Subglacial sediments as a control on the onset and location of two Siple Coast ice streams, West Antarctica

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[1] Laterally continuous subglacial sediments are a necessary component for ice streaming in the modern onset regions of the ice streams draining the Siple Coast of West Antarctica on the basis of new seismic data combined with previous results. We present geophysical results from seismic reflection and refraction experiments in the upper reaches of ice streams C and D that highlight continuous sedimentary basins within and upstream of the current onset regions of both ice streams, with streaming ice overlying these sedimentary packages. The subglacial environment changes from no-sediment to discontinuous-sediment to continuous-sediment cover along a longitudinal profile from the ice sheet to tributary C1B. Along this same profile, we observe a speedup of ice flow and then full development of the ice stream tributary. Ice stream D flows above a thick sedimentary package with an uppermost low-seismic-velocity zone indicative of soft till, and the upglacier and lateral extensions of ice stream D are tightly constrained by the extent of continuous sediments. The inland termination of these sediments suggests that future migration of high-velocity, low-shear-stress ice flow in these regions appears unlikely.

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1. Introduction

[2] The West Antarctic Ice Sheet (WAIS) is of special interest because of the possibility that acceleration of its flow could cause rapid sea level rise [*Hughes*, 1977]. The thick inland ice toward the center of the ice sheet flows slowly, mostly by deformation within the ice, and feeds thin, rapidly flowing ice streams that move by sliding over or deforming soft sediment beds [*Alley et al.*, 1986, 1987]. These ice streams typically feed floating extensions, called ice shelves, with changes in ice stream discharge contributing to sea level fluctuations. Increased melting under ice shelves has been shown to almost immediately increase the flow speed of adjacent ice streams, contributing to sea level rise [*Joughin et al.*, 2004a; *Rignot et al.*, 2004; *Dupont and Alley*, 2005]. However, because most ice streams today are thin, they cannot contribute greatly to global sea level rise. Inland, nonstreaming ice is the main contributor to potential sea level rise, but as long as basal velocities remain low in the interior of the WAIS, any contribution will be relatively slow. Understanding the controls on the transition from slow flowing inland ice to the fast flowing ice streams is key to modeling the future of the WAIS.

[3] Of special concern is the possibility that portions of the inland ice reservoir could switch to ice streaming through increased basal lubrication. This is the likely explanation of the Heinrich events from the Laurentide ice sheet during the last ice age [*MacAyeal*, 1993; *Alley and MacAyeal*, 1994]. For these events, the leading hypothesis is that rapid basal motion was geologically allowed, owing to unconsolidated or smooth sediments in Hudson Strait [*MacAyeal*, 1993; *Alley and MacAyeal*, 1994]. Rapid basal motion was glaciologically prevented most of the time, though, because the ice was frozen to the bed or lacked sufficient basal meltwater to initiate basal sliding or till deformation. Rapid glaciological switching occasionally enabled catastrophic drainage contributing to sudden sea level rise. Had ice streaming been geologically difficult due to a rough bedrock bed, then such rapid basal switching would not have been possible.

[4] Knowledge of the geological as well as glaciological conditions under West Antarctic ice is thus a prerequisite to

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assessing how rapidly the WAIS would contribute to sea level rise. The underlying questions are how strongly subglacial geology influences the transition from slow flow via internal deformation to streaming ice flow due to rapid basal motion, and how far the potential for ice streaming extends into the inland ice reservoir of West Antarctica. Previous work has shown that streaming ice in West Antarctica is underlain by sedimentary packages [*Blankenship et al.*, 1986, 1987, 2001; *Rooney et al.*, 1991; *Smith*, 1997; *Anandakrishnan et al.*, 1998; *Bell et al.*, 1998; *Doake et al.*, 2001; *Studinger et al.*, 2001; *Anandakrishnan*, 2003; *Vaughan et al.*, 2003], that a soft bed is present immediately beneath the ice to allow rapid basal motion [*Engelhardt et al.*, 1990; *Engelhardt and Kamb*, 1998; *Tulaczyk et al.*, 1998; *Kamb*, 2001], and that subglacial sediments appear to shape the inland extent of streaming ice flow [*Anandakrishnan et al.*, 1998; *Bell et al.*, 1998; *Blankenship et al.*, 2001; *Studinger et al.*, 2001]. However, much geological heterogeneity has been found across West Antarctica, indicating that the mechanisms for producing streaming ice flow may vary. Thus additional geophysical surveys are required to provide tighter constraints on the role of subglacial geology in influencing streaming ice flow across West Antarctica.

[5] To that end, we have conducted seismic surveys in the inferred onset regions of ice streams C and D, as determined from *Bindschadler et al.* [2001], and we have refined geophysical techniques to better constrain the influence of subglacial geology on the onset of fast ice flow and the potential for inland migration of streaming ice. Using reflection and refraction seismic methods, we imaged a zone of low-seismic-velocity material, which we interpret to be layered sedimentary basins, beneath the fast flowing ice in these regions (For clarification, we discuss both the velocity of seismic waves and the velocity of flowing ice. Following usual terminology, ‘‘low-velocity zones’’ and ‘‘velocity-depth profiles’’ apply to seismic velocities, whereas “ice velocity” or “ice flow velocity” refers to the motion of the ice.). Our seismic reflection experiment along ice stream D provides a detailed profile of several sedimentary layers beneath the ice. Our transverse profile of ice stream D was, for safety reasons, located upglacier of the extensive open crevasses associated with especially fast ice flow. This profile thus crossed either an upstream extension of ice stream D, or a tributary of ice stream D; the difference is probably more semantic than visual. The profile crossed a feature that flows faster than does ice on the surrounding ridges, but which is narrower and flows slower than does the main body of ice stream D farther downglacier [*Joughin et al.*, 1999; *Bamber et al.*, 2000]. This upstream extension of D overlies a thick sedimentary package, and the margins of the ice stream and sedimentary package are coincident, suggesting that the limited extent of continuous sediment cover may restrict the full development of streaming ice flow. Our longitudinal profile of tributary C1B images a sedimentary basin pinching out in the upstream direction of the ice stream, in agreement with an upstream increase in driving stress and basal shear stress, yet an upstream decrease in ice flow velocity. This result suggests that streaming ice flow may have reached its inland extent in this region, in agreement with a previous aerogeophysical survey across the head of tributary C1B [*Bell et al.*, 1998]. We believe that the subglacial extent of these sediments and the presence of a dilatant till are fundamental to understanding the evolution of ice streaming in West Antarctica and the future of the WAIS.

2. Background

[6] Changes in the flow speed of the ice streams that drain the WAIS influence the mass balance of the WAIS [*Vaughan et al.*, 1999; *Rignot and Thomas*, 2002; *Joughin et al.*, 2002] and can have significant impacts on global sea level [*Oppenheimer*, 1998]. *Mercer* [1978] suggested that this ice sheet, largely grounded below sea level, could shrink rapidly in response to small perturbations in ocean water temperature or in the position of the grounding line. This potential for ice sheet instability is enhanced by the presence of fast flowing ice streams, whose rapid short-term fluctuations could have long-lasting effects on ice volume change in West Antarctica [*Whillans and van der Veen*, 1993].

[7] The Siple Coast ice streams drain the WAIS through the Siple Coast of West Antarctica into the Ross Ice Shelf (Figure 1a; for simplicity, we continue to refer to Whillans, Kamb, and Bindschadler ice streams by their former names, ice streams B, C, and D, respectively). These fast flowing ice streams are conspicuous in satellite images [*Stephenson and Bindschadler*, 1990; *Scambos and Bindschadler*, 1991, 1993; *Hodge and Doppelhammer*, 1996] and ice velocity maps [*Joughin et al.*, 1999, 2002], with well-defined lateral margins, and extend several hundred kilometers into the ice sheet. Seismic imaging [*Blankenship et al.*, 1986, 1987; *Rooney et al.*, 1991; *Anandakrishnan et al.*, 1998] and borehole analysis [*Engelhardt et al.*, 1990; *Tulaczyk et al.*, 1998; *Kamb*, 2001] have shown that, in places, streaming ice is coincident with subglacial sediments. The upper meters of these sediments are soft till, with the potential for active deformation [*Alley et al.*, 1986, 1987; *Engelhardt et al.*, 1990]. Gravity and magnetic anomalies across the Siple Coast have been interpreted as thick sedimentary packages, which are believed to have a strong influence on streaming ice flow [*Bell et al.*, 1998; *Blankenship et al.*, 2001; *Studinger et al.*, 2001].

[8] Numerous ice streams drain the WAIS, and previous geophysical studies of the West Antarctic ice streams have revealed distinct spatial variations in their ice flow dynamics. The trunks of the Siple Coast ice streams have especially well-lubricated beds where ice flows over flat basal topography in comparison to ice streams draining into the Filchner-Ronne Ice Shelf [*Smith*, 1997; *Doake et al.*, 2001; *Vaughan et al.*, 2003] and Pine Island Bay [*Thomas et al.*, 2004]. The Rutford Ice Stream, for example, experiences fast flow in the confines of a subglacial trough bounded by the Ellsworth Mountains on one side and by ice on the other [*Doake et al.*, 2001], while the Siple Coast ice streams show a seaward increase in ice flow velocities across a relatively smooth bed topography. The ice streams draining into the Filchner-Ronne Ice Shelf possess spatial variations that influence streaming ice flow, such as those observed along the Rutford Ice Stream [*Smith*, 1997] and between neighboring ice streams [*Doake et al.*, 2001; *Vaughan et al.*, 2003]. No ground-based geophysical surveys have been

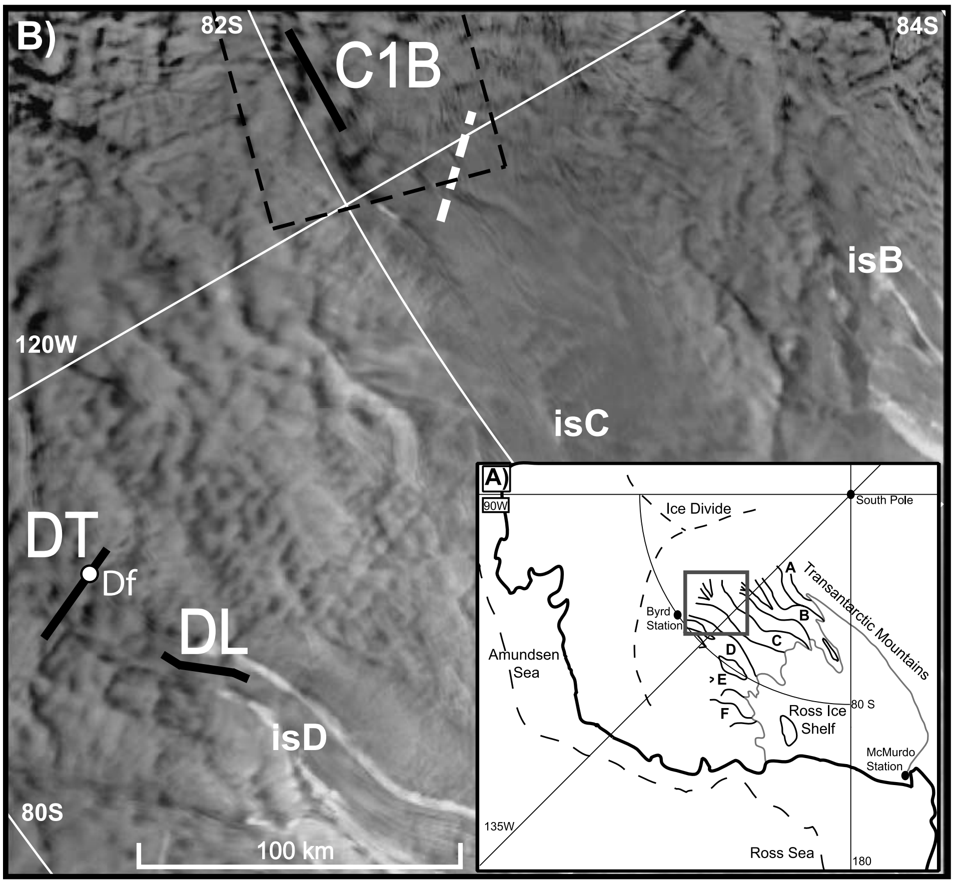

Figure 1. (a) Schematic map of the Siple Coast of West Antarctica. The Siple Coast ice streams (labeled A–F: A, Mercer Ice Stream; B, Whillans Ice Stream; C, Kamb Ice Stream; D, Bindchadler Ice Sream; E, MacAyeal Ice Sream; F, Echelmeyer Ice Stream) are outlined along the Siple Coast to the grounding line of the Ross Ice Shelf. The box outlines the boundaries of the enlarged map of the field sites (Figure 1b). (b) RADARSAT image of the Siple Coast [*Liu et al.*, 2001], highlighting the seimic reflection and refraction experiments performed along ice streams C and D. C1B and DT are the locations of the deep seismic refraction experiments. Df is the shallow refraction experiment, and DL represents the seismic reflection experiment. The details of each of these experiments are explained in the text. The black dashed box to the top of the map is the location of the aerogeophysical survey of *Bell et al.* [1998], and the white dashed line is the seismic refraction study of *Anandakrishnan et al.* [1998]. The study region of *Studinger et al.* [2001] covers most of this map, and the aerogeophysical survey of *Blankenship et al.* [2001] covers the catchments of ice streams B and C. Other seismic and borehole experiments performed along the Siple Coast are located closer to the grounding line along the ice streams [*Blankenship et al.*, 1986, 1987; *Rooney et al.*, 1987a, 1987b, 1991; *Engelhardt et al.*, 1990; *Munson and Bentley*, 1992; *Engelhardt and Kamb*, 1998; *Tulaczyk et al.*, 1998; *Kamb*, 2001].

performed in the Pine Island Bay region, leading to limited knowledge of the influence the bed has on streaming ice flow and drainage into the Amundsen Sea.

[9] Spatial variations in ice flow velocity along the Siple Coast of West Antarctica have been observed through ice velocity measurements [*Joughin et al.*, 1999, 2002], which seismic and radar studies have attributed to the basal conditions of these ice streams [*Blankenship et al.*, 1986, 1987; *Rooney et al.*, 1987a, 1991; *Anandakrishnan et al.*, 1998; *Bentley et al.*, 1998; *Jacobel et al.*, 2000; *Anandakrishnan*, 2003; *Catania et al.*, 2003]. These observations have largely been limited to ice streams B and C, revealing sediments beneath both ice streams. Ice stream B reaches ice flow velocities of greater than 800m/yr, and ice stream C likely experienced high velocities before its recent slow down in ice flow [*Retzlaff and Bentley*, 1993; *Anandakrishnan and Alley*, 1997; *Anandakrishnan et al.*, 2001].

[10] Reconnaissance geophysical surveys of subtill geology have also been conducted on the Siple Coast. These have especially targeted the trunks of ice streams B and C [*Blankenship et al.*, 1986, 2001; *Rooney et al.*, 1987b, 1991; *Engelhardt et al.*, 1990; *Munson and Bentley*, 1992; *Clarke et al.*, 1997; *Anandakrishnan et al.*, 1998; *Bell et al.*, 1998; *Kamb*, 2001; *Studinger et al.*, 2001], although limited coverage has been extended to the trunk of ice stream D and to the onset regions [*Kamb*, 2001; *Bindschadler et al.*, 2001; *Studinger et al.*, 2001]. The sparseness of coverage with high-resolution subglacial imaging in the onset regions of the Siple Coast ice streams has led to poor geological constraints on physical controls of the onset of ice streaming.

[11] Identification of the onset of streaming ice flow is problematic, especially in a region such as the Siple Coast of West Antarctica. The "easiest" definition is that inland ice begins streaming where flow passes over a sharp downstep into a deep channel or depression in bedrock [*McIntyre*, 1985]. However, the Siple Coast ice streams largely lack such features. A gradual transition in ice flow is often inferred, from internal deformation upglacier to basal sliding downglacier [*Joughin et al.*, 1999]. In such situations, the onset of ice streaming may be defined as the maximum in driving stress along the flow line from the ice divide to the grounding line [*Alley and Whillans*, 1991; *Bindschadler et al.*, 2001].

[12] On the basis of aerogeophysical surveys of the Siple Coast, previous workers have proposed that subglacial sediments are necessary for streaming ice flow [*Bell et al.*, 1998; *Studinger et al.*, 2001], although the occurrence of sediments under both fast moving and slow moving ice suggests that sediments alone are not sufficient to cause ice streaming. Seismic constraints in the onset regions of the Siple Coast have been limited to one location along tributary C1B of ice stream C, where the boundary of a sedimentary basin has been imaged, coincident with the margin of streaming ice flow [*Anandakrishnan et al.*, 1998]. Coupled with current knowledge of heterogeneities in both ice flow and subglacial character across West Antarctica [*Bennett*, 2003; *Vaughan et al.*, 2003] and the lack of seismic observations in the onset regions of streaming ice flow, better subglacial constraints on the onset of streaming ice flow are necessary to fully understand the dynamics and drainage potential of the WAIS.

3. Field Experiments

[13] We performed a series of seismic experiments during the 2002–2003 Antarctic field season to image subglacial geology and determine subglacial properties in the upstream reaches of ice streams C and D (Figure 1b). To optimize scientific return, we conducted a field campaign to characterize the firn, ice, any shallow sedimentary deposits, and deeper geology. We sampled both parallel and transverse to ice flow to better characterize geological conditions that may influence streaming ice flow.

[14] The particular seismic experiments we performed are (Figure 1b) (1) a shallow seismic refraction experiment at location Df on ice stream D, to produce a velocity-depth profile of the firn and ice, (2) a multichannel seismic reflection experiment parallel to ice flow on ice stream D, to image subglacial layers and determine geological structure beneath the ice stream, and (3) two seismic refraction experiments, one along ice stream C (the tributary C1B profile) and one across the upstream reaches of ice stream D (the DT profile), to determine subglacial geology.

3.1. Shallow Seismic Refraction

[15] Our shallow seismic refraction experiment was designed to characterize the seismic nature of the firn and the ice. We used 120 vertical 40 Hz geophones in our array, with the source located 1 m from the nearest geophone. The 60 geophones closest to the source were spaced only 1 m apart to resolve the steep gradient in the upper firn, while the far geophones had a 5 m spacing to characterize the deeper firn where the seismic velocity gradient is small. We measured both the compressional wave (P wave) and vertical shear wave (S_V wave) velocities in the firn, using explosives and hammer blows as sources for the experiment.

3.2. Seismic Reflection

[16] Seismic reflection data were collected along a 16.7 km longitudinal profile of ice stream D, oriented parallel to ice flow and placed near the center of the ice stream (DL in Figure 1b). We used a receiver array of 120 vertical 40 Hz geophones, with a 10 m spacing between geophones. Each shot sequence consisted of three explosive charges, with 150 g and 454 g charges detonated in succession 30 m from the nearest receiver, and then another 454 g charge detonated 1230 m from the nearest receiver, giving a maximum source-receiver offset of 2420 m. This shot sequence was repeated at 300 m intervals along the profile to produce fourfold data with a common midpoint spacing of 5 m. The charges were detonated at 60 m depth, below the firn-ice transition, to improve the source coupling and frequency content of the data compared to shots at shallow depths within the firn [*Clarke et al.*, 1997].

3.3. Subice Seismic Refraction

[17] We conducted a 31.2 km longitudinal seismic refraction experiment along ice stream C, at the head of tributary C1B (C1B in Figure 1b). Explosive charges were detonated at nine shot locations along the line, each at a spacing of 3900 m and a depth of approximately 20 m. We deployed two seismic receiver arrays along the line, one near the upstream end and the other near the downstream end of the profile, producing a fully reversed profile of the line. The upstream receiver array consisted of 20 three-component 4.5 Hz geophones, with a 200 m spacing, based on the optimal receiver spread for deep seismic refraction in Antarctica determined by *Clarke et al.* [1997]. The downstream receiver array was composed of 120 vertical 40 Hz geophones, with a 10 m geophone spacing. Vertically oriented geophones to record P waves were alternated with horizontally oriented geophones to record S_V waves.

[18] Our ice stream D seismic refraction experiment consisted of a 34.5 km transverse profile extending entirely across the upper reaches of ice stream D (DT in Figure 1b). We were able to survey across the ice stream margins because shear there was insufficient to produce crevassing. Explosive charges were detonated at a depth of approximately 60 m with a 4800 m spacing along the line, and their seismic energy was recorded by two receiver arrays, one at the southern end and one near the center of the line. The southern end receiver array consisted of 25 three-component 4.5 Hz geophones separated by 200 m. The midstream array was identical to the high-frequency array used on the ice stream C profile (120 40 Hz geophones at 10 m spacing, with alternating vertical and horizontal geophones).

4. Data Analysis

[19] The goal of the seismic experiments in the upper reaches of ice streams C and D was to characterize the ice column and subglacial geology at and above the onset of streaming ice flow. The shallow refraction experiment

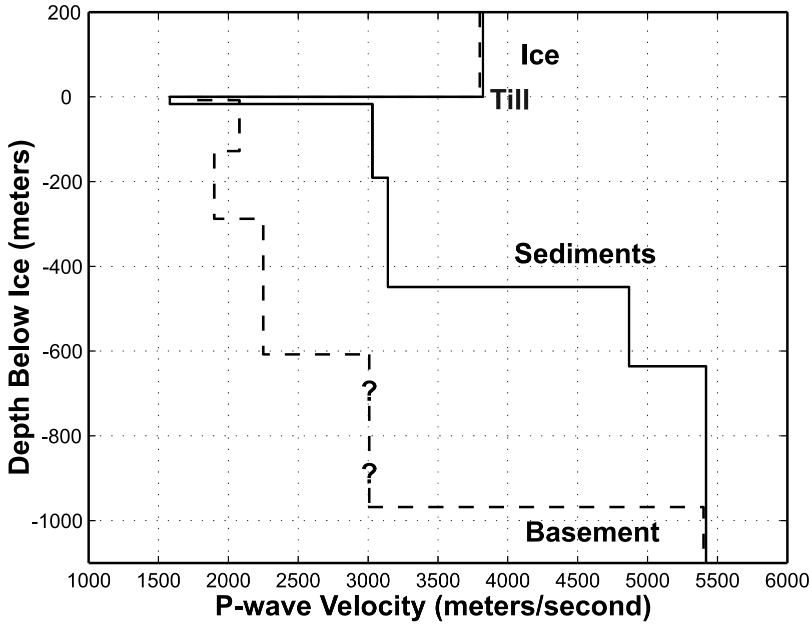

Figure 2. Subglacial velocity-depth profiles along the Siple Coast of West Antarctica. The solid line represents the profile produced from a seismic reflection experiment along ice stream D (DL) during the 2002–2003 field season, and the dashed line marks the profile determined by *Rooney et al.* [1991] along ice stream B. The zone of question marks along the dashed line of *Rooney et al.* [1991] represents the portion of their profile that could not be resolved. The velocity-depth profiles are based on *P* wave seismic velocities.

produced a high-resolution profile of the firn to constrain near-surface effects on seismic travel times. Seismic reflection data were used to determine the seismic velocities of the layers directly beneath the ice, including any low-velocity zones to which refraction seismic profiling is less sensitive. Our seismic refraction experiments mapped the extent of these low-velocity zones beneath the ice stream C and D profiles, both of which extended from slow to rapidly flowing ice.

4.1. Shallow Refraction

[20] The firn, although thin, has a considerable effect on seismic travel times because of its extremely low seismic velocities. The refracted arrivals from the shallow refraction experiment were used to calculate a velocity-depth profile of the firn, allowing us to correct the seismic data that propagate to greater depths. Previous results suggested that firn velocities are nearly independent of horizontal direction [*Bentley and Kohnen*, 1976], vary only slowly laterally [*Bentley and Kohnen*, 1976; *Alley and Bentley*, 1988], and increase monotonically with increasing depth [*Blankenship et al.*, 1987]. This permits us to use the Herglotz-Wiechert equation [e.g., *Shearer*, 1999, chapter 5] to invert for firn seismic velocity from the refracted arrivals, and to apply these results to our other seismic experiments.

[21] Near-surface velocities are low ($V_P \sim 600$ m/s, $V_S \sim 100$ m/s), increasing rapidly in the upper 10 m and then more slowly with depth. This depth profile is due almost exclusively to firn densification [*Alley and Bentley*, 1988]. Both seismic velocities monotonically approached typical values for West Antarctic ice stream ice ($V_P \sim 3810$ m/s, $V_S \sim 1850$ m/s [*Blankenship et al.*, 1987]); we found seismic velocities of $V_P = 3800 \pm 10$ m/s at 64 ± 1 m depth and $V_S = 1840 \pm 10$ m/s at 57 ± 1 m depth from our experiment with the depths approximating the firn-ice interface.

4.2. Seismic Reflection

[22] The seismic reflection data were used to calculate the seismic velocities of the ice and subglacial rock layers. We processed the data using standard multichannel seismic methods, including band-pass filtering, predictive deconvolution, and migration, to improve the signal-to-noise ratio and isolate subsurface reflectors. Twenty neighboring common midpoint (CMP) gathers were averaged to improve the fold of the data, at the expense of horizontal resolution.

[23] A velocity analysis of the CMP gathers assumed that the reflectors had hyperbolic moveout (the so-called small-spread approximation), which is justified for our target depths ($\sim$2 km) and maximum offset (2.4 km). We employed a grid search method to determine the best fitting stacking velocity and zero-offset two-way travel time for each reflective layer. Application of the Dix equation [*Dix*, 1955] then gave us the interval velocities and layer thicknesses of our observed subglacial layers. Our velocity-depth profile is produced from five seismic reflectors, forming four subglacial layers. From top to bottom, the interval velocities and layer thicknesses are: $V_P = 1580 \pm 150$ m/s, $h = 17$ m; $V_P = 3030 \pm 40$ m/s, $h = 170$ m; $V_P = 3140 \pm 40$ m/s, $h = 260$ m; $V_P = 4870 \pm 40$ m/s, $h = 190$ m; and the basement velocity is $V_P = 5420 \pm 50$ m/s (Figure 2).

[24] Errors in the final seismic velocity and thickness calculations for each layer are due to uncertainties in the seismic velocity of the firn, uncertainties in shot hole depths, and anisotropy in the ice. The velocity-depth profile

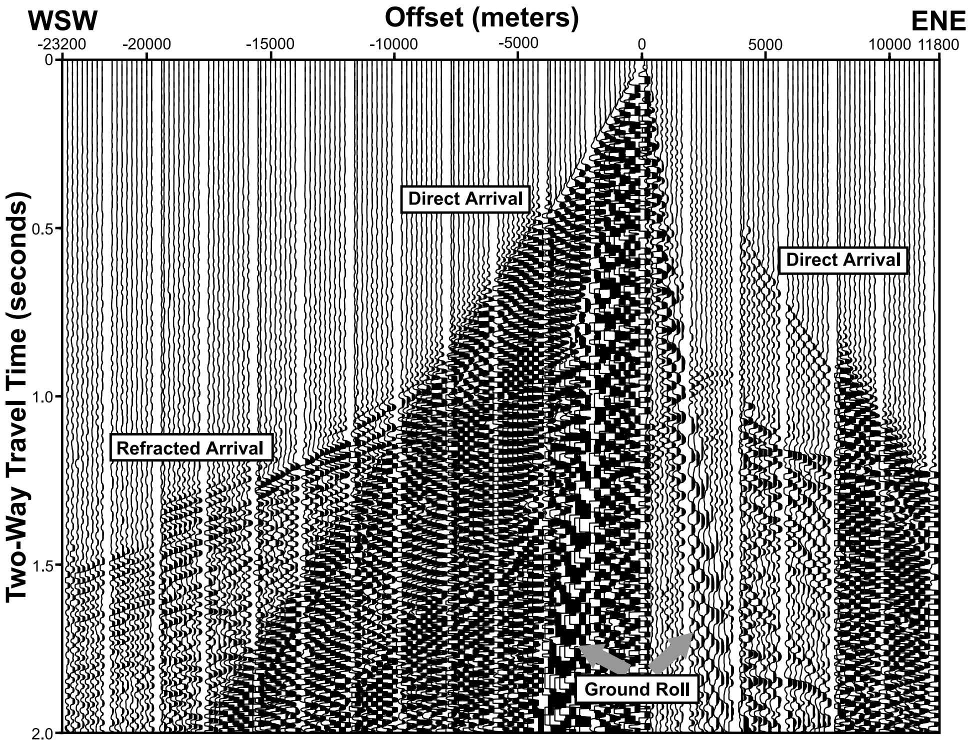

Figure 3. Seismic refraction shot record for P wave component of the C1B profile. The receiver array used to produce this record consisted of twenty 4.5 Hz, three-component geophones covering 3800 m. Gaps in the data are due to a malfunctioning geophone. The variations in amplitude along the shot record are due to differences in the energy (size) of the source at each shot hole. Three arrivals are observed in the data: (1) the refracted arrival ($V_P \sim 6400-6700$ m/s), (2) the direct arrival ($V_P \sim 3900$ m/s), and (3) the ground roll ($V \sim 1900$ m/s). A linear moveout (LMO) correction of $V_{\mathrm{LMO}} = 6500$ m/s was applied to the entire shot record to flatten the refracted arrival.

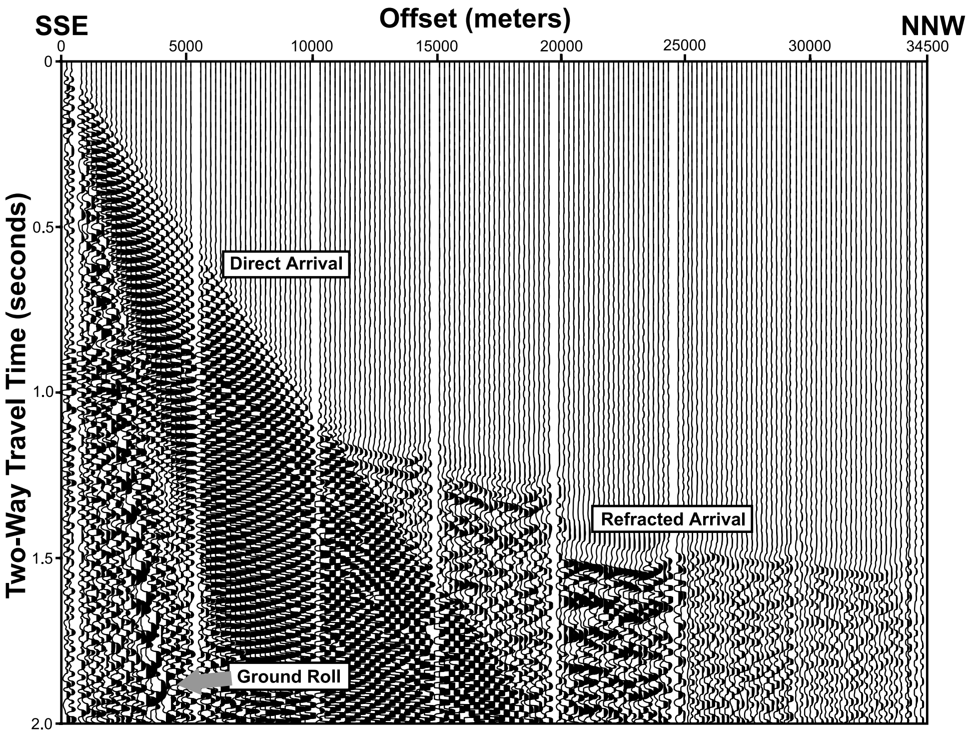

Figure 4. Seismic refraction shot record for P wave component of the DT profile. The receiver array used to produce this record consisted of twenty-five 4.5 Hz, three-component geophones covering 4800 m. All other parameters are the same in Figure 3.

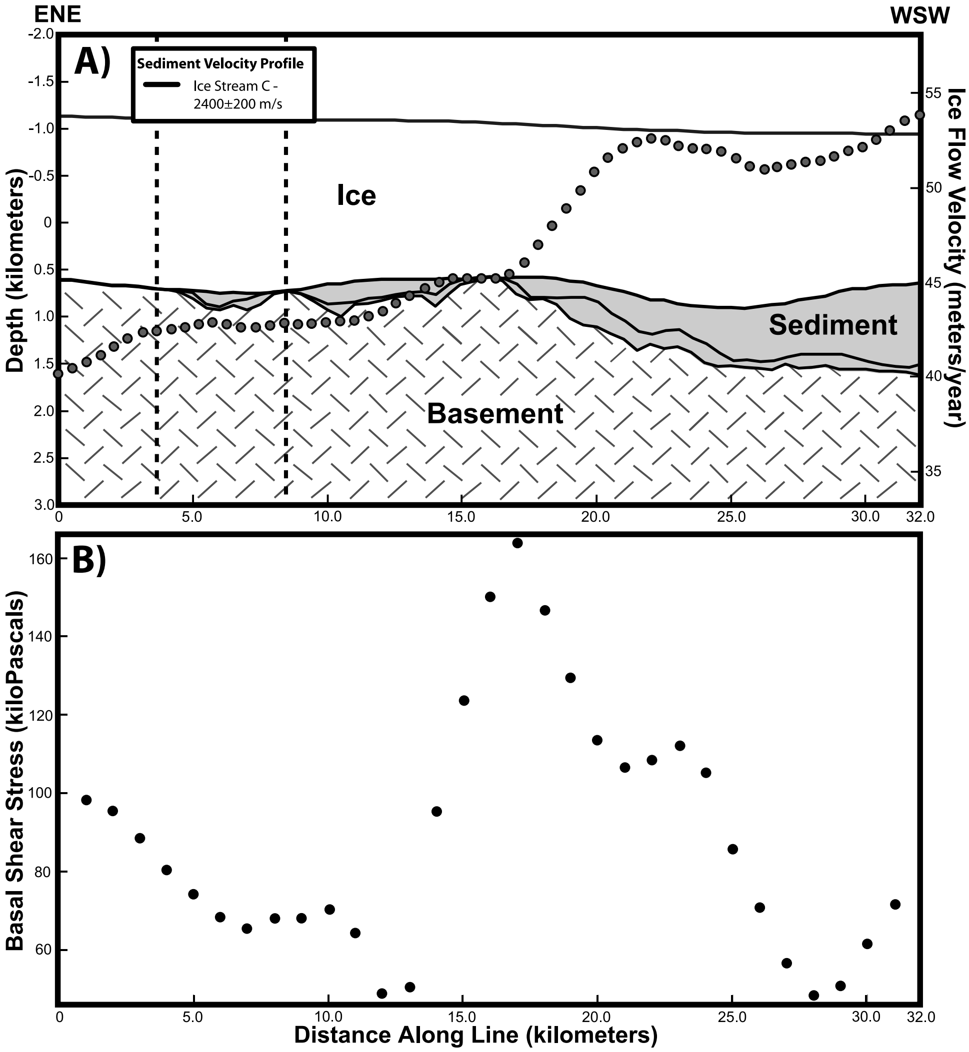

Figure 5. Inversion model and measurements for the C1B profile. (a) Best fit inversion results of the C1B seismic refraction profile. A three-layer model was used, consisting of (1) the ice, with good thickness constraints [*Lythe et al.*, 2001], (2) a sedimentary unit, with an average *P* wave seismic velocity of V_P = 2400 ± 200 m/s (taken from the preliminary results of a seismic reflection experiment along a portion of this profile, with the two dashed vertical lines highlighting the area covered), and (3) the basement. The two solid lines for each sediment thickness represent the error bars that apply to the best fit of the low and high sediment velocities. The left-hand side lists depth below sea level in kilometers, and the right-hand side gives ice flow velocity in meters per year. The circles mark ice flow velocities taken from the ice velocity map of *Joughin et al.* [2002], where uncertainties are less than 2 m/yr in this location. (b) Diagram of basal shear stress measurements [*Joughin et al.*, 2004b] estimated along our C1B profile, each with uncertainties of 5 kPa or less.

of the firn had a depth uncertainty of σ_Z = 1 m and a velocity uncertainty of σ_V = 10 m/s, creating a travel time uncertainty of σ_{t1} = 1 ms through the firn. The shot hole depths had an uncertainty of σ_Z = 1 m, corresponding to a travel time uncertainty of σ_{t2} = 0.5 ms. Anisotropy in the ice can contribute up to a 3% uncertainty in travel times [*Bentley*, 1972], producing an uncertainty of $\sigma_{t3} \leq 30$ ms. Each of these uncertainties in travel time was propagated through our calculations to yield the overall seismic velocity and thickness uncertainties in the final velocity-depth profile.

4.3. Long-Offset Refraction for Subglacial Structure

[25] We applied the two-dimensional (2-D) inversion methods of *Zelt and Smith* [1992] to the *P* wave refracted arrivals from our two seismic refraction profiles (C1B and DT, Figures 3 and 4, respectively) to invert for the thickness of the low-velocity zone beneath the two profiles. We created a four-layer model for each profile, consisting of (1) the firn (with seismic velocities and thicknesses from the short-refraction experiment), (2) the ice (with thicknesses from ice-penetrating radar [*Lythe et al.*, 2001]), (3) a one-layer low-velocity sedimentary package (with an average seismic velocity from the reflection work), and (4) crystalline basement rock (with seismic velocity from the reflection work and previous refraction experiments [*Munson and Bentley*, 1992; *Anandakrishnan et al.*, 1998]). We allowed the thickness of each sedimentary layer to vary until the model

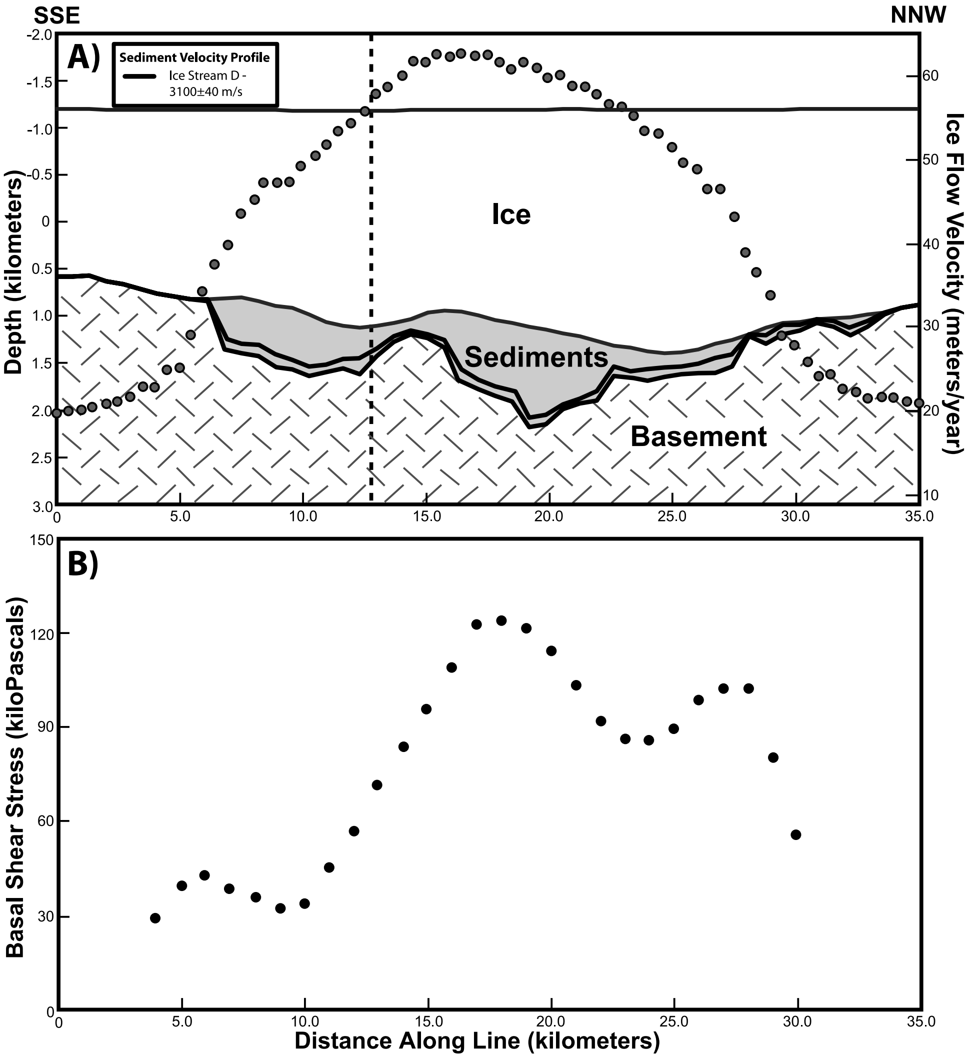

Figure 6. Inversion model and measurements for the DT profile. (a) Best fit inversion results of the DT seismic refraction profile. All parameters are the same as in Figure 5, with the exception of the seismic velocity of the sedimentary unit (an average *P* wave velocity of V_P = 3100 ± 40 m/s is applied, derived from the seismic reflection results of the DL profile; the dashed vertical line marks the approximate upstream intersection with our profile.). (b) Diagram of basal shear stress measurements [*Joughin et al.*, 2004b] estimated along our DT profile, each with uncertainties of 5 kPa or less.

produced a best fit between the observed and modeled arrival times.

[26] The best fit inversion models yield low-velocity zones beneath the ice where streaming ice is present. We imaged a low-velocity zone with an average seismic velocity of V_P = 2400 ± 200 m/s beneath our C1B profile that pinches out in the upstream direction and becomes discontinuous to absent by the end of the profile (Figure 5). This low-velocity zone thickens in the downstream direction, with the velocity function for this zone taken from preliminary results of the seismic reflection experiment along a portion of the C1B profile (highlighted in Figure 5), where we obtained the average seismic velocity applied to this low-velocity zone. We also detected a low-velocity zone across our transverse profile of ice stream D (DT; Figure 6), coincident with increased ice flow velocities. We applied the velocity-depth profile of our ice stream D seismic reflection experiment to the low-velocity zone, which had an average seismic velocity of V_P = 3100 ± 40 m/s, since the two experiments were less than 50 km apart and at similar bed elevations.

[27] The small and nonsystematic residuals calculated for both the C1B and DT inversion models show that our models provide good fits to the observed travel times. The C1B model has a travel time residual of Δt_{RMS} = 7 ms and a normalized χ^2 = 11.3, based on 185 data points and a travel time picking uncertainty of σ_t = 2 ms (Figure 7). Most of the travel time residuals are less than 10 ms, with a maximum travel time residual of Δt_{RMS} = 43 ms observed at an offset of x = 11500 m, which produces an uncertainty of 0.85% of the total travel time. A travel time residual of Δt_{RMS} = 9 ms and a normalized χ^2 = 21.1 are calculated from 311 travel time measurements along our DT profile, each with a picking uncertainty of σ_t = 2 ms (Figure 8). The maximum

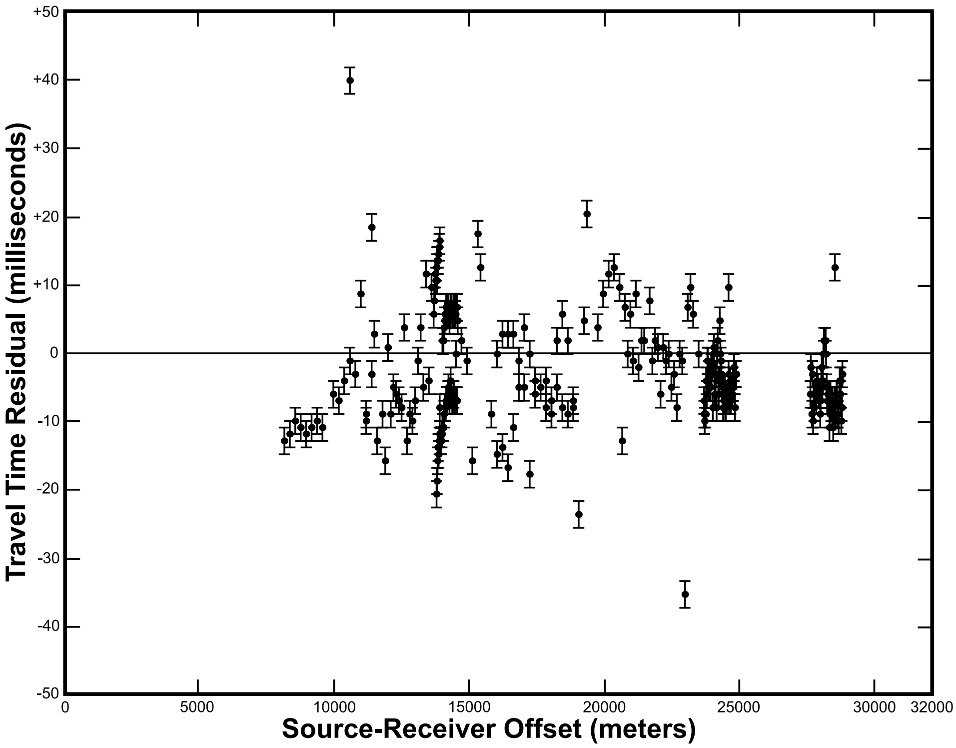

Figure 7. Residual plot for the C1B profile. The best fit model has an RMS residual of Δt_{RMS} = 7 ms and a normalized χ^2 = 11.3, based on 185 data points and an estimated picking uncertainty of σ_t = 2 ms (this model is shown in Figure 5a). A maximum Δt_{RMS} = 43 ms observed at an offset of 11,500 m, which is approximately 0.85% of the total travel time, while most of the residuals are less than 10 ms.

travel time residual observed along the DT profile corresponds to about 0.6% of the total travel time. The differences between the modeled and observed travel times may be due to (1) anisotropy in the ice that could create up to a 3% variation in travel time as a function of offset [*Bentley*, 1972], (2) uncertainties in the thickness of the ice (<50 m [*Lythe et al.*, 2001]), (3) uncertainties in the overall seismic velocity of the sedimentary layer and the assumption that it is laterally homogeneous, and (4) lateral variations in basement velocity, which we estimate to be as large as 3% [*Clarke et al.*, 1997].

5. Discussion

[28] Identification of the onset of streaming ice flow and the factors that control its position within the ice sheet is fundamental in assessing the future of the WAIS. We have imaged thick low-velocity zones underlying the upper reaches of ice streams C and D, which we interpret to be layered sedimentary basins. These sedimentary basins are closely linked to streaming ice flow, as observed from measurements of ice velocity [*Joughin et al.*, 1999, 2002], with the extent of these basins influencing the location of streaming ice flow. We believe that the inland termination of these sedimentary basins may provide an important control on the potential migration of streaming ice flow and rapid drainage of the WAIS.

[29] If the past serves as an indicator for the subglacial influence on modern streaming ice flow, then the imprint left from the paleo-ice streams of the northern hemisphere ice sheets should provide clues to controls on the drainage system of the WAIS. Most evidence points to topographic controls and soft-sediment cover in controlling the location of the paleo-ice streams that drained the Laurentide and Scandinavian ice sheets [*Stokes and Clark*, 2001], though recent work in Finland [*Punkari*, 1995] and northern Canada [*Stokes and Clark*, 2003] has revealed paleo-ice streams that once flowed over a crystalline bed that lacked topographic controls. The modern ice streams of the WAIS have yet to show any indication of streaming ice flow over a crystalline bed, with all previous observations pointing to soft sediments beneath the streaming ice [*Blankenship et al.*, 1986, 1987, 2001; *Rooney et al.*, 1987a, 1987b, 1991; *Smith*, 1997; *Anandakrishnan et al.*, 1998; *Bell et al.*, 1998; *Studinger et al.*, 2001; *Vaughan et al.*, 2003]. Our seismic results also link streaming ice flow to soft, subglacial sediments and nonstreaming ice to a crystalline or discontinuous sediment bed. This is not to say that streaming ice cannot exist over a bed lacking sediment cover or topographic lows, but current subglacial conditions point to the necessity and preference for laterally continuous soft sediments in the onset regions and main trunks of streaming ice flow in West Antarctica.

[30] Subglacial topography has been interpreted to play an important role in streaming ice flow, both among the paleo-ice streams of the northern hemisphere ice sheets [e.g., *Laymon*, 1992; *Hodgson*, 1994; *Kaplan et al.*, 1999] and the modern ice streams of the WAIS [*Smith*, 1997; *Anandakrishnan et al.*, 1998; *Bell et al.*, 1998; *Blankenship*

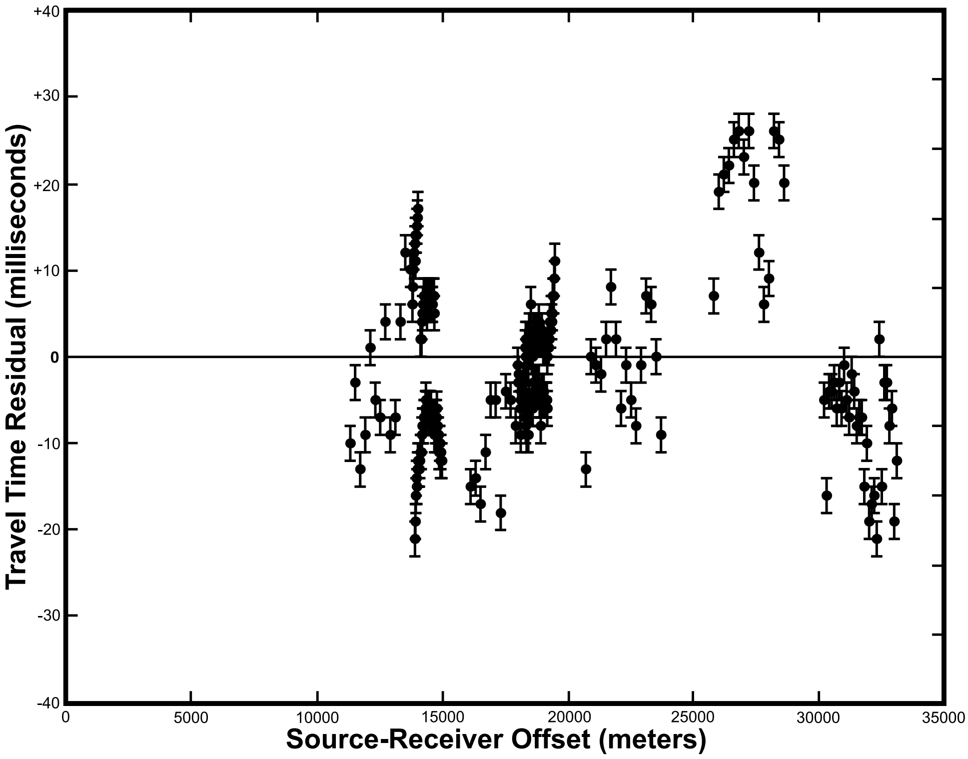

Figure 8. Residual plot for the DT profile. The best fit model has an RMS residual of $\Delta t_{\mathrm{RMS}} = 9$ ms and a normalized $\chi^2 = 21.1$, based on 311 data points and an estimated picking uncertainty of $\sigma_t = 2$ ms (this model is shown in Figure 6a). A maximum $\Delta t_{\mathrm{RMS}} = 30$ ms observed at an offset of 27,600 m, which is approximately 0.6% of the total travel time, while most of the residuals are less than 15 ms.

et al., 2001; *Studinger et al.*, 2001]. Several authors have also placed emphasis on the close tie between the onset of streaming ice flow and a step in subglacial topography [*McIntyre*, 1985; *Joughin et al.*, 1999; *Hulbe et al.*, 2000], and our results are in agreement with this interpretation, revealing an indirect influence of subglacial topography on the initiation of streaming ice flow along the upper reaches of ice streams C and D. Seismic evidence points to depressions in bedrock topography that have been filled by subglacial sediments, probably from marine deposition in a preglacial setting [*Scherer et al.*, 1998; *Tulaczyk et al.*, 1998]. Ice flow velocity increase in the flow direction of C1B occurs with no identifiable step in bed topography (Figure 5), and the enhanced velocity of ice stream D occurs over a subtle low in basal topography (Figure 6). However, the enhanced flow in both cases overlies lows in bedrock topography that have been filled with sediments to give a smooth bed topography. Thus one way in which topographic lows favor ice streaming is by focusing sediment deposition during times of deglaciation.

5.1. A Sedimentary Basin Pinching Out in the Upstream Reaches of Tributary C1B: Sediments and the Onset of Streaming Ice Flow

[31] A key to understanding streaming ice flow along the Siple Coast and across the WAIS in general lies in constraining the subglacial requirements for initiating and enhancing basal sliding, thereby establishing the drainage system of the inland ice reservoir of West Antarctica. Our C1B results reveal a tight coupling between subglacial sediments and ice streaming. The base of the ice shows little relief in the downstream direction; continuous basement and then mixed basement sediments give way to continuous sediments as the basement steps downward (Figure 5).

[32] Ice velocity increases slowly downglacier to the onset of continuous sediments, where it increases rapidly before plateauing at a high value. The basal shear stress for ice flow is high where the ice rests on continuous basement, exhibits local minima over sediment ‘‘puddles’’ where sediment is discontinuous, reaches a large maximum on the last outcrop of basement, and then falls rapidly downglacier as the velocity increases (Figure 5b). Some care is required in interpretation owing to the possibility of third dimension variations and to the effects of longitudinal stress variations on flow, but the tie between the presence of subglacial sediment and efficient basal lubrication appears clear and strong. The two local regions of ice basement contact between sediment regions appear to be acting as poorly lubricated ‘‘sticky spots,’’ similar to those observed farther downglacier on ice stream C [*Anandakrishnan and Alley*, 1994, 1997; *Anandakrishnan et al.*, 2001]. Lubrication between these sticky spots may arise from generation of till from the sediments, or simply from a smoother upper surface of sediment than of (presumably crystalline) basement. We interpret the upstream margin of this continuous sedimentary basin to mark the onset of the channelization of ice

into streaming ice flow, as well as the maximum inland extent of streaming ice flow, along tributary C1B.

[33] The inland extent of the sedimentary drape across West Antarctica may outline the maximum possible extent of streaming ice along the Siple Coast, as previously suggested by *Studinger et al.* [2001]. Deposition of the sedimentary drape across West Antarctica is inferred to have occurred in preglacial times or during periods of ice sheet collapse [*Scherer et al.*, 1998; *Studinger et al.*, 2001], accumulating in depressions in bedrock topography [*Studinger et al.*, 2001]. Our seismic results reveal a sediment drape infilling bedrock lows along the upstream reaches of ice streams C and D, in agreement with the interpretation of regional gravity and magnetic surveys [*Bell et al.*, 1998; *Blankenship et al.*, 2001; *Studinger et al.*, 2001], which may be providing a lubricating till layer that is important in establishing the onset of streaming ice flow along the Siple Coast of West Antarctica.

5.2. A Sedimentary Basin Coincident With Streaming Ice Along Ice Stream D: Sediments and Streaming Ice Flow

[34] We imaged a thick sedimentary basin beneath the full width of ice stream D, with the lateral margins of continuous sediment cover coincident with enhanced ice flow (Figure 6). Given the obvious benefits of keeping field party members out of crevasses, our DT profile across ice stream D and its shear margins was located safely upstream of the zone of prominent marginal crevassing as mapped by *Shabtaie and Bentley* [1987]. Profile DT is thus upstream of the peak in driving stress along ice stream D, which can be taken as the onset of streaming ice flow [*Joughin et al.*, 1999, 2002; *Bindschadler et al.*, 2001]. The peak ice flow speed reached near the center of our profile is barely over 60 m/yr (Figure 6), slow for ice streaming on the Siple Coast. Nonetheless, we crossed a region of faster flowing ice flanked by much slower flow, and that faster flowing region continues directly into ice stream D as usually defined. Also, the ice flow velocity across the profile is substantially too high to arise solely from internal deformation of the ice, and indicates substantial basal lubrication (calculations following *Patterson* [1994, Chapter 11]). Thus some workers might prefer to refer to our DT profile as crossing a tributary of ice stream D or an upstream extension of ice stream D; for simplicity, here we say that our profile crosses ice stream D.

[35] Fast ice flow is observed across the entire width of ice stream D, although significant variability in sediment thickness is observed for the modeled sedimentary basin (Figure 6). The surface flow field is relatively uniform within the ice stream, suggesting that only the absence or presence of sediments is important in fast ice flow, not the total thickness, a physically reasonable result. Continuous sediments thus appear to be a necessary condition (though not sufficient; see *Parizek et al.* [2002, 2003] for the meltwater contribution) for streaming ice flow, as hypothesized in previous studies along the Siple Coast [e.g., *Alley et al.*, 1986, 1987; *Engelhardt et al.*, 1990; *Tulaczyk et al.*, 1998].

[36] The inversion results of *Joughin et al.* [2004b] show generally reduced basal shear stress with enhanced ice flow velocity in the ice stream near our DT profile, in comparison to the ice stream flanks. Within the ice stream, the calculated basal shear stress shows strong spatial variations with cross-flow bands of high drag looking something like speed bumps. Our profile lies in a region of locally high basal drag, as shown in Figure 6, but very close to a region of very low basal drag. Given possible errors, and the smoothing effect of the basal drag inversion, we are hesitant to draw strong conclusions. (This also raises some questions about detailed matching of basal drag and sediment coverage in the discontinuous sediment region of our profile along ice stream C, which exhibits a slightly banded basal drag structure in the inversion of *Joughin et al.* [2004b]. Detailed assignment of basal drag to sticky spots at spatial scales on the order of the ice thickness or less will likely require targeted 3-D modeling and surveys.)

[37] Because the sedimentary package imaged across the width of our DT profile is relatively thin, pinch-out seems possible along flow, and may contribute to the banding in basal drag. However, we lack the necessary data, and other explanations are also possible (e.g., ‘‘waves’’ of basal lubricant [*Gray et al.*, 2005]). The most robust conclusion about basal lubrication at DT is that the ice stream, with enhanced basal velocity, overlies an $\sim$25 km wide region of continuous basal sediments, and ends where those sediments end on one side and become discontinuous on the other side. Thus, as for our longitudinal profile of ice stream C, we find that streaming ice flow and continuous basal sediments occur together.

[38] We note further that ice velocities across DT may be less than in the downstream reaches of the Siple Coast ice streams in part because the sedimentary fill, and hence the ice stream, is rather narrow at DT. *Raymond* [2000] showed that as basal drag is reduced, side drag comes to control streaming ice flow, and centerline velocity approaches a fourth-power dependence on ice stream width. The ice stream width at DT is half or less of more typical values downstream. The region of discontinuous sediments just outboard of the ice stream at DT clearly includes sediment patches too narrow to allow independent streaming.

5.3. Profile of a Layered Sedimentary Basin

[39] Subglacial sediments are a necessary condition for streaming ice flow under low basal shear stress, with a meters thick dilatant till layer enhancing basal lubrication and allowing for rapid basal motion [*Blankenship et al.*, 1986, 1987, 2001; *Engelhardt et al.*, 1990; *Alley and Whillans*, 1991; *Rooney et al.*, 1991; *Smith*, 1997; *Anandakrishnan et al.*, 1998; *Bell et al.*, 1998; *Engelhardt and Kamb*, 1998; *Tulaczyk et al.*, 1998; *Doake et al.*, 2001; *Kamb*, 2001; *Studinger et al.*, 2001; *Vaughan et al.*, 2003]. Our longitudinal profile on ice stream D, DL, imaged a layered sedimentary basin near the proposed onset of streaming ice flow [*Joughin et al.*, 1999; *Bindschadler et al.*, 2001]. Differences between the velocity-depth profiles of this basin and of the sediments under ice stream B [*Rooney et al.*, 1991] (Figure 2), suggest considerable variability in sedimentary character along the Siple Coast of West Antarctica.

[40] The uppermost layer of the sedimentary basin along DL is interpreted to be a till-water system, providing a template for rapid basal motion and increased ice flow. Its low seismic velocity of $V_P = 1580 \pm 150$ m/s suggests that

this material is highly porous, saturated, and has water pressure almost as large as the ice overburden pressure (possesses a low effective pressure [*Blankenship et al.*, 1987]). These characteristics of the till layer are in agreement with previous seismic results along ice stream B, where a dilatant till with a seismic velocity of V_P = 1550 ± 150 m/s and a thickness of less than 10 m was imaged [*Blankenship et al.*, 1986, 1987; *Rooney et al.*, 1987a]. Similar results were obtained from borehole observations along ice stream D that indicate the presence of a till layer and water conduit system beneath the streaming ice [*Kamb*, 2001]. Our seismic results show that this till layer is continuous along the entirety of our line, varying in thickness from less than 10 m to about 17 m.

[41] Beneath this deformable till are three sedimentary layers, which become more lithified with depth. This succession of sedimentary packages may be the result of episodic sedimentation during periods of open water across West Antarctica [*Scherer et al.*, 1998; *Studinger et al.*, 2001]. Our study area along ice streams C and D would be below sea level during times of complete deglaciation, even after isostatic compensation [*Drewry*, 1983], allowing marine deposition of a sediment drape across the submerged basement [*Studinger et al.*, 2001]. This interpretation is supported by our two seismic refraction profiles, which image sediments filling lows in basement topography.

[42] The variability in subglacial geology between ice streams B and D, as determined from seismic reflection, indicates that there are significant subglacial heterogeneities across the Siple Coast. Both locations have sedimentary basins, but our ice stream D profile shows more rapid lithification of sediments with depth than at ice stream B [*Rooney et al.*, 1991] (Figure 2). Similarity in ice flow between the two ice streams likely arises from the meters thick dilatant till in each profile. This similarity suggests that streaming ice flow is largely influenced by interactions at the ice bed interface and the presence of a soft bed to allow rapid basal motion [e.g., *Alley et al.*, 1986, 1987; *Engelhardt et al.*, 1990; *Alley and Whillans*, 1991]. As long as the sediments underlying this till can be scoured by the overriding ice, the deformable till can be replenished and fast ice flow can continue to drain the WAIS.

6. Conclusions

[43] Mapping the extent of subglacial sediment cover across the Siple Coast of West Antarctica is key to understanding the possible advancement of streaming ice flow into the WAIS and the future stability of the WAIS. We have imaged sedimentary basins beneath the upstream reaches of ice streams C and D, with fast flowing ice overlying these sedimentary packages. Seismic imaging along ice streams C and D (along with previous work performed along ice stream B [*Rooney et al.*, 1991]), coupled with driving stress, basal shear stress, and ice velocity measurements, reinforces previous observations that streaming ice flow and the location of the Siple Coast ice streams are controlled by basement topographic lows that were filled with sedimentary packages when the WAIS was absent [*Scherer et al.*, 1998; *Studinger et al.*, 2001]. The upper few meters of the sedimentary basin we imaged beneath ice stream D consists of a soft till-water system, creating a lubricated bed that enables streaming ice flow along the ice stream.

[44] The limited extent of these sedimentary basins beneath the WAIS may serve to limit low-shear-stress ice streaming. Although our data set remains very small, we find that low-shear-stress, high-speed flow requires continuous subglacial sediments, and that streaming ice flow is already occurring on laterally extensive sediments (In downstream regions, some sedimentary rocks host slow ice flow because the ice bed interface is subfreezing [*Kamb*, 2001].). Pending further data in modeling the future changes around the onsets of current Siple Coast ice streams, it seems reasonable to retain inefficient basal lubrication. Similar studies elsewhere, and particularly around the rapidly changing glaciers feeding Pine Island Bay [*Thomas et al.*, 2004] would be of considerable interest.

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