Magnetic Till Fabric: Applications of anisotropy of magnetic susceptibility (AMS) to subglacial deformation of till and ice

Nathan Russell Hopkins
Lehigh University

Follow this and additional works at: http://preserve.lehigh.edu/etd
Part of the Environmental Sciences Commons

Recommended Citation
Hopkins, Nathan Russell, "Magnetic Till Fabric: Applications of anisotropy of magnetic susceptibility (AMS) to subglacial deformation of till and ice" (2016). Theses and Dissertations. 2638.
http://preserve.lehigh.edu/etd/2638
Magnetic Till Fabric: Applications of anisotropy of magnetic susceptibility (AMS) to subglacial deformation of till and ice

1. Till kinematics and the origins of drumlins, New York State
2. Till kinematics of the Baltic Ice Stream, Sweden
3. Characterization of deformation within the Basal Stratified Ice of the Matanuska, Glacier, Alaska

by

Nathan Russell Hopkins

A Dissertation
Presented to the Graduate and Research Committee
of Lehigh University
in Candidacy for the Degree of
Doctor of Philosophy

in
Earth and Environmental Science

Lehigh University
January, 2016
Approved and recommended for acceptance as a dissertation in partial fulfillment of
the requirements for the degree of Doctor of Philosophy

Nathan Russell Hopkins
Magnetic Till Fabric: Applications of anisotropy of magnetic susceptibility (AMS) to
subglacial deformation of till and ice

November 17, 2015
Defense Date

Edward B. Evenson, Advisor

Approved Date

Committee Members:

Kenneth P. Kodama

Claudio Berti

Joan Ramage

Francisco Gomez, external
ACKNOWLEDGMENTS

Many people have contributed, in one way or another, to this dissertation. First and foremost, I’d like to thank my advisor, Ed Evenson, for sheparding me throughout the many ups and downs of academic research. Along with his wife, Laura Cambiotti, Ed has been a great source of counsel and enjoyment over the last 4 years. I’d also like to thank my committee – Ken Kodama, Joan Ramage, Claudio Berti, and Paco Gomez. Many mentors and friends within the Earth and Environmental Sciences department, including Frank Pazzaglia, Johanna Blake, Michael Clifford, Josh Stachnik, Chris Dempsey, Daniel Minguez, Jen Schmidt, Ryan McKeon, Jien Zhang, Allison Teletzke, and Nancy Roman, have joined me in interesting intellectual discussion, troubleshooting, and necessary distraction.

The research enclosed here was shaped by many academic contributors, including Andrew Kozlowski and Brian Bird (New York Geological Survey), Johan Kleman (Stockholm University), Grahame Larson (Michigan State University), Dan Lawson (CRREL), Andres Meglioli (Mountain Pass LLC), and Pat Burkhart (Slippery Rock University). Much of this work would have been impossible without the willingness of three men to allow a young glacial geologist to interrupt their daily affairs for the sake of science – tremendous thanks to Larry Waterman of Cato, New York, Leo Kramar of Dalby, Sweden, and Bill and Kelly Stevenson of the Matanuska Glacier, Alaska.

Finally, no words, no matter how colorful, can express the immeasurable contributions of my lovely wife, Brittany Hopkins.
## TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>List of Figures</td>
<td>vi</td>
</tr>
<tr>
<td>Abstract</td>
<td>1</td>
</tr>
<tr>
<td>Introduction</td>
<td>3</td>
</tr>
<tr>
<td>Chapter 1: An anisotropy of magnetic susceptibility (AMS) investigation of the till fabric of till drumlins: Support for an accretionary origin</td>
<td>11</td>
</tr>
<tr>
<td>Chapter 2: An anisotropy of magnetic susceptibility (AMS) fabric record of ice flow and till kinematics within a Late Weichselian Baltic Ice Stream Till, southern Sweden</td>
<td>36</td>
</tr>
<tr>
<td>Chapter 3: Magnetic fabric and the distribution of shear within stratified basal ice, Matanuska Glacier, Alaska, USA</td>
<td>65</td>
</tr>
<tr>
<td>Conclusion</td>
<td>94</td>
</tr>
<tr>
<td>Vita</td>
<td>97</td>
</tr>
</tbody>
</table>
# LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>Weedsport Drumlin Field, New York</td>
<td>16</td>
</tr>
<tr>
<td>1.2</td>
<td>Sampling Scheme, Cato Drumlin</td>
<td>20</td>
</tr>
<tr>
<td>1.3</td>
<td>AMS fabrics results</td>
<td>23</td>
</tr>
<tr>
<td>1.4</td>
<td>AMS P’-T plot</td>
<td>25</td>
</tr>
<tr>
<td>1.5</td>
<td>Fabric parameters vs. elevation</td>
<td>25</td>
</tr>
<tr>
<td>1.6</td>
<td>Inferred patterns on ice flow</td>
<td>27</td>
</tr>
<tr>
<td>1.7</td>
<td>Bulk fabric of the Cato Drumlin</td>
<td>27</td>
</tr>
<tr>
<td>2.1</td>
<td>Ice streams of the Fennoscandinavian ice sheet</td>
<td>40</td>
</tr>
<tr>
<td>2.2</td>
<td>Geomorphology of SW Skåne</td>
<td>42</td>
</tr>
<tr>
<td>2.3</td>
<td>Stratigraphy of the Dalby Quarry</td>
<td>43</td>
</tr>
<tr>
<td>2.4</td>
<td>Sample scheme and AMS fabrics</td>
<td>47</td>
</tr>
<tr>
<td>2.5</td>
<td>Stratigraphic variation of fabric</td>
<td>49</td>
</tr>
<tr>
<td>2.6</td>
<td>AMS P’-T plot</td>
<td>50</td>
</tr>
<tr>
<td>2.7</td>
<td>Thermal demagnetization plots</td>
<td>52</td>
</tr>
<tr>
<td>2.8</td>
<td>pARM spectra</td>
<td>53</td>
</tr>
<tr>
<td>2.9</td>
<td>AMS fabrics of the surface diamicton</td>
<td>54</td>
</tr>
<tr>
<td>2.10</td>
<td>Ice flow interpretations</td>
<td>57</td>
</tr>
<tr>
<td>2.11</td>
<td>AMS fabric and boulder line development</td>
<td>59</td>
</tr>
<tr>
<td>3.1</td>
<td>Stratified Basal Ice, Matanuska Glacier</td>
<td>69</td>
</tr>
<tr>
<td>3.2</td>
<td>Location map of south-central Alaska</td>
<td>70</td>
</tr>
<tr>
<td>3.3</td>
<td>Hillshade of the Matanuska terminus</td>
<td>71</td>
</tr>
<tr>
<td>3.4</td>
<td>Sample scheme</td>
<td>75</td>
</tr>
<tr>
<td>Figure</td>
<td>Description</td>
<td>Page</td>
</tr>
<tr>
<td>--------</td>
<td>--------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>3.5</td>
<td>AMS P’-T plot</td>
<td>78</td>
</tr>
<tr>
<td>3.6</td>
<td>Bulk Susceptibility and Debris content</td>
<td>78</td>
</tr>
<tr>
<td>3.7</td>
<td>Composite AMS fabrics</td>
<td>80</td>
</tr>
<tr>
<td>3.8</td>
<td>Bulk AMS fabric</td>
<td>81</td>
</tr>
<tr>
<td>3.9</td>
<td>AMS fabric of sub-sampled slabs</td>
<td>82</td>
</tr>
<tr>
<td>3.10</td>
<td>Anisotropy parameters vs. fabric</td>
<td>84</td>
</tr>
<tr>
<td>3.11</td>
<td>Bulk fabric of clustered datasets</td>
<td>84</td>
</tr>
</tbody>
</table>
Abstract

Subglacial processes are significant contributors to the dynamics and sediment transport of glaciers and ice sheets, drive much of the observed variability in modern systems, and are responsible for much of the landscape in glaciated terrains. Subglacial deformation (i.e., deformation of the glacier substrate) is now recognized as a fundamental process of motion and sediment transport for warm-based glaciers; however, the spatial and temporal variability of this deformation is poorly understood, owing in part to the inaccessibility of the modern subglacial environment. The Pleistocene sedimentary record provides a complementary view, and many processes are recorded within the sediments. Herein, the variability of subglacial deformation is assessed in three distinct glacial settings using anisotropy of magnetic susceptibility (AMS) fabric analysis. AMS is a robust, quantitative, volume-averaged, and objective measure of the orientations of the axes of magnetic susceptibility, which is predominantly controlled by grain shape such that the orientation of the maximum susceptibility axis parallels the long axis of the grain. Thus, AMS provides a useful proxy for the orientation of magnetic grains within sediments, from which direction of ice flow and the magnitude of deformation can be inferred.

First, I assess the spatial variability of ice flow and till transport around streamlined bedforms (drumlins) and demonstrate relationships between internal drumlin fabric and the drumlin surface morphology, indicating these landforms cannot be erosional remnants, but are likely formed as till is transported to the drumlin locality and syndepositionally streamlined. Next, I evaluate stratigraphic (i.e., temporal) variation as
recorded in a late-Weichselian Baltic Ice Stream (BIS) till in southern Sweden. Our analysis records systematic and dramatic changes in ice flow direction and bed deformation, demonstrating the dynamic nature of the BIS and allowing for the discrimination of intra-till kinematic zones. Finally, I address the distribution of deformation within the debris-rich basal stratified ice of the Matanuska Glacier, a modern temperate glacier in southern Alaska and a type-locality of stratified basal ice. Basal ice that is debris-rich relative to englacial ice and interacts with the glacier bed is common to many glaciers and ice sheets, but its genesis and contributions to glacier motion are poorly understood. This analysis reveals basal stratified ice has experienced significant shear, and that shear appears concentrated in debris-rich layers in the form of simple shear along sub-horizontal shear planes. Debris-poor layers possess ‘compaction’ fabrics indicative of pure shear driven by the force of the overlying englacial ice. Thus, the debris-rich basal ice is characterized by rheological inhomogeneities resulting from the competing factors of debris-content and ice crystal size, among others. These results indicate that deformation is pervasive within the subglacial environment, extending from the unfrozen substrate into the debris-rich basal ice, but that this deformation is characterized by high spatial and temporal variability in both magnitude and direction.
Introduction

At present, approximately 10% of land area of Earth is covered by ice in the form of glaciers and ice sheets. These ice bodies contain approximately 70% of total freshwater, and are considered among the most effective agents of sediment transport. Much of today’s glaciers reside in sparsely populated alpine (e.g. Alaskan mountain ranges) and polar regions; Greenland and Antarctica collectively represent 9.7%. As a result, the direct influence of glaciation is commonly overlooked in general society; however, modern glaciers and ice sheets are expected to respond to anthropogenic climate change in potentially dramatic ways, and often with wide-reaching consequences.

At the height of Pleistocene glaciation the extent of glaciers and ice sheets was approximately three-times larger than present (Benn & Evans 2010) as a result of the expansion of alpine glaciers and the rise of continental ice sheets, such as the Laurentide of Canada and the northern United States and the Eurasian, which spread from Ireland to eastern Russia and into mainland Europe. Driven by Milankovitch cycles, the cyclic growth of these ice sheets and their ancestors is the defining characteristics of the Quaternary period, and is associated with wide-reaching environmental impacts. Additionally, glaciers and ice sheets leave a lasting impact on the landscape, and control and/or influence the morphology, distribution of sediments, and hydrology.

Because of their influence on the earth system and human society, significant efforts have been taken to understand modern glacial processes and to observe and predict their response to climate change. The scientific community is making progress regarding the distribution, mass balance, and dynamics of modern glacial systems. One
key aspect of glacial dynamics is glacial flow. Modern observations of the surface velocity of ice bodies reveal dramatic spatial and temporal variability largely driven by basal processes. The best demonstration of the spatial variability is the satellite-derived surface velocity model of the Antarctic Ice Sheet produced by Rignot et al. (2011), which clearly illustrates the organization of ice into discrete drainage networks culminating in fast-flowing ice streams. From the drainage divide to the ice stream terminus, ice velocity increases by three orders of magnitude. Many of these streams – the Ice stream-type ice streams of Truffer & Echelmeyer (2003) or the Ross-type ice streams of Benn & Evans (2010) common to the Siple Coast region – are characterized by low surface slopes and weak deformable beds that contribute significantly to the ice stream motion. Significant temporal variability is most clearly demonstrated by surge glaciers. Surge-type glaciers are common throughout alpine and polar environments, including Alaska, the Yukon, Iceland, Svalbard, and Greenland (Benn & Evans 2010). Motion of these glaciers occurs in two phases; a slow flow quiescent phase and a fast flowing active phase. This unstable flow regime arises from internally-driven oscillations in the hydrologic conditions of the glacier bed (Meier & Post 1969; Sharp 1988).

The above examples demonstrate variability in glacier motion arising from basal processes. For the purposes of this discussion and the following chapters I provide here a brief review of the basal processes, and bed deformation in particular. Generally speaking there are three mechanisms of glacier motion and the occurrence and efficiency of each mechanism is highly influenced by the thermal state and bed conditions of the glacier. The mechanism of motion common to all glaciers is glacier flow – glacier ice undergoes internal deformation in response to applied stresses according to Glen’s flow
law (Glen 1955). For glaciers whose base is below the pressure-melting point and frozen to its substrate (i.e., cold-based glaciers) this is the only mechanism of glacier flow. However, it is common that the beds of many alpine glaciers and parts of ice sheets are at temperatures equal to or above the pressure-melting point (i.e., warm-based glaciers). In this case, the additional processes of basal sliding and bed deformation are possible. Basal sliding is simply the motion allowed by the glacier sliding along the ice-bed interface. If a warm-based glacier is overlying soft sediment, that sediment is capable of deforming and contributing significantly to glacier motion.

The first observations of bed deformation were made within subglacial tunnels excavated into the margins of Breiðamerkurjökull, Iceland (Boulton 1979; Boulton & Hindmarsh 1987). Within this tunnel, researchers inserted strain markers in vertical profiles into the till resurveyed them after 136 hours. The resulting patterns of displacement demonstrated that subglacial tills deformed and that bed deformation could represent a major portion of glacier motion (90% for Breiðamerkurjökull at the time of the experiment). These observations dramatically influenced research in glaciology and glacial geology for 15-20 years; however, questions remain regarding the rheology and geological consequences of bed deformation. Boulton & Hindmarsh (1987) suggested the deformation of till was non-linear and viscous. Subsequent laboratory testing and new, yet limited, field observations have demonstrated that till actually possess a Coulomb-plastic rheology (Kamb 1991, Inverson et al. 1998; Hooyer & Iverson 2000; Tulaczyk et al. 2000). Rheology aside, there are many unanswered questions: How common are/were deformable beds? How does the distribution of deformation vary
spatially and temporally? What are the implications of bed deformation for the subglacial transport of sediment?

Modern, direct observation has limited potential to answer these questions. There are two key limitations: (1) modern observations are unable to observe the dynamic behavior of ice sheets at timescales comparable to past glacier change and (2) modern observations are limited in number due to the inaccessibility of glacial environments, particularly the subglacial environment.

Glacial landscapes are the product of basal processes of erosion, transport, deposition, and the terrain and sediments are an archive of these processes. Glacial geologists use this archive as a window into the subglacial environment and investigate the chronologies and processes of past glaciations and to make inferences and predictions regarding the modern subglacial environment. A common tool assessing history of ice flow within glacial till is ‘till fabric’ – the preferred orientation of clasts embedded within the till matrix. First coined by Holmes (1941), countless studies have employed this technique, and the results have provided valuable insight into glacial processes. However, this technique was always limited by the lack of a quantitative relationship between fabric and the processes that may lead to its development, as well as the opportunity for unintentional user bias.

In a series of recent papers, Hooyer et al. (2008) and Iverson et al. (2008) have established a laboratory basis for the application of anisotropy of magnetic susceptibility (AMS) fabric analysis to till. AMS has a storied history in the field of structural geology (e.g. Ferré et al. 2014); however, this technique was only infrequently applied to glacial tills prior to the publication of the articles above, and was commonly limited by an
incomplete understanding of the subglacial processes that would lead to the development of a magnetic fabric. Most sediments contain an assemblage of one or more magnetic grains, including ferrimagnetic, paramagnetic, and diamagnetic grains. One property of magnetism is magnetic susceptibility, the proportionality between an applied field and the field induced within the magnetic material. For AMS analysis, the magnetic susceptibility of a given material is measured multiple orientations (typically 9 or 15) and fit to a symmetric second rank tensor that can be graphically described by an ellipsoid. That ellipsoid is defined by three axes: \( k_1 \) (maximum susceptibility), \( k_2 \) (intermediate susceptibility), and \( k_3 \) (minimum susceptibility). The AMS of non-equant magnetic grains is dominantly controlled by shape, such that the \( k_1 \) axis parallels to the long axis of the magnetic grains (Tarling & Hrouda 1993). Thus, we can use AMS as a proxy for the orientation of microscopic magnetic grains and characterize the magnetic fabric of tills.

Hooyer et al. (2008) and Iverson et al. (2008) completed extensive laboratory experiments evaluating the kinematics of till deformation under simple shear. Their experiments utilized natural tills and a mechanical ring-shear device to apply progressively higher amounts of strain. After application of a given amount of strain the authors would then measure the AMS of 25 samples and characterize the fabric. Their result can be summarized by the following:

1. Under simple shear, elongate grains within till rotate such that their long axes parallel shear direction and plunge up-shear;

2. This gives rise to an AMS fabric where the maximum axes of magnetization \( (k_1) \) parallels shear direction and plunges up-shear
relative to the shear plane, intermediate axes ($k_2$) lie within the shear plane, and the minimum axes ($k_3$) are orthogonal to the $k_1$-$k_2$ plane;

3. The clustering of the $k_1$ axes increases progressively under increasing strain until the fabric becomes saturated.

There is now an active group of researchers applying AMS fabrics to problems in glacial geology and glaciology. Topics of research include dynamics of ice lobes (Shumway & Iverson 2009), origins of landforms (Gentoso et al. 2012; Ankerstjerne et al. 2015; Hopkins et al. 2015; Vreeland et al. 2015), and the rheology of modern glaciers (Fleming et al. 2013), among others. This dissertation builds upon this active body of research, demonstrates the effectiveness of AMS fabrics in diverse glacial settings, and addresses the fundamental issue of spatial and temporal distribution and variability of subglacial deformation. Chapter One details an investigation into the patterns of ice flow and till kinematics within a drumlin, enigmatic landforms common in streamlined glacial landscapes, and was published in BOREAS in July, 2015. Chapter Two addresses the stratigraphic variation in ice flow and bed deformation beneath a late-Weichselian Baltic Ice Stream till in southern Sweden, and was submitted for publication in BOREAS in September, 2015. Finally, Chapter Three explores the distribution of deformation within the basal stratified ice of the Matanuska Glacier, Alaska, and will be submitted to the Journal of Glaciology in December, 2015.

References


Chapter 1:

An anisotropy of magnetic susceptibility (AMS) investigation of the till fabric of till drumlins: Support for an accretionary origin

Nathan R. Hopkins¹*

Edward B. Evenson¹, Kenneth P. Kodama¹, Andrew Kozlowski²

¹Lehigh University Department of Earth and Environmental Sciences

²New York State Geological Survey, Albany, NY, USA

Published in BOREAS, July 2015
Abstract

This paper describes the results of a spatially dense anisotropy of magnetic susceptibility (AMS) till fabric study of a single drumlin in the Weedsport Drumlin Field, New York State, USA. AMS till fabrics provide a robust, quantitative, and unbiased approach to assess subglacial till kinematics and infer ice flow dynamics. The drumlin selected for this detailed investigation was systematically sampled at 18 locations to evaluate the patterns of ice flow and associated till kinematics within a drumlin and to test erosional versus depositional models for their formation. AMS till fabric analysis yielded strong fabrics that increase in strength towards the drumlin crest, indicating bed deformation occurred during till deposition and that deformation within the drumlin was greater than that in the interdrumlin low. Fabric orientations reveal drumlin convergent, divergent, and parallel ice flow paths that illustrate a complex interaction between ice flow and the drumlin form; fabric strength and shape reveal systematic differences in bed deformation between the interdrumlin and drumlin regions. These observations are inconsistent with purely erosional models of drumlin genesis; instead, these observations are more consistent with syndepositional streamlining of till transported, likely locally as a deforming bed, from the interdrumlin low towards the drumlin locality.
Introduction

Drumlins are subglacial bedforms whose origins have intrigued geologists and glaciologists since their earliest recognition. As W. M. Davis stated “The arched hills of glacial drift that have been called drumlins by the Irish geologists are among the most peculiar results of the action of land ice-sheets … but just how they were constructed is still an open question” (Davis, 1884). Drumlins are widespread landforms in glaciated terrains with a strikingly consistent morphology. The classic drumlin form is often described as an asymmetric, elongate hill with a blunt up-ice nose and a long tail oriented parallel to ice flow; however, the accuracy of this characterization has been questioned (Spagnolo et al. 2010). Drumlins are probably part of a suite of subglacial bedforms, including mega-flutes and mega-scale glacial lineations (Spagnolo et al. 2014). Despite their morphologic similarity, great controversy exists regarding drumlin origins, owing in large part to the wide variability in their composition and internal structure. Drumlins may be composed entirely of one or more units of basal till, subglacial stratified sediments, proglacial sediments, or bedrock. Drumlins not cored by till are typically mantled with a layer of till. A variety of internal structures have been described, including massive homogeneous basal till, layered diamictons conformable to drumlin form, truncated stratigraphy, and deformed sorted sediments (e.g. Muller 1974; Menzies 1987).

The large drumlin fields comprised of drumlins composed almost entirely of basal till are particularly problematic, because there exists no apparent obstacle or asperity from which to nucleate drumlin formation and ice sheet conditions are largely unknown. It is clear that the mechanism of drumlin formation in these large drumlin fields is the
result of some interaction between ice flow and the substrate at the glacier sole. A direct consequence of this interaction and associated substrate (i.e., bed) deformation is the generation of fabrics – preferred orientations of non-equant particles – within the substrate. Till fabrics have a long history of application in glacial geology and have been commonly used to interpret the origins of subglacial landforms. In general, there exists striking commonalities in the fabric of till drumlins globally as well as within drumlin fields. Fabric long axes typically align sub-parallel to the drumlin long axes, particularly along, or near, the center of the drumlin (Wright 1957; Evenson 1971; Hill 1971; Walker 1973; Stanford & Mickelson 1985; Stea & Brown 1989; Johnson et al. 2010; Gentoso et al. 2012; Vreeland et al. 2015). The fabric long axis typically plunges up-ice between 10 and 30 degrees (Harrison 1957; Evenson 1971; Stanford & Mickelson 1985; Stea & Brown 1989; Johnson et al. 2010; Gentoso et al. 2012; Vreeland et al. 2015). Azimuthal deviation of the fabric from the drumlin long axis increases away from the drumlin centerline (Evenson 1971; Hill 1971; Walker 1973; Stanford & Mickelson 1985). When sampling is conducted at a high spatial density, fabrics have been shown to parallel topographic contours, indicating flow divergence (i.e., flow away from drumlin centerline) around the nose of the drumlin, and parallel to convergent flow (i.e., flow towards the drumlin centerline) along the drumlin flanks (Savage 1968; Hill 1971; Walker 1973). Fabrics measured within the drumlin tail are usually either contour parallel (convergent) or oriented obliquely down-slope (divergent) (Savage 1968; Walker 1973).

Models for till drumlin genesis invoking different relative magnitudes of erosional and depositional processes make specific predictions for the fabric of the drumlin
interior. A purely erosional genesis of till drumlins (e.g. Boyce & Eyles 1991; Kerr & Eyles 2007; Vreeland et al. 2015) would predict internal fabrics that predate drumlin formation and thus have no genetic relationship to its surface morphology, unless serendipitously eroded by ice flow parallel to ice flow during earlier till deposition. A depositional origin (e.g. Smalley & Unwin 1968; Evenson 1971; Boulton & Hindmarsh 1987; Boulton 1996; Stokes et al. 2013) would lead to internal fabrics generated simultaneously with the drumlin form and thus predicts some direct relationship between drumlin fabric and shape. To test these predictions one drumlin (43.126°N, 76.625°W) in the vicinity of the village of Cato, New York, USA, within the Weedsport Drumlin Field (WDF) was selected for a detailed anisotropy of magnetic susceptibility (AMS) fabric analysis. This drumlin, hereafter called the “Cato drumlin”, was chosen due to its ideal representation of the drumlins in New York and access provided by a local landowner. Our goal is to, for the first time, systematically analyze the AMS till fabric of one till drumlin of the WDF using a high-density sampling scheme to determine the likely processes by which it developed, and to speculate on the origins of till drumlins in general.

**Field setting**

The Weedsport Drumlin Field (WDF, Fig. 1.1) spans the area from the south shore of Lake Ontario to the Allegheny Plateau, occupying the physiographic region known as the Lake Ontario Lowland. Drumlins in this field possess a relatively uniform composition; most of the WDF drumlins are composed of silty, sandy till with identical grain size distributions throughout the suite of bedforms (Stahman 1992; Stahman et al.}
Figure 1.1 - Central portion of the Weedsport Drumlin Field north of Cayuga Lake shown by a 10-m digital elevation model made available by the NYSDEC and U.S. Geological Survey. The study area SW of Cato is indicated by the black rectangle.

These tills were mapped along the Lake Ontario bluffs by Calkin & Muller (1992) as the Furnaceville and Somerset Tills. These units are classic subglacial tills – relatively fine grained, massive, pink to brown, overcompacted diamictons with abundant striated and faceted clasts of both local and exotic lithologies. The thickness of the till is variable.
throughout the Lake Ontario Lowlands, but is sufficiently thick in the study area that no interaction between the ice, deforming bed, and the Paleozoic sedimentary bedrock could occur during drumlin formation (Hopkins 2013). Additionally, no change in drumlin morphology associated with till thickness is observed in this field (Hess & Briner 2009). Throughout much of this field, it is commonly observed that the bedforms display a down-ice trend in increasing bedform elongation. Drumlins with a length:width (L:W) ratio of approximately 3:1 dominate near the Lake Ontario shore and transition to flutes with L:W ratios up to 63:1 near the southern edge of the WDF (Stahman 1992). However, the transition is not manifested in a linear progression in elongation; the transition occurs through an intermediate bedform referred to as “complex” or “fluted” drumlins (Stahman 1992). This morphological transition has been interpreted as evidence for topographically driven ice streaming (Briner 2007; Hess & Briner, 2009). The composition of the till is constant throughout this suite of landforms (Stahman 1992).

Previous till fabric studies within the WDF yield results markedly consistent with those from elsewhere. The following description of past investigations into the fabrics within the WDF is intended to illustrate the commonalities between these drumlins and drumlins composed of till elsewhere, and demonstrate the likelihood of a common genesis for drumlins composed almost entirely of basal till. Fabrics within the drumlins of this region have been reported irregularly beginning in the late 1960’s as part of investigations into glacial processes and glacial chronologies, and the application and detail of the fabric analyses reflects the motives of each investigation. In an early study assessing the applicability of a plastic flow analysis to drumlins, Savage (1968) conducted a systematic study of macrofabrics (pebbles) within one till drumlin near
Syracuse, New York. His sixteen fabrics reveal fabric maximum orientations broadly subparallel to the drumlin long axis, with patterns suggestive of divergence and convergence around the drumlin form. Most strikingly, the four fabrics recovered from the nose and tail display nearly contour-parallel divergence and convergence, respectively, which agree well with subsequent observations elsewhere (Hill 1971; Walker 1973). Krall (1977) noted E-W, drumlin parallel fabrics within drumlins of the Mohawk Valley, except for those drumlins modified by a later southward advance of the Lake Ontario lobe; fabric within the core of modified drumlins preserved relict E-W fabrics associated with initial deposition and formation of the drumlins, as was shown in others areas with complex ice flow histories (Stea & Brown 1989; Newman & Mickelson 1994).

Hart (1997) reports four fabrics taken within the stoss and flank of one drumlin in New York; however, these fabrics are typically weak (mean $S_1$ eigenvalues of 0.59) and display drumlin parallel and convergent fabrics in close proximity. Menzies & Brand (2007) report relatively strong ($S_1 > 0.7$) drumlin parallel macrofabrics for two near-surface sites measured within one drumlin near Port Byron, New York. Strong agreement between near-surface fabrics and drumlin orientation has been recorded in numerous other locations, as described above; however, the Port Byron drumlin possess a complex internal stratigraphy of proglacial sorted sediments that appears to have been remolded by subsequent ice advance (Menzies & Brand 2007). Thus, the Port Byron drumlin is likely genetically distinct from the majority of the WDF.

Gentoso et al. (2012) conducted the first systematic application of AMS to drumlins and flutes in the WDF. They examined five drumlins and five flutes and
compared the AMS fabrics with traditional pebble macrofabrics measured at the sample place as the AMS samples were collected. The resulting AMS fabrics displayed variable strengths ($S_1$), but fabric orientations were consistently drumlin-parallel and plunged up-glacier (north). Additionally, the mean fabric of all 250 AMS samples closely matched the orientation of the pebble macrofabrics. Collectively, these results indicate that AMS fabric faithfully records ice flow directions in the New York Drumlín Field; additionally, it is clear that bed deformation occurred during the deposition of tills within the WDF and that deformation was pervasive (i.e., affecting all grain sizes), as has been reported by previous workers (Savage 1968; Boulton 1979; Boulton & Hindmarsh 1987; Alley 1991; Menzies et al. 1997; Boulton et al. 2001).

Methods

The Cato drumlin and adjacent interdrumlin low was sampled at 18 locations (Fig. 1.2) in an attempt to accurately capture any potential relationship between the form of the drumlin and localized ice flow (i.e., divergence or convergence around the drumlin form). Due to logistical restrictions (e.g. forest cover on the eastern side) and the desire for a higher density of sample sites, only the western half of the drumlin was sampled. While this sampling strategy is unable to capture the patterns of ice flow and till kinematics around the entire drumlin, it instead allows these patterns to capture in greater detail and to assess the differences between the drumlin and interdrumlin low. At each site, a backhoe pit approximately 1 m wide and 2.5 m deep was excavated.
Figure 1.2 - Sampling locations (Rows 1 – 5) on the Cato drumlin near Cato, NY, and half-rose diagrams of associated AMS fabric analysis. Contour lines are presented in m a.s.l. with a 5 m contour interval. For location see Fig. 1.

At each of the 18 fabric sites, between 20 and 29 oriented 8 cm$^3$ samples of unweathered till were carved from the pit face at approximately 2 m depth using a putty knife and oriented with a Brunton compass. A total of 441 samples were collected from the 18 fabric sites. GPS location was recorded for each pit and the elevation obtained from a 10-m DEM produced by New York State Department of Environmental Conservation and the U.S. Geological Survey.

Gentoso et al. (2012) identified maghemite as the dominant magnetic mineralogy in this region of the WDF. Stepwise thermal demagnetization of samples collected from
the Cato drumlin further confirmed the presence of maghemite, so no further investigation into the magnetic carrier was undertaken.

Magnetic susceptibility is the proportionality tensor between an applied and induced magnetic field. Anisotropy within a sample arises when the ‘easy’ axes of mineral crystals become preferentially aligned; the alignment of these axes can be characterized by a second-rank tensor, where the maximum ($k_1$), intermediate ($k_2$) and minimum ($k_3$) axes of preferential magnetization of a sample define the AMS ellipsoid (Tarling & Hrouda 1993). The AMS ellipsoid of nonequant (elongate) mineral grains is dominantly controlled by shape, such that the $k_1$ axis is aligned parallel to the long axis of the grain. AMS has been applied with limited success and understanding to till fabric analysis sporadically since the 1960s (Fuller 1962; Stupavsky et al. 1974; Stupavsky & Gravenor 1975; Boulton 1976; Gravenor & Stupavsky 1976; Gravenor et al. 1984; Eyles et al. 1987; Gravenor & Wong 1987; Principato et al. 2005). However, in a series of recent papers a laboratory basis for the development, interpretation and application of AMS till fabrics was established using a ring-shear device and modified natural tills (Hooyer et al. 2008; Iverson et al. 2008). Shear-parallel fabrics were shown to develop as a function of total shear strain; initially randomly oriented grains were quickly reoriented to develop a strong shear-parallel fabric ($S_1 \sim 0.90$) at shear strains of approximately 20 (Hooyer et al. 2008; Iverson et al. 2008). Continued shear to higher strains has little effect on fabric strength. In effect, once sheared to sufficiently high strains the intensity of fabric becomes saturated.

Our AMS measurements were completed using a KLY-3S Kappabridge housed at the Lehigh University Paleomagnetism Laboratory. For each sample, susceptibility was
measured in 15 static positions and fitted by a least-squares method to a second rank tensor to derive the AMS fabric ellipsoid. Fabric orientation is determined by the principal eigenvector orientation (V₁) calculated from the orientations of the 25 k₁ susceptibility axes, and the degree of alignment of the k₁ susceptibility axes (fabric strength) is characterized using the eigenvalue method (Mark 1973), where the principal eigenvalue (S₁) equals 1 when k₁ axes are perfectly aligned, and 0.33 when random. The degree of anisotropy within a sample is characterized by P’. The shape factor, T, indicates whether AMS ellipsoids are oblate (positive values) or prolate (negative values).

**Results**

*Fabric strength and shape*

AMS fabrics of the Cato drumlin (Fig. 1.3) are typically strong to very strong (mean S₁ of 0.882), with S₁ ranging from 0.772 and 0.964. Fabrics from the adjacent interdrumlin low (e.g., 3E, F) are weaker (mean S₁ of 0.741) than the drumlin fabrics. All fabrics exceed the critical strength (S₁ > 0.57) used by others (e.g. Vreeland *et al.* 2015) to indicate the till experienced shear sufficient to record ice flow direction. Mean anisotropy (P’) and shape (T) for the 16 drumlin and 2 interdrumlin fabrics show considerable variability (Fig. 1.4). Anisotropy (P’) ranges from 1.07 to 1.18, while T ranges from -0.29 to 0.35. T typically displays much greater variability within a given
Figure 1.3 (previous page) - AMS $k_1$ fabrics and half-rose diagrams for the 18 drumlin fabric sites. Lower-hemisphere stereographic projects of AMS $k_1$ orientations (black dots), $V_1$ orientation (red diamond), number of samples (N), and fabric statistics ($V_1$, $S_1$) are shown

site, likely reflecting the irregular distribution of paramagnetic clays and ferrimagnetic grains within the till. In general, fabrics are dominantly triaxial.

Fabric strength increases non-linearly as a function of elevation (Fig. 1.5A). A similar relationship is found for anisotropy ($P'$), with the drumlin crest possessing much greater anisotropy than the interdrumlin low (Fig. 1.5B). Anisotropy and fabric strength are strongly linked, and the similarity of the relationship is expected. $P'$ is high where the maximum susceptibility axes of grains are aligned within the sample; high $P'$ should generate higher $S_1$ values for a given site assuming shear is approximately uniformly distributed at the decimeter scale. No clear relationship between $T$ and elevation is observed, although fabrics near the drumlin crest are likely to be more strongly triaxial than those at lower elevations.

**Fabric orientations**

With few exceptions, AMS fabrics (Fig. 1.3) are oriented sub-parallel to the drumlin long axis orientation (345°), with a notable increasing deviation from the drumlin axis with distance from the drumlin center line. All mean principal eigenvector orientations lie within 54 degrees of the drumlin orientation, and most (11 of 18) lie within 30°. In all cases, principal eigenvectors plunge northward in agreement with the known past ice-flow. The spatial patterns of localized flow (Fig. 1.6) determined by the
Figure 1.4 - Mean degree of anisotropy (P') and shape (T) factors and 95% confidence intervals from each fabric site. Positive (negative) T indicates oblate (prolate) AMS ellipsoids.

Figure 1.5 - Mean fabric strength (A) and anisotropy (B) as a function of elevation, illustrating the relationship between drumlin form and the AMS fabric.
AMS $V_1$ vary systematically. Fabrics at the far up-ice end (Row 1) record SSE to SE ice flow directions. Uniformly southwesterly oriented fabrics characterize the nose of the drumlin (Row 2), while south-southeast oriented fabrics occur along drumlin flanks (Row 3). Down-ice fabrics (Rows 4 and 5) are dominated by oblique down-slope (SSW) directions. Fabrics along the drumlin centerline crest are all strong and near parallel to drumlin long axis.

When all AMS $k_1$ axis orientations are evaluated without respect to location, the resulting fabric is expectedly weaker ($S_1$ value of 0.644); however, the principal eigenvector ($347^\circ$, $24^\circ$) is aligned parallel to drumlin long axis and plunging up-ice (Fig. 7). The principal axis orientations plot an asymmetric bimodal distribution around the drumlin long axis, with a clear maximum deflected approximately $20^\circ$ east of the drumlin, and a secondary mode deflected $45^\circ$ to $60^\circ$ west (Fig. 1.7). These apparent deflections reflect the complex ice flow paths inferred from the individual fabric orientations; ice flow only parallels the drumlin long axis along its centerline. The relative magnitudes of these deflections from the drumlin orientation are largely the product of the sampling distribution; we sampled only the western half of the drumlin.

**Discussion**

The sub-parallel to parallel nature of individual fabrics (Figs. 3, 6A) and total fabric (Fig. 7) within the Cato drumlin is in complete agreement with the fabrics of other drumlins within the Weedsport Drumlin Field (Krall 1977; Menzies & Brand 2007; Gentoso *et al.* 2012). Additionally, the spatial variation of fabric orientation around the drumlin form reported here is similar to that of Savage (1968). Based upon these
Figure 1.6 - Principal eigenvector orientation ($V_1$) of each fabric (A) and ice flow paths (B) inferred from the AMS fabric orientations.

Figure 1.7 - Lower-hemisphere stereographic projection of the AMS $k_1$ orientations (black dots) from all 16 drumlin sites with 2-sigma Kamb contours. Drumlin trend ($345^\circ$) is shown by the black arrow, and fabric orientation ($V_1$) is indicated below.
consistencies, we consider this drumlin to be representative of this drumlin field. Lastly, these observations are largely consistent with fabrics of till drumlins measured elsewhere; thus, we consider this drumlin to be representative of till drumlins more broadly (Evenson 1971; Hill 1971; Walker 1973; Stanford & Mickelson 1985).

Fabrics from the Cato drumlin are very strong compared to other regions (e.g. Vreeland et al. 2015), indicating that significant bed deformation occurred during the deposition of till at the sampling depth. All of our samples were taken at approximately 2 m depth so we cannot assess fabric variation with depth. Most of the drumlins in this region have likely experienced some combination of tree fall, bioturbation, and human activities at depths shallower than 2 m, which would negatively impact the preservation of microscopic till fabric; however, based upon the lack of vertical heterogeneities within the exposed till profiles, we expect bed deformation to have been significant throughout the upper 2 m. The increasing trends of fabric strength and anisotropy from the interdrumlin low to the drumlin crest clearly illustrate a relationship between the drumlin form, AMS fabric, and ice flow. Based upon comparison with ring-shear experiments (e.g. Iverson et al. 2008) greater fabric strengths ($S_1 > 0.9$) at the drumlin crest indicate greater degrees of shear, while lower fabric strengths ($S_1 < 0.75$) of the interdrumlin lows indicate smaller degrees of shear. The direct conclusion is that bed deformation was not enhanced in the drumlin lows as would be suggested for erosional models that invoke ice streaming and enhanced erosion through interdrumlin lows (e.g. Boyce & Eyles 1991); instead, bed deformation within the interdrumlin low is slight in comparison to the drumlin proper.
V₁ orientations of the k₁ axes display systematic variability, with orientations that indicate ice flow paths that appear to interact with the drumlin form in a complex manner (Fig. 1.6). The sequence of observed divergence around the drumlin nose, followed by convergence in the drumlin midsection, and resumed divergence at the drumlin tail has not been observed previously; however, some elements of this pattern, particularly the divergence of fabrics around the nose of the drumlin, were observed by Savage (1968) and Walker (1973). The herringbone pattern of fabric convergence at the midsection has been observed elsewhere (Andrews & King 1968), and is predicted in some models (Hooke & Medford 2013). The consistent divergence at the tail of the drumlin is unique to this study and does not match predictions (e.g. Boulton & Hindmarsh 1987; Hooke & Medford 2013).

Any hypothesis for the genesis of till drumlins must satisfactorily explain the observations outlined above. For the Weedsport Drumlin Field, these observations are inconsistent with an erosional origin for the following reasons: internal fabric conforms to the morphology of the drumlin and hence cannot pre-date its formation; weak interdrumlin fabrics with respect to the drumlin crest are inconsistent with ice-streaming through interdrumlin lows or erosion of a preexisting well-developed fabric such as those in the drumlin proper. These fabrics are more consistent with an accretionary model wherein till is transported towards the drumlin, likely from the adjacent interdrumlin lows, and deposited at the drumlin surface and simultaneously streamlined by the flow of the ice around a forming obstacle. In this case, the growing obstacle is the proto-drumlin. This suggests that drumlin relief grows over time with continuous or episodic deposition at the drumlin surface.
We cannot infer the mechanism for drumlin nucleation from these results; however, the spatial relationship between drumlins and longitudinal crevasse clusters in Iceland (Johnson et al. 2010) and spatially variable temperature conditions of the bed (Hooke & Medford 2013) provide possible explanations. Additionally, we do not suggest that this process for drumlin formation by accretion applies to drumlins possessing different compositions and internal structures; the streamlined nature associated with such drumlins may in fact result from net erosion at the glacier sole. The work of Stokes et al. (2013) neatly summarizes how the competing effects of erosion and deposition may change in relative magnitude as a function of location and subglacial conditions, and how the drumlinoid form may be produced under both net-erosive and net-depositional conditions. However, it is clear that our drumlins possess fabrics indicating that deposition dominated during drumlin formation.

Our discussion has focused on drumlins composed of till; however, the term “drumlin” is commonly used in the literature to describe a suite of morphologically similar landforms without regard to their composition, internal structures, and spatial characteristics. Thus, it is likely that the drumlinoid form is not unique to a single process, but that drumlins of different characteristics can result from either a suite of distinct processes or a continuum of processes occurring at the sole of the ice sheet (Hart 1997). If such a continuum exists, it is likely the result of competing erosional and depositional processes (Hart 1997). Recent work on the application of flow instability theory to drumlin formation shows great promise in explaining the consistency of the drumlin form despite the observed variability in structure and composition (e.g. Stokes et
al. 2013); however, the subglacial kinematics associated with this theory and the predicted till fabrics are still unknown.

Conclusions

Detailed AMS till fabric analysis of one till drumlin in the Weedsport Drumlin Field indicates this drumlins is the product of syndepositional streamlining of till deposited at the drumlin locality by way of complex ice flow paths resulting from interaction of the ice body with the protodrumlin form – a growing obstacle at the ice-till interface. We presume this to be true of all the drumlins in the WDF. The strength, orientation, and consistency of the till fabrics of drumlins is further confirmation that deforming bed conditions existed at the time of till deposition and drumlin formation and that AMS till fabrics reliably and accurately record local ice flow conditions. Because of the striking consistency of till fabrics measured within till drumlins from distant regions and differing glacial histories, it is possible that the Cato drumlin is broadly representative of most drumlins composed of deforming bed till.

Acknowledgements

First, we would like to thank Dr. Neal Iverson and his students, for without their exceptional work developing AMS till fabric analysis for modern research in glacial geology this article would have been impossible. We’d also like to thank Larry Waterman of Cato, NY, for his tremendous generosity in the field and for sacrificing a portion of his soybean crop for science. Additional thanks to Dr. Brian Bird (New York State Geological Survey), Michael Clifford (Desert Research Institute) and Dr. Dan E.
Lawson (Cold Regions Research and Engineering Laboratory). This work was funded, in part, by the Lehigh University Earth and Environmental Sciences Department, the New York State Geological Survey, Tom Pasquini, and ExxonMobil. Finally, we’d like to thank the reviewers for their assistance in improving this article.

References


Chapter 2:

An anisotropy of magnetic susceptibility (AMS) fabric record of ice flow and till kinematics within a Late Weichselian Baltic Ice Stream Till, southern Sweden

Nathan R. Hopkins

Johan Kleman, Edward B. Evenson, Kenneth P. Kodama

1Department of Earth and Environmental Sciences, Lehigh University

2Department of Physical Geography and Quaternary Geology, Stockholm University,

Stockholm, Sweden

Submitted for publication in BOREAS, September 2015
Abstract

Herein we report on the results of an anisotropy of magnetic susceptibility (AMS) till fabric case-study of a late Weichselian Baltic Ice Stream (BIS) till exposed in a bedrock quarry in Dalby, Skåne, southern Sweden – the first application of AMS fabric analysis to Swedish till. Understanding the temporal variability of the subglacial system beneath ice streams is critically important for assessing the dynamics of past ice sheets and for predictions of dynamics and stability of modern ice sheets. AMS till fabrics are robust indicators of ice flow history and till kinematics, and provide a unique tool to investigate till kinematics within and between till units. The till section investigated here contains ~8 m of the Dalby till – a dark grey silt-clay rich BIS till – overlain by ~1.5 m of the regional surface diamicton. AMS fabrics within the Dalby till record a coherent rotation in ice flow direction from east-northeasterly to southwesterly occurring coincident with a boulder line approximately 3 m from the base of the till. Additionally, this rotation is coincident with changing trends in fabric strength marked by an excursion at the boulder line; the lower (upper) section records increasing (decreasing) shear separated by a pronounced weak fabric at the boulder line. We interpret these fabrics to record shifting ice flow and bed conditions at the margins of the Young BIS, prior to a short-lived readvance of the Fennoscandian Ice Sheet recorded by the surface diamicton.
**Introduction**

Ice streams – narrow zones of fast-flowing ice within ice sheets – are widely recognized as fundamental and significant components of modern and paleo-ice sheets (Bennet 2003). Despite their typically small size relative to the overall ice sheet, modern ice streams (e.g. Siple Coast, Antarctica) are thought to act as the principal agent of ice discharge, and thus regulate ice sheet mass balance and, ultimately, stability (Bamber et al. 2000; Alley & Bindschadler 2001; Bennett 2003; Rignot et al. 2011). The causes, mechanics, and dynamics of these large, possibly episodic ice streams without obvious topographic drivers, termed “Pure Ice Streams” by Stokes & Clark (1999) have been the focus of much debate over the last three decades.

Some combination of enhanced basal sliding and bed deformation is typically regarded as the primary mechanism for the enhanced surface velocities of ice streams; however, the depth of subglacial deformation (i.e. thickness of the deforming layer) is debated and probably variable, but estimates are typically less than 1 m (e.g., Engelhard & Kamb 1998; Kamb 2001; Alley 2000). Modern ice streams are commonly underlain by a clay-rich subglacial till with a high porosity and uneven degrees of deformation (Alley et al. 1986; Alley et al. 1987; Tulaczyk et al. 1998; Tulaczyk et al. 2001). Ice stream beds are also characterized by high basal water pressures, approximately equaling the pressure of the overlying ice, which serve to enhance deformation and basal sliding (Engelhard & Kamb 1998; Kamb 2001).

The investigation of paleo-ice streams offers a complementary view of the ice-stream bed. Paleo-ice streams are commonly identified by the presence of glacial lineations and landform assemblages indicative of pervasive bed deformation, including
drumlins, flutes, and mega-scale glacial lineations (e.g. Patterson 1998; Cofaigh et al. 2002; Stokes & Clark 1999, 2002, 2003; De Angelis & Kleman 2007; De Angelis & Kleman 2008; Evans et al. 2014). These landform assemblages and, often, well-preserved stratigraphy provide a valuable window into the bed of an ice stream; as such, analysis of these sediments and landforms has the potential to elucidate the bed conditions and dynamics of ice streams at a range of spatial and temporal scales.

Geomorphic analysis is particularly useful for assessing distribution, controls, and timing of paleo-ice streams at spatial extents of $10^1$-$10^6$ km$^2$ and larger; assessing local variations at a finer spatial, and ultimately temporal, resolution requires detailed investigation within the sedimentary record (e.g., Christofferson & Tulaczyk 2003; Ross et al. 2006; Lusardi et al. 2011; Narloch et al. 2012; Evans et al. 2012).

Numerous paleo-ice streams have been identified for the Weichselian Fennoscandian Ice Sheet (FIS; Fig. 2.1). The Baltic Ice Stream (BIS) is the largest of these. It transported ice and sediment south from the Bay of Bothnia towards the SW margins of the Baltic Basin, terminating in southern Sweden, Denmark, Northern Germany, Poland, and the Baltic nations (Boulton et al. 2001; Houmark-Nielson & Kjær 2003). Although there is agreement of the existence of repeated ice streaming in the southern Baltic, most of the evidence for individual events comes from terrestrial sections and lobe outlines in their distal areas. Less is known about ice flow pathways within the Baltic basin, and the development during the individual streaming events. The BIS was active during at least two periods surrounding the last glacial maximum (LGM) – the pre-LGM Old Baltic Advance (~35-33 kyr BP) and the post-LGM Young Baltic Advance
(~19-17 kyr; Houmark-Nielsen & Kjær, 2003). In Denmark and southern Sweden, the Young Baltic

Figure 2.1 - Deglaciation and ice streams of the Fennoscandian Ice Sheet (FIS). Ice margins are represented by solid or dashed black lines with corresponding ages in kyr. Ice streams are indicated by arrows, and ice stream corridors are outlined in bold. The inset box marks the domain of repeated ice streaming in the lower Baltic Sea extending into southern Sweden, Denmark, Germany and Poland. Figure provided courtesy of Dr. Johan Kleman.
Advance can be further separated into two advances or phases; the East Jylland Advance was the furthest advance of Young Baltic ice, which was shortly followed by the Baelthav Advance. Herein, we investigate the stratigraphic evolution of till kinematics, ice flow, and bed conditions of the Baltic Ice Stream (BIS) in Skåne, Southern Sweden, as preserved in one stratigraphic section using anisotropy of magnetic susceptibility (AMS) till fabric analysis.

Field Setting

We investigated a till exposure within a bedrock quarry outside the village of Dalby, 15 km ESE of Lund, Skåne County, Sweden (Fig. 2.2). Dalby is centrally located on the Romeleasan ridge, a NW-SE trending horst predominantly formed of Precambrian crystalline rocks and thin (<20 m) Quaternary cover separating the adjacent Vomb and Lund basins (Lidmar-bergstrom et al. 1991). In the vicinity of Dalby and northwards the bedrock ridge is streamlined in the NE-SW direction. To the east, the Vomb basin is dominated by proglacial lacustrine and glacifluvial sediments deposited during Weichselian deglaciation.

The glacial deposits exposed at Dalby overlie highly fractured gneissic bedrock common in the central and southern portions of the Romeleasan Horst (Fig. 2.3A). Immediately above the bedrock is a dark grey silt-clay rich diamicton containing very few cobbles and boulders (Fig. 2.3A, B); however, the coarse fraction of materials is uniformly faceted and striated when present. Clast lithology is dominantly Paleozoic carbonates and to a lesser extent Precambrian igneous and metamorphic rocks (Berglund and Lagerlund, 1981); however, there appears to be no incorporation of the local bedrock
Figure 2.2 - Lidar elevation map of Western Skåne, southern Sweden, with

generalized glacial geomorphology indicated, including bedforms and lineations
(short black lines), eskers (light blue lines), lake plains and proglacial drainage
(dark blue).

into the till. The till possesses a well-developed fissility. Based upon these characteristics
and provenance of the clasts, we agree with past researchers (Berglund & Lagerlund
1981; Ringberg 1988; Ringberg 1992) and identify this unit as a subglacial traction till of
the Baltic Ice Stream. The thickness of this unit is variable, as it is preserved largely in
bedrock depressions; in some localities this unit is not present between the bedrock and
the upper diamicton. Within the area of our analysis, the unit ranges from 6 – 10 m thick.
Figure 2.3 - A. The stratigraphic section on the southern wall of the quarry where the Dalby Till was sampled.  B. Close up of the clay-rich Dalby Till at sample site 2.  C. Close up of a polished, striated boulder within the boulder line at sample site 4.  D. The surface diamicton sampled on the northern wall of the quarry.

The coarse sediments (i.e. cobbles to boulders) are concentrated in a moderately developed boulder line approximately 3 m above the bedrock contact, occasionally associated with discontinuous, poorly sorted sand lenses ranging from 0.1 – 10 cm thick (Fig. 2.3C). Laterally, an additional, lower boulder line is exposed in an adjacent bedrock depression.
The BIS till is overlain by a weakly-compacted, friable, sand-rich diamicton containing abundant igneous clasts with poorly developed faceting and striations interpreted as a supraglacial till by Ringberg (1992) (Fig. 2.3D). In bedrock highs where the lower till is absent, this upper till directly overlies and appears to incorporate bedrock. Where present, the contact between this diamicton and the lower unit is weakly gradational, and shows some evidence of reworking.

Previous investigations at this and nearby sites have produced contradictory naming conventions and correlations. Berglund and Lagerlund (1981) proposed to name the lower till the Dalby Till, which the authors subdivided into 4 members based upon grain size and lithology. Ringberg (1992) separated the lower till into two units separated by the boulder line, which he described as a discontinuous sand horizon; he referred to the lower and upper sections of this unit the Alnarör and Önnemo Till, respectively. In an earlier publication, similar till units at a nearby exposure are named the Boksbacke and Hardeberga tills, respectively (Ringberg 1988). Counter to the observations of past researchers (Berglund and Lagerlund 1981; Ringberg 1992), we observed no sedimentologic evidence at Dalby that would indicate that the material above and below the boulder line represents two separate till units. Lithologic and textural changes across the boulder line are minor and consistent with what one would expect associated with an intra-till boulder line. OSL age dates of the quartz grains taken from this sandy layer confirm subglacial deposition and indicate no subaerial exposure (Kjær et al. 2006). This interpretation is also consistent with till kinematic evidence discussed below. Herein, we use the accepted naming convention of Berglund & Lagerlund (1981) and refer to the silt-
clay BIS till as the “Dalby till”, and the overlying sand-rich diamicton as the “surface 
diamicton.”

Similarly, the ages of the units exposed here are poorly controlled and, due to the lack of any quantitative age data, are largely determined by correlating ice flow indicators within the section with known regional ice flow of different ice advances. Previous investigations report changing ice flow directions within the Dalby till, as measured by macrofabric, striations, and clast lithologies. Berglund & Lagerlund (1981) report striation measurements and clast lithology at this locality that indicate initial ice flow from the SE followed up-section by flow from the NE, and, in an additional nearby exposure, subsequent resumed ice flow from the SE, and suggests that the Dalby Till was deposited continuously beneath an Old Baltic Advance, the LGM advance, and subsequently by the Young Baltic advances (Lagerlund 1987). The observations of Berglund & Lagerlund (1981) were largely repeated by Ringberg (1992). He identified the lower section (his Alnarör till) as corresponding to the Old Baltic Advance and the upper section to the main LGM advance. At a nearby locality, the LGM till appears to have been subsequently tectonized from the SW, likely as a result of Young Baltic ice flow from the SW (Ringberg, 1988); however, there is no deposition of the till at that time. More recently, the Dalby till has been assigned solely to the LGM advance and the subsequent East Jylland phase of the Young Baltic Advance (Anjar et al. 2014). No matter the age(s) of all or parts of the Dalby Till exposed here, it is clear that the unit was deposited by ice from the Baltic basin, and therefore, transported and deposited by the Baltic Ice Stream.
Methods

Anisotropy of magnetic susceptibility (AMS) provides a quantitative, robust, and unbiased assessment of the orientation of non-equant magnetic mineral grains, and has been demonstrated as a useful and reliable method of measuring till fabric and interpreting ice flow history and till kinematics in numerous field and laboratory settings (e.g., Iverson et al. 2008; Hooyer et al. 2008; Shumway & Iverson 2009; Gentoso et al. 2012; Vreeland et al. 2015; Hopkins et al. 2015). For non-equant grains, AMS is controlled largely by shape, where the maximum susceptibility lies along the long axis of the grains (Tarling & Hrouda 1993); thus, AMS provides an effective proxy for the measurement of grain orientation. Most sediments contain magnetic grains, such that AMS analysis of one 8 cm$^3$ sample has the potential to measure the volume-averaged orientation of numerous mineral grains. Additionally, evaluation of the AMS ellipsoid is useful for determining overall shape and magnitude of the anisotropy and assessing deformation of the till.

The tills exposed at Dalby were systematically sampled in September of 2014 and May of 2015 to capture potential changes in AMS fabric orientation and character throughout the section. At the primary exposure, the lower till was sampled in 8 sites (Fig. 2.4). At each site, a fresh-face was exposed and twenty five 8 cm$^3$ sampled were hand-carved from the till using an aluminum putty knife and oriented using a Brunton compass. Due to the friable nature of the surface till which prevented the collection of small oriented cubes, five oriented 10cm x 6.5cm x 5cm Kubiena tins were collected and impregnated with resin at Brock University. The impregnated blocks were then subsampled into 25 to 30 cubes for AMS analysis. Due to difficulty sampling the steep
face, only one surface till sample (site 7) was collected at the primary exposure; an additional four samples were collected at an exposure of surface till directly overlying bedrock on the opposite quarry wall. A total of 329 samples – 202 from the Dalby Till; 127 from the surface diamicton – were collected for AMS analysis.

Figure 2.4 - Sampling scheme (left) and AMS fabric results (right, not spaced to scale). For each fabric site, the k1 susceptibility axes (black dots) are plotted on lower-hemisphere stereo projections (rotated to near horizontal) and contoured using 2σ-Kamb contours. Sample number, S₁ eigenvalue, and principal eigenvector orientation (V₁) are provided. Black arrows indicate ice flow direction indicated by fabric. North is to the top of the page for all steronets.
AMS measurements were conducted at Lehigh University using a KLY-3S Agico Kappabridge. For each sample, magnetic susceptibility was measured in 15 orientations and fitted by least-squares method to an ellipsoid defined by the maximum \( (k_1) \), intermediate \( (k_2) \), and minimum \( (k_3) \) susceptibility axes. Following the methods used previously (e.g., Gentoso et al. 2012; Hopkins et al. 2015), we have chosen to characterize the AMS fabric orientation using the \( k_1 \) axes. Fabrics were plotted using Stereonet 9 (Allmendinger et al. 2012; Cardozo & Allmendinger 2013). For each site, fabrics were characterized using the eigenvalue method of Mark (1973), wherein a tensor is fitted to the 25 \( k_1 \) axes to solve for the principal eigenvector orientation \( (V_1) \) and its associated eigenvalue \( (S_1) \). \( S_1 \) is used to measure fabric strength, such that \( S_1 \) equals 0.33 when the \( k_1 \) axes are randomly oriented and 1 when perfectly aligned. The relationship between \( S_1 \) and shear strain has been observed experimentally, confirming that \( S_1 \) is a useful indicator of relative shear magnitude, at least for low strains (e.g., Iverson et al. 2008; Hooyer et al. 2008).

Additional experiments undertaken to identify magnetic carrier grain size and composition were a partial anhysteretic remanence magnetization (pARM) spectrum and thermal demagnetization experiment. pARM exploits the grain-size dependence of magnetic coercivities to characterize the grain size of the magnetic minerals within the sample (Jackson et al. 1988). Step-wise thermal demagnetization of a 3-axis isothermal remanent magnetization in three different fields (1 T, 0.5T, and 0.1 T) allows for the identification of the magnetic mineralogy (Lowrie 1990). Six samples distributed throughout the Dalby till were selected for these additional experiments to identify any changes within the mineralogic character of the unit throughout the exposed section.
Results

*Dalby Till*

Fabric orientation (V₁) varies significantly throughout the till section (Fig. 2.4). The lower section (Sites 1 – 3A) is dominated by ENE fabrics. The plunge of the V₁ axes is either shallow (~5°) or typical of till fabrics (~20-25°), and in three of the four fabrics plunge is towards the east. The fabric from the boulder line (Site 4) is oriented SSE (169°) with a 20° plunge towards the south. The upper section (Sites 4A – 6) shows greater variability in the V₁ orientation than the lower section, ranging from ENE (Site 4A) to SW (Site 5) and W (Site 6). All fabrics within this section plunge at values typical of basal till fabrics (15-30°).

![Figure 2.5 - Variation of fabric strength, anisotropy, and shape with stratigraphic position. Note the increasing and decreasing trends in fabric strength below and above the boulder line separated by the weak boulder line fabric.](image)
Fabric strength ($S_1$) is also variable (Fig. 2.5). The fabrics are weakest near the base (Site 1; $S_1 = 0.728$), boulder line (Site 4; $S_1 = 0.665$), and top (Site 6; $S_1 = 0.734$) of the Dalby Till. Fabrics are strongest just below (Site 3A; $S_1 = 0.962$) and above (Site 4A; $S_1 = 0.925$). Therefore, fabric strengths increase from both the base and the top of the section towards the boulder line, where these trends are separated by a markedly weak fabric.

![Figure 2.6](image)

**Figure 2.6** - Mean $P'$ (degree of anisotropy) and $T$ (shape factor) with 1 standard deviation error bars for each of the Dalby Till and surface diamicton fabrics. Note that the 5 surface diamicton samples cluster towards the left (lower anisotropy), but display largely similar $T$ values as the Dalby Till samples.

Sample anisotropy of the Dalby till tends to be triaxial to oblate (Fig. 2.5, 2.6). The mean shape factor, $T$, for each site varies from -0.256 (Site 3A) to 0.463 (Site 2), but
as a whole is weakly oblate (average $T = 0.155$). Samples tend to be more oblate, yet also more variable, within the lower section, and rather uniform throughout the upper section. $S_1$ does not appear dependent upon $T$, indicating that $k_1$ is a useful indicator of fabric orientation even when AMS ellipsoids are dominantly oblate. Site mean anisotropy ($P'$) displays less variability, ranging from 1.07 to 1.11 throughout the section; however, $P'$ is strongest and least variable in and around the boulder line (Sites 3A-4A).

**Magnetic Mineralogy**

Thermal demagnetization experiments (Fig. 2.7) display consistent patterns throughout the Dalby Till. Sharp declines in magnetization of low (0.1 T) and moderate (0.5 T) coercivities at 200 - 300 °C are consistent with maghemite, and final demagnetization at approximately 600°C is indicative of magnetite. Based upon the relative magnitudes of magnetization loss, maghemite is the dominant magnetic carrier. Maghemite, a more oxidized form of iron oxide, has been found in sediments associated with cold environments (Kodama 1982), and has been identified in AMS analysis of tills elsewhere (Gentoso et al. 2012; Hopkins et al. 2015). pARM spectra (Fig. 2.8) are also consistent throughout the Dalby Till, and reveal a peak in magnetization in the 30-40 mT range. Assuming maghemite and magnetite possess similar coercivities, we can use the coercivity – grain size relationships of Jackson et al. (1988), which indicate a grain size of approximately 2 microns for peak coercivities in this range.

**Surface Diamicton**

Fabrics of the surface diamicton are dominantly (4 of 5) oriented towards the NNE, and shallowly plunging (plunge ≤ 15°) (Fig. 2.9). Fabric strengths are variable, but
Figure 2.7 - Representative thermal demagnetization plots of the Dalby Till. Note the dramatic increase of slope of the low (0.1 T) and moderate (0.5 T) coercivity ranges at 200-300°C.
characteristically weak ($0.5098 < S_1 < 0.7425$). Similar to the Dalby till fabrics, mean sample shape factor indicates weakly oblate to triaxial anisotropy (mean $T = 0.172$); however, the magnitude of the anisotropy within the surface diamicton is low (mean $P' = 1.052$) when compared to the Dalby Till samples (mean $P' = 1.098$) (Fig. 2.6). AMS characteristics of the sample collected within the primary till section of investigation (Site 7) and those sampled at the second exposure (ST 1 – 4) are consistent.

Figure 2.8 - Partial Anhysteretic Remanent Magnetization (pARM) acquisition plots for the lower, upper, and boulder line sections of the Dalby Till. Note the consistent peak at the 30-40 mT range.
Discussion

As discussed above, till fabrics are a measure of the preferred orientation of grains and reflect local direction of shear, till transport and, by extension, local ice flow direction. It has been demonstrated that, in some locations, local ice flow direction (as
measured by azimuth of AMS fabric $V_1$) approximates regional ice flow, with slight variability (~30°) possible associated with bedforms or “sticky spots” (e.g. Gentoso et al. 2012; Hopkins et al. 2015). Therefore, it is reasonable to assume that the local flow directions recorded by AMS fabrics here approximate regional ice flow. Additionally, till fabrics in most field settings plunge in the up-ice direction approximately 15-30° (e.g. Harrison 1957; Evenson 1971; Gentoso et al. 2012; Hopkins et al. 2015); the same is observed in laboratory experiments where the shear plane is known to be horizontal (Iverson et al. 2008; Hooyer et al. 2008). Thus, we assume a horizontal to sub-horizontal shear plane. We interpret our AMS fabric results such that the $V_1$ of each fabric parallels regional ice flow direction at the time of deposition of the till horizon sampled, and that the direction of plunge indicates the up-ice direction when magnitudes of plunge are significant (greater than 10°). In cases where $V_1$ plunge is less than 10°, we attempt to identify up-ice direction based upon vertically neighboring fabrics.

The fabric orientations recorded in this section suggest changing ice flow direction during deposition of the Dalby till. The lower 3 meters records ice flow from the ENE. The boulder line fabric (Site 4) records ice flow from the S, above which ice flow appears to be dominated by flow from the SW. Site 4A, immediately above the boulder line, possesses a well-developed fabric plunging towards the NE. Due to the S and SW orientations of the fabrics above and below site 4A it is possible that the plunge in this fabric is not recording regional ice flow; instead, this plunge may be the result in a local change in the attitude of the shear plane or glacier bed. Therefore, ice flow direction, as recorded by the AMS fabrics, rotates from easterly to westerly initiating at or just below the boulder line. These patterns of ice flow variation differ somewhat from
the clast fabrics of Berglund & Lagerlund (1981); AMS fabrics suggest ice flow from the SW in the upper section of the Dalby till where they report flow from the NE. In this investigation we have sampled the Dalby till at a finer resolution than in previous investigations, and, as a result, our results display a more coherent pattern of ice flow change. It is possible that the Berglund & Lagerlund (1981) fabric is coincident with our site 4A; however, that site is inconsistent with neighboring fabrics. Ringberg (1988, 2003) argues for ice flow from the SW corresponding to the Young Baltic Advance, likely the Baelthav phase, in SW Skåne. Therefore, our fabrics suggest that at least the upper section of the Dalby Till was deformed by Young Baltic (likely Baelthav phase) ice flow from the SW sector.

Based upon the fabric orientations and the known Baltic provenance of this till it is possible that this Dalby till exposed in this section is all Young Baltic in age (Fig. 2.10). In this case the lower section dominated by ice flow from the ENE would represent an earlier phase (East Jylland), while the upper section records flow from the SW associated with the Baelthav phase. The dramatic rotation is unlikely to reflect changing flow or orientation of the ice stream as a whole; instead, it likely records changing ice flow along the margins of the ice lobe during readvance. The site is very close to a NW-SE oriented discontinuity in the ice flow pattern, with previous authors (e.g., Lidmar-Bergström et al. 1991; Ringberg 1988 figure 11E) portraying a reentrant developing in the ice margin immediately NW of the Dalby site during deglaciation. Hence, the site may have been located within the ice stream proper during early stages, but later occupied an extreme right-lateral position in the extremely divergent terminal lobe. This shifting location within the ice stream/lobe system may fully explain the
drastic shifts in flow direction. Retreat of the main FIS covering the Swedish mainland and bordering the BIS to north appears to have continued during the Baelthav phase (Houmark-Nielsen & Kjær 2003).

The coincidence of the boulder line with the shifting direction of ice flow is perhaps indicative of changing bed conditions near the margin of the BIS at this locality. Additionally, the boulder line is coincident with a changing pattern in fabric strength described above (Fig. 2.6). It has been demonstrated in experimental work (Iverson et al. 2008; Hooyer et al. 2008) that the $S_1$ eigenvalue is a useful indicator of shear strain.

Figure 2.10 - Two possible interpretations of the glacial history recorded in the AMS fabrics of the Dalby Quarry. The Dalby till could be entirely Young Baltic in Age and record changing ice flow direction during deposition, possibly as a consequence
Figure 2.10, cont. - of the marginal position of Dalby with respect to the Baltic Ice Stream (Left). Alternatively, the lower section of the Dalby Till could be associated with the Old Baltic or LGM advances and be separated from the Young Baltic by the boulder line (an erosional unconformity).

Boulder lines within till are lag deposits marking the former position of the ice-till interface (Benn & Evans 2010). It has been postulated that intra-till boulder lines separate an advance phase till from the retreat phase till, and arise due to rapid descent (erosion) of the deforming bed when the glacier was at or near its maximum extent (Boulton 1996). The patterns of till deformation recorded here may support this interpretation for the Dalby Till (Fig. 2.11); the increasing and decreasing shear indicated by the $S_1$ trends may indicate ice thickening (advance) and thinning (retreat), respectively.

Alternatively, it is possible that boulder line represents an erosional unconformity between Young Baltic and LGM or older till (Fig. 2.10); past researchers have attributed parts of the Dalby Till to these older advances (Berglund & Lagerlund 1981; Ringberg 1992; Anjar et al. 2014). Our fabrics indicate two phases of flow: an earlier phase dominantly from the ENE followed by flow from the SW; therefore, it is unlikely that the Dalby Till records Old Baltic, LGM, and Young Baltic advances, as suggested by Berglund & Lagerlund (1981). It is possible that the lower section is Old Baltic in age (Ringberg 1992); however, that would require the boulder line to represent a large unconformity (~17ka) and the absence or erosion of an LGM till.
Figure 2.11 (previous page) - Schematic illustration of the ice stream till kinematics that could produce the boulder line and AMS fabric characteristics. Note, the advance and retreat phases of ice behavior are likely occurring along the margin of the BIS at Dalby.

In this case, the apparent rotation can be explained as an “overprinting” or reorganization of the preexisting fabric. However, the lack of any compositionally different material between these two till units, and the lack of an explanation for the observed trends in fabric strength discussed above is troubling. Finally, it is possible that the lower section
of this exposure is associated with the Baltic Ice during LGM advance, as suggested by Anjar et al. (2014); however, this interpretation lacks an explanation for the till kinematics inferred from the fabric strengths.

Our results are consistent with an interpretation of the surface diamicton as a subglacial till weakly deformed by ice flow from the NNE. Deformation and transport distance was low, preventing the generation of faceted, striated clasts; however, significant erosion must have occurred at the base of the unit to erode the Dalby Till such that it is only now present in bedrock depressions. This unit indicates that following the Baelthav phase of the Young Baltic Advance, a subsequent, short-lived ice advance of FIS from the NNE covered the region in the proximity of Dalby and truncated the Dalby Till at this location.

Conclusions

AMS fabric analysis of the Dalby Till deposited by the Baltic Ice Stream in Skåne, southern Sweden, reveals a rotation of ice flow from the ENE to the SW coincident with a boulder line. This pattern reflects changing ice flow direction of the Baltic Ice Stream during one or multiple ice advances. Additionally, the magnitudes of deformation recorded within the till (fabric strengths) are consistent with expected patterns of ice advance and retreat, separated by an intra-till boulder line. The surface diamicton at this locality possesses weak fabrics; however, these fabrics are consistently oriented from the NNE. Further investigation into the surface diamicton’s magnetic properties and deformation kinematics is needed to explain the variable strengths of the fabrics and to assess its function as a record of the basal mechanics of the FIS. Detailed reconstruction
of the glacial history is difficult, as the regions stratigraphy likely records complex interactions the Baltic Ice Stream and the main body of the Fennoscandian Ice Sheet. However, this study demonstrates that AMS is a useful tool for assessing ice flow history as well as till deformation within individual till units and, thus, can provide insight into the temporal evolution of subglacial conditions at the bed of paleo-ice streams.

**Acknowledgements**

We thank Leo Kramár of Dalby-Sydsten and Dr. Grahame Larson for support in the field, and Christopher Geist for support in the laboratory analysis. This work was funded by a grant to Johan Kleman from the Swedish Research Council, in addition to gifts to the Lehigh University Dept. of Earth & Environmental Science from Tom Pasquini and ExxonMobil.

**References**


Chapter 3:
Magnetic fabric and the distribution of shear within stratified basal ice, Matanuska Glacier, Alaska, USA

Hopkins, Nathan R.¹

Edward B. Evenson¹

¹Earth and Environmental Sciences, Lehigh University

Intended for publication in Journal of Glaciology:
anticipated submission, December, 2015
Abstract

Herein we present the results of an anisotropy of magnetic susceptibility (AMS) investigation of the fabric of the stratified facies of the debris-rich basal zone which is common to many glaciers and ice streams - most notably at the Matanuska Glacier, Alaska, where it was first described. The debris-rich basal zone is a potentially significant component of glacier motion, and is thought to be rheologically distinct owing to its debris content. At the Matanuska glacier, stratified basal ice occurs as a typically thin (1 to 15m, where stacked by thrusting) zone at the base of the glacier. It is best exposed at or near the terminus during glacial advance in the winter. This facies, which is enhanced by winter sublimation, is composed of distinct, semi-continuous, alternating layers of debris-poor and debris-rich ice that typically dip up-glacier on the order of 15°. It is thought to be genetically related to debris-rich frazil ice forming subglacially as silt-laden supercooled water emerges from overdeepenings at the glacier bed and underplates the overlying glacial ice; however, the mechanism by which the randomly distributed and randomly oriented debris within the frazil ice becomes separated from the ice and organized into sub-parallel up-glacier dipping strata is not well understood. It is well-known that AMS fabric is a useful indicator of shear, and as such has the potential to elucidate the processes leading to the formation of the distinct stratification present in this zone. In the winter of 2014-2015 we collected oriented samples of the stratified basal ice from a vertical column for AMS investigation to assess the presence and distribution of deformation within the stratified facies. Fabrics indicate rheological inhomogeneities associated changes in debris-content. Fabrics indicative of simple shear along sub-horizontal shear planes are dominant and associated with the debris-rich
horizons. In contrast, fabrics in debris-poor horizons indicate either no shear or pure shear. Interestingly, fabrics do not consistently reflect the plane of the stratification. These results suggest variation in debris-content and ice crystal size influence the rheological behavior of stratified basal ice, and that the stratified basal ice observed at glacier margins is the product of shear deformation following freeze-on.
Introduction

A basal zone of ice that interacts with the glacier bed is common to many glaciers and ice streams. This zone is commonly debris-rich and free of air bubbles, relative to the englacial ice derived from snow, indicating its subglacial origin. The basal zone also commonly possesses internal structures, such zonation, banding, and folds. Basal ice is derived from subglacial waters – both glacial melt and surface-derived water – which freezes and incorporates debris by processes of regelation or hydraulic supercooling (Alley et al. 1998). Within the basal zone, there is commonly a layer several meters thick that exhibits a banding – termed “stratification” by Lawson (1979) – that arises from alternating layers of debris-poor and debris-rich ice (Fig. 3.1). The mechanism by which this stratification arises is poorly understood, and might either reflect pseudo-seasonal hydraulic fluctuations at the glacier bed or be a foliation resulting from post-freeze-on segregation and deformation within the basal zone (Alley et al. 1998). The origin of this stratification has implications for the velocity structure of glaciers and ice streams, and processes of subglacial sediment entrainment and transport.

Fabric – the preferred orientation of clasts, grains, and particles – is a common tool used to inform depositional processes and post-depositional deformation. In glacial geology, “macro-fabric” refers to the orientation of macroscopic grains, commonly pebbles, and is an easy, low-cost, and useful tool for investigating glacial processes. However, this tool is limited by grain size (i.e., cannot measure fabric of fine-grained sediments) and the potential for unintentional user-bias. Anisotropy of magnetic susceptibility (AMS) has a long history of application in structural geology as an indicator of strain in rocks (e.g., Borradaile 1988; Pares 2004; Ferre et al. 2014). AMS is
Figure 3.1 - Basal ice exposure in the vicinity of our sample location (see figure 3.2). Photograph looking towards the SE. Note the abrupt transition from the clean englacial ice to the debris-rich stratified ice, and the consistent up-glacier dip of the debris-rich horizons.

dominantly controlled by shape of magnetic grains (Tarling & Hrouda, 1993), and thus provides a robust, volume-averaged, and objective alternative to macrofabric for the investigation of unconsolidated fine-grained sediments. Laboratory experiments (Hooyer et al. 2008; Iverson et al. 2008) have clearly demonstrated that sheared sediments (tills) rapidly develop magnetic fabrics that reflect the orientation and magnitude of shear, as well as the state of strain. Numerous field studies have confirmed the laboratory results in subglacial tills in diverse geologic settings (Shumway & Iverson 2009; Thomason & Iverson 2009; Gentoso et al. 2012; Ankerstjerne et al. 2015; Hopkins et al. 2015; Vreeland et al. 2015), and Fleming et al. (2013) demonstrated the application of AMS to
basal ice of a polythermal surge-type glacier in Svalbard. Herein, we evaluate the vertical distribution of AMS fabric within the stratified basal ice of the Matanuska glacier, Alaska, U.S.A., to address the presence, distribution, magnitude, and state of strain within the basal ice.

Figure 3.2 – Location map of south-central Alaska, including the Anchorage lowland and the Matanuska valley.
Figure 3.3 – Lidar-derived hillshade map of the terminus of the Matanuska glacier. The blue shaded region indicates the active, debris-free portion of the glacier, and the prominent overdeepenings of Lawson et al. (1998) are indicated by the black, hatched lines. Sample locality is indicated by the black circle.

Field Setting

The Matanuska glacier (Fig. 3.2), located ~150 km NE of Anchorage, AK, is an approximately 48 km long temperate glacier which drains NNW from the ice fields of the Chugach Mountains. The Matanuska is the ideal location for this study as it represents a type-locality for stratified basal ice, provides excellent exposures, and has a particularly well-developed stratification. The glacier has been remarkably stable throughout much of the Holocene (Williams & Ferrians 1961). The terminal lobe of the Matanuska (Fig. 3.3) is approximately 4 km wide, the eastern portion of which is presumably stagnating and
covered with superglacial debris; the western, debris-free portion is active and is the location of this study.

During winter months, ice advance exposes the base of the glacier and provides access to the basal stratified ice. This phenomenon provides excellent exposure through the englacial ice, the basal stratified ice, and subglacial sediments. The basal zone of the Matanuska was first described in detail by Lawson (1979a, 1979b). The Matanuska is a remarkably well-studied glacier, so the discussion that follows is largely a summary of the relevant details from Lawson (1979a, 1979b, 1981), as well as additional relevant observations published in the literature.

The basal zone of the Matanuska ranges from 1-15 meters thick when exposed; however, the thicker exposures are packages of stacked ice. It has been observed, at one time or another, along the entire width of the active portion of the glacier, and has been detected geophysically and traced up-glacier approximately 300 meters (Arcone et al. 1995; Lawson et al. 1998; Baker et al. 2003). Lawson (1979a, 1979b) conducted detailed descriptions of the englacial and basal zone, and identified ice facies useful for characterizing the ice. The englacial ice – the debris free glacier ice constituting most of the glacier – can be divided into a diffuse and banded facies; however, the englacial zone is outside the scope of this work and can be largely ignored for this study. The basal zone is characterized by two facies: dispersed and stratified. The dispersed facies is commonly found directly below the englacial ice, and the transition between the two is abrupt. This facies can be thin (~ 0.2 m) to thick (8 m), and in some localities nonexistent. Debris within the dispersed facies is uniformly distributed and rarely clustered; debris typically represents 8% of the sample by mass.
The stratified facies (Fig. 3.1) underlies the dispersed facies, and the contact is well-defined, but irregular. The defining characteristic of this facies is a stratification which arises from alternating layers of debris-poor and debris-rich ice. Debris-poor layers are composed of fine grained ice, and are generally bubble-free; when present, air bubbles are elongate in the direction of ice flow. The sedimentology of stratified basal ice was described in detail by Larson et al. (in press). Within these layers, debris is commonly present in irregularly shaped silt aggregates. The debris-rich layers are composed predominantly of silt to coarse sand. In some cases layers are ice-supported, in others ice is only present in the interstices between clasts. Gravels and cobbles are occasionally present, and are mostly subround and lack striations. Layers are generally semi-continuous at the scale of the exposure, and commonly dip up-glacier 10 to 15°. Debris layers are commonly 1 - 3 cm thick, and extend for several meters.

The freeze-on process for the basal zone of the Matanuska has been clearly identified as hydraulic supercooling of subglacial waters rising from an overdeepening near the glacier terminus (Alley et al. 1998). An overdeepening beneath the active portion of the glacier is readily identified by a zone of heavy crevassing immediately up-glacier of the modern glacier terminus, and has been identified geophysically (Lawson et al. 1998; Baker et al. 2003). In the summer months, frazil ice can be observed forming at vents near the glacier terminus (Lawson et al. 1998; Evenson et al. 1999). This frazil ice is produced as pressure is released from supercooled water as it rises out of the overdeepening, resulting in rapid ice crystal growth and entrainment of suspended and, in some cases, bedload sediments (Alley et al. 1998). Isotopic studies of the basal zone of the Matanuska indicate that rainfall runoff, in lieu of glacial melt, is an important
component of the basal waters, and that the basal zone ice post-dates the production of bomb-produced tritium in the atmosphere (Lawson and Kulla 1978; Strasser et al. 1998; Titus et al. 1999).

**Methods**

The basal zone of the Matanuska Glacier was sampled in November 2014 at one location on the western front of the active portion of the glacier. This location provided an approximately 1.5 meter thick section of the stratified facies underlying a thin (∼4cm) section dispersed facies. Here, the layering within stratified facies displays a relatively uniform up-glacier dip of approximately 15°. At the section sampled, the basal zone was capped by approximately 0.5 meters of clean (debris-free), englacial ice.

Samples were collected for AMS analysis and determination of debris content from one vertical column extending from the upper contact of the basal zone to the base of the exposure (Fig. 3.4). Large (15 cm * 15 cm * 10 cm) blocks were cut using a concrete saw, oriented using a Brunton compass, and extracted from the exposure using a chisel and hammer. Orientation marks were carved into the ice blocks, and the samples were tightly wrapped with plastic wrap and bagged to minimize sublimation during storage. Each large block was subsampled into two to three parallel, sub-horizontal slabs, and each slab sampled into 8 cm³ cubes and inserted into standard paleomagnetic sample boxes at Lehigh University. Fabric was analyzed for each slab, and a composite fabric was produced for each sampled block. A total of 267 samples were collected for AMS analysis.
Figure 3.4 – Sample strategy for the approximately 1.5 m section of basal ice. The snow-covered upper portion is debris-free englacial ice.

AMS was measured at Lehigh University using a KLY-3S Agico Kappabridge. For each sample, susceptibility was measured in 15 static orientation and fit by least-squares method to an AMS ellipsoid defined by the maximum ($k_1$), intermediate ($k_2$), and minimum ($k_3$) susceptibility axes. The anisotropy within a sample is characterized using the anisotropy ($P'$) and shape ($T$) factors (Hrouda 1982):
\[ P' = \exp \left( 2 \sum (\ln K_i - \ln K_m)^2 \right) \]

\[ T = \left[ \frac{2 \ln K_2 - \ln K_3}{\ln K_1} \right] - 1 \]

Following the convention within glacial geology, the fabric orientations are characterized using the eigenvalue method of Mark (1973), wherein a symmetric second-rank tensor is fit to the distribution of a given susceptibility axis (e.g., \( k_1 \)). That distribution of axes is then characterized by principal eigenvector orientations (\( V_1, V_2, \) and \( V_3 \)), and their associated eigenvalues (\( S_1, S_2, S_3 \)) indicate the degree of clustering about that axis. \( V_1 \) represents the mean orientation of the axes, \( V_2 \) lies \( 90^\circ \) to \( V_1 \) within the best-fit plane through the distribution, and \( V_3 \) represents the pole to that plane. In the fabric analysis of till, it is common to analyze fabric using the \( k_1 \) orientation, which becomes oriented parallel to the shear direction and possesses a characteristic up-glacier plunge relative to the shear plane (e.g., Hooyer et al. 2008; Iverson et al. 2008). However, for fabrics that are poorly clustered, fabric can be additionally characterized utilizing the additional (\( k_2 \) and \( k_3 \)) susceptibility axes. Fabric shapes can be characterized by their eigenvalues using the ternary plots of Benn (1994) using the following indices: fabric isotropy (\( I = S_3/S_1 \)) and fabric elongation (\( E = 1 - S_2/S_1 \)). Characterizing fabrics using the eigenvalue method applied to each axis independently may lead to erroneous principal axis orientations for poorly clustered and girdled fabrics; therefore, \( V_1 \) orientations are only reported for clustered (\( E > 0.5 \)) data sets, following the methods of Ankerstjerne et al. (2015). We apply this analysis to each of the three axes to determine fabric shape and to evaluate the state of strain recorded in the sample, following methods previously applied to subglacial tills (Ankerstjerne et al, 2015).
Results

Anisotropy

Individual sample anisotropy (Fig. 3.5) varies considerably in both degree ($P'$) and shape ($T$), as has been observed in AMS investigations previously (Gentoso et al. 2012; Hopkins et al. 2015). $T$ is dominantly triaxial - oblate (positive), but range from nearly perfectly prolate (-0.96) to perfectly oblate (0.908). Within site variation of $T$ is also large (mean standard deviation of 0.4). $P'$ shows significantly less variability, except for those samples with extremely low (less than $10^{-5}$) to negative susceptibilities which possess erroneously high $P'$ (Hrouda 2004). For a given site with broadly similar susceptibilities and debris concentrations, variability of $P'$ is low (standard deviations $< 0.05$).

Bulk Susceptibility and Debris Content

Average debris content within this section is 40.6 ±17.3%; however, individual samples range from 0.08 to 74%. The distribution of debris within the basal zone (Fig. 3.6) is consistent with previous observations (Lawson 1979): Low debris within the englacial and dispersed facies which abruptly increases into the stratified facies. The minor variability within the stratified facies likely reflects varying density, thickness, and morphology of the debris-rich layers. Bulk susceptibility displays only minor variability throughout the section, except for at sites with significant variation in debris content. This likely reflects the relative dominance of ferrimagnetic debris and diamagnetic ice contributing to the AMS signal.
Figure 3.5 - Distribution of degree of anisotropy ($R'$) and shape ($T$) for each sample analyzed. Positive (negative) $T$ indicates oblate (prolate) AMS ellipsoids.

Figure 3.6 – Bulk susceptibility (left) and debris content (right) for each sample block. Individual debris content measurements are indicated using open squares. Sample block means and standard deviation presented in closed squares.
Fabric

Composite AMS fabrics for each sampled block are presented in Figure 3.7. Fabric site 1, the uppermost fabric spanning the englacial, dispersed, and stratified facies of ice possesses a near random fabric for all the susceptibility axes. This site is also characterized by low debris content (17 ± 18 wt. %). All lower sites are exclusively within the stratified facies of ice, and possess anisotropic fabrics. Four of these 6 fabrics display significant clustering of their k₁ axes (S₁ > 0.6, E > 0.5), and the V₁ trend for these fabric parallels to sub-parallels known ice flow direction (NW). These sites generally display insignificant plunge (≤ 5°) down-glacier, except for the lower-most fabric (Site 7). Site 7 possesses the strongest fabric of this section (S₁ = 0.8565) and plunges 16° up-glacier.

Fabrics shapes, characterized using the ternary plots of Benn (1994) are shown in column 5 of figure 3.7. Three of the four fabrics discussed above possess significant clustering about all three axes, with Site 4 (for which k₂ is unclustered) being the exception. The two sites discussed above (3, 6) whose k₁ axes do not parallel ice flow display girdle fabrics within the k₁-k₂ plane. All fabrics within the stratified facies possess significant clustering about the k₃ axis, and in all cases this axis is near-vertical.

The bulk k₁ fabric of all samples (Fig. 3.8) in this section parallels ice flow (V₁ = 316°, 2°), but displays weak clustering (S₁ = 0.5506). Fabric shape (I = 0.213; E = 0.403) indicates that the k₁ axes are weakly girdled in the horizontal plane, which is to be expected from averaging the clustered, girdled, and isotropic fabrics displayed above. On
Figure 3.7 - Composite AMS fabrics for each sample block. Fabrics are presented in lower-hemisphere stereographic projections, with north towards the top of the page. Individual AMS $k_1$ (red), $k_2$ (green), and $k_3$ (blue) axes are displayed using small squares; axis $V_1$ orientations are presented using large squares for axes with $E > 0.5$. Column 2 (from left) presents $k_1$ orientations with 2-sigma kamb contours. Column 3 presents $k_1$ rose diagrams. Fabric $k_1$ statistics are shown in Column 4.
Figure 3.7 (continued) - Fabric shape diagrams are presented in Column 5. Simple site descriptions are shown in Column 6.

Figure 3.8 - All AMS $k_1$ orientations (small squares) without respect to sample locations and contoured using 2-sigma kamb contours, $V_1$ (large red square), and ice flow direction.

closer inspection, the girdle is better described as a bimodal distribution, with one cluster parallel to ice flow ($315^\circ$) and another oriented perpendicular to flow. If the samples from the nearly isotropic, dispersed facies ice is excluded, anisotropy in the sample is improved ($I = 0.152$), while there is little change in the fabric elongation ($E = 0.438$).

Each composite fabric (sites 1 – 7) is composed of 2 – 3 slabs, and investigating fabric at the slab-level allows us to evaluate the variability of fabric (and thus, shear) at a finer resolution (Fig. 3.9). Ignoring plunge direction and treating $V_1$ as a line, 14 of 19 fabrics possess $V_1$ orientations subparallel to ice flow. The distribution of $k_1$ axes within each horizon is anisotropic ($I < 0.5$), and for 12 of the 19 fabrics anisotropy is significant ($I < 0.1$). The $k_1$ axes of twelve fabrics are significantly clustered ($E > 0.5$).

Interestingly, all fabrics that do not approximate ice flow within 45° are clustered,
Site 1-1
n = 12
V_1 = 290\degree, 26\degree
S_1 = 0.513
I = 0.242
E = 0.292

Site 1-2
n = 14
V_1 = 225\degree, 44\degree
S_1 = 0.463
I = 0.411
E = 0.250

Site 1-3
n = 11
V_1 = 174\degree, 26\degree
S_1 = 0.524
I = 0.416
E = 0.508

Site 2-1
n = 17
V_1 = 315\degree, 05\degree
S_1 = 0.842
I = 0.078
E = 0.891

Site 2-2
n = 20
V_1 = 306\degree, 12\degree
S_1 = 0.677
I = 0.028
E = 0.552

Site 2-3
n = 18
V_1 = 92\degree, 14\degree
S_1 = 0.525
I = 0.194
E = 0.288

Site 3-1
n = 12
V_1 = 272\degree, 12\degree
S_1 = 0.557
I = 0.049
E = 0.255

Site 3-2
n = 13
V_1 = 56\degree, 01\degree
S_1 = 0.700
I = 0.039
E = 0.610

Site 3-3
n = 12
V_1 = 31\degree, 12\degree
S_1 = 0.637
I = 0.089
E = 0.520

Site 4-1
n = 7
V_1 = 100\degree, 04\degree
S_1 = 0.785
I = 0.003
E = 0.729

Site 4-2
n = 15
V_1 = 347\degree, 00\degree
S_1 = 0.510
I = 0.023
E = 0.062

Site 4-3
n = 15
V_1 = 329\degree, 05\degree
S_1 = 0.825
I = 0.006
E = 0.798

Site 5-1
n = 23
V_1 = 123\degree, 01\degree
S_1 = 0.660
I = 0.075
E = 0.559

Site 5-2
n = 16
V_1 = 306\degree, 01\degree
S_1 = 0.721
I = 0.011
E = 0.624

Site 6-1
n = 12
V_1 = 28\degree, 17\degree
S_1 = 0.795
I = 0.058
E = 0.780

Site 6-2
n = 10
V_1 = 331\degree, 43\degree
S_1 = 0.603
I = 0.173
E = 0.515

Site 6-3
n = 8
V_1 = 295\degree, 38\degree
S_1 = 0.694
I = 0.151
E = 0.711

Site 7-1
n = 16
V_1 = 146\degree, 16\degree
S_1 = 0.883
I = 0.010
E = 0.877

Site 7-2
n = 15
V_1 = 141\degree, 16\degree
S_1 = 0.830
I = 0.026
E = 0.821
Figure 3.9 (previous page) – Slab fabrics arranged according to location within column. Stereographic projections plotted as in figure 6. Principal eigenvectors and eigenvalues for the k₁ axes are provided, as well as fabric shape factors.

while several fabrics that parallel ice flow possess low E values (E < 0.5) indicating girdling within the V₁-V₂ plane. Fabric shape parameters do not correlate with shape and/or magnitude of the sample anisotropy (Fig. 3.10), as has been observed elsewhere. In general, a given composite block contains slabs of better and lesser oriented slabs, and therefore the composite signal is weaker. For example, site 4 contains two slabs with well clustered fabrics (S₁ > 0.75, E > 0.7) that agree with ice flow direction and one girdled fabric (S₁ = 0.51, E = 0.062). The resulting composite fabric remains parallel to ice flow; however the fabric strength is decreased.

As k₁ for well-clustered datasets is expected to approximate direction of shear, we can omit the girdled and isotropic fabric from Table 1. Figure 3.11 displays the V₁ orientation of the k₁ axes for each slab-level fabric with E > 0.5 (A), and the V₁ orientations of the k₃ axes for each slab-level fabric with E < 0.5. For the clustered k₁ fabrics, it is apparent here that the dominant orientation is parallel to ice flow (NW). A second, lesser group of axes clusters orthogonal to ice flow (NE). The V₁ of this fabric is 319°, 05° (S₁ = 0.6480). This distribution is clearly anisotropic (I = 0.147) and reasonably clustered (E = 0.605). When k₁ axes are girdled, k₃ axes are consistently vertical.

83
Figure 3.10 - Sample anisotropy parameters as a function of $k_1$ fabric strength ($S_1$).

Note the lack of correlation.

Figure 3.11 – Lower-hemisphere stereographic projection of $V_1$ orientations of $k_1$ axes for 14 slabs possessing clustered ($E > 0.5$) $k_1$ fabrics, contoured using kamb 2-sigma contours. Note the flow-parallel and flow-perpendicular fabric components.
Discussion

The presence of fabric is an indicator of strain. This dataset is dominated by triaxial fabrics oriented sub-parallel to ice flow direction, indicating simple shear along sub-horizontal shear planes within the stratified basal ice. These results are consistent with macrofabrics from the basal zone of the Matanuska glacier (Lawson 1979; Hart 1995), and in general agreement with the AMS fabric results of Fleming et al. (2013), who reported triaxial fabrics consistent with simple shear parallel to ice flow direction within stratified basal ice of Tunabreen, Svalbard. Macrofabrics of the Matanuska are commonly moderately strong and plunge up-glacier, commonly within the plane of the stratification (Lawson 1979), as would be expected under simple shear and is commonly observed in subglacial tills. Up-glacier plunge is also observed in the magnetic fabrics of Fleming et al (2013). At the Matanuska, significant up-glacier plunge is only observed for the strongest fabrics. Simple shear within the basal zone has been suggested previously based upon macroscopic structures, including folds, boudinage, pressure shadows, and rotational clasts (Hubbard & Sharp 1989; Hart 1998). These features are rarely observed at the Matanuska Glacier. Fabrics where the k₁ & k₂ axes are girdled (3 out of 19 fabrics) likely indicate pure shear parallel to the orientation of the k₃ axes. In the basal ice, the direction of pure shear appears to be nearly vertical. These fabrics resemble compaction fabrics measured in sedimentary rocks (e.g., Parés et al. 1999), and have been observed in subglacially deformed till (Ankerstjerne et al. 2015).

The patterns in fabric shape and orientation observed here are suggestive of debris-dependent rheological inhomogeneities within the stratified basal ice. Debris-poor samples possess random to girdle fabrics (pure shear), while debris-rich sampled possess
well-developed, triaxial fabrics (simple shear). The magnitude of shear within debris-rich layers is perhaps variable; however, in most cases it is not insignificant. Fabrics strengths reported here are comparable to AMS fabric strengths of deforming till beds (Shumway & Iverson 2009; Gentoso et al. 2012; Hopkins et al. 2015; Vreeland et al. 2015). Experimental studies to quantify, in relative terms, magnitudes of pure shear using fabric strengths are not presently available.

The rheological behavior of basal ice is significant for glacial dynamics, including subglacial sediment transport and glacier velocity/deformation profiles. However, contributing factors to deformation within ice-debris mixtures are numerous and poorly understood (Hubbard & Sharp 1989; Benn & Evans 2004; Moore 2014). This complication arises due to numerous contributing and competing mechanisms of creep in a mixture of debris and ice (Moore 2014). It is commonly held that an ice-debris mixture is stronger than its pure end-member components (Ting et al. 1983); however, a thorough review of the literature demonstrates the relationship between debris-content and material strength is complex (Moore 2014). Significant factors include debris content, ice crystal size, ice fabric, and the presence of unfrozen water. While some researchers have presented field evidence suggesting that increasing debris-content weakens ice and allows for increased and more rapid deformation (Echelmeyer & Zhongxian 1987; Brugman 1983; Cohen 2000), laboratory studies often indicate an apparent “hardening” of ice associated with increasing debris (e.g., Hooke 1972; Baker 1979; Nickling & Bennet 1984). This discrepancy potentially reflects the competing relationships between the three variables discussed above. Crystal size impacts ice strength, such that fine-grained ice tends to deform more readily, and appears inversely related to debris content (Baker
Therefore, increasing debris-content requires progressively smaller ice crystals such that ice in debris-rich layers can be largely interstitial. In fact, the strain rate of ice appears directly related to debris content (Fisher & Koerner 1986). Microstructures identified within the debris-rich basal zone of the Taylor Glacier suggest the importance of unfrozen water, which may allow slip across interfacial water films (Samyn et al. 2005, 2009). Finally, a well-developed ice fabric is known to dramatically increase the ability of ice to deform as much as 100 to 1000 times (Hooke 2005); however, ice fabric is logistically difficult to observe in debris-rich ice. Thus, it is possible that deformation is concentrated in the debris-rich layers of the stratified basal ice as a result of decreased ice grain size and greater availability of unfrozen water associated with high debris content.

There is no apparent trend in fabric strength throughout the column sampled here, which suggests that, although deformation appears concentrated along debris-rich horizons, deformation is approximately equally distributed throughout its thickness, contrary to some past observations (Hart 1995). This is consistent with a lack of macroscopic variation throughout the column. Lastly, the orientation of these fabrics are largely sub-horizontal and do not tend to lie within the plane of the stratification, except for the most well-developed fabrics (Site 7). This discrepancy is at odds with our interpretation of localized simple shear along debris-rich layers; however, these fabrics are generally only weakly to moderately developed. Under continued shear, we expect fabrics to better reflect the stratification, as at site 7.

As the basal zone is deforming, it is unlikely that the stratification is a primary feature of the freeze-on process. Therefore, we suspect that the stratification is a foliation
that develops in response to deformation and debris-migration post-freeze-on of debris-rich frazil ice. If this is the case, that transition from frazil to stratified ice takes place over, at most, a few hundred meters once the basal ice makes contact with the bed and is transported out of the overdeepening (Lawson et al. 1998; Baker et al. 2003). Though this distance at the Matanuska is short, as the overdeepening is immediately adjacent to the glacier margin, debris-rich and stratified basal ice has been observed at numerous glaciers and ice streams, and its contribution to glacier motion and sediment transport/entrainment is poorly understood. If deformation within the ice is dependent upon debris-content, then ice deformation (velocity) may vary spatially and temporally depending upon sediment and water availability at the ice-bed interface.

Conclusions

Anisotropy of magnetic susceptibility (AMS) fabric analysis within stratified basal ice of the Matanuska glacier confirms previous suggestions and hypotheses regarding the occurrence of simple shear within the basal zone common to many glaciers and ice streams. Furthermore, our analysis indicates rheological inhomogeneity predominantly controlled by the distribution of debris. Simple shear parallel to ice flow appears to dominate debris-rich horizons, whereas pure shear is present in debris-poor horizons. Pervasive deformation within the basal ice precludes preservation of primary structures associated with the freeze-on process, and suggests that the stratification results from deformation following freeze-on. Further work is needed regarding the physical distribution and morphology of debris and ice to constrain variables controlling the rheology of stratified basal ice.
Acknowledgements

This work benefited tremendously from helpful suggestions and discussion with Ken Kodama, Daniel Lawson, Grahame Larson, Josh Stachnik, Claudio Berti, and Tom Pasquini. Additional thanks to Tom Pasquini and ExxonMobil for partially funding this research. Endless thanks to Bill and Kelly Stevenson for their graciousness in providing lodging, logistical help, and entertainment.

References


Hart, J. K. 1998: The deformed bed/debris-rich basal-ice continuum and its implications for the formation of glacial landforms (flutes) and sediments (melt-out till). *Quaternary Science Reviews* 17, 737-754.


Conclusion

Bed deformation is a key process of glacial motion and the erosion, transport, and deposition of till. This is evident from observations at modern glaciers and ice sheets, despite their limited number. Our knowledge about the present spatial and temporal distribution of bed deformation is less clear, and even more so for previous glaciations. However, the application of new and quantitative methods, such as anisotropy of magnetic susceptibility (AMS), and detailed analysis of the Pleistocene sedimentary record has the potential to provide valuable insight into these problems. At present, the research community is engaged in great efforts to assess the basal dynamics of Wisconsinan and modern glaciers and ice sheets through remote sensing, terrain analysis, sedimentologic analysis, and modeling.

This dissertation detailed three research projects seeking to assess the kinematics, distribution, and variability of deformation in the basal environment of glaciers and ice sheets. Through unique sampling strategies in three unique glacial settings, I have demonstrated the strength of AMS fabric analysis for the characterization of till kinematics. In the Weedsport drumlin field of New York, AMS fabrics capture the spatial variability of till kinematics and bed deformation associated with obstructions at the ice-till interface (i.e., bedforms). Furthermore, we now have significant evidence that these drumlins are the product of syndepositional streamlining of till deposited at the drumlin locality by way of complex ice flow paths resulting from interaction of the ice body with the protodrumlin form. AMS fabrics were also capable of characterizing systematic stratigraphic variation in till kinematics within the Baltic Ice Stream’s Dalby till. In addition to producing a history of ice flow direction, I was able to characterize
systematic changes in fabric strength and, thus, shear. Lastly, AMS fabrics were used to characterize strain within the stratified debris-rich basal ice of the Matanuska Glacier, Alaska. The results demonstrate the presence of shear, and suggest that the basal ice is rheologically controlled by the distribution and morphology of debris.

Collectively, the above results illustrate a complex basal environment where shear is pervasive, yet spatially and temporally variable in direction, magnitude, and nature. This research adds to a body of literature that demonstrates the deforming bed is significant, yet the system is complicated and poorly understood. However, this dissertation also illustrates the power of AMS, and through the greater application of this tool and the integration of localized studies of till kinematics, broad analyses of glacial geomorphology, and numerical modeling we may be able to more effectively capture and explain the kinematics and distribution of bed deformation and to further assess its significance for the dynamics of glaciers and ice sheets.

Significant questions remain regarding the process of bed deformation, as well as for the AMS technique when applied to glacial sediments. First, more work is needed on characterizing the controls on, and the carrier of, the AMS signal of glacial sediments. For example, most AMS fabric studies in glacial geology and glaciology have investigated tills dominated by a ferrimagnetic mineralogy. However, glacial sediments may often contain high proportions of clays, and thus the sediments may be dominated by a paramagnetic mineralogy. We do not yet have the experimental evidence for the development of fabric within these tills. Deconvolving the AMS signal is crucial in order to discriminate between intrinsic magnetic properties of the sediment and those resulting
from deformation, and without this knowledge we may be prone to erroneous conclusions.

The community also needs to continue to address the controls on the deformation of tills. We should expect factors such as grain size, compaction, basal hydrology, and thermal conditions to influence the deformation of till and the resulting fabric. Without this knowledge it will remain difficult to address the greater question of the spatiotemporal variability of bed deformation, particularly over broad regions. Ultimately, that is the most significant question that AMS fabric analysis of tills has the potential to address: at what spatial and temporal scales does the nature of bed deformation vary and what is its geologic consequence?
Vitae

Nathan R. Hopkins
Earth and Environmental Sciences
Lehigh University
1 W. Packer Ave
Bethlehem, PA 18015

Education

Ph.D. Candidate, Earth and Environmental Sciences
Lehigh University, Bethlehem, PA
Advisor: Dr. Edward Evenson
Magnetic Till Fabric: Applications of anisotropy magnetic susceptibility
(AMS) to subglacial deformation of till and ice

Bachelors of Science in Geological Sciences, Aug. 2011
University of Missouri
Graduated Cum Laude with Honors
University of Missouri, Columbia, MO
Advisor: Dr. Francisco Gomez
Senior Thesis: Interferometric Synthetic Aperture Radar (InSAR) Analysis
of North American Periglacial Phenomena

Research Interests

1. Dynamics of the subglacial environment and the kinematics of subglacial deformation
2. Landforms as a record of earth surface processes and environmental change
3. Surficial processes of sediment production, transport, and landscape evolution
4. Geologic and environmental applications of geophysical and remote sensing techniques
5. Integration of modern geophysical techniques with traditional field methods

Research Experience

1. Quantitative assessment of subglacial sediment deformation and transport using environmental and paleomagnetic techniques, including anisotropy of magnetic susceptibility.
   a. Origins of subglacial bedforms, New York State
   b. Temporal variability of ice flow and bed deformation beneath a paleo-ice stream, southern Sweden.
   c. Deformation within the basal stratified ice of a modern glacier, Matanuska Glacier, Alaska
2. Interferometric Synthetic Aperture Radar (InSAR) analysis of rock glacier deformation in the high Andes, San Juan Province, Argentina
   a. Assessment of environmental controls, flow kinematics and sediment transport utilizing linux-based SAR and interferometry software packages.

3. Select Field Experience:
   a. Field mapping of bedrock and surficial geology in the Rocky Mountains
   b. Field mapping of the Surficial Geology of the Cato Quadrangle and surrounding regions of Upstate New York, in association with the New York State Geological Survey (Summer 2012).
   c. Field Investigation, sedimentologic description, and sampling of subglacial tills and in New York (Summer 2012) and Sweden (September 2013, May 2014, September 2015)
   d. Surficial, Geophysical, and Hydrological Investigations of Rock Glaciers in San Juan Province, Western Argentina (January 2013)

**Teaching Experience**

**Teaching Assistantships:**

EES 022: *Exploring Earth Laboratory.*
- Introductory earth and environmental science laboratory
- Coordinating TA for five semesters
- Lab Development:
  1. Ecogeocaching: Field-based introduction to observation, measurement, and geospatial skills utilizing tablet GPS software.
  2. Topographic Maps, Coordinate Systems, and Geospatial Information from paper maps to digital elevation models.

EES 027: *Natural Hazards*

EES 341: *Lehigh University Field Camp*
- Five week, mobile geology field camp in the Rocky Mountains
- Two summers experience (2013, 2015)

**Department Service**

EES Department Graduate Student Travel Fund Coordinator (2011-present)

**Awards**

1. College of Arts & Sciences, Lehigh University, Dean’s Summer Research Fellowship (2015)
3. Lehigh University Doctoral Travel Grants for Global Opportunities: *Deformation kinematics within a subglacial thermal boundary zone of the Scandinavian Ice Sheet, western Sweden*
5. Outstanding Undergraduate: Dept. of Geological Sciences, University of Missouri (2011)
6. University of Missouri Undergraduate Research Mentorship Program: InSAR Analysis of North American Periglacial Phenomena

Professional Affiliations

American Geophysical Union
Geological Society of America

Publications


Manuscripts in preparation


Conference Presentations with Published Abstracts


**Field Trip Guidebook Contributions**


**Invited Presentations**

Department of Geography, Geology and the Environment, Slippery Rock University, Slippery Rock, PA. April 9, 2015.

Dept. of Environmental Engineering & Earth Sciences, Wilkes University, Wilkes-Barre, PA. September 26, 2014.