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Spatiotemporal variations of englacial scattering of radar within Bench Glacier a temperate glacier in coastal Alaska

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SPATIOTEMPORAL VARIATIONS OF ENGLACIAL SCATTERING OF RADAR 
WITHIN BENCH GLACIER, A TEMPERATE GLACIER IN COASTAL ALASKA

by

Joel Brown

B.A. University of Montana, 2003

presented in partial fulfillment of the requirements

for the degree of

Master of Sciences

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July 2006

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Committee Chairman

Dean, Graduate school

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Date

7/26/06
Spatiotemporal Variations of Englacial Scattering of Radar within Bench Glacier, a Temperate Glacier in Coastal Alaska

Committee chairman: Joel T. Harper

Temperate glaciers are thought to be homogeneous bodies of ice. Radar surveys of temperate glaciers are thought to have scattering events to the surface of the glacier. Recent radar surveys of Bench Glacier, a temperate alpine glacier in the Chugach Range of southeastern Alaska, show a transparent layer within the glacial body. This transparent layer has been imaged in previous radar surveys of temperate glaciers but it has not been recognized as evidence of the inhomogeneity of temperate glacial ice.

I have determined the extent of the layer in three different years: 1999, 2003, and 2005. The layer is not constant through space or time. Phase analysis of the reflection events from the lower boundary of the transparent layer suggests that the englacial scattering is due to filled or partially filled voids within the glacial ice, the size of these voids is \(-0.168\) m or greater in diameter. There is no evidence of voids within the transparent layer of Bench Glacier nor is there evidence of upward fining of englacial voids within the layer of scatterers. It is likely that the transparent layer within Bench Glacier is getting smaller through time. It is possible that the change in layer is related to stress/strain relations within the glacier and may be an indicator of changes to the dynamics that govern the movement of the glacier.
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Chapter 1 - Introduction:

Basal decoupling due to the presence of water at the ice/rock boundary may be directly related to recent glacial thinning and rapid flow rates of outlet glaciers in Greenland. The basal water probably originates as melt water at the surface of the glacier [Krabill et al., 1999]. Since 70 m of sea level is held in glaciers and ice sheets, understanding hydrodynamic processes within glaciers is paramount in predicting effects of the warming climate. A major part of the glacial hydrological process is the mechanism by which surface water is routed through the ice mass to the bed of the glacier.

Current understanding of englacial hydrology is limited by the lack of direct observations of hydrological processes. It is known through direct observation that water is introduced to the englacial environment via moulins and crevasses [Fountain and Walder, 1998]. It is also known that water flows at the base of glacial bodies [Fountain, 1993]. The means of transport of water from the surface to the base of the glacial body is not well known [Fountain, 1993].

In the accumulation zone of glaciers, water is held in firn tens of meters thick and covering the nearly impermeable glacial ice. The water is thought to create a water table that drains into the glacial body by means of crevasses [Fountain and Walder, 1998]. As surface melt occurs on the relatively impermeable ice of the ablation zone, water moves down-gradient creating channels and streams on the surface. This system of channels and streams eventually leads to moulins and crevasses.

The current hypothesis for the transportation of water within glaciers is via a system of conduits [Fountain and Walder, 1998]. It is thought that these conduits stay
open under the high englacial pressure because the rate of melting along the surface area
of the conduits is in balance with the radially inward flow rate of the englacial ice as
determined by the internal stress field (Figure 1.1).

![Diagram of balancing forces](image)

**Figure 1.1** Illustration of balancing of forces between inward pressure and outward melting of an englacial channel.

Although there is evidence supporting the existence of englacial channels [Stuart et al., 2003], the formation process is not understood. One theory of englacial channel formation suggests that water is held in spaces between triple junctions of ice crystals [Nye and Frank, 1973]; a small amount of water flows from one junction point to another creating and transferring heat. The heat eventually becomes great enough to melt and maintain a water-transporting conduit. Shreve [1972] suggests that the englacial transport of water follows an arborescent network of steeply dipping channels (Figure 1.2). The formation of such channels would follow the theory of intergranular flow of liquid water through ice. Dye tracer injections in crevasses empirically support the existence of englacial channels transporting surface melt water to the bed of glaciers [Fountain, 1993]; direct observations of englacial conditions through borehole video [Harper and Humphrey, 1995; Pohjola, 1994] reveal englacial voids that Pohjola [1994] interprets as shallowly dipping channels. Harper and Humphrey [1995] found at least
one conduit that was dipping at 65°. The near-vertical nature of boreholes causes them to preferentially miss steeper dipping channels.

Figure 1.2 Example of arborescent drainage system as described by Shreve [1972] and adapted by Fountain and Walder [1998].

Fountain and Walder [1998] conjecture that water flowing in a crevasse would form a conduit. They state that water flowing along the base and walls of a crevasse will create enough thermal energy to carve a channel into the glacial body. This channel is then closed off at the top by the plastic nature of the ice thus becoming an englacial conduit. This theory is based on the previous existence of englacial channels or on water flowing through microfractures between crevasses; it does not deal with the formation of previous englacial channels or the ability of microfractures to drain crevasses at the rate at which they are filled. It is possible for crevasses to reach the base of the glacier, but only if the crevasse is constantly full [Robin, 1974]. The weight of water in a crevasse will overcome the tensile strength of the englacial ice causing the crevasse to propagate to the base of the glacier. Calculations of surface melt rates suggest that there is not enough water to continuously keep the crevasse filled.

Reeve [personal communication, 2006] has shown that there are orders of magnitude more moulins on a glacier than commonly assumed; most of the surface melt,
however, is transmitted to subglacial channels via only a small number of very large moulins.

1.1 Methods of studying englacial features

Englacial properties are elusive primarily due to the extreme difficulty of obtaining direct measurements. Direct measurements and observations of englacial properties have been conducted in two ways: by descending into crevasses and moulins and by drilling boreholes. Crevasses do not typically reach depths exceeding 25-30 m, and moulins become too small to safely negotiate and study before they reach the glacial bed. Boreholes can be continuous from the surface to the bed of a glacier and are extremely useful for studying englacial environments. Measurements of ice temperature [Harrison, 1974], water pressure [Fudge et al., 2005], and stress regimes [Pfeffer et al., 2000], as well as video analysis of the structure boreholes are examples of ways that boreholes are used to determine the englacial environment.

Indirect ways of studying englacial conditions include dye trace experiments and radar imaging. Dye trace experiments are used to study relationships between surface melt input and glacial stream outflow; the means of englacial transport are inferred from this relationship, but they cannot be separated from subglacial transport. Radar imaging (radio echo sounding) experiments have been used extensively to map the depth of ice sheets and glaciers [Steinhage et al., 2001]. It has also been used to study the internal structure of polythermal glaciers and cold ice masses such as those found in Greenland and Antarctica.

Ice sheets have internal stratigraphy. Layers of snow are buried by subsequent years' snowfall and turn into glacial ice with yearly boundaries having slight differences
in electrical properties due to variations in grain packing and the amounts of dust settling on the snowpack. There is very little lateral flow within the center of ice sheets; the stratigraphy within the body of the ice mass is retained because the flow of ice is near normal to the boundaries. Radar surveys of ice sheets reveal the englacial stratigraphy in great detail [e.g., Steinhage et al., 2001].

Polythermal glaciers occur in cold, relatively dry climates where loss of heat from the glacial surface is not balanced by the internal advection of temperate ice to the surface of the glacier [Petterson et al., 2003]. Polythermal glaciers have a layer of cold ice overlying temperate ice that sits at the pressure melting point. The temperature of the cold layer is below the pressure melting point of ice; water within this layer does not remain in a liquid state. The use of radar on polythermal glaciers reveals layering where a cold transparent layer of ice with no water held in its mass overlies a layer of temperate ice that has many sources of scatterers assumed to be proportional to the amount of water held within the ice.

Within temperate glaciers, the entire glacial body is at a temperature very near the pressure melting point of ice. The annual stratigraphy that occurs in ice sheets is not preserved in temperate valley glaciers. Melting occurs over the entire glacial body in a temperate glacier, including the accumulation zone. Liquid water from the melting snow in the accumulation zone travels vertically through the layers of seasonal snow and previous years' firm. The liquid water disturbs horizontal layering of the firm before it becomes glacial ice. Layering in the accumulation zone of temperate glaciers is weak because of this mixing of layers. Valley glaciers have a complex flow field with shear both in cross-valley and along valley directions. The complex flow field of temperate
valley glaciers completely distorts the weak, broken layering of the glacial ice. Due to this apparent lack of englacial structure or layering, the use of radar in temperate glaciers has largely been confined to mapping the base of the glacier.

The use of higher frequency radar systems and the filtering of data aimed at bringing out englacial features has potential to reveal englacial structures that are a large part of englacial hydrology. Some benefits of radar surveys are: 1) they are not limited to the upper 25 m of a glacier (as direct observation of crevasses are), 2) they do not change the englacial environment (as boreholes do), and 3) they can be used to locate water bodies that are not currently directly connecting the surface to bed hydrological network. Limitations of radar surveys include: 1) limited resolution (nominally 1/4 to 1/10 of the wavelength of the frequency used), the attenuation of the signal, and the variation of englacial radar velocities [Bradford and Harper, 2005].

1.2 Study site

Bench Glacier is a temperate glacier located in the Chugach Mountains in southeastern Alaska (Figure 1.3) [Harper and Bradford, 2005]. The glacier is approximately 8 km in length and is 600 m (near the terminus) to 1.5 km (in the accumulation zone) wide. The glacial body is 150-200 m thick through most of the glacier. The geometry of Bench Glacier is relatively simple. It has no tributary glaciers and the slope is a relatively constant 10° over the length of the glacier. There is an icefall located approximately 4.1 km from the terminus.
1.3 Review of Previous Work on Bench Glacier:

1.3.1 Boreholes on Bench Glacier:

In 2002 and 2003 over 50 boreholes were drilled to the base of Bench Glacier [Harper et al., 2005]. The boreholes were fitted with instruments to measure the water pressure and flow velocity at the base of the glacier. Measurements showed that the water pressure at the base of the boreholes was not constant. The basal water pressure changed over seasons and fluctuated diurnally in the late summer. Cameras lowered into boreholes have revealed near vertical water filled planar cracks and voids on the order of centimeters to meters [Harper and Bradford, 2005]. Time-lapse photography of the englacial cracks and voids suggests that they are dynamic features [McGee et al., 2003]. These voids do not appear to be confined by depth or location on the glacier. It is important to note that during the drilling of over 50 boreholes in Bench Glacier no englacial rocks were encountered.
1.3.2 Interconnectivity of boreholes

Recent studies conducted by Harper and Bradford [2005] on Bench Glacier in Alaska have identified features within the glacier which are a possible alternate means by which water flows to the base of the glacier. Harper et al. [2005] drilled more than 50 boreholes to the bed of Bench Glacier in 2002 and 2003, including a set of 16 holes on a 20 x 20 m grid [Fudge et al., 2005; Harper et al., 2005]. By monitoring the water levels within completed boreholes while drilling new boreholes, they found hydrological interconnectivity [Harper et al., 2005]. This inter-hole communication in conjunction with the lack of conduits imaged via borehole videos, suggests interconnectedness independent of a Shreve [1972] type system of conduits [Harper and Bradford, 2005].

1.3.3 Layering within Bench Glacier

While mapping the base of Bench Glacier with 25 MHz GPR, Harper and Bradford identified two distinct glacial facies: a radar-transparent facies (herein referred to as the transparent layer), and a facies defined by large numbers of reflection events (scattering layer) [Harper and Bradford, 2005] (Figure 1.4). In each radar profile that has both of these facies; the transparent layer overlies the scattering layer. Englacial reflection (and refraction) events are due to changes in the dielectric permittivity of the media through which the radar waves are propagating (thus indicating a change in propagation velocity and the index of refraction (n) of that media).
1.3.4 Migration velocity analysis of Bench Glacier

Migration velocity analysis (MVA) is a technique of determining the velocity of radio waves through a medium by migration of reflection hyperbolas at varying velocities.
in order to determine the velocity that collapses the hyperbola to a point source [Bradford and Harper, 2005; Stuart et al., 2003]. Bradford and Harper [2005] used MVA to determine the spatial variation of englacial radar propagation over the area of a single transect from Bench Glacier. They determined that there is a distinct difference between velocities of the radar-transparent and radar-opaque layers [Bradford and Harper, 2005]. The transparent layer had an average velocity of .171 m/ns and the opaque layer had an average velocity of .152 m/ns. Radio waves propagating through temperate ice have a velocity of .168 m/ns. Bradford and Harper [2005] assume that air filled voids increase the average velocity within the transparent layer and that englacial water accounts for the decreased propagation velocities in the opaque layer.

1.4 Statement of problem:

The existence of the transparent layer within Bench Glacier is not expected based on current understanding of temperate glaciers. In order to try to understand the transparent layer within Bench Glacier I have attempted to answer three questions. 1) Where is the layer located? By determining what regions of the glacier the layer occupies and where there is no evidence of layering, I have attempted to determine if the layer is continuous over large areas, or if it is a discontinuous, random feature. 2) Does the layer vary over time? I have tried to determine if the layer is present over multiple years and if it is to determine if it is dynamic and over what period it may change. I have also attempted to determine if any fluctuations are permanent or just seasonal variations. 3) What is the layer? I have tried to find the source of the dielectric variations within the glacial body. These are new questions in glaciology and I do not expect to fully answer any of them. The data and interpretations within this thesis do, however, put bounds on
the answers to these questions. It is likely that a new description of temperate ice is needed. The information presented within this thesis will need to be explained by any new models.

The overarching goal of this study is to improve understanding of englacial structure and the related glaciological processes. Better understanding of englacial structure will lead to better understanding of englacial hydrology, rheology, and thermal exchanges which in turn will lead to a better understanding and prediction of glacial response to climate changes.
Chapter 2 - Methods

The data used in this thesis were collected over three field campaigns with three different radar systems. I collected the most recent data in late July and early August of 2005. Other workers collected the two earlier data sets. The data from 1999 and 2003 were collected with the intention of mapping the bed of Bench Glacier. The 1999 and 2003 data were not intended to address the problem herein. The 2005 data had a secondary purpose of mapping the bed of Bench Glacier in further detail.

2.1 July/August 2005 data collection

2.1.1 Impulse Radar System

Data collected July 24-August 3, 2005 employed 5 MHz, 10 MHz, and 20 MHz ice penetrating radar systems. The length of the antennae is 10 m for the 5 MHz antenna, 4.2 m for the 10 MHz antenna, and 2.1 m for the 20 MHZ antenna. For each system the transmitting antenna was attached to a Narod Geophysics-type broadband transmitter [Narod and Clarke, 1994]. The transmitter has a peak power of 24 kW, a rise rate of <2 ns, and has a repetition rate of 512 Hz. The receiving antenna was attached to a Fluke ScopeMeter 190 series oscilloscope. A 3 ft. length of coax wire connected the receiving antenna to an impedance matcher that was attached to a signal splitter (T). The T split the signal between the A channel and the B channel. The T was attached directly to the A channel input and a short (~7 cm) length of coax was attached to the B channel. The oscilloscope was attached to a laptop computer via an optically isolated RS-232 adapter/cable. The length of coax cables was kept to a minimum to decrease signal interference. The oscilloscope recorded wave traces through channel A and was triggered to start recording via the airwave received through channel B.
The oscilloscope was used in this survey only as a signal detection and averaging device. The oscilloscope recorded voltage vs. time and was not tuned to any specific frequency. This was only possible because of the remote nature of glaciers. There is minimal background noise and interference from other man-made radio signals. A laptop computer was needed to record and save each averaged trace signal. I adapted and combined two programs to make a new program (RADAR1.BAS) that remotely accesses the oscilloscope and stores the values as a vector in ASCII format. RADAR1.BAS runs as a QBASIC program. It downloads the information that the oscilloscope has on-screen when prompted. The information is received as a vector 600 numbers long. Each number corresponds to a voltage received at a time. RADAR1.BAS creates an ASCII file to which it saves the first trace of a transect, each subsequent trace in a transect is then added to the end of the file. The traces can be imaged in the field by a simple Matlab program that divides the long vector into a matrix [600 x number of traces] and plots it as a grayscale image.

2.1.2 Narod Geophysics-type broadband transmitter wave/pulse generation

The Narod Geophysics type broadband transmitter employs a complimentary pair of three-transistor avalanche breakdown pulsers triggered simultaneously by current spikes from pulse transformers [Narod and Clarke, 1994]. The pair of three-transistor avalanche breakdown pulsers generates symmetric 550 V bipolar square pulses. The spikes from the pulse transformers cause one transistor from each string to breakdown. The timing of the pulses is controlled by digital logic. A crystal oscillator creates the trigger pulses that are delivered to the transformers. A square pulse is generated by the use of a charge line with impedance that is greater than the combined load and transistor.
impedances. The transistors are turned off quickly and cleanly by the negative reflection resulting from the open created by the source of the large impedance (a resistor) [Baker, 1991]. The duration of the pulse is equal to the two-way travel time of the output cable (antenna in this case). The dependence of duration of the pulse on the resonant properties of the antennae allows a single transmitter to be used with multiple different wavelength antennas.

The pulses are sent to the resistively loaded antennae (Figure 2.1). The resistance of the resistors increases nonlinearly from antenna feed point to antenna outer end, this controls ringing in the system. The center frequency and final shape of the emitted wave are controlled by the resonant properties of the antenna [Watts et al., 1976].

![Figure 2.1 Schematic of basic relation between transmitter and antennas. The Narod geophysical type transmitter sends symmetric 550 V bipolar pulses (A) to resistively loaded antennas (B). The duration of the pulse is equivalent to the two-way travel time (TWT) of the signal through the antenna (C). Notice that the longer antenna (2) has longer pulse duration than the shorter antenna (1).](image-url)
2.2 Data collection:

2.2.1 System settings

The window setting for the oscilloscope was set at 200 ns/division a total of 2400 ns was recorded for each trace. Channel B was set to a sensitivity of 100 mV/division and channel A was set to a sensitivity of 20 mV/division. In order to decrease the influence of random noise, an average of 8-64 traces was taken for each trace record.

2.2.2 Traces and transects

Since the 2005 field season occurred during the late summer, the snowpack in the ablation zone was gone (Figure 2.3) and the transects were collected over bare ice. I collected 12 transects orthogonal to the major axis of the glacier. Each transect was comprised of traces recorded every 5 m. The path of each transect was mapped with a handheld GPS. The receiving antenna runs parallel to the transmitting antenna. For this survey, separation of antennas was kept constant at 20 m for the 5 MHz and 10 MHz by attaching a safety rope between the two persons pulling the antennas. A separation distance of 5 m was used for the 20 MHz system in an attempt to counteract the effects of

![Figure 2.2 Picture of Radar transmitter showing sled and 10 MHz antenna. This picture was taken during a failed attempt at conducting a common-midpoint velocity analysis. Note the bare ice and slightly hummocky surface. Dave Schuler for scale.](image)
signal attenuation. Three people operated the radar system. The transmitter was placed in a small sled and pulled along the surface of the glacier by means of a rope attached to the sled (Figure 2.2). The front end of the antenna was attached to the rope and the trailing end of the antenna dragged behind the sled. The receiving antenna was also attached to a rope by which it was pulled along the glacial surface. Trace spacing was kept constant by markings on the ropes used for dragging the antennas. It is easiest (and safest) for the person operating the computer and oscilloscope to signal when the 5 m have been covered.

Figure 2.3 Picture of Bench Glacier looking up-glacier (roughly south); note the bare ice over most of the glacial surface.
2.3 June 2003 data collection

In June of 2003 Harper and Bradford used a Sensors and Software PE100A with a 1000 V transmitter and 25-100 MHz antennas to collect 6 cross glacier profiles and one axial profile. The system was dragged in a sled behind a snow machine traveling at constant velocity. Traces were collected at a trace spacing of 1.14 m. A wheel dragged behind the radar setup triggered the system once every revolution. The path of each transect was mapped using a Trimble differential GPS system. Since the 2003 field season occurred during the early summer, there was still snow covering the surface of the glacier (less than 3 m). The transects were collected on top of the snow layer.

2.4 June 1999 data collection

In early June of 1999, Harper collected 15 cross-glacier profiles with a Narod

![Figure 2.4](image.png)

Figure 2.4 Figure showing banded and normalized transects from 2005 field season. (10 MHz) transmitter and a 5 MHz antenna. The traces were recorded using a Fluke 100 oscilloscope and a laptop computer. The beginning and end of each transect were marked with bright flags, the positions of which were surveyed in the following month. Trace separations were either 13 m or 15 m. These distances were measured using a length of rope. Since the 1999 field season occurred during the early summer, there was
still snow covering the surface of the glacier (less than 3 m). The transects were collected on top of the snow layer.

2.5 Data migration/filtering techniques

2.5.1 Normalizing traces

Imaging transects in the field during the 2005 field season revealed a banded effect within the gray-scale plots (Figure 2.4 (A)). This banding is due to variability in the average amplitude of the signal received by the receiving antenna. Inconsistent coupling of both the transmitting and receiving antennas with the glacial ice likely causes this variability. Decimeter scale hummocks present on the surface of Bench Glacier during late summer made it impractical to have a consistent level of coupling for each trace. In order to eliminate this banding effect, I wrote a program, normalize.m, which normalizes each trace to zero. The result of this normalization can be seen in Figure 2.4 (B).

2.5.2 Geo-referencing traces:

Data collected in the 2005 and 1999 did not have GPS locations for each trace. Geo-referencing of individual traces was needed to correctly locate the true location of each trace.

2.5.2.1 Geo-referencing 1999 data:

Transects collected in 1999 were collected without the use of a GPS tracking system of any kind. The approximate starting location of each transect was located by surveying techniques after the transects were obtained. Thirteen of the fifteen transects were located, the relative position of fourteen of the transects were recorded (benp15 is in the upper basin and was not located in detail). In order to determine the location of each
trace, I determined the distance to the terminus for each transect and found the position of that distance along the 2003 axial profile. I then set the location of the middle trace for each transect to the appropriate location along the axial transect and determined the positions of the rest of the traces by assigning them positions at the proper trace spacing for each transect. The strike of the transects was defined by the mean of the range of strikes as defined by the transects from 2003 and 2005.

2.5.2.2 Geo-referencing 2005 data:

Since the mapping of trace locations was done by a tracking feature of a handheld GPS system instead of taking a reading at each trace location, the true location in x-y-z coordinates (corresponding to latitude, longitude, and elevation, respectively) of each trace along a transect needed to be determined from the GPS track. For every transect the number of trace locations and the number of points for the corresponding track log were different. I wrote a program, GarminToREFLEX.m, which inputs a matrix form of the original ASCII transect file (as is created by normalize.m) and a *.txt file containing the track log positions. It finds the total distance traversed in the x-y plane and divides that distance by the number of traces. Each trace is then assigned a location in x-y-z space. For example, trace #2 is located between log points 1 and 2; GarminToREFLEX.m assigns trace #2 a location along the segment that connects points 1 and 2, distance from point 1 to the location of trace #2 ($D_{loc}$) is: $D_{loc} = \frac{N_{trace} D_{segtot}}{N_{tot}}$ where $D_{segtot}$ is the total distance of all segments between track points, $N_{trace}$ is the trace number, and $N_{tot}$ is the total number of traces along the transect (Figure 2.5). The distances between the
locations of traces assigned by GarminToREFLEX.m are consistently close to the desired trace separation of 5m.

![Figure 2.5 Illustration of georeferencing 2005 traces; the open circles represent traces assigned to locations between GPS points (filled circles). Note that the GPS points do not have constant separation but trace locations do.]

2.5.3 Migration and filtering using REFLEX

The migration and filtering of data for this project were done using REFLEX, an industry seismic/GPR processing software package. Migration is widely used to determine proper geometry in echo-sounding data sets. The theory behind the use of migration is discussed by many [Moran et al., 2003; Moran et al., 2000; Stuart et al., 2003; Welch et al., 1998; Yilmaz, 2001]. Kirchoff migration was used in the processing/interpretation of images for this project.

Filters used to enhance radar images include band-pass filtering, subtracting the mean trace, and DC filtering. Gain adjustments to the images were minimal since the project focuses on the internal structure of the glacier as defined by layering within the top third of the glacial body. Figure captions state the level and type of data enhancement.

2.6 Radar trace phase analysis

By modeling synthetic radar travel paths within englacial channels Stuart et al. (2003) determined a theoretical means of differentiating between ice/air and ice/water interfaces. If the reflector is large enough to be fully resolved by the frequency of the
radar \( d > \frac{1}{4} \lambda \), then the polarity can be used to determine if the void is filled with water or air [Stuart et al., 2003]. A void filled with water will have a reverse polarity whereas an air filled void will have a normal polarity. Snell's law and simplified Fresnel equations show that:

\[
\frac{n_r}{n_i} = \frac{\sin \theta_i}{\sin \theta_r}; \tag{1}
\]

\[
\text{phase}_r = -\frac{\sin(\theta_i - \theta_r)}{\sin(\theta_i + \theta_r)}; \tag{2}
\]

\[
\text{phase}_i = \frac{2\sin \theta_i \cos \theta_r}{\sin(\theta_i + \theta_r)\cos(\theta_i - \theta_r)}; \tag{3}
\]

where \( \theta_i \) is the angle of transmission, \( \theta_i \) is the angle of incidence [Hecht, 2002]. Both angles are measured from normal to the plane of incidence. The \text{phase} equations are for the portion of the wave with the electric field perpendicular to the plane of incidence defined by the incident and reflecting waves [Hecht, 2002]. From these equations it is apparent that transmitted waveforms will be in phase with the incident wave. However, a reflected waveform will only be in phase when \( n_t < n_i \), when \( n_t > n_i \) the reflected wave will be \( \pi \) radians out of phase (negative).

The received waveform is a complex combination of background, direct transmission, and reflected radiation. The background radiation is theoretically random in both phase and amplitude and is eliminated by averaging many wavetraces. Direct transmission of the radar signal from the transmitting antenna to the receiving antenna occurs via the air (airwave) and the ice (surface wave). The arrival time and shape of these waves can be predicted and are distinguishable when antenna separation is great
enough [Walfrod et al., 1986]. Reflected radiation is received from all angles and is the addition of all of the waveforms that arrive at the receiving antenna at a specific time. Due to the non-unique solution to the shape of any particular waveform, analyzing waveforms that arrive at times greater than the first arrival of a reflected waveform is unreliable. In this thesis I analyze the first reflection event at the base of the radar transparent layer to determine the relative index of refraction of the reflecting body.
Chapter 3 – Results:

The results presented herein are based on the analysis of individual radar traces and the profiles they make up. Traces are plots of electric potential versus time. Profiles are many traces (between 31 and 3605) plotted next to each other. Data included in each individual trace is the distance to a reflecting object (inverted from average velocity of transmission), the amplitude of reflection, and the phase of the reflection. Overall englacial structure is more easily interpreted from profiles. Most individual traces (and therefore the profiles they are part of) have identifiable peaks. Theoretically, an air wave is received first, then a surface wave, then a large peak that can be interpreted as a reflection off of the bed of the glacier (Figure 3.1). Gray scale images of the radar profiles may show the bed reflection even when the individual traces are difficult to interpret (Figure 3.2).

![Graph showing radar reflections](image)

**Figure 3.1** Figure showing obvious air, surface, and bed reflections; note the absence of scattering events between 20 and 40 m. (5 MHz)
Figure 3.2 The bed reflection in trace 58 from B10M10 collected in 2005 is not easily picked out of the single trace location. By stacking many traces next to each other structure is seen and the bed reflection is identifiable. (10 MHz unfiltered)

3.1 Transparent layer within Bench Glacier

Apparent in some of the profile images of Bench Glacier is a transparent layer defined by a notable lack of reflections at time periods significantly longer than the surface wave arrival. For each profile I have interpreted the image and defined the surface, bed, and where applicable, the transparent layer (Figure 3.3).

The radar surveys of Bench Glacier show layering in profiles collected between the icefall and the terminus of the glacier. I consider the transparent layer to have less than 3 scattering events per wavelength. This definition allows for a point source within the transparent layer. Layering is not seen in every transect (Figure 3.4). There are also some pull up features within the transparent layer (Figure 3.5). These pull up features are
Figure 3.3 Figure showing raw (unfiltered) data with surface, layer, and bed interpretations; notice the few scattering events in the layer interpreted as transparent. These ‘uplift’ events of the scattering layer are interpreted as crevasses.

Figure 3.4 Absence of layering within transect B10M3, 10 MHz data from 2005 field season. This image has been band-pass filtered (2, 8, 12, 18).
imaged as many point source reflections stacked on top of each other; I am interpreting these features as crevasses. For a scattering event to occur an inconsistency of the ice mass must be present, the size of these inconsistencies must be roughly 0.1 times the wavelength in diameter (perpendicular to the plane of incidence) to reflect any detectable amount of the wave (this is dependant on attenuation, distance to scattering event, and sensitivity of detector) [Jacobel and Raymond, 1984] and must be 0.25 the length of the wavelength of the incident wave to fully resolve the anomaly [Reynolds, 1997]. The thickness of a scattering event must be 0.03 times the wavelength (in the direction of propagation) to reflect the incident wave [Sheriff, 1991]. The inconsistency in the ice mass can be due to changes in crystalline form, inclusions of water, inclusions of air, or inclusions of rock. The transparent layer within Bench Glacier is lacking inconsistencies in the ice mass that are large enough to cause scattering.
3.1.1 June 1999

In June 1999, fifteen cross-glacier transects were collected by J. Harper [unpublished]. The purpose of these traces was to determine the depth and geometry of the bed of the glacier for future drilling of boreholes. The approximate locations of the traces are shown by Figure 3.6. Twelve of the transects have evidence of a radar-transparent layer. These traces were obtained using a 5 MHz system. If the average propagation velocity of the radar wave is 0.168 m ns$^{-1}$, the theoretical resolution of the images is 8.4 m. Jacobel and Raymond [1984] state that strong scattering events occur when the source of scattering is greater than 0.1 times the wavelength of the incident wave, 3.36 m for 5 MHz at 0.168 m/ns.

Figure 3.7 is an example of the data obtained in 1999. The image is shown with and without interpretation. Interpretations of each transect collected in 1999 and used in this thesis are shown in Figure 3.8. The 99 data show a large volume of radar transparent ice with a layer depth up to 85 m. In section A of Figure 3.9 there is little to no layering apparent in the transects. Section B shows a deep layer that rises very quickly at the intersection of sections A and B. The layer is quite deep throughout the length of section B. Section C encompasses lines benp3, benp4, and benp2. Line ben3p has three large uplift events in the center of the transect, the layering in benp4 is dipping to the southeast, benp2 has very weak evidence of layering (Figure 3.10) but layering can still be inferred for the image. Section D includes lines benp12, benp13 and benp14. This section has a transparent layer between 30 m and 50 m thick with the center of the glacier having thicker layering than the edges of the glacier. The line benp15 was acquired in the accumulation area of the glacier. There is no evidence of layering in this area.
Figure 3.6 Relative locations of 1999 transects on bench glacier note the positions of the accumulation area (A), the icefall (B), the glacial tongue entering from the west (C), and the terminus (D).

Figure 3.7 Example of 5 MHz unfiltered data from the 1999 field season, line benp12 is unfiltered and is shown with and without interpretations.
Figure 3.8 Interpretations of all 1999 data used in this study. All of the interpretations are to scale except benp15, the transect from the upper basin. The transects get closer to the terminus from top to bottom, left column to right column.
Figure 3.9 Contour of depth of scattering layer in 1999; the terminus of the glacier is to the northwest the coordinates are UTM. Notice: 1) that there is a large and continuous layer throughout most of the ablation zone. 2) The layer is deepest at benp8 and quickly disappears. This feature dips 10-20 degrees up glacier as measured from the surface of the glacier. 3) The terminus has scattering to the surface. 4) Layering is highly variable in regions C and D.
Figure 3.10 Benp2 (5 MHz, unfiltered) with interpretation of layer; the evidence of layering within this transect is fairly weak, but I believe that the profile shows that layering is present between 30 m and 50 m. The length of the radar wave almost obscures the layering.

3.1.2 June 2003

In June 2003, seven cross-glacier transects and one axial transect were collected by Harper and Bradford [2005] using a 25 MHz GPR system. The locations of each of these transects on Bench Glacier is shown in Figure 3.11. For an average wave velocity of 0.168 m ns\(^{-1}\), the 0.25 wavelength resolution of the 25 MHz data is 1.68 m. For the same velocity, the 0.1 wavelength of the 25 MHz is 0.672 m. In general, the transparent layer is much thinner in the 2003 data than the 1999 data.

Figure 3.12 shows the interpretations for lines 1-6 from the 2003 data. Section A of Figure 3.13 is a region where no evidence of layering is present. Layering is present only at the edges of the glacier in section B. In section C the transparent layer is present all of the way across the glacial body. The depth of the transparent layer is between 30 m and 50 m in section C.

The long axis image (Figure 3.14) shows layering from about 1800 m from the terminus to 4100 m from the terminus (where the transect ends). There are regions where
heavy ringing of the radar signal occurred. This ringing is likely due to instrumented boreholes with wires extending to the base of the glacier. The layering in the axial transect shows that the transparent layering within Bench Glacier is continuous between cross glacier transects. The axial transect shows that the transparent layer is continuous for over 2 km from the base of the icefall to a distance of approximately 1800 m from the terminus. The maximum depth of the layer is approximately 50 m. A transect collected in the accumulation area of Bench Glacier shows no evidence of layering in that region of the glacier.

In each transect from 2003 that show evidence of layering there are point-source reflections within the region that I have defined as a "radar-transparent" layer. These single point sources are not included within the "scattering" layer because of their relative spatial isolation.

![Relative locations of 2003 transects on bench glacier](image)

**Figure 3.11** Relative locations of 2003 transects on bench glacier note the positions of the accumulation area (A), the icefall (B), the glacial tongue entering from the west (C), and the terminus (D).
Figure 3.12 Interpretations of all 2003 data used in this study. All of the interpretations are to scale except Upper_basin, the transect from the upper basin. The transects get closer to the terminus from top to bottom, left column to right column.
Figure 3.13 Contour of depth of scattering layer in 2003; the terminus of the glacier is to the northwest the coordinates are UTM. Notice: 1) reflections to the surface throughout region B. 2) The layer is located only in regions C, D, and E. 3) Layering dips at a ~2° toward the southwest. 4) The transparent layer is deepest in line_6 and line_2. 5) The location of the deepest part of the layer in line_2 is the location where the glacial tongue meets Bench Glacier.
Figure 3.14 Axial transect from 2003 with interpretation of layer. Notice 1) that the layer is continuous for 2300 m from 1800 m from the terminus to the end of the profile. 2) There are ringing events that reach below the bed of the glacier; these are likely due to instrumented boreholes. 3) The layer ends relatively abruptly (‘ramps up’ to the surface).
3.1.3 July/August 2005

In July and August of 2005 cross-glacier radar imaging was conducted over 8 different transects. The locations of these transects are shown in Figure 3.15. Of these eight transect locations only two had areas of radar-transparent ice. Locations 7 and 10 had a small (relative to 2003 and 1999) transparent layer. The interpretations of the images from 2005 are shown in Figure 3.17. At location 10 there is a transparent region from about 200 m to 550 m that reaches a maximum depth of \(-40\) m. Within this transparent region there is an up-lift event.

The image of location 7 (Figure 3.16) shows an area of transparent ice. The transparent area is approximately 350 m long and reaches a depth of \(-30\) m at the northwest side of the glacier where the transparent layer dips toward the edge of the glacier. Figure 3.18 shows that the layering in 2005 is very limited and is not spatially consistent.

![Figure 3.15 Relative locations of 2005 transects on bench glacier note the positions of the accumulation area (A), the icefall (B), the glacial tongue entering from the west (C), and the terminus (D).](image-url)
Figure 3.16 Grayscale image of location 7 from the 2005 field season. Notice the transparent layer (above dotted line) in this region of Bench Glacier.
Figure 3.17 Interpretations of all 2005 data used in this study. All of the interpretations are to scale. The transects get closer to the terminus from top to bottom, left column to right column.
Figure 3.18 Contour of depth of scattering layer in 2005; the terminus of the glacier is to the northwest the coordinates are UTM. Notice: 1) reflections to the surface throughout region B and most of region C (layering less than 5 m is due to non zero picking of surface). 2) The layer is only located in transects B10M7 and B10M10 3) Scatterers are present to the surface where glacial tongue enters Bench Glacier.
3.2 Time variability of transparent layer

The transparent layer within Bench Glacier was not constant over the study interval. The accuracy of each contour image is related to the number and spacing of the transects taken for each year as well as the accuracy of the system used. The data collected in 1999 is the least accurate of the three sets of data. This is due to the lack of true spatial positioning of each trace and the long wavelength used in the surveys. The vertical resolution of the 2003 data are better than the 1999 and 2005 data as is the accuracy of the positioning of each trace, however, the number of cross glacier transects is low. The 2005 data set has spatial resolution that is accurate to ~0.168 m (0.1 wavelength resolution).

Figure 3.19 Comparison of contours of depths from 1999 data (on the left) and 2003 data (on the right). The 1999 layer is much deeper and scattering only reaches the surface at the terminus and in region D (southwestern side of glacier). The 2003 layer is much smaller, both in area and depth. Notice that the ridge in the 1999 data near the boundary between regions C and D and the region of no layering in region C of the 2003 data suggests preferential regions of scattering.
3.2.1 1999 compared to 2003

The depth of the transparent layer was much greater in 1999 than in 2003 (Figure 3.19). In both 2003 and 1999, there was no evidence of a transparent layer in region A. The maximum depth of the layer in 99 was over 80 m, this depth occurred in region B. Four years later this region had no evidence of layering whatsoever. The total volume of the ice changing from transparent to scattering in region B alone was well over $3 \times 10^7 \text{ m}^3$. The total disappearance of the transparent layer in this region seems to be caused by the appearance of englacial voids that are filled with water. It should be noted that the difference in the layer depth in region B is not likely due to an upward fining of water-filled bodies. A comparison of data collected at different frequencies (Section 3.2.2) shows that upward fining is not the source of the change in depth and extent of the transparent layer between 2003 and 2005 data. It is likely that the source of the changes in the extent of the transparent layer between 1999 and 2003 is similar to the changes in the extent of the transparent layer between 2003 and 2005.

Region C has some similarities between the two years. There is an area of thinner transparent layer (with respect to the surrounding area) in the 1999 data just south of the C-B boundary. A lack of layering at the same location in the 2003 data has the same basic shape (in map view) as the relatively thin area in the 1999 data. Both data sets also show that the layer is thicker in the southwestern side of the glacier than it is in the northeastern side throughout region C. Over the width of the glacier there is an average dip of $\sim 2^\circ$ to the southwest in region C of both data sets. It should be noted that the dip of the layer is an average over the width of the glacier and is nearly flat. Regions D and
E show vague similarities in depth, however, the lack of data in 2003 for region D and in 1999 for region E make further analysis of similarities and differences impossible.

3.2.2 2003 compared to 2005

The 2003 and 2005 data (Figure 3.20) only overlap in regions B and C. They both lack evidence of layering in region B. The transparent layer is similar between 2003 and 2005. The general structure is fairly consistent with thicker layering further up glacier (south). The main difference between the two layers is that the 2003 layer is 30 m thick at the southwestern edge of region C whereas the same area of the 2005 data has no evidence of layering. This englacial change effected $7.5 \times 10^6$ m$^3$ of the glacier. The disappearance of the transparent layer occurred preferentially on the southwest half of the glacier in region C. The fact that the resolution of the 2005 data is less than that of the

![Figure 3.20 Comparison of contours of depths from 2003 data (on the left) and 2005 data (on the right). The 2003 layer is between 20 and 30 m in depth for much of regions C, D, and E, including the region where the tongue of the glacier enters Bench Glacier. The 2005 data shows scattering to the surface in this region.](image-url)
2003 data suggests that upward fining of englacial voids is not the cause of this apparent filling of the transparent layer. It should be noted that the change in the layering could be due to the different seasons during which the data were collected. The 2005 data were collected in the fall while the 2003 data were collected in the spring. It is interesting to note that the depth of the transparent layer in all three years is around 20 m at the boundary between region C and region D. This could further suggest that there is an underlying structure within Bench Glacier that controls the filling of the transparent layer.

### 3.2.3 Diurnal variations of englacial layers

In 2005 data were collected over a large range of hours of the day for 11 days. The earliest transect was collected between 6:30 and 7:10 in the morning and the latest transect was collected between 8:26 and 9:01 in the evening. Two transects were mapped both in the early morning (when melt rates are lowest) and in the early evening (when melt rates are high). One of the AM/PM transects was conducted over a region with no apparent layering (Figure 3.21). The other AM/PM transect was conducted over the area that had the greatest amount of layering in 2005 (Figure 3.22). There is no discernable difference in layering between the AM and PM transects at either location. This suggests that the transparent layer does not change diurnally over a scale great enough to detect with 10 MHz radar. The lack of large, erratic variations in the transparent layer suggests that the layer was stable over the duration of the 2005 field season (this is also true for 2003 field season for similar reasons). It should be noted that the weather during the 2005 field season had very minimal diurnal changes. The conditions were \( \sim 4 \, ^\circ C \) to \( 5 \, ^\circ C \) and raining both day and night.
Figure 3.21 AM/PM transects at location 5 of 2005. The transects were collected 14 hours apart. There is no discernable difference between the two transects. Notice that scattering is evident to the surface of the glacier. (10 MHz, unfiltered)

Figure 3.22 AM/PM transects at location 10 of 2005. The transects were collected 11 hours apart. There is no discernable difference between the two transects. Notice that the layering evident in the transects is the same in size and shape for both transects. (10 MHz, unfiltered, interpreted)
Figure 3.23 Comparison of 5 MHz and 10 MHz data from location 2 in 2005. Notice: 1) scattering is to the surface in both transect, indicating that there is not upward fining of englacial scattering events. 2) Resolution/intensity differences between the two frequencies of radar; the 5 MHz data are not as accurate but it experiences less attenuation. 3) The 5 MHz data contain the same waveform for 8 traces; this operator error was not eliminated because the traces give the proper geometry to the figure. (Both figures are unfiltered)

3.2.4 Comparing 5 MHz data with 10 MHz data over the same profile:

In order to assure that the untested 10 MHz radar system could image the bed of Bench Glacier, one profile was collected with both 5 MHz and 10 MHz antennas. The profiles were collected 21 hours apart from each other. Both profiles show scattering to the surface of the glacier (Figure 3.23). This suggests that a transparent layer did not form from one day to the next. This also suggests that there is not an upward fining of englacial scatterers with diameters between the ranges of 1.68 m to 3.36 m.

3.3.1 Phase analysis of layer boundary

I have conducted a limited phase analysis of the first reflection event that defines the lower boundary of the transparent layer in the axial transect of the 2003 data. The section of the waveform that shows the first reflection event is out of phase with the airwave. This observation indicates that the scattering layer has an upper boundary that is defined by inclusions of water within the glacial ice. These inclusions of water have a diameter (perpendicular to the plane of incidence) greater than ~0.672 m and a thickness
greater than \(-0.2\) m. It is reasonable to assume that the first reflections from 1999 and 2005 are also due to inclusions of water in the glacial ice.

Figure 3.24 Wiggle plot showing the relative phase of the air wave and the first reflection events at the base of the transparent layer. Notice that the first reflection events are out of phase with the air wave. This indicates that water bodies are located in the scattering layer.

3.3.2 Phase analysis of scattering events within the transparent layer

I have analyzed apparent point sources within the transparent layer of Line 1 from 2005. Two of the three point sources labeled in Figure 3.25 are point sources with no apparent out-of-plane interference. The third (middle) point source is located near the surface where interference from the surface wave makes phase analysis unreliable. The phase of the point source reflectors is out of phase with the incident wave. This indicates that the point sources are due to water filled bodies within the glacial ice. These point sources have a diameter (perpendicular to the plane of incidence) greater than \(-0.672\) m and a thickness greater than \(-0.2\) m.
Figure 3.25 Wave traces typical of scatterers located in the transparent layer, notice that many of the first scattering events are out of phase with the airwave. This is evidence that the scattering layer is due to englacial water.

3.4 Summary of important findings

* Transparent layer is found in data collected in the spring of 1999 and 2003 and the fall of 2005
  
  o The transparent layer overlies the layer of high-density scattering events everywhere it is found.
  
  o Transparent layer lacks numerous of water filled voids.
  
  o Depth of scattering layer is consistent with borehole video analysis revealing the presence of water filled cracks.
  
  o Diameter (perpendicular to the plane of incidence) of reflecting bodies is greater than \( \sim 0.1 \lambda \) (\( \sim 0.672 \) m for 25 MHz, \( \sim 1.68 \) m for 10 MHz, and \( \sim 3.36 \) m for 5 MHz); thickness of reflecting bodies is greater than \( \sim 0.03 \lambda \) (\( \sim 0.2 \) m for 25 MHz, \( \sim 0.5 \) m for 10 MHz, and \( \sim 1 \) m for 5 MHz).
- A few reflection events are present in the transparent layer.
- Upward fining englacial reflections is not apparent in different wavelength data.

acional reflections in different wavelength data.

- Transparent layer was not constant over the study interval.
  - The depth and geometry of the transparent layer changes on time scales less than 2 years and greater than hours to days.

- 1999 and 2003 data suggest the transparent layer ramps up to surface abruptly towards the terminus.

- The 2003 axial transect proves that the layering in Bench Glacier is continuous between cross glacier profiles.

**Table 3.1 Depth of transparent layer in Bench Glacier**

<table>
<thead>
<tr>
<th>Year (Season)</th>
<th>Maximum depth</th>
<th>Average depth</th>
<th>Depth in accumulation area</th>
<th>Depth near terminus</th>
</tr>
</thead>
<tbody>
<tr>
<td>1999 (Spring)</td>
<td>85 m</td>
<td>40-50 m</td>
<td>No layering</td>
<td>No layering</td>
</tr>
<tr>
<td>2003 (Spring)</td>
<td>45 m</td>
<td>10-30 m</td>
<td>No layering</td>
<td>No layering</td>
</tr>
<tr>
<td>2005 (Fall)</td>
<td>35 m</td>
<td>~20 m</td>
<td>No data</td>
<td>No layering</td>
</tr>
</tbody>
</table>
Chapter 4 – Discussion

Radar surveys of Bench Glacier reveal an englacial layer with very few scattering events (transparent layer) overlying a layer of scatterers. This layering has been imaged with different radar systems and with frequencies ranging from 5 MHz to 100 MHz. This study focused on three separate years of data acquisition: 1999, 2003 and 2005. The results of this study show that layering was present during all three field seasons. The location of the transparent layer is spatially restricted and varies over time.

4.1 Layering not due to thermal stratification

It has been assumed that temperate glaciers should have water inclusions to the surface of the glacier and would therefore not have a transparent layer. Transparent layers have been observed in polythermal glaciers [Bamber, 1987; Petterson et al., 2003; Stuart et al., 2003]. Polythermal glaciers have a transparent layer because the top layer of ice is too cold for water to remain liquid. There are many reasons that Bench Glacier is not a polythermal glacier. 1) The glacier is at a low elevation (~1000 m) in coastal Alaska where the climate is relatively mild. 2) 4 m to 6 m of insulating snow accumulates during the winter [Harper, 2006 personal communication]. 3) Observations in over 50 boreholes show no evidence of polythermal stratification. 4) These boreholes only freeze very near the surface of the glacier during the coldest months of winter [Harper, 2006 personal communication]. 5) The geometry of layering described by this thesis changes over periods of a few months to a few years, this time period is too short to be due to a change of thermal régime within the glacial body. 6) There are a few reflection events within the transparent layer of Bench Glacier that are due to pockets of liquid water being
present within the transparent layer (Figure 3.25). If the transparent layer within Bench Glacier were due to thermal layering, the water in this layer would freeze. 7) There is scattering to the surface near the terminus and in the accumulation zone of Bench Glacier. Polythermal glaciers have cold, radar-transparent ice below layers of firm in the accumulation zone. Cold, radar-transparent ice is also apparent in the near surface of polythermal glaciers close to the terminus.

4.2 Scattering layer

I believe scattering events within Bench Glacier are due to water-filled voids whereas the transparent layer lacks these water-filled voids. Evidence supporting the theory that scattering events are caused by water filled voids includes: 1) Phase analysis of englacial scattering events reveal that there is a change of phase between incident and reflected waveforms. This indicates a transmission from a lower index of refraction to a higher index of refraction. 2) The transparent layer is rapidly changing in depth over large areas in short amounts of time (possibly up to 20 m/year). This precludes the layer of scattering being due to thermal layering or englacial debris. 3) Borehole video analysis has directly observed englacial voids in the region of large numbers of englacial scatterers. There is a general lack of voids within the volume of ice that corresponds to the transparent layer [Harper and Bradford, 2005].

The water bodies that make up the scattering layer have a diameter (perpendicular to the plane of incidence) greater than 0.672 m – 3.36 m (depending on the frequency used to detect them). The thickness of the water bodies is greater than 0.2 m – 1 m (also depending on the frequency used to detect them. These are lower bounds on the size of
the scattering body only. The shape and orientation of the bodies has not been
determined by this study.

4.3 Transparent layer

The transparent layer is a layer that lacks quantities of liquid water great enough
to change the dielectric properties of the englacial ice mass. A scattering/partial
reflection event occurs where there is a change in the index of refraction of the medium
that the incident wave is traveling through. The number and intensity of the scattering
events imaged englacially is directly related to the number and size (respectively) of
inhomogeneous inclusions within the ice. In natural glacial environments there are three
known sources of materials with indices of refraction that differ from ice, they are: rock,
air, and liquid water. Over 50 boreholes drilled to the bed of Bench Glacier, suggest that
rocks are virtually nonexistent within the ice mass. If the scattering events were being
caused by englacial rock debris, the glacier would be filled with rocky material. The
internal pressure of glaciers precludes the inclusion of large air filled voids.

4.4 Spatiotemporal variability of the englacial layering

The transparent layer within Bench Glacier has not remained constant through
space and time. There has been a general trend of the transparent layer getting smaller.
The transparent layer has changed into a layer filled with many scattering events starting
at the terminus and moving up-glacier. There are some features of the geometry of the
transparent layer that are fairly constant through the three different data sets implying
there is englacial structure that controls the layering to some extent. These features could
be related to the englacial stress field or the spatial distribution of surface water input.
It is important to consider that each data set represent a snapshot of the englacial conditions during that particular field season. This means that the data represent three points on a curve, making an interpretation of the shape of that curve ambiguous. The actual rate of englacial change has not been determined by this study. A lower boundary to the rate of change has been determined. Data collected in 2005 indicate that the layer changes over time periods greater than days. Comparisons of all data used in this thesis indicate changes of the englacial dielectric characteristics of Bench Glacier over periods of years. There is then, an upper and lower bound on the time over which the layer is changing. Data collected in during the spring of 2006 [J. Bradford and J. Harper, personal communication] may suggest that changes of the geometry and depth of the transparent layer in Bench Glacier may be seasonal fluctuations. Ongoing work will likely determine the rate at which the transparent layer changes.

4.5 Relationship of variation of transparent layer with down-valley advection and long-term thinning rates:

In the regions of Bench Glacier with the highest down valley advection rates, the average surface velocity from September 1999 to August 2000 was approximately 2.8 cm d⁻¹ or 10.22 m yr⁻¹. Using this velocity, it is reasonable to project that between 1999 and 2003 a point in this highest velocity area moved approximately 41 m down-glacier. The same point would move approximately 21 m further down glacier between the 2003 and 2005 field seasons. The scale of this movement is shown in Figure 4.1. The annual surface velocity is a measure of an upward limit of down valley advection within Bench Glacier. The emergence velocity of englacial ice in Bench Glacier is on the order of ~1 m yr⁻¹ whereas the vertical (smaller) component of the dielectric change discussed herein
is much more than twice this rate (~20 m yr\(^{-1}\)). Surface and emergence velocities are not
great enough to fully explain the change in the geometry of the transparent layer.

Bench Glacier has been thinning at an average rate of approximately 1 m yr\(^{-1}\) [Shuler, 2006, personal communication]. Using this rate, the total thinning was
approximately 6 m over the time period that this study encompassed. The change of the
thickness of the transparent layer is 4 to 14 times the total thinning over most of the
glacier. Therefore, although the thinning of the transparent layer within Bench Glacier is
likely affected by the flow and volume changes, the scale of the change of the transparent
layer that has occurred over the time period of this study suggests that the flow and
thinning are not first order causes.

4.6 Other glaciers:

The transparent layer within Bench Glacier is not unique. There is evidence of a
transparent layer within at least three other glaciers. Transparent layering occurs in
figures published in Welch [1998], Arcone [1995], and Blindow and Thyssen [1986].
The glaciers are (respectively) Worthington Glacier, Matanuska Glacier, and the
Vernagtferner. Worthington Glacier and Matanuska Glacier are temperate glaciers in
southeast coastal Alaska. The Vernagtferner is a temperate glacier in the Austrian Oetztal
Alps. The transparent layer within Matanuska Glacier is about 30 m thick [Arcone et al.,
1995]. The layers within Worthington Glacier and the Vernagtferner are each about 80 m
thick [Blindow and Thyssen, 1986; Welch et al., 1998]. It is likely that there are many
other radar studies of temperate glaciers where a radar transparent layer is present but not
identified as such.
Figure 4.1 Figure of 2005 contour with a line 50 m (to scale) in length, notice that the scales over which the englacial layering has changed are vast.

4.7 Implications:

The transparent layer within Bench and other glaciers in coastal Alaska may potentially constitute the need for a new paradigm of englacial hydrology. In the
currently accepted Shrevian model, there is an upward fining of englacial conduits.

There is no upward fining of reflection events within Bench Glacier; data collected with 5 MHz antennas duplicate the scattering imaged by 10 MHz antennas. This shows that the Shrevian model of englacial hydrology is, at best, incomplete.

Models of rheology of temperate glaciers may also need to be rethought. The englacial environment of temperate glaciers is not homogenous. Even though the ice is very nearly at the same temperature between different regions of the ice mass, there is a large disparity between the amounts of liquid water held within voids. Harper and Humphrey [1995] suggested that inhomogeneities within Worthington Glacier constituted a small enough percentage of the englacial environment that bulk rheology would not be affected. However, additional work on the same glacier [Harper et. al., 2002; Marshal et. al., 2003] suggested that the relation between stress and strain within Worthington Glacier was stratified. Near the surface of the glacier the constitutive law had a linear region and the region below was highly non-linear. The boundary between the two regions was abrupt and when driving stresses and strain rates were calculated over the same length scale the depth of this transition zone was ~75 m. It is not clear if there is a cause/effect relationship between flow dynamics and the depth of the transparent layer (imaged by Welch [1998] identified by me) within Worthington Glacier or if the similarity in depths is just coincidence.

4.8 Unanswered questions:

Future research should focus on answering these questions:

1) Why are there no water-filled voids near the surface yet there is an abundance of them deeper within temperate glaciers?
2) What is the geometry of the englacial water-filled voids?

3) What glaciological processes create and/or destroy the water filled voids.

4) This study only addressed scattering events that occur within the transparent layer
   or at the lower boundary of this layer, do water filled voids vary within the
   scattering layer?

   Future work that should be conducted in order to solve the main questions of why
the transparent layer exists and what causes it is: 1) radar transects need to be repeated
over time scales of weeks, months, and years; 2) three dimensional radar grids in layered
regions of glaciers need to be conducted to determine the geometry of the water-filled
voids; 3) direct observations of englacial environment through borehole video analysis
should be taken in areas of layering. Future models of englacial hydrology and rheology
of temperate glaciers need to account for the findings of this study.
Chapter 5 - Conclusions

There is a radar transparent layer within Bench Glacier. This layer has not been recognized before and does not fit currently accepted ideas of temperate glaciers. The layer does not extend throughout the entire area of the glacier and it is not constant through time. By mapping the spatial extent of the transparent layer for each year and comparing the results it has been shown that the layer has changed over the duration of this study. The geometry of the layer changes over time periods less than years and greater than weeks.

By employing phase analysis techniques, it has been shown that the change of the dielectric properties within the englacial environment is likely due to the inclusion of water-filled cracks into a volume of ice that does not have preexisting cracks or voids. The volumes of reflection events are bounded by the wavelength of the radar system used to image them. Data collected in 1999 has a minimum diameter of scatterer (normal to the plane of incidence) of ~3.36 m and the 2003 data can show scattering from sources with diameters as small as ~0.67 m. The thickness (in the direction of propagation) of the scattering events must be ~0.2 m for the 2003 data and ~1 m for the 1999 data.
Appendix A

A.1.1 Basic theory of propagation of electromagnetic (EM) waves through dense media

Simplifying assumptions:
- Transmitting antenna acts as a point source
- Ice is considered homogeneous infinite half-space
- Wave packet is a square wave

Using these simplifying assumptions the theory of wave propagation through ice is fairly basic. The wave packet travels as a constantly expanding hemispherical wave front. Assuming that the media is homogeneous, the amplitude and phase of the wave are independent of the direction of travel, they depend solely on the distance traveled. The distance \( d(\tau) = v(t_x - t_0) \). The velocity \( v \) of an EM wave through a dense media is described by \( v = \frac{c}{n} \) where \( n \) is the index of refraction of the material and \( c \) is the speed of light.

The wave packet has initial amplitude \( (A_0) \). As the wave propagates, \( A_0 \) is spread over the surface area of the wave front. The result of this spreading is that the amplitude of the wave at any point decreases as \( d^{-1} \); this in addition to energy being transferred to individual molecules accounts for the attenuation of the radar wave as it propagates through dense media.

A.1.2 EM propagation through heterogeneous media

The propagation of EM waves through heterogeneous media (in this case ice, air, and liquid water) is not as mathematically simple as homogeneous media. Ice, air, and water have different indices of refraction (in fact, inconsistencies of each media can vary...
n values within that media). The index of refraction of each media is determined by the EM propagation velocity, which is dependent on the electromagnetic properties of that media (dielectric permittivity \((\varepsilon^*)\) and permeability \((\mu)\)).

\[
v = \frac{1}{\sqrt{\varepsilon^* \mu}}
\]

Dielectric permittivity is related to the capacitance of the material by \(E = \frac{Q}{\varepsilon^* A}\) where \(E\) is the strength of electrical field between a parallel plate capacitor, \(Q\) is the charge on the capacitor, and \(A\) is the area of each plate.

At every media interface there is a partial reflection and partial transmission and/or refraction of the incident wave packet; the ratio of reflection to transmission is dependent on the angle of incidence \((\theta_i)\) (as measured from normal to the plane of incidence) and the \(n\) values for the two media.

A.2 Using englacial reflection events to ‘image’ glacial bodies

The radar systems used in this study operate by sending a wave packet through a transmitting antenna and recording the amplitude of EM waves received for a short period (~100-200 ns) after the wave is sent. The output from this process is a vector of numbers; each number is a value of amps received by the receiving antenna at a certain time. The values amplitude values for each wave trace can be plotted against time. There is no way to determine the position of a single reflection event from a single wave trace, only the two-way travel time (TWT) is known for each amplitude value in the vector. If a plane normal to the surface of the glacier is defined by the radar transect, a singular reflection event within that plane will define a hyperbola in a wiggle plot image (side by side evenly spaced individual time/amplitude traces); out of plane reflectors will
have no geometrical coherence within the wiggle plot. The source of in-plane reflection events can be assumed to be at the location of the peak of the hyperbola. By assuming an average EM propagation velocity for the glacial body, one can determine the relative geometrical position of each reflection event.

Recent dye-trace analysis at Bench glacier indicates two types of connectivity between surface introduction of water and basal out flow (personal communication, Jonathan Reeve). The two types of connectivity suggested are a direct link to subglacial flow paths and an indirect, possibly diffuse connection to subglacial flow paths.

A.3 Comparing 5, 10, and 20 MHz data sets using Icefield instruments Narod geophysics type broadband transmitter

During the 2005 field season, I surveyed transect position #2 with three different frequencies of radar: 5 MHz, 10 MHz, and 20 MHz. The 20 MHz system collected completely indecipherable data (figure A1). The 5 MHz and 10 MHz systems both collected good reflection images.

Both radar image plots have a large reflection event at the glacial bed (figure A2). This reflection event is much stronger in the 5 MHz image than it is in the 10 MHz image. The spatial resolution of the 5 MHz bed reflection is less precise than the 10 MHz equivalent. Some englacial
Figure A.5.1 5 MHz (top) and 10 MHz (bottom) transects over the same line. Notice that the bed profile in the 10 MHz transect is more accurate but less apparent than the 5 MHz transect.

reflections are discernable as point source reflections within the 10 MHz image; in the 5 MHz image the reflections appear as non-descript scattering. An operator error produced the false data on the 5 MHz image from 330 m to 380 m; this error is due to the same trace being recorded over multiple trace locations. This error has been left in the data set to preserve true locations of all other traces in the image.
References:


