus of Urumqi Glacier in China. Interest in till deformation was stimulated by the evidence for a water saturated till layer beneath Ice Stream B (Blankenship and others, 1986) that was interpreted to be soft and deforming (Alley and others, 1986)[2]. In a subsequent drilling effort samples of the underlying till were recovered and high basal water pressures — within about 160 kPa of the ice overburden pressure — were measured (Engelhardt and others, 1990).

Although the importance of subglacial till in glacier motion has been recognized, there remains controversy about its rheology. Boulton and Hindmarsh (1987) concluded that till was deforming like a nearly linear viscous fluid. Kamb's (1991) tests on Ice Stream B samples suggested a highly non-linear behavior. Iverson and others (1998) summarized several shear tests carried on to high strains in a ring shear apparatus. They show a shear strength of the material that is essentially strain rate independent, i.e. a Coulomb-plastic rheology. The consequences of such a rheology for subglacial material has been examined by Tulaczyk (in press) and Tulaczyk and others (subm.). They and Iverson and others (1998) also pointed out that a Coulomb-plastic rheology can produce distributed motion in a till layer.

Several instruments have been developed for in-situ studies of subglacial sediments: borehole cameras (Harrison and Kamb, 1973), tiltmeters, pore pressure transducers, drag spools, and ploughmeters (Blake and others, 1992, 1994; Fischer and Clarke, 1994). Work with a combination of such instruments has suggested that — at least in some cases —

[2]In this paper we will call all unlithified subglacial material 'till', irrespective of its origin.

episodes of fast motion are caused by sliding of ice over till and not by till deformation (Hooke and others, 1997). Engelhardt and Kamb (1998) used a tethered stake to conclude that a majority of the observed surface velocity at Ice Stream B was due to sliding of ice over till, or possibly deformation of a very thin (3 cm thick) till layer.

Observations of subglacial till have also been made in tunnels cut into the ice in the terminal area of glaciers. Boulton and Hindmarsh (1987) observed distributed motion in the uppermost 50 cm of the till that accounted for 80-95% of the observed glacier motion. Echelmeyer and Zhongxiang (1987) showed that 60-85% of the motion of sub-freezing Urumqi No. 1 Glacier in China was due to basal processes: enhanced deformation of ice-laden till, motion across discrete shear planes, and basal sliding at the ice-till interface. Motion across shear bands and planes accounted for 10-25% of the total glacier motion. Sliding of ice over bedrock or till, till deformation and faulting within the till, and perhaps even sliding of till over underlying bedrock have all been suggested as mechanisms of basal motion (e.g. Knight, 1999, his Fig. 7.1), although the relative magnitudes of these processes remain uncertain.

In this paper we use data from one location on Black Rapids Glacier, Alaska, to investigate the contribution of basal motion to the observed surface motion of the glacier. Detailed measurements allow us to ascertain where in the basal layer this component of motion is occurring. The implications of our findings are then discussed in terms of till deformation and also surge initiation.

3.2 SETTING

Black Rapids Glacier is a 40 km long surge-type glacier in the central Alaska Range. It last surged in 1936/37. The glacier has been investigated since the early 1970s by the U.S. Geological Survey, the University of Alaska Fairbanks and the University of Washington. Heinrichs and others (1996) describe the glacier in its quiescent state. More recent work includes a seismic study (Nolan and Echelmeyer, 1999a, 1999b) in which changes in the seismic return signal on very short time scales — as little as half an hour — were observed. They could only explain this observation with drastic changes of the overburden pressure on a water saturated till layer, at least 5 m thick. These changes occurred as a result of the draining of marginal lakes.

In spring 1997 a wire line drill rig was set up about 15 km from the head of the glacier in an effort to sample basal ice, subglacial till and underlying bedrock (Truffer and others, 1999). This is the area of the glacier where the velocities are high, where large seasonal and annual velocity variations are observed (Heinrichs and others, 1996), and where the above mentioned seismic study was carried out. The glacier's maximum thickness is over 600 m at that location, and the surface slope is about 2°. Drilling to bedrock was successful in two places, where we measured a thickness of the till layer of 7.5 m at N1 and 4.5 m at Center (Fig. 2.1, Truffer and others, 1999). Basal ice, containing some dirt layers, was clearly separated from the underlying till. The upper few centimeters or decimeters did contain some ice, however. A recovered sample of till matrix from the top of the till layer underwent several soil engineering tests (see 3.4.3).

A string of three tiltmeters and one piezometer was installed in borehole N1A (Fig. 2.1). At that location the ice thickness was 499 m. This paper concentrates on the results of this borehole experiment.

3.3 METHODS

In this section we discuss the methods used to measure surface motion, ice deformation, motion of the underlying till, and water pressure. The results are reported in the next section.

3.3.1 Surface motion and deformation of the ice

The glacier's surface motion at the drill site was measured over various time scales. The annual displacement of the borehole marker was measured by resection with a theodolite. In June 1997 a GPS receiver recorded the position of a nearby stake four times a day for three weeks. Upstream of the drill site a camera has taken one picture a day since 1986 to record the motion of a pole.

Internal ice deformation in borehole N1 (located 10 m to the north of N1A) was measured with a retrievable inclinometer that was equipped with two mutually perpendicular tilt sensors and a compass. The inclinometry was performed on three occasions: 16 May, 1

June, and 28 June 1997.

3.3.2 Measurements within the till

Three tiltmeters were installed near the bed and in the till; the uppermost at the ice-till interface, and the others at 1 m and 2 m below the interface (Fig. 3.1). This distance was measured from the top of the borehole down, as well as from the bottom up. The total length of the borehole was very well known from the wireline drilling. The two methods of measurement were in excellent agreement. The elastic properties of the instrument cable were measured prior to installation, and the 1.5 m stretch of the cable due to a 20 kg load at its end was taken into account. The buoyant weight of the cable was 20 kg. Potential errors result from additional stretching of the cable due to its own weight, the buoyancy of the steel weight in a drill mud of a density unknown to us, and from the possibility of caving in of parts of the borehole before instrument insertion. These errors add up to a maximum of 0.5 m in the position of the instruments.

The instruments were mounted on a cable of 1.3 cm diameter (Cortland Cable Company) that was designed to stretch up to 30% before breaking. We used dual-axis electrolytic tiltmeters (Fredericks) with a measuring range of 0° to 30°. A datalogger (Campbell 21X) recorded tilt along both axes, which was converted to total tilt and azimuth (Fig. 3.2). Changes in azimuth cannot be distinguished from rotation of the instrument around its own axis, because no compass was used. We assume that such rotation is small because it would induce torsion in the short section of cable between the individual instruments. The tiltmeters were mounted in 16 cm long cylindrical pressure cases, 7 cm in diameter.

Pore water pressure was measured with a vibrating wire piezometer (Geokon) 0.5 m below the ice-till interface. The instrument had a range of 7 MPa and a resolution of 3 kPa. Borehole water levels were also recorded with a pressure transducer at 150 m below the glacier surface. This transducer was less accurate and subject to systematic errors.

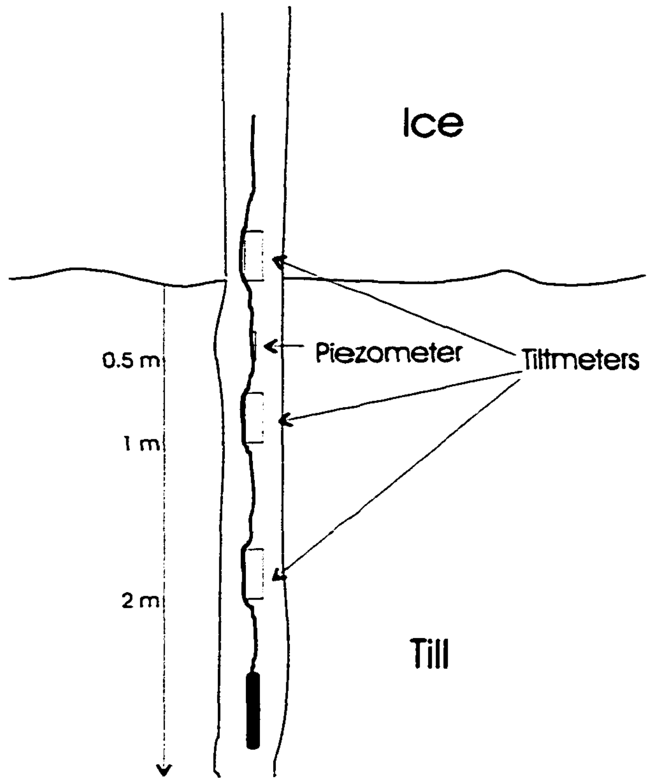

Figure 3.1: Postion of the instruments. Schematic drawing of the borehole instruments and their position in the borehole, as discussed in the text.

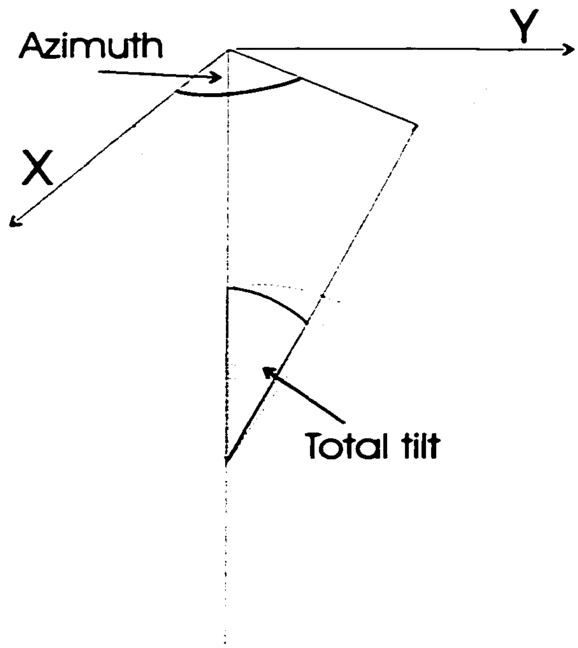

Figure 3.2: Dual axis tilt measurement. Tilt is measured along two axis (dotted lines). It is then converted into total tilt and azimuth (bold lines).

3.4 RESULTS

3.4.1 Surface displacement and ice deformation

The average annual surface velocity was about 60 $\mathrm{m\,a^{-1}}$ at the borehole location. This implies a surface displacement of almost 70 m for the 410 days of borehole data collection. The position of a marker 300 m to the south of the drilling site was measured with a theodolite and an EDM until JD 155, and then four times a day using GPS methods. Fig. 3.3a shows the horizontal velocity. The spring speed-up on 7 June and several other events are prominent features of the record. The one on 21 June was a result of the drainage of a marginal lake upstream of the drill site, observed by an automatic camera. These events are very well resolved, as the GPS derived velocities have an error of less than 6 $\mathrm{cm\,d^{-1}}$ (see error bar in Fig. 3.3a).

Only the lowermost 100 m of the inclinometry record could be interpreted because, apparently, the borehole was too wide in its upper part to yield meaningful results. Under the assumption that the upper part of the borehole does not contribute significantly to the

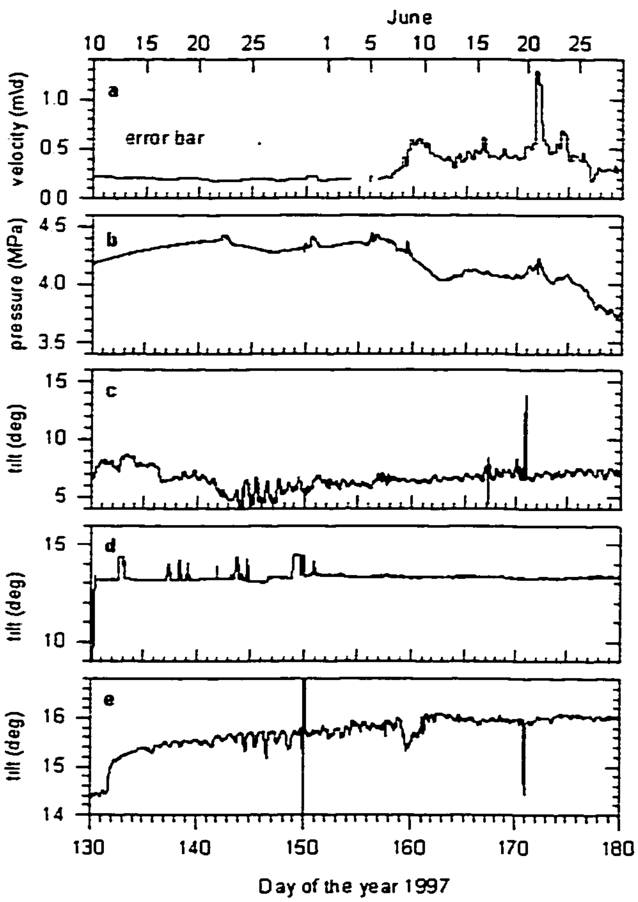

Figure 3.3: Velocity, water pressure and tilt. a) Velocity measured with theodolite and EDM (before JD 155) and GPS methods (thereafter). b) Water pressure measured 0.5 m below the ice-till interface. Total tilt of the tiltmeters at c) the interface, d) 1 m and e) 2 m below it. Note the different scales on the ordinate axes.

total deformation, we derived deformational speeds of 19 $\mathrm{m\,a^{-1}}$ for the period between 16 May and 1 June, and of 23 $\mathrm{m\,a^{-1}}$ for the period between 1 June and 28 June. On 7 June 1997 the annual spring speed-up occurred and the surface velocity of the glacier rose by a factor of three (Fig. 3.3a). This did not significantly affect the results of the inclinometry. A second estimate of deformational speed can be obtained from winter velocities. Minimum observed velocities amount to 35 $\mathrm{m\,a^{-1}}$. This is an upper estimate for the deformational speed, because there is reason to believe that basal motion is occurring year round (Heinrichs and others, 1996). A third estimate of about 30 $\mathrm{m\,a^{-1}}$ was obtained by using a model based on Kamb and Echelmeyer's (1986) longitudinal averaging method. For the remainder of the paper we will assume that 20 to 30 $\mathrm{m\,a^{-1}}$ is due to deformation of the ice. This implies basal motion of 30-40 $\mathrm{m\,a^{-1}}$.

3.4.2 Measurements in the till

Tiltmeter and piezometer data were recorded from 8 May 1997 until 22 June 1998, when water flooded the datalogger box. Data gaps resulted from using up the storage module's memory in summer 1997 and loss of battery power in winter 1998 (Fig. 3.4). We had expected the cable to break during one of the early speed up events in spring 1997. The fact that data were still being recorded well over a year after initiation of the experiment came as a major surprise.

Fig. 3.5 shows the tiltmeter records in a projection of the tiltmeter axis on a horizontal plane. The total change in tilt between the first and the last measurement is 3.4° for the uppermost, 4.9° for the second, and 1.8° for the lowest tiltmeter, implying 13 cm of basal motion over the uppermost 2 m of till.

The small values of total tilt, the fact that the cable was still intact after 410 days, and the estimate of 35 to 45 m of basal motion during that time interval lead us to conclude that the majority (50 to 70%) of glacier motion occurred *below* the lowermost tiltmeter, i.e. below 2 m in the till (Fig. 3.6).

At times there appears to be larger tilt but this tilting seems to be a reversible phenomenon. The tiltmeter at 1 m below the interface shows one jump of about 15° on JD 207, and then several excursions later on, all in the same direction (Fig. 3.4 and 3.5). All of the tiltmeters show diurnal variations at some times. No permanent tilting is associated with

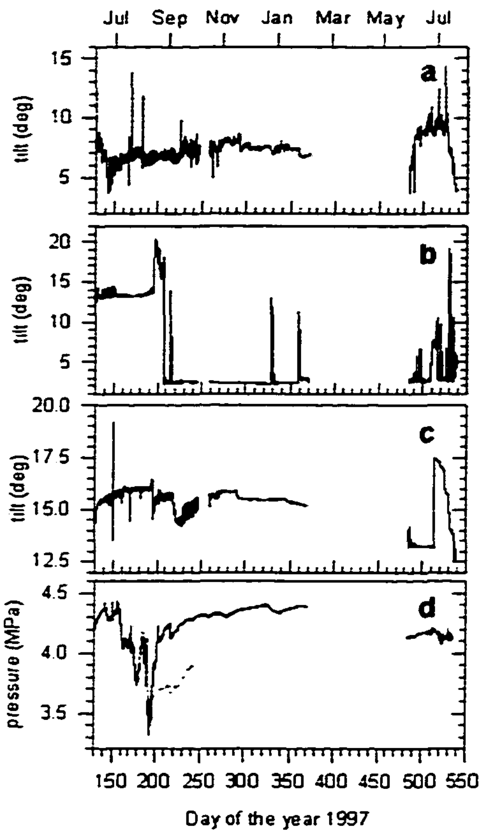

Figure 3.4: Total tilt and water pressure. Total tilt a) at the interface, b) 1 m below, and c) 2 m below. Azimuth is not shown in this graph. d) Water pressures measured at 0.5 m below the interface (solid line), and at 150 m below the glacier surface (dotted line). The borehole waterlevel record was converted to pressure at the position of the piezometer in the till. They overlap from JD 180 to JD 195, when a shift in the waterlevel record occurs.

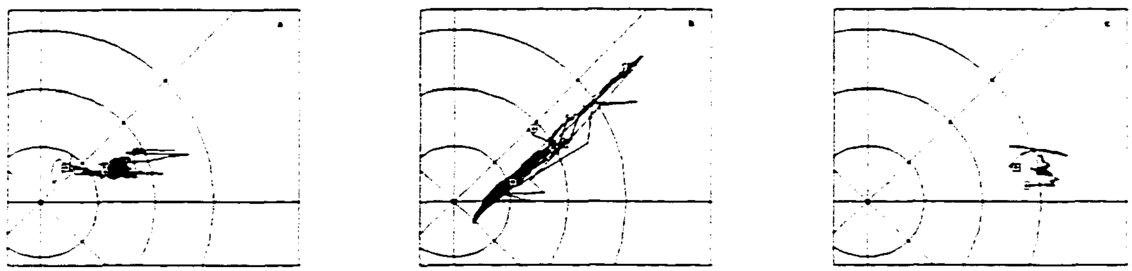

Figure 3.5: Stereo plots of tilt. The total path of the tiltmeter axis projected on a horizontal plane (i.e. the x-y plane in Fig. 3.2. a) Tiltmeter at the interface, b) 1 m, and c) 2 m below. The open circles mark the start and the open squares the end of the record. The circles are contour lines of total tilt.

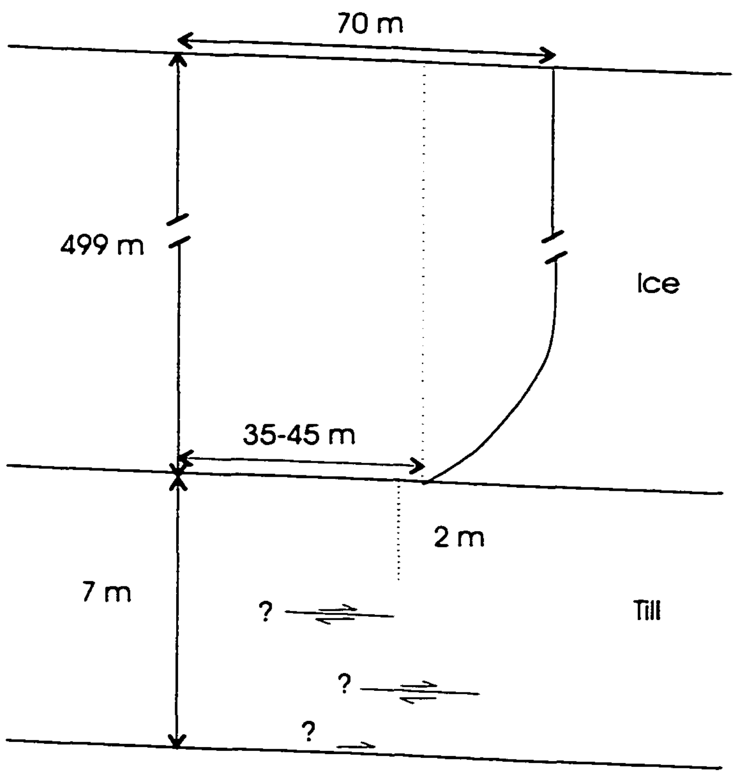

Figure 3.6: Inferred distribution of glacier motion. The diagram shows the inferred distribution of glacier motion over the 410 day measurement period. 50-70% of the observed motion occurs below 2 m in the till, either on a series of shear bands or at the till-bedrock interface.

speed-up events (Fig. 3.3), although the lowermost tiltmeter seems to react to the spring speed-up, and the uppermost and lowermost instruments show strong, but reversible spikes prior to the large speed-up event on 21 June (Fig. 3.3).

Fig. 3.4d shows the pore water pressure record. For about two months water pressures were simultaneously measured in the borehole at 150 m below the glacier surface. These measurements show that the piezometer in the till stayed hydraulically connected to the borehole for at least that long, as no phase lags and no attenuation of high frequency variations were observed. The transducer at 150 m below the surface is subject to systematic errors which explains the shift after JD 195.

3.4.3 Engineering properties of a till sample

A cylindrical till matrix sample of 11 cm length and 5 cm diameter was recovered from the top of the till layer in hole N1A. The sample contained some ice and a water content of 29% was measured after letting it melt. Sieving showed that only 6% of the sample's weight is in the silt and clay fraction (Truffer and others, 1999). The remoulded till was tested in a triaxial apparatus and an oedometer. In addition, an Atterberg liquid limit of 10% was determined in a liquid limit testing device (Bowles, 1992). Table 3.1 summarizes the results. The friction angle ϕ and the apparent cohesion c_a define a linear relationship (Coulomb friction law) between the sample strength τ and the effective pressure σ':

$$\tau = c_a + \sigma' \tan(\phi) \tag{3.1}$$

(e.g. Lambe and Whitman, 1979). The low apparent cohesion is expected for a granular material, but previously measured friction angles on glacial till were typically smaller than 30° (Iverson and others, 1998). Our high value reflects the low silt and clay content (Lambe and Whitman, 1979) and the high angularity (data by D.B. Simmons shown in Julien, 1995, his Fig. 7.2), although it might have been caused by a few large particles that were not removed before the test. The in-situ diffusivity of the till is likely to be lower than that measured in the oedometer due to the inferred presence of ice in the uppermost part of the till (Truffer and others, 1999).

Ring shear testing was done by Slawek Tulaczyk at the California Institute of Technology to test the strain rate dependence of the shear strength. The shear strength increased by

soil density	2200 $\mathrm{kg\,m^{-3}}$
water content	29%
liquid limit	10%
friction angle	40°
apparent cohesion	1.3 kPa
hydraulic diffusivity	$1.5 \cdot 10^{-5}\,\mathrm{m^2\,s^{-1}}$

Table 3.1: Engineering properties of a till sample from borehole N1A

only 2 kPa (from 21 to 23 kPa), as the strain rates were varied over two orders of magnitude, reflecting a highly non-linear nature of the till. The slight increase in shear stress may reflect strain rate induced variations of effective pressure, and not a viscous effect (Kamb, 1991; Tulaczyk and others, in press).

3.5 DISCUSSION

In this section we will first discuss alternative interpretations of our data, and then give a tentative explanation for short term fluctuations in the record. After discussing a mechanism that can explain motion on shear layers deep within the till, we go on to discuss the glaciological consequences of these findings, in particular on erosion and the surge mechanism.

3.5.1 Alternative interpretations of the data

In the above section we have concluded that the majority of the observed surface motion originates from deeper than 2 m below the ice-till interface. We have considered several other possibilities, such as failure of the instruments, pulling of the instruments out of the till, and misinterpreting a dirty basal ice layer as water-saturated basal till. None of these possibilities can satisfactorily explain the results. If, for some reason, the instruments were placed in the basal ice rather than in the till, a steady tilting of much higher magnitude would be expected. The most likely alternative interpretation is that a substantial amount of basal motion occurs as sliding at the ice-till interface (Fig. 3.7). This could be the case under three conditions. First, the uppermost tiltmeter would have to be placed far enough from the ice-till interface not to be affected by the sliding motion. This is a possibility due to the uncertainties in the emplacement (see section 3.3). Second, the sliding motion would not have destroyed the cable. In principle, this is a possibility. If the induced stretch could be taken up over the entire length of the cable, it could stretch up to 150 m, substantially more than the 35 to 45 m required from basal motion. It seems rather unlikely, however, that the cable survived such an intense shearing environment for well over a year, a view shared by the manufacturer (Jack Dower, Cortland Cable Company, pers. comm. 4/99). And, third, the instruments would have to be firmly emplaced in a very strong till. If they were not firmly emplaced, they would be pulled out of the till and into the ice. Although ice deformation accounts for only half or less of the total surface motion, tilting rates at the bottom of the glacier are expected to be significantly higher than those recorded by the tiltmeters. If they had been pulled out of the till, one after the other would have recorded higher tilting rates. If the till was not strong, the instruments would be dragged through the deformable till, again causing appreciable amounts of tilting in one instrument after the other. This is not observed. On the other hand, the tiltmeter at 1 m below the interface, exhibits several excursions of up to 15° of tilt. Also, the lowermost tiltmeter jumps by about 5° on 25 May 1998 (Fig. 3.4). At about this time the spring speed-up, which was not recorded in 1998, is expected to happen (Fig. 3.3). The other tiltmeters do not record this event, or at least not as clearly. This observation fits very well with the idea of shear layers just below the instruments, but not so well with the idea of a strong till and a sliding interface between the ice and the till.

3.5.2 Short term fluctuations

We tested the idea whether a diffusing water pressure wave affected the till strength (Fischer and others, 1998) and thus some of the structure in the tiltmeter record (Tulaczyk, 1998; Tulaczyk and others, in press). We made the assumption that the pore water pressures in the till are driven by a fluctuating water pressure at the ice-till interface. The water pressure beneath a valley glacier normally drops to a minimum in summer, once a good drainage system is established (Fig. 3.4). This increases the effective pressure in the till and makes it stronger. Underlying till will still be at higher pore water pressures and therefore

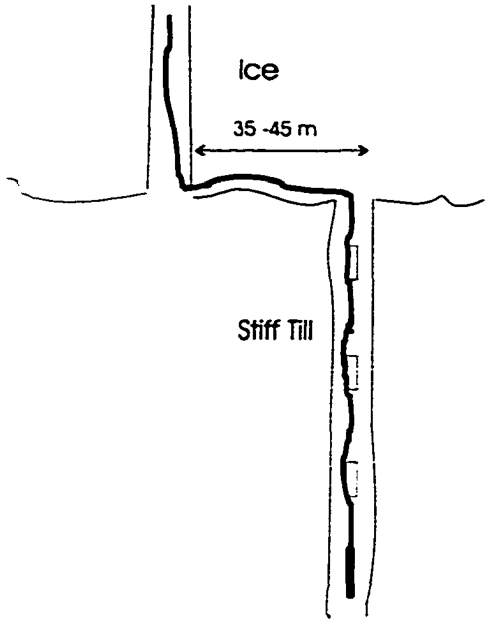

Figure 3.7: Position of the instruments if sliding is dominant. Most of the basal motion would occur on a discrete sliding plane at the ice-till interface or in a thin layer of sediments just below this interface.

lower effective pressures. It will deform more readily. We modeled pore water pressures in the till by numerically solving the one dimensional diffusion equation

$$\frac{\partial p}{\partial t} = C_v \frac{\partial^2 p}{\partial z^2} \tag{3.2}$$

(de Marsily, 1986), where p is the pore water pressure and C_v the hydraulic diffusivity of the till. The measured water pressures were prescribed at the top of the till layer — interpolating through the data gap and assuming the pressure has a periodicity of one year. A zero gradient of hydraulic head was assumed at the bottom. We then compared the modeled results with a smoothed record of measured tilts (Fig. 3.8). The largest changes in tilt are observed at the time when the low water pressure wave reaches the corresponding tilt meter, but otherwise the correlation between the two curves is not strong. The best agreement between the curves was obtained when a hydraulic diffusivity of $3 \cdot 10^{-7}\,\mathrm{m^2\,s^{-1}}$ was used, and five day averages of tilts were plotted. The diffusivity is somewhat smaller than that measured on a sample ($10^{-5}\,\mathrm{m^2\,s^{-1}}$), but due to the inferred presence of ice in the uppermost part of the till layer, we expect that the lab test overestimated the hydraulic diffusivity (Truffer and others, 1999).

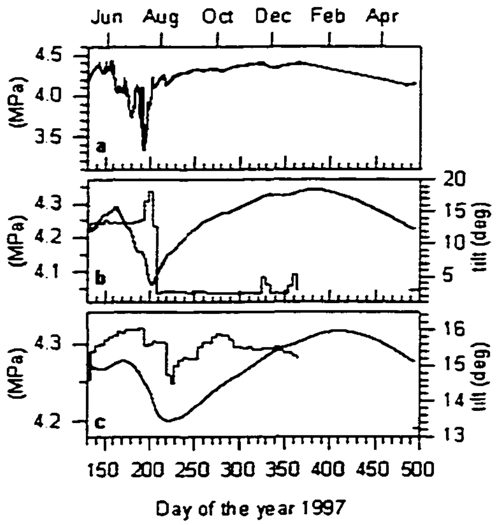

Figure 3.8: Modeled pore pressure and tilt. a) Observed water pressure used as input for the model. b) and c) Modeled pore water pressures and averaged total tilt (step plot) 1 m and 2 m below the interface.

Although the overall measured tilt rate is negligibly small to affect surface motion significantly, each of the tiltmeters indicates periods of high strain rates, both positive and negative. Such variations have been observed before (Blake, 1992), and have been interpreted as elastic effects, due to changing local basal shear stress (Iverson and others, 1999) or effective stress (Tulaczyk, in press). Such elastic effects are predicted by an elastic-plastic soil model if stress changes are not large enough to exceed a previously reached maximum (e.g. Wood, 1992). Nolan and Echelmeyer (1999b) concluded that lake drainage events cause hydraulic jacking and temporary changes in local ice overburden pressure. Together with changes in the pore water pressure and local shear stresses, rather dramatic switches in the stress state should be expected.

It should be kept in mind that the small scale features in the tiltmeter record are much harder to interpret because the tiltmeters are connected with a cable. The cable allows for a certain amount of interaction between the instruments and some of the tilting could be due to pulling on the cable.

The tiltmeter record indicates that basal motion due to shearing in the upper 2 m of the till amounts to only 13 cm in 410 days. 50 to 70% of the glacier's surface motion is occurring more than 2 m beneath the ice-till interface. The total tilts reported above are somewhat fortuitously small, since tilts as large as 20° are observed (Fig. 3.4). A uniform tilt of 20° would still only amount to 0.7 m of basal motion. a negligible amount.

3.5.3 Motion deep in a till layer: other observations

Direct observations of the distribution of motion in till layers have been made in tunnels near the glacier terminus (Boulton and Hindmarsh, 1987; Echelmeyer and Zhongxiang, 1987). The latter observed shear bands in the frozen subglacial till beneath Urumqi Glacier in China, and concluded that 10-25% of the total surface motion occurred on such bands. The present study concludes that an even larger amount of motion can occur deep in the till layer of a relatively fast moving temperate glacier. In retrospect, several other observations also fit into this picture. Harrison and others (1986) used borehole TV on surge-type Variegated Glacier, and were surprised that they did not manage to measure significant motion across the ice-till interface. Iverson and others (1999) attributed the lack of significant till deformation in the upper 20 cm of the till layer to sliding of the ice over the underlying till, but did allow that it could also be due to motion deeper in the till.

In a study of the Puget glacial lobe Brown and others (1987) concluded that 'basal motion was confined to the ice-bed interface or to distinct faults within the substrate', and that 'shearing along discrete shear zones is evident and may amount to significant displacement' (Brown and others, 1987, page 8994). Hiemstra and Van der Meer (1997) also concluded that glacial deposits from Wijnjewoude in The Netherlands contained discrete horizontal to sub-horizontal shear zones, along which strain was focused. Astakhov and others (1996) reported observations of shear bands in West Siberian permafrost that was formerly overlain by glaciers. The occurrence of such discrete shear zones in glacial deposits has thus become textbook knowledge in glacial geology (e.g. Menzies and Shilts, 1996, page 48), although it has not been possible to show the amount of displacement that occurs in such zones.

Alley (1991) presented arguments for widespread till deformation beneath the Laurentide ice sheet. This contrasts with the view of Clayton and others (1989) who argue against pervasive till deformation on the basis of till stratigraphy. It is also often assumed that coherent thrust sheets imply a frozen bed (e.g. Mickelson and others, 1983). The results presented in this paper might help resolve these paradoxes by allowing slip across discrete planes deep within unfrozen till layers. Such deep-seated slip may also help explain observations of large till sheets (up to 1000 km^2) which have been displaced up to 250 km with little apparent internal deformation (references on page 328 in Van der Wateren, 1995).

Discrete shear zones are also observed in shear tests on granular materials (e.g. Mandl and others, 1977).

3.5.4 Is till deformation at a depth of over 2 m physically plausible?

The question of till deformation is intimately linked with that of the strength of the ice-till interface (Brown and others, 1987; Alley, 1989; Tulaczyk, 1998; Iverson, 1999). These studies show that this strength depends on the size of clasts and the grain sizes of till, and on the effective pressure at the interface. An interface between ice and coarse clast-rich till, such as the one encountered beneath Black Rapids Glacier, is expected to have a high strength, thus making till deformation more probable. The inferred presence of regelation ice in the upper part of the till (Truffer and others, 1999) only serves to make this coupling stronger.

Tulaczyk (1998) described three different models of basal motion over a water saturated till: sliding over the till, shear deformation at a depth that is determined by the minimal effective pressure (overburden minus pore water pressure), and shear deformation at a depth determined by the minimum of the past maximum effective pressure. The last model, 'the perfectly over-consolidated till', takes into account the memory of a granular material. When subjected to a shear load, a granular material will exhibit a peak strength that is determined by the maximum effective pressure it has experienced since it was last disturbed. If this maximum is higher than the current effective stress, the till is called overconsolidated (Clarke, 1987). The maximum effective pressure is the difference between the overburden pressure, which increases linearly with depth in the till, and the minimum in water pressure over a given time period at a certain depth. This minimum water pressure increases with depth due to diffusion. To model the depth-dependence of the maximum effective pressure throughout the year, the water pressure was calculated by solving Equation (3.2) for a 7 m thick till of hydraulic diffusivity $C_v = 3 \cdot 10^{-7} \mathrm{m}^2 \mathrm{s}^{-1}$ and density $\rho_{\mathrm{till}} = 2700 \, \mathrm{kg} \, \mathrm{m}^{-3}$. Figure 3.9 shows the resulting annual maximum effective pressure as a function of depth. A strong decrease in the first 2 m is followed by a broad minimum at around 3.5 m and an asymptotic decrease that reflects the increasing overburden pressure. The position of this minimum in the annual maximum effective pressure reflects the position of the smallest value of peak strength, and marks a likely place for a shear layer. Using a hydraulic diffusivity of $C_v = 10^{-5} \, \mathrm{m}^2 \, \mathrm{s}^{-1}$ (Table 1) places such a layer even deeper in the basal zone, at 4.6 m below the ice-till interface.

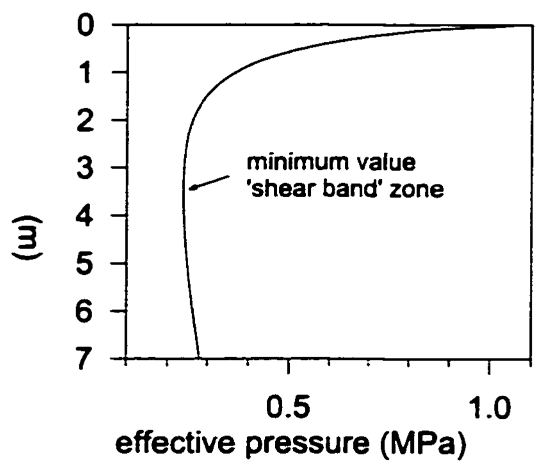

Figure 3.9: Maximum effective pressure as a function of depth. Overburden pressure increases linearly with depth. The annual minimum of the water pressure was calculated by solving a diffusion equation.

In a proper model of basal motion, the strength of the till-bedrock interface has to be considered as well. It is possible that this interface is rather smooth, and that it is the weakest point in the basal ice - till - bedrock system. Sliding of till over bedrock has been discussed by Cuffey and Alley (1996) and Hindmarsh (1996).

3.5.5 Erosion and till mass balance

The glacier's main accumulation area lies to the south of the Denali Fault from where it crosses into the main fault valley about 3 km upstream of the drill site. In Truffer and others (1999) we concluded that about 85% of the till sampled in the boreholes was derived from the northern side of the fault. On the northern side or along this fault the glacier extends between 7 km (ice divide) to 12 km (head of the westernmost tributary) upstream from the drill site (Fig. 2.1). If between 2 m and 7 m of till were to be carried along at the rate of the basal motion, the average upstream erosion rate would have to be 4-30 $\mathrm{mm\,a^{-1}}$. This is somewhat higher than what is thought to be typical of alpine glaciers (e.g. Bennett and Glasser, 1996), but agrees well with the high rates found in Southeast Alaska (Hallet and others, 1996). Humphrey and Raymond (1994) found that the erosion rate on Variegated Glacier was directly proportional to the ice velocity. They calculated a dimensionless erosion rate — the ratio of erosion rate to ice velocity — on the order of 10^{-4} for Variegated Glacier, and they showed that a similar rate was found for Glacier d'Argentière and Breidamerkurjökull. If this ratio applied at Black Rapids Glacier, an erosion rate of 6 $\mathrm{mm\,a^{-1}}$ would result, just at the lower limit of the above estimate. It is possible that sliding of till over the underlying bedrock is the erosional mechanism for supplying this substantial amount of till (Cuffey and Alley, 1996; Hindmarsh, 1996). The presence of large clasts and boulders in the till (Truffer and others, 1999) shows that abrasion cannot be the only erosional mechanism, however, and plucking has to occur as well.

3.5.6 Implications for the surge behavior

Black Rapids is a surge-type glacier and naturally the question arises if our observations can bring us a step closer towards understanding surge behavior and initiation. The fast motion during the 1982/83 surge of Variegated Glacier was accompanied by water pressures close to overburden (Kamb and others, 1985). Dye tracing experiments indicated a complete switch from a fast to a slow drainage system between non-surge and surge condition (Brugman, 1986), and large volumes of water were released at times of decreasing velocities (Humphrey and Raymond, 1994). This has also been observed at the surge termination of the Kolka Glacier (Krenke and Rototayev, 1973) and the West Fork Glacier (Harrison and others, 1994). Kamb's (1987) model of a switch from a drainage system consisting of R-channels to a linked cavity system can account for these observations. It assumes that the glacier is underlain mainly by clean bedrock, however, most likely a significant oversimplification (Harrison and others, 1986; Humphrey and Raymond, 1994; Truffer and others, 1999).

We suggest that a switch in the drainage system is not the cause but rather an effect of fast glacier motion. Surges could be initiated by the large scale mobilization of subglacial till. This would happen as basal shear stresses exceed a critical value (Boulton and Jones, 1979; Clarke and others, 1984). Increasing shear stresses have been observed on Variegated Glacier prior to the 1982/83 surge (Raymond and Harrison, 1988). An efficient drainage system, such as one consisting of R-channels, will presumably be destroyed by high basal motion, causing a switch to a slowly draining system, such as the distributed 'canals' suggested by Walder and Fowler (1986), with a substantial amount of water storage and high basal water pressures. This makes sustained fast motion possible. The sudden release of the stored water will cause an abrupt slow down or an end to the surge. The interaction between the subglacial hydraulic system and basal motion in the context of till mechanics is not understood well enough to elaborate further.

The mechanism suggested here is in accord with observations made on the surging Variegated Glacier, and it emphasises the role of a subglacial till layer. It establishes the presence of a highly erodible glacier bed as a necessary, but not sufficient condition for glacier surging, as has been proposed before (Wilbur, 1988; Hamilton and Dowdeswell, 1996). A glacier geometry that allows the build-up of high basal shear stresses is possibly a second necessary condition.

3.6 CONCLUSION

Results from surface motion measurements, borehole inclinometry, and tiltmeters installed in the till beneath Black Rapids Glacier reveal that between 50% and 70% of the observed surface motion occurs at depths in the till layer greater than two meters. This study was conducted in the most active area of the glacier, where the observed surface motion is large.

and where a 7 m thick subglacial till layer exists. It is possible that the basal motion is occurring on a system of discrete shear zones or at the till-bedrock interface. Shear zones at relatively large depths can be explained with a simple model of 'perfectly over-consolidated material' and diffusion of water pressure into the till.

The small-scale structure of the tiltmeter record can be tentatively explained, although the interpretation is much more difficult and involves more assumptions. At any rate, the majority of the observed tilting events are reversible, and the total measured tilt corresponds to only 13 cm of basal motion (within the 2 m sampled) over the 410 days of measurements, as compared to the 70 m observed at the surface.

The observation that the upper part of the till layer is not deforming substantially is difficult to reconcile with a viscous model of the till which would require decreasing deformation rates with depth. The heterogeneity of the subglacial till (Truffer and others, 1999) cautions against applying simple continuum mechanical concepts to the study of this material.

The results of this field study have implications for the interpretation of the geologic record, such as far-traveled till sheets and the pervasive deformation versus soft till debate on the southern Laurentide Ice Sheet.

Our observations suggest that the large-scale mobilization of subglacial sediments plays a dominant role in the surge mechanism. Such mobilization could occur as the glacier reaches a 'surge geometry' at which a critical basal shear stress is reached. Taking a broader view, the mechanisms of velocity variation on several time scales observed on Black Rapids (Heinrichs and others, 1996) and many other glaciers (Willis, 1995) also need to be examined within the framework of our observations, but to do so is beyond the scope of this paper.

This study, among many others, shows the importance of basal processes in the understanding of glacier dynamics. It could be argued that a knowledge of till rheology is at least as important as that of ice rheology, and that the till mass balance, perhaps conventionally considered within the realm of glacial geology and geomorphology, is as important to the dynamics of the glacier as is the more familiar surface mass balance. More specifically, our study shoes that it is important to consider the full extent of the till layer, and not just the uppermost few centimeters. This is a challenge to both field and theoretical studies.

3.7 ACKNOWLEDGEMENTS

We would like to thank Slawek Tulaczyk, then at the California Institute of Technology, for helping with the various soil tests, and Ronald F. Scott of the Division of Engineering and Applied Science, also at Caltech, for allowing us to use his soil testing equipment. Dr. Less Fruth and the management of Earth Technology Inc. in Irvine, California, made their triaxial testing laboratory available. Thanks to all the helpers in the field. We appreciate the valuable reviews by Neal Iverson, Kurt Cuffey and Garry Clarke who pointed us to some relevant references, and to Almut Iken who read an earlier version of the manuscript. This research was funded by grant OPP 9423477 of the National Science Foundation.

Chapter 4

Implications of till deformation on glacier dynamics

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ABSTRACT[1] The dynamics of glacier motion are governed, to a large extent, by the properties of the basal interface. In this paper we address the interaction of a glacier with a layer of till at its bed in an attempt to test whether our physical understanding of till is sufficient to explain general features of observations made on Black Rapids Glacier, Alaska. We also investigate whether or not a till layer leaves a clear surface-observable signature in the dynamics of the glacier. Towards this end we use a finite element ice-flow model with the basal boundary condition determined by a Coulomb failure criterion for the till layer. We show that simple 'till physics' can be used to describe secular (decadal), seasonal, and short term (hours to days) velocity variations. Expansion and contraction of an extensive till layer ($\sim$ 5 m thick) can possibly explain surface uplift events and dye tracing results. The data and our analysis, however have not resulted in the identification of a clear signature of the till layer in the surface dynamics of the glacier.

[1]This chapter will be submitted to the Journal of Glaciology

4.1 INTRODUCTION

4.1.1 Motivation

Observations on many glaciers suggest that till may be common beneath them, yet most models of glacier behavior are based on the physics of a sharp ice-bedrock interface. We know that till is present under parts of many valley glaciers and ice streams, and that it is often actively deforming. It is therefore important to put forward a glacier model that is based on a consistent picture of till physics. Towards this goal, we investigate whether our understanding of till is sufficient to explain at least the general features of surface observations. We also attempt to identify a possible signature of a till layer, that is, a characteristic surface observation that would allow a clear distinction between a glacier underlain by deforming till from one underlain by clean bedrock.

4.1.2 Previous work

Field studies of valley glaciers underlain by till have been concentrated on Trapridge Glacier and Storglaciären. Hooke and others (1997) and Fischer and others (1999) have shown that the coupling of ice and till becomes weaker during times of faster basal motion. However, we have not found this to be the case beneath Black Rapids Glacier, Alaska (Truffer and others, in press). On this glacier the differential motion across the ice-till interface could have been, at most, a small fraction of the required 30 to 40 m annual basal motion.

Modeling studies involving subglacial till have often been based on the assumption that basal till is deforming viscously (see Paterson, 1994 for many references). This is in direct contrast to deformation studies on small samples that indicate a Coulomb-plastic rheology (Kamb, 1991; Iverson and others, 1998; Tulaczyk and others, in press). These latter authors also offer some explanations of observed features in tiltmeter records in terms of an elastic-plastic till model. Clarke (1987) laid out a frame work to describe till in the context of soil mechanics. Our observations beneath Black Rapids Glacier support the laboratory studies, and provide additional evidence that, at least in some cases, till does not follow a viscous rheology (Truffer and others, in press). In this paper we will use a perfectly plastic rheology for till, consistent with our observations. We therefore assume that there is no till deformation if the basal shear stress is below the failure strength of the till, and the strength of the till is independent of the deformation rate. We will explore such a model in the context of glacier motion and hydrology.

A whole spectrum of basal conditions, ranging from several meter thick till layers to relatively clean ice-bedrock interfaces, are probably encountered in nature. Many of the observations discussed in this paper have previously been discussed in the context of a clean ice-bedrock interface. We are addressing the other end of the spectrum in this paper.

4.1.3 Black Rapids Glacier

Black Rapids Glacier is a 40 km long glacier in the central Alaska Range. It is a surge-type glacier that is presently in its quiescent phase. It has been investigated in detail since 1973, with measurements of mass balance and velocities (Heinrichs and others, 1996), hydrological studies (Cochran, 1995; Raymond and others, 1995), radio echo sounding (Gades, 1998), seismic studies (Nolan and Echelmeyer, 1999a and b), a drilling program (Truffer and others, 1999) and a till deformation study (Truffer and others, in press). The drilling program has revealed a 5 to 7 m thick till layer and the seismic study has suggested that it is at least locally abundant, the seismic signature being about 40,000 m^2 in extent. The seismic study also found that the properties of the till layer can change rapidly throughout the till layer. The deformation study showed that a majority of the measured surface motion occurs at the bed and that this basal motion is occurring at a depth of more than 2 m below the ice-till interface. Recent observations also include high resolution (four times a day) surface velocities and vertical motion at three poles along the glacier, measured with GPS methods.

Most of these studies were focused on the area of the drill sites, because large velocity variations have been observed there. Figure 2.1 shows a map of the glacier and the location of the boreholes. The names of the boreholes used in this paper are N1, Center, and S1, from north to south (Fig. 4.2). The depth at the center of the channel is 620 m.

4.2 FLOW THROUGH A GLACIER CROSS SECTION

In this section we will discuss the flow through a transverse section of the glacier simulated using a finite element (FE) model. The modeled transverse bed geometry was derived from a radio echo sounding profile at 15.5 km from the glacier head, close to drill sites. This model is used to calculate the glacier's response to changes in driving stress and subglacial water pressure.

4.2.1 Boundary conditions

The shear strength τ_s of a till layer along the transverse perimeter of the glacier channel is related to the effective pressure $\bar{\sigma}$ through a Coulomb failure criterion:

$$\tau_s = c_a + \tan\phi \cdot \bar{\sigma} \tag{4.1}$$

where ϕ is the angle of internal friction and c_a the apparent cohesion. The effective pressure $\bar{\sigma} = p_o - p_w$ is the difference between the overburden p_o and the water pressure p_w. The pore water pressure depends on the water pressure at the ice-till interface and changes within the till, as discussed below. The apparent cohesion c_a is neglected here. It is usually a negligible quantity, on one sample from Black Rapids Glacier we measured 1.3 kPa (Truffer and others, in press). The name 'apparent cohesion' results from the observation that Equation 4.1 is not strictly valid near zero effective pressure, at which the shear strength of a granular material is actually zero (Wood, 1990).

Equation 4.1 defines a relationship between the shear strength and the effective pressure. The material properties of the till are determined by the friction angle ϕ (and the cohesion). In the remainder of this paper we will call a till 'stiffer' or 'weaker' if its shear strength changes as a result of changes in effective pressure. No change in material properties is implied with this terminology.

The distribution of water pressure along the glacier bed is determined by the nature of the subglacial drainage system. Walder and Fowler (1994) have proposed that water drains through shallow 'canals' in the till. They obtained a direct relationship between water flux and pressure which favors a distributed drainage system. This relationship, however, is a direct consequence of their assumption that canals have a depth independent of their width, an assumption open to question (Hooke, 1998). Nevertheless, we expect a distributed drainage system in the study area, which is in the upper ablation area near the maximum retreat of the snowline. Drainage is expected to be more distributed in the upper regions of the ablation area than lower in the ablation area, where it should be more channelized (Fountain and Walder, 1998).

Data from boreholes show that water pressure can vary greatly over short distances (Hubbard and others, 1995; Murray and Clarke, 1995). Hubbard and others (1995) suggested that lower, but more variable pressures were measured along a 'preferential drainage axis', where drainage was dominated by flow through a subglacial channel. Higher and more constant pressures were found in adjacent areas, where flow was interpreted to occur through a distributed system.

No such drainage axis was inferred on Findelengletscher, where Iken and Bindschadler (1986) found very nearly equal levels of the water table in eleven connected boreholes, except that the water table dropped slightly towards the margins.

On Black Rapids Glacier we have water pressure records from two boreholes. Center and N1 (Fig. 4.1) of more than 300 days. The sensor at N1 was measuring pore water pressure in the uppermost part of the till, while the sensor at Center was located a few meters above the glacier bed and could have been affected by borehole closure. Water levels are generally higher at Center. A shorter record from a borehole farther south (S1) shows water levels closer to those measured at Center. The differences are relatively small, and as a first model we assume a constant depth to the water table across the glacier surface. Our model requires that the water table be between the levels measured at N1 and Center, as shown below.

We assume that changes in basal water pressure affect the bed evenly. While many measurements show this to be incorrect (as mentioned above), it is still a reasonable assumption when considering changes that happen over time scales that are longer than a few days. We will therefore not use this assumption when discussing velocity variations on shorter time scales.

The assumption of a constant depth to the water table across the section (Fig. 4.2) leads to an effective pressure of

$$\bar{\sigma}(x) = p_o - p_w = \rho_{\text{ice}} g h(x) - \rho_w g(h(x) - H) \tag{4.2}$$

where $h(x)$ is the ice thickness as a function of the transverse coordinate x, and H is the depth of the water table below the glacier surface. The glacier surface was at a nearly constant elevation across the transverse section.

Reynaud (1973) used a finite difference model with the same type of Coulomb friction law

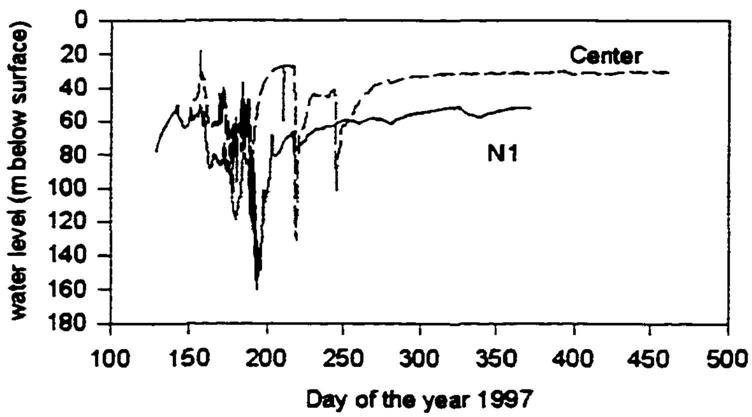

Figure 4.1: Water pressure. Water pressure in holes N1 (solid line) and Center (dotted line)

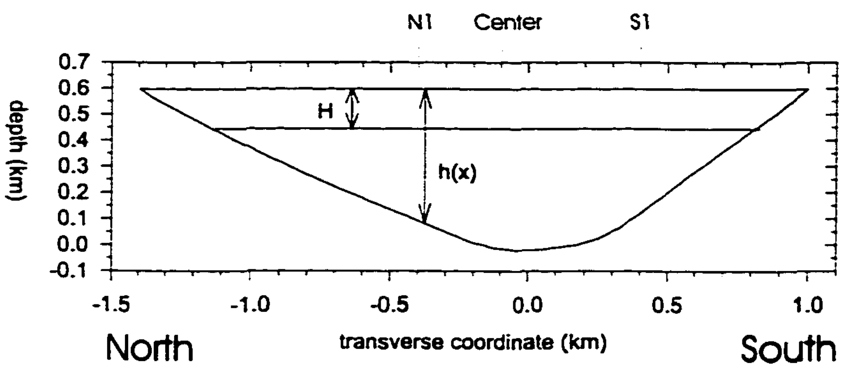

Figure 4.2: Cross section of Black Rapids Glacier used in the FE study. H is the depth of a hypothetical water table below the glacier surface. The location of the boreholes is also indicated.

(Eqn. 4.1) for the boundary condition in a flow model of a transverse section of Athabasca Glacier. He specified this boundary condition as an ad hoc solid friction law in an attempt to reproduce the flow measured by Raymond (1971) in this section. We approach it from a much different direction, using the knowledge that till is present under some glaciers and that it deforms as a Coulomb plastic material. Some of Reynaud's (1973) results are, however, directly applicable to our till-based model.

4.2.2 The finite element method

The finite element (FE) code developed by Echelmeyer (1983) is employed for modeling power law creep. This model allows the calculation of all three components of the velocity field in a plane section, but is restricted to cases where out-of-plane stress and deformation gradients are zero. The code has been extensively tested against analytical solutions and against Nye's (1965) numerical results (Echelmeyer, 1983), and these tests have been repeated by different users.

This FE code was used to calculate the out-of-plane velocities for a cross section of Black Rapids Glacier, restricting the in-plane velocities to be zero, that is no transverse or vertical velocities ($\mathbf{u} = (0, 0, u)$). A power law rheology was used with an exponent $n = 3$ and flow parameter $A = 6.79 \cdot 10^{-24}\,\mathrm{Pa^{-3}s^{-1}}$ (Paterson, 1994). The glacier geometry was derived from radio echo sounding (Gades, 1998 and Gades, pers. comm.). Figure 4.3 shows the grid used. The till layer was simulated by adding a thin layer of elements to the lower boundary of the FE grid. The bottom nodes of these boundary elements were kept at zero velocity. In a first run the boundary elements were kept several orders of magnitude stiffer than ice. The resulting bed-parallel shear stress at the bottom of the ice was then compared to the calculated failure stress of the till (Eqn. 4.1). If the shear stress exceeded failure, the corresponding boundary elements were made softer. Iteration continued until the shear stress was at or below failure along the entire perimeter of the glacier channel. The stiff elements at places where the basal stress was below the failure strength resulted in zero basal motion. This then effectively made a mixed boundary condition, a Neumann condition (u=0) along parts of the bed (the 'margins') and a Dirichlet condition ($\frac{\partial u}{\partial n} = 2\mathcal{A}\tau_s^3$) along the remainder of the bed.

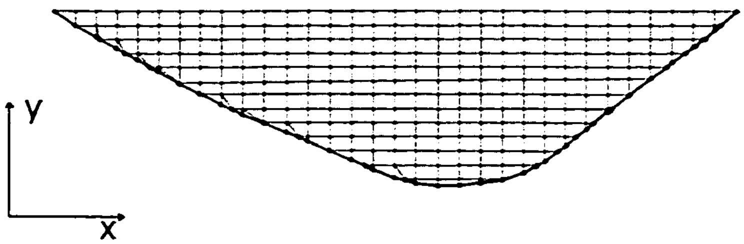

Figure 4.3: Finite element grid. The thin layer of boundary elements along the glacier bed cannot be seen clearly at this resolution. z-direction is into the plane.

4.2.3 Calculated flow through a cross section

Base model

The FE model was run for different values of the water table H, friction angle ϕ and the surface slope α. The goal was to choose a best-fit model that would serve as a base to calculate the effects of different forcings, such as changes in water table or driving stress. This 'base' model was chosen to match the measured mean annual surface velocity close to the center of the glacier, which was 60 $\mathrm{m\,a^{-1}}$ in 1997/98, and the measured deformation at the bottom of the N1 borehole. The best fit was obtained for a surface slope of 1.40° (or 0.024), a water table of 55 m below the surface and a friction angle of 30°. Figure 4.4 shows a contour plot of velocity for the base model. Calculated ice deformation in the N1 borehole is shown in Figure 4.5 which also includes measured deformation rates at the bottom of the hole. Meaningful inclinometer measurements could only be made in the lowest 100 m of the borehole (Truffer and others, in press). Agreement between measured and calculated centerline speed, and between measured and calculated borehole deformation is ok.

The measured surface slope is somewhat larger than the one from our base model. Heinrichs (1994) reports about 1.5° and GPS measurements in spring 1999 gave an average slope of 1.75° over a distance of 4 km. The lower value used here reflects to some degree the compressive longitudinal stresses that act to reduce flow velocities. These stresses are attributed to the inflow of a major tributary (Locket tributary) 7 km downstream and show up as a decrease in speed as the tributary is approached from above.

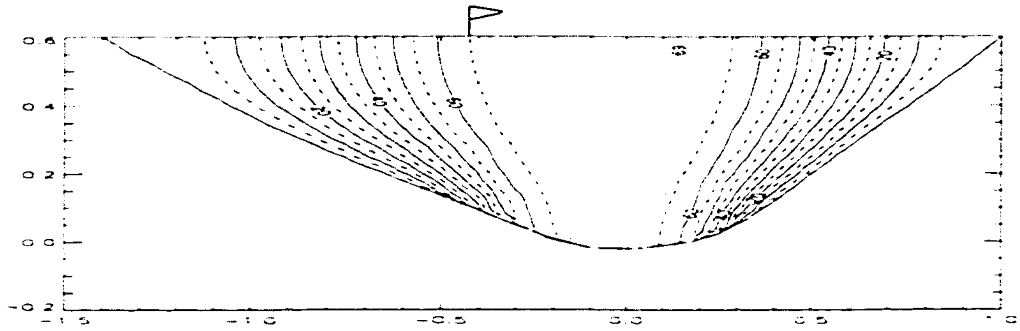

Figure 4.4: Velocity contour plot. Velocity (in $\mathrm{m\,a^{-1}}$) contour plot for $\alpha = 1.40°$, H=55 m and $\phi = 30°$. The flag shows the approximate position of the N1 borehole. Dashed contours are at 5 $\mathrm{m\,a^{-1}}$) increments, dotted at 1 $\mathrm{m\,a^{-1}}$ increments from 65 to 69 $\mathrm{m\,a^{-1}}$. The units on the axis are km.

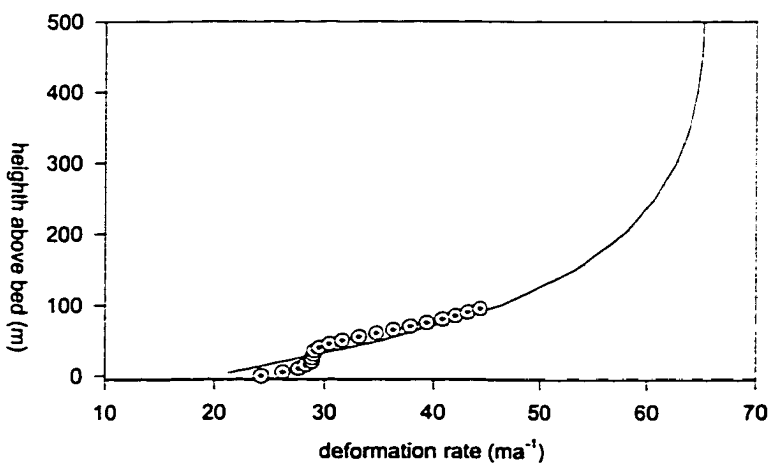

Figure 4.5: Calculated ice deformation at the N1 location. The dots show measured values at the bottom of the borehole.

Our best-fit value of 55 m for the depth to the water table below the surface is about 5 m higher than that measured during winter at N1, and it is about 17 m lower than that measured at Center (Fig. 4.1). This places our best-fit water depth in between the two measured winter-time average values. Note that we have not investigated the effect of a water table that is at a variable depth across the glacier surface, because our water pressure coverage was so sparse.

The best-fit friction angle was 30°. This is somewhat lower than the 40° measured in the lab on a sample of Black Rapids till (Truffer and others, 1999), but is somewhat higher than the 27° that has been reported as typical for valley glacier tills (Iverson and others, 1998). It is noteworthy, however, that the difference in surface velocities resulting from variations of ϕ within these values is small, as discussed below.

Figure 4.6 shows a transverse profile across the glacier surface. A velocity profile measured in late winter 1993 was scaled to the mean annual speed at the center (Fig 4.6). This scaling is not inappropriate, because the shape of the transverse velocity profile does not change through periods of varying velocity (Fig. 4.7).

The measured and the calculated transverse velocity profiles diverge significantly, especially at the south side. This is partly due to a tributary that enters the glacier at that location from the south. Another tributary enters the glacier upstream of the study area from the north. The model does not account for this. However, we are only interested in general features of the flow, and in the response to different changes. For this purpose our base model is sufficient. We will now discuss the values of these model parameters.

4.2.4 Sensitivity

Table 4.1 shows the percentage change of the centerline velocity in response to variations in surface slope, depth of water table and friction angle.

The large sensitivity of the surface velocity to changes in slope is due to the low value of the overall surface slope. The flow law exponent $n = 3$ results in a percentage velocity response that is about three times as large as the percentage change in driving stress ($\sim$7% in table 4.1). This is borne out in the model. The response is actually slightly larger than that because an increase in driving stress also results in some additional till failure.

The response of the surface velocity to changes in depth to water table will be discussed

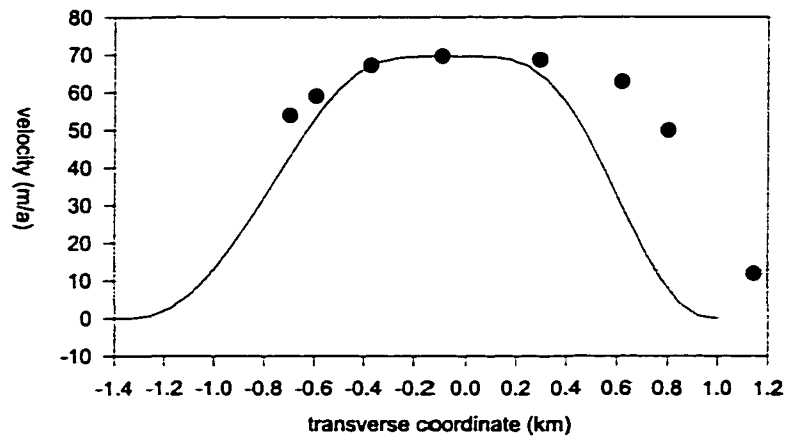

Figure 4.6: Transverse velocity profile. Modeled transverse profile of velocities is shown by the solid line. The dots are measurements from summer 1993 scaled to the modeled centerline velocity.

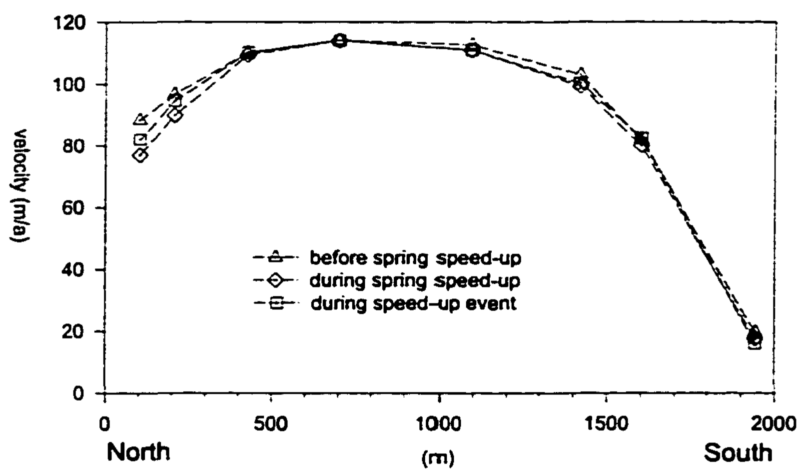

Figure 4.7: Measured scaled transverse velocity profiles. Scaled transverse profiles measured in spring 1993, before and during the spring speed-up and during a lake drainage event. The shape of the profiles does not change significantly. Profiles are scaled to the centerline velocity of the spring speed-up event.

surface slope α	+0.1°	+23%
$\alpha \sim 1.4°$	-0.1°	-22%
depth of water table H	-5 m	+18%
$H \sim 55$m	+5 m	-24%
friction angle ϕ	-5°	+3%
$\phi \sim 30°$	+5°	-4%

Table 4.1: Sensitivity of modeled velocity. Sensitivity of the maximum surface velocity (right column) to changes in surface slope. depth of water table and friction angle.

below as a possible explanation for seasonal velocity changes.

Table 4.1 shows that the friction angle is not a critical parameter. It is possible, however. that deformation is occurring on clay-rich shear layers, which would have a considerably smaller friction angle. This possibility is not further investigated here.

We arbitrarily defined a 'plug' flow width parameter as the transverse distance over which the velocity is within 10% the centerline velocity. The sensitivity analysis showed no appreciable changes in the plug width occur due to changes in any of the parameters discussed above. It was therefore not possible to match the measured plug width within the parameter limits of this model (Fig. 4.6). A more variable plug width could possibly be obtained by lifting the restriction of a constant water table or by relaxing the no-slip boundary condition at the glacier margin. However, this was not investigated.

4.3 VELOCITY VARIATIONS

In this section we will discuss velocity variations at time scales of decades, years, and hours to days. At each scale, we will present observational data from Black Rapids Glacier and then suggest possible mechanisms in terms of simple till physics, using the model velocities calculated above.

4.3.1 Secular variations

The positions of several stakes on Black Rapids Glacier have been determined once or twice a year since 1973. Figure 4.8 shows the resulting mean annual velocity at 14 km from the glacier head, some two kilometers upstream from the study area. It is an updated plot from Heinrichs and others (1996). The mean annual velocity did not steadily increase during the quiescent phase, as it has done on the archetypal surge-type Variegated Glacier (Raymond and Harrison, 1988). Instead, the velocity went through an eleven year cycle with changes up to 40%. The current trend is one of decreasing speed. This secular variation is related to changes in driving stress, but the magnitude of it is too large to be explained by changes in deformational speed only (Heinrichs and others, 1996). Other markers at 8, 11, and 20 km show similar changes, the one in the accumulation area (8 km) with a significantly reduced amplitude. There is, of course, no evidence that these velocity variations will continue in a cyclical fashion.

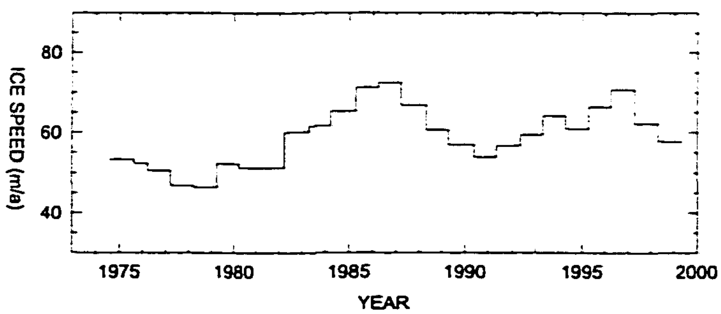

Figure 4.8: Secular velocity changes. Mean annual velocities at a site close to the drilling area. This is an update of the plot from Heinrichs and others (1996).

Figure 4.9 shows the calculated centerline velocity as a function of depth of water table for different values of the surface slope. It demonstrates that an increase in surface slope, and thus driving stress, has a larger effect when some of the till is at or near failure. When the water table is at a depth of 85 m, then none of the till is failing and the flow is purely deformational. An increase in surface slope from 1.4° to 1.6 ° results in a velocity increase of 16 m a^{-1}. The same increase at a depth of water table of 55 m results in a velocity increase of 36 m a^{-1}. A large velocity response can thus occur — at the surface and at the bed — due to geometry variations, without changes the subglacial hydrology. This is because none of the additional stress can be accommodated in areas where the till is already at failure, which causes the additional stress to be transferred to the sides, increasing the deformation rates there. These ideas are discussed in more detail in section 4.5.1.

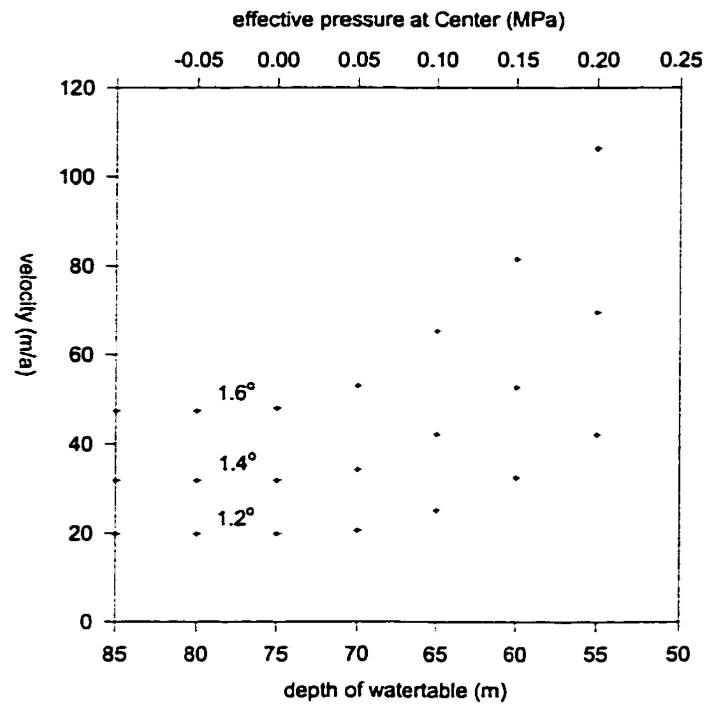

Figure 4.9: Velocity vs. depth of water table. Surface velocity is plotted as a function of depth of water table (or effective pressure at Center). The curves are drawn for different surface slopes.

4.3.2 Seasonal variations

The surface velocity of many temperate glaciers varies seasonally (Willis, 1995). On Black Rapids Glaicer the seasonal pattern was measured at several locations on the glacier by using time lapse photography, following methods of Harrison and others (1992). The step plot in Figure 4.10 was obtained by averaging all the available monthly velocity averages of a pole about 2 km upstream from the study area to obtain a 'typical' seasonal variation. Similar variations were observed 6 km downglacier and — at a reduced amplitude — 6 km upglacier. The camera was in operation intermittently between 1986 and 1996, but for each individual month, only four to six different monthly average speeds exist.

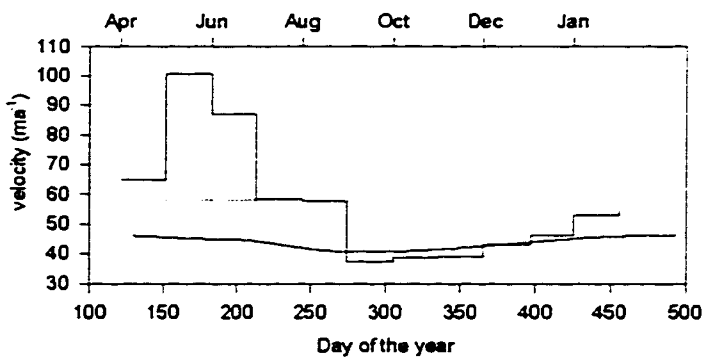

Figure 4.10: Calculated basal motion. The step plot is a compilation of camera data at a site 2 km upglacier from the study area. The dotted line results if speed-up events are left out of the record, as explained in the text. The smooth curve shows velocities calculated from modeled water pressures at 4 m below the top of the till.

Typically, velocity increases in spring and early summer, as melt and rain water enters the glacier's hydraulic system and reaches the glacier bed. In late summer the speed decreases, and it reaches a minimum in early winter. Thereafter the speed slowly increases throughout the rest of the winter. The high average monthly velocities in early summer are caused by a series of short-lived but high amplitude speed-up events, such as those shown in Figure 4.11. These events are responsible for about 10% of the annual glacier motion, as determined from GPS measurements in early summer 1997. We will discuss them separately in the following section, and first address the 'low frequency' behavior, or the seasonal velocity variation without the speed-up events (dotted line in Fig. 4.10).

In Truffer and others (in press) we presented a model of a perfectly over-consolidated till (Clarke, 1987) that predicted till failure 4 m below the ice-till interface. If failure is indeed occurring at that depth, then the failure strength of the till τ_s should vary with water pressure, p_w, at that same depth in the till (Eqn. 4.1). This variation in strength will then result in changing surface velocities as indicated in Figure 4.9.

To investigate the effects of changing water pressure at depth in the till following changes in p_w at the ice-till interface, we numerically solved the one-dimensional diffusion equation (de Marsily, 1986)

$$\frac{\partial p_w}{\partial t} = C_v \frac{\partial^2 p_w}{\partial z^2} \tag{4.3}$$

where p_w is the water pressure, z the vertical coordinate positive downward, and t the time. C_v is the hydraulic diffusivity, which was taken to be $3 \cdot 10^{-7}\,\mathrm{m}^2\,\mathrm{s}^{-1}$ (Truffer and others, in press). We assumed a hydrostatic pressure gradient at the bottom of the 7 m thick till layer. At the top of the till (ice-till interface) we applied the measured water pressure at N1 (Fig. 4.1) shifted up by 0.1 MPa, or 10 m of water level, as discussed in section 4.2.3 and 4.2.1. Following our conceptual model of water input and pressure, we assumed that p_w was a harmonic function of time with a periodicity of one year. Peak water pressures at the ice-till interface occur in spring and early summer when the drainage system occasionally gets overloaded. Later in the summer, as a good drainage system is developed, the annual minimum is reached. Water pressure stays high and varies little through late fall and winter.

The assumptions in this model are that water pressure in the till is governed by a purely diffusional process, hydraulic properties are isotropic and constant with depth in the till and laterally, and that the water pressure is uniform over large horizontal distances compared to the thickness of the till layer. This allows us to use the one-dimensional model given by Equation 4.3.

The water pressures obtained at 4 m depth in the till were then converted to depth to the water table, and a time-varying effective pressure was calculated. The corresponding surface velocity was obtained by using a quadratic fit to the results shown in Figure 4.9 for depth to water table versus calculated surface velocity. A surface slope of 1.4° was used.

The resulting velocity curve is shown as the smoothly-varying solid line in Figure 4.10. The seasonal velocity curve (without the speed-up events) has a larger amplitude than the modeled curve, but the timing of the velocity minima in late fall coincide. Deformation deep in a till layer together with diffusion of water pressure in the till is thus a reasonable mechanism to explain the late fall velocity minimum, at a time when water pressure at the bottom of the ice is generally high (Fig. 4.1).

4.3.3 Speed-up events

In late May or early June Black Rapids Glacier typically goes through a spring speed-up event during which velocities increase by about a factor of three (Nolan and Echelmeyer. 1999a; Truffer and others, in press; Fig. 4.13). In subsequent weeks, several other speed-up events occur, such as the one shown in Figure 4.11. Our observations indicate that they are usually — if not always — a result of enhanced water input, such as the drainage of a marginal lake or a pothole, or a long period of rainfall.

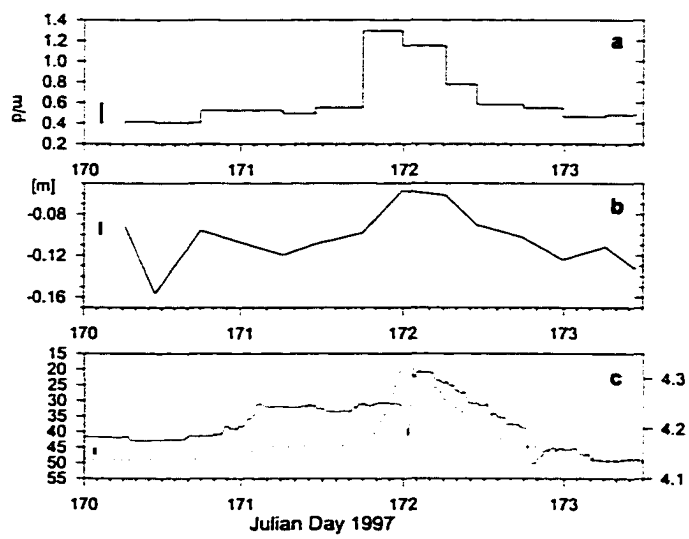

Figure 4.11: Lake drainage event. A speed-up event caused by the drainage of a marginal lake. a) velocity measured with GPS methods, b) vertical position minus the downhill movement, also measured with GPS methods, and c) water pressures measured in boreholes N1 (solid line) and S1 (dotted line). The bars at the left show the respective errors.

A seismic study on Black Rapids Glacier has shown that a till layer of at least 5 m thickness became seismically transparent during a speed-up event (Nolan and Echelmeyer, 1999a,b). These authors attributed this to a temporary change from a fully water saturated till to a saturation of about 95%, caused by a temporary decrease in local overburden pressure. This decrease occurred because a marginal lake drained, over-pressuring the drainage system and the glacier was locally lifted up, reducing the overburden pressure in adjacent areas through an elastic bridging effect.

This scenario has the following effects on the glacier-till system. In areas where the subglacial drainage system is over-pressurized the ice and till are completely de-coupled. In adjacent areas the overburden pressure decreases and thus the effective stress decreases, reducing the local shear strength of the till (Eqn. 4.1). Because the driving stress still has to be balanced at the 'wetted' perimeter, basal shear drag is transferred to other areas, such as the sides of the glacier, or possibly areas up- or downstream.

Thus, a possible scenario for the speed-up is the following: The central part of the channel is subject to a decreased shear strength of the till (as p_o and therefore $\bar{\sigma}$ decreases), transferring drag to the margins. This increased marginal shear stress leads to enhanced ice-flow through the non-linear rheology, and thus higher velocities (see 4.5.1). The weaker central part is due to a loss of ice-till coupling in places where the water drains, and it is due to lower overburden pressure and thus strength in adjacent areas. This elastic bridging effect can only last for relatively short times. The sudden change in overburden pressure is reflected in the sudden change in seismic properties (Nolan and Echelmeyer, 1999a,b), which revert to 'normal' conditions after a period of hours.

4.4 TILL VOLUME CHANGES

A granular material will react to changes in effective pressure by expanding or contracting, and storing or releasing of water. These till thickness changes, if large enough, should be noticeable at the glacier's surface. In the following sections we will test whether till expansion can be related to upward motion (uplift events) and discuss dye tracing results from Black Rapids Glacier in this context.

4.4.1 Uplift events

Speed-up events on Black Rapids Glacier are often accompanied by periods of upward vertical motion (Fig. 4.11). The magnitude of these events is on the order of 10 cm of uplift. Vertical motion events have been reported from many other glaciers as well (see summary by Willis, 1995).

The amount of expansion Δz of an element of length L_z is determined by the compressibility coefficient α_v:

$$\Delta z = -\alpha_v L_z \Delta\bar{\sigma} \tag{4.4}$$

where $\Delta\bar{\sigma}$ is the change in effective pressure (de Marsily, 1986). Typical values of compressibility for a till are between $10^{-6}\,\mathrm{Pa}^{-1}$ for a clay-rich till to $10^{-9}\,\mathrm{Pa}^{-1}$ for a sandy till (de Marsily, 1986). If the till layer is divided into m layers of height $(L_z)_i$, the change in thickness of the entire layer (of thickness $L(t)$) in a time step Δt is an integral approximated by the following sum

$$\Delta L(t) = \sum_{i=1}^{z=m} (\Delta z)_i(t) = -\alpha_v \sum_{1}^{m} (L_z)_i(t) \Delta\bar{\sigma}(z_i, t) \tag{4.5}$$

where $(\Delta z)_i(t)$ is the thickness change at depth z_i. To obtain the actual till thickness as a function of time, $\Delta L(t)$ has to be integrated over time.

To calculate the till thickness changes, $\bar{\sigma}(z, t)$ and hence $p_w(z, t)$ has to be known. We used the same diffusion model described above (4.3.2) and the water pressure measured in the Center borehole (Fig. 4.1) as a boundary condition at the ice-till interface, and calculated the water pressure distribution in time and depth of till. The calculated variation of till thickness is shown in Figure 4.12 and compared to the measured vertical position of a surface marker with the downslope component removed. This was obtained with GPS methods and measured in the vicinity of the Center borehole. Note that the variation of till thickness is only shown for the short period of time when vertical motion was recorded, although it was calculated for the entire year.

The uplift events associated with lake drainages could potentially be explained by the expansion of a till layer, mostly of its uppermost part, because the water pressure variations do not diffuse far into the till. However, this requires a compressibility of $10^{-6}\,\mathrm{Pa}^{-1}$, which

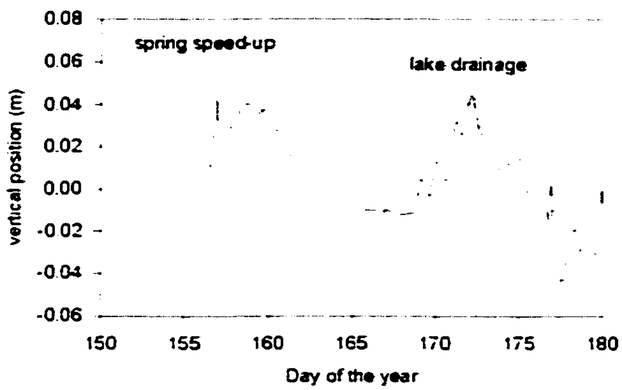

Figure 4.12: Expansion of the till layer. De-trended vertical position (solid line) of a pole at the center of the channel and calculated expansion and contraction of the underlying till layer (dotted line). A compressibility of $\alpha_v = 10^{-6}\,\mathrm{Pa}^{-1}$ was used.

is high considering the granular nature of recovered till (Truffer and others, 1999). The spring speed-up event could not be matched with the till expansion model. This might be an indication of its different nature: While lake drainage events happen a short distance upstream from the study site, the spring speed-up is triggered downglacier and propagates up the glacier (Fig. 4.13).

The total amount of till expansion is proportional to the compressibility of the till (Eqn. 4.5). The range of probable compressiblities spans three orders of magnitude. A high compressibility could account for the entire vertical motion of a lake drainage speed-up event, whereas the lower end of the plausible range would render till expansion negligible.

We discussed the possible drastic reduction in overburden pressure in some areas (section 4.3.3) . This would also result in a reduction of effective pressure and therefore in till expansion. For a 5 m thick till layer (found at Center) and a very low compressibility ($\alpha_v = 10^{-9}\,\mathrm{Pa}^{-1}$, a temporary reduction in local overburden of 2 MPa would result in 1.0 cm of instantaneous till expansion.

Uplift events could also be caused by vertical strains. The question is difficult to settle, because vertical strain rates can change with depth in magnitude and sign (Balise and Raymond, 1985; Gudmundsson, 1997; Gudmundsson and others, 1997). Longitudinal strain

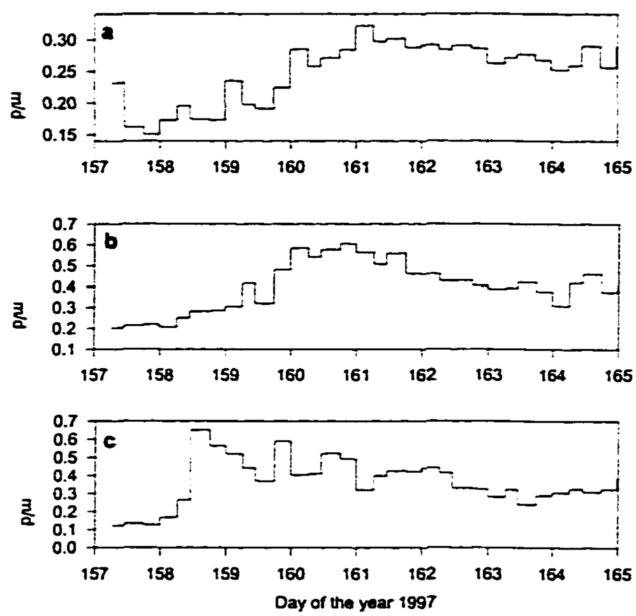

Figure 4.13: Spring speed-up. The spring speed-up at 13 km, 16 km, and 22 km from the glacier head. The event travels upglacier. 16 km is close to the location of the drill site. Velocities were measured with GPS methods.

rates measured at the surface of Black Rapids Glacier were an order of magnitude too small to account for the observed uplift. This, together with more detailed measurements on other glaciers (Iken and others, 1996), leads us to conclude that vertical straining of the ice is unlikely to explain the entire observed vertical motion.

4.4.2 Dye tracing

Dye was introduced into a draining lake during a drainage event on Black Rapids Glacier in spring 1993. Cochran (1995) sampled the proglacial stream for dye concentration. She found an initial peak in dye return, but only a small fraction of the total dye input was output during that event. Several other peaks in dye concentration followed in the next three weeks, always after another speed-up/lake drainage event had occurred.

The model of till expansion and contraction presented above offers the following explanation for this observation: As water pressures rise (and the glacier speeds up), till is expanding and storing water, and with it some dye. When the till contracts again, water and with it some dye is released and drains into a channel. This mechanism of water storage in the till could probably be substantiated with more detailed measurements in the proglacial stream, such as electric conductivity or a chemical analysis that would show extensive contact with till.

4.5 DISCUSSION

We will first discuss the transfer of shear stress along the glacier perimeter, a crucial idea in this paper. In a second section we will shortly discuss how similar surface measurements have been explained in an ice-bedrock model.

4.5.1 Transfer of basal shear stress

We have put forward a model of glacier-till interaction based on a simple Coulomb-plastic failure criterion for till, defined broadly as unlithified granular material. This allows us to explain velocity variations on different time scales. In our model all of these variations are caused by increasing bed-parallel shear stresses at the glacier margin, causing larger deformation rates. Three mechanisms for such a stress increase have been identified: (a) an increase in driving stress, (b) an increase in water pressure, and (c) short term changes in local overburden pressure. They are illustrated in Figure 4.14 and explained in the following.

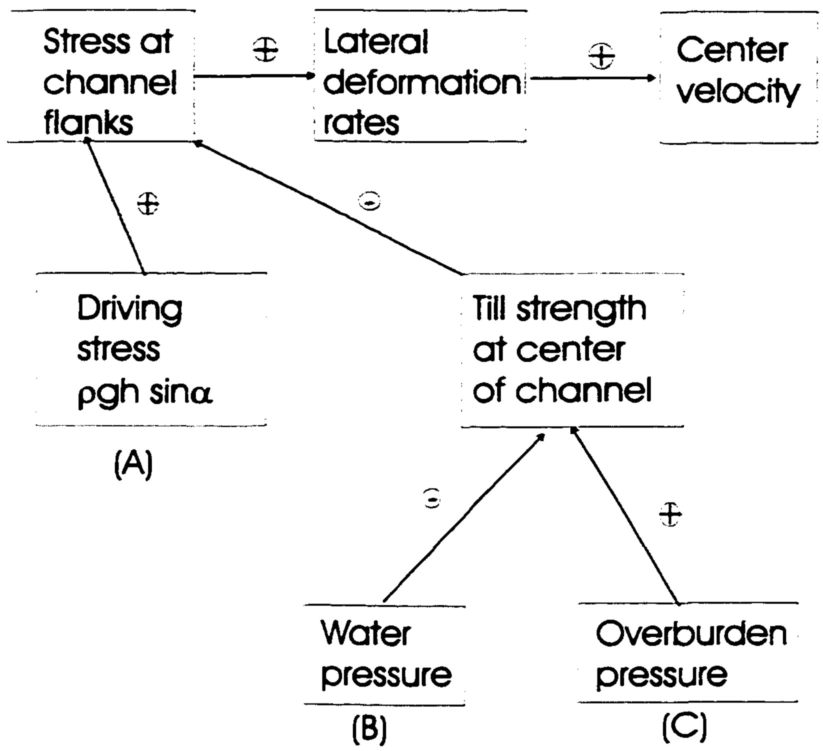

Figure 4.14: Diagram of stress re-distribution. The diagram shows stress transfer mechanism as explained in the text. '+' signs indicate a positive correlation and '-' signs a negative correlation, e.g. an increase in water pressure leads to a decrease in till strength at channel center. (A), (B) and (C) correspond to the three causes for velocity changes, as described in the text.

A characteristic property of a perfectly plastic material is that it cannot accommodate any stress above its failure strength. If a perfectly plastic till is at failure under parts of the glacier and the driving stress is increased, the additional stress has to be accommodated at other parts of the bed, such as the glacier margins. We propose that velocity changes on decadal time scales are caused by such variations in driving stress. The extreme case of this scenario is if marginal stresses get large enough to exceed the breaking strength of the ice. The glacier would presumably shear off its margin and a surge could be initiated.

We also propose a mechanism for seasonal velocity variations. In Truffer and others (in press) we showed that till deformation is happening several meters below the ice-till interface. If the water pressure distribution in the till is governed by diffusion, then a few meters below the top of the till it will by dominated by a cycle with a period of one year and a minimum in early winter. This water pressure cycle causes a cycle of till strength variation. If water pressure rises, the till gets weaker. Some of the driving stress is then transferred from the locations of the weaker till to the margins, or possibly some combination of the margins and any remaining 'sticky spots'. This will again cause higher marginal deformation rate and therefore a higher centerline velocity. Figure 4.15 shows the stress redistribution and the resulting velocity contours when the depth to water table is raised from 70 m to 55 m. The stresses drop in the center and rise at the flanks of the channel.

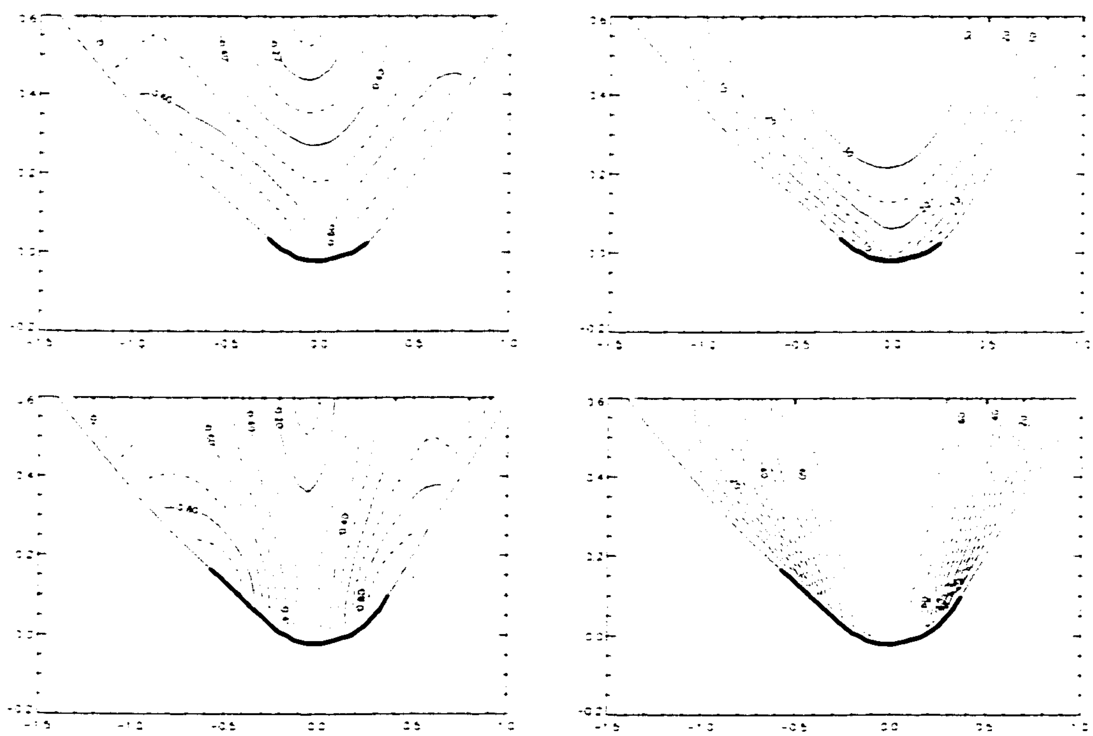

Figure 4.15: Stress re-distribution. The top panels show the stress (left) and velocity (right) contour lines for a depth to water table of 70 m. The bottom panels show the same contours for a depth to water table of 55 m. The bold line indicates the area of till at failure. The units on the axes are kilometers. Velocities are in $\mathrm{m\,a^{-1}}$ and stresses in bar (1 bar = 100 kPa).

Velocities also vary on time scales of hours or days. We know from a seismic study (Nolan and Echelmeyer, 1999a,b) that the till layer undergoes rapid changes in seismic properties. Following these authors, we suggest that the glacier is locally lifted up because parts of the drainage system are over-pressurized. This reduces the overburden pressure in adjacent areas through viscoelastic bridging effects and makes the till weaker. Again, stress is transferred to the sides, causing larger deformation gradients there and higher surface velocities.

Traditional analysis has treated velocity variations by considering deformational and basal motion to be linearly independent conditions to overall motion, that is $u_{\text{total}} = u_{\text{def}} + u_{\text{basal}}$. Our analysis shows that this is not correct. Changing basal stress conditions (both $\bar{sigma}$ and τ_s) can produce greater basal motion at the center of the channel and — at the same time — higher deformational velocities at the margins. This situation is analogous to that encountered at the well-lubricated Siple Coast ice streams in West Antarctica, where a very weak basal till layer causes a transfer of the overall resistive drag to the ice stream margins (e.g. Echelmeyer and others, 1994).

4.5.2 Till versus bedrock

Observations of fluctuations in surface speed and elevation similar to the ones discussed in this paper have been made on many other glaciers (e.g. Willis, 1995). Some of these glaciers probably have a relatively clean ice-bedrock interface, while others may have a substantial till layer at their base. While we have focused on the explanation of these phenomena in terms of changes within the till layer found beneath one section of Black Rapids Glacier, it is also possible to explain some of them in terms of a model in which clean glacier ice overlies bedrock. Such explanations have been developed in the past, and we discuss a few of them here.

Secular velocity variations have been explained by long term changes in the subglacial drainage system, such as the inter-connectedness of cavities (Iken and Truffer, 1997). Such changes could occur as a reaction to geometry changes. Kamb (1987) has introduced a stability parameter that describes the stability of a distributed versus an arborescent drainage system. This parameter depends — among many other things — on the basal shear stress. Geometry changes could therefore trigger a switch in drainage systems and with it variations in basal velocities. He proposed this as a possible mechanism for surge initiation on Variegated Glacier. It should be noted, however, that borehole drilling in that glacier indicated that the ice-bedrock interface was not clean of debris (Harrison and others, 1986).

Hodge (1974) related water storage to seasonal velocity changes. Kamb and others (1994) also found that water storage was a good control variable for basal motion on shorter time scales, such as the speed-up events on Columbia Glacier. This has been explained in terms of bed separation and cavity formation (Lliboutry, 1968). Cavity formation and water storage at the glacier bed is commonly assumed to be responsible for uplift events (Iken and others, 1983, 1996), although the contributing effects of vertical strain can be difficult to estimate from surface measurements alone (Balise and Raymond, 1985; Gudmundsson, 1997; Gudmundsson and others, 1997).

This shows that similar observations find explanations in an ice-bedrock as well as an ice-till model. We have not succeeded in identifying a characteristic signature of a basal till layer in surface measurements. As far as the velocity distribution in a transverse section is concerned, this can already be recognized by comparing modeling results of Reynaud (1973) and Harbor (1992). They used a friction law and a sliding law, respectively, and obtained similar results to ours. One therefore needs to exercise care when making inferences about the ice substrate from glacier dynamics.

4.6 CONCLUSION

Black Rapids Glacier and possibly many other glaciers are underlain by till. We have shown that treatment of till as a Coulomb-plastic material can, at least in principle, explain many of the glacier dynamics features observed on Black Rapids Glacier and elsewhere. These include velocity variations on time scales of decades, years, and hours to days.

We have also calculated the expansion and contraction of a till layer in response to water and overburden pressure variations. This expansion can possibly explain observed vertical motion events, although the uncertainties of the in-situ till compressibility are large. Water storage and release in response to this expansion and contraction can explain dye tracing results obtained in an earlier study on Black Rapids Glacier (Cochran, 1995).

We have not managed to identify a clear surface-observable signature of a till layer in the dynamics of a glacier. Many surface observations seem to find plausible explanations in an ice-till model, but they can also be explained in terms of an ice-bedrock model with a changing hydraulic system. An appealing feature of our glacier-till model is that glacier velocity can change appreciably without requiring dramatic changes in the subglacial hydraulic system. Such changes are often difficult to explain.

Glacier velocity changes will always be a result of changing basal stress conditions, such as those outlined in section 4.5.1. This is independent of the model for the glacier substrate. It is therefore not legitimate to treat deformational and basal motion as independent variables.

4.7 ACKNOWLEDGEMENTS

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