Journal of Glaciology, Vol. 39, No. 133, 1993

Laterally varying basal conditions beneath Ice Streams B and C, West Antarctica

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ABSTRACT. The study of the phase of reflections of P waves off the base of Ice Streams B and C, and Ridge BC, indicates that acoustic impedances of the beds of both ice streams vary laterally. In some places, the impedance is higher than in the ice (a high-impedance bed) and in some places it is less (a low-impedance bed). The estimated impedances in a dilated bed (porosity 0.4) and in a model of the lowermost ice that takes into account the relatively low P-wave speed in ice at or very near the melting point are nearly the same. Whether the impedance in the bed is greater or less than in the ice could depend on minor changes in the nature of the sediments composing the bed, or the physical state of the bed (e.g. porosity) that could occur laterally. Lateral variations of this kind provide a ready explanation for the observations on Ice Stream B. The bed under a substantial part of Ice Stream C that exhibits a low-impedance bed also must be dilated. The evaluation of the state of the bed under the rest of Ice Stream C and on Ridge BC requires further analysis, which is in progress.

INTRODUCTION

The dynamic behavior of that part of the West Antarctic ice sheet that flows into the Ross Ice Shelf is dominated by the dynamic behavior of the so-called "Ross" ice streams (Bentley, 1987). These ice streams flow much more rapidly than the surrounding ice and carry over 90% of the ice flux (Shabtaie and Bentley, 1987). A major objective of the Siple Coast Project, begun in 1983 by the University of Wisconsin (UW), The Ohio State University (OSU), NASA, the University of Chicago, California Institute of Technology and the University of Alaska has been to gain a thorough understanding of the processes that form, maintain and destroy them, which is essential to an understanding of the dynamics of the West Antarctic ice sheet as a whole.

The initial UW work on this joint project was conducted in the 1983–84 field season on fast-moving Ice Stream B (Fig. 1). A wide-angle seismic-reflection experiment indicated the presence of a low-wave-speed layer below the ice. Results from the velocity analysis suggested that P-wave and S-wave speeds of the layer were $1550 \pm 150\,\mathrm{m\,s^{-1}}$ and $145 \pm 15\,\mathrm{m\,s^{-1}}$, respectively (Blankenship and others, 1987). From these speeds, it was estimated that the basal layer has a porosity of about 40%, and that the effective pressure in it is $50 \pm 50\,\mathrm{kPa}$, i.e. that the pore pressure in the basal layer is 99% of the overburden pressure (Blankenship and others, 1987). Later seismic profiling across a wide area showed the basal layer was almost universally present in the survey area; thickness ranged from 2 m or less (i.e. too small to be resolved) to 12 m (Rooney and others, 1987). Alley and others (1987) concluded from theoretical studies that the basal layer under Ice Stream B is probably deforming and hypothesized that a basal deforming till is the cause of the high velocity of ice movement observed at UpB ($450\,\mathrm{m\,year^{-1}}$; Whillans and others, 1987). The physical characteristics deduced from seismic observations were confirmed in the 1988–89 and 1989–90 field seasons from drilling through the ice stream. A dilatant layer (40% porosity, density $1.99\,\mathrm{Mg\,m^{-3}}$), with an effective pressure in the range of $30 \pm 100\,\mathrm{kPa}$, was indeed found at the bed in the holes (Engelhardt and others, 1990).

In 1988–89, the UW project moved to UpC on Ice Stream C (Fig. 1), which is nearly stagnant, i.e. moving at only a few meters per year (Whillans and others, 1987), to investigate whether a significant difference between the beds of Ice Streams B and C could be discerned. In this study, we use the phase of the basal reflections to investigate the physical properties of these beds. The phase of a reflection from a boundary between two media is determined by the contrast in acoustic impedances (product of wave speed and density) across the boundary. The reflection coefficient for a wave in medium 1 incident on an interface with medium 2 is given by $(Z_2 - Z_1)/(Z_2 + Z_1)$, where Z_1 and Z_2 are the acoustic impedances in media 1 and 2, respectively. Reflection of a P wave from a medium with a lesser acoustic impedance ($Z_2 < Z_1$) results in a phase reversal, whereas reflection from an acoustically "harder" layer ($Z_2 > Z_1$) will cause no phase change. This is true regardless of angle of incidence because the wave speed in the beds is in any case less than in the ice (cf. Richards, 1988, fig. 3a).

We are justified in assuming real impedances in both the ice and the bed because loss factors in both media are small. For the loss factor in the ice, see Bentley and Kohnen (1976). The best evidence for a low loss factor in the subglacial sediments is the recording of reflections

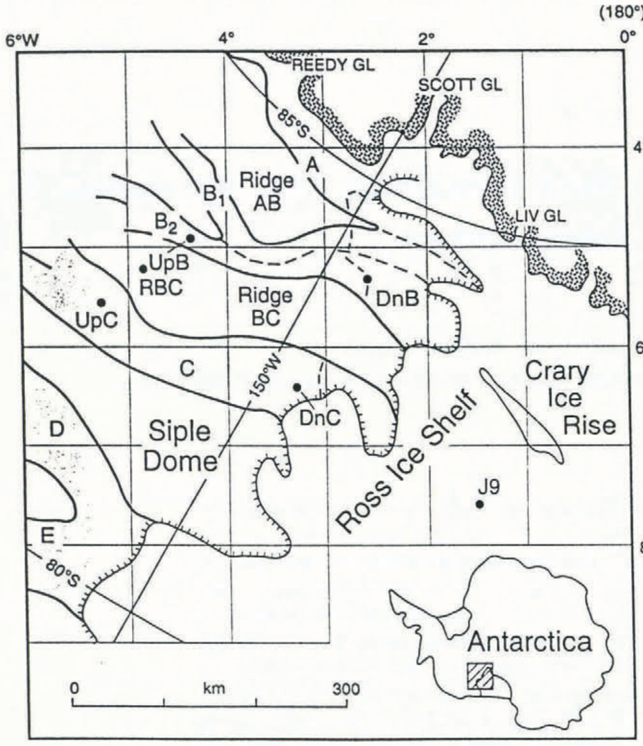

Fig. 1. Location of the Siple–Gould Coast region. Camp locations indicated are Upstream B (UpB), Downstream B (DnB), Upstream C (UpC), Ridge BC (RBC), Downstream C (DnC) and the Ross Ice Shelf Project drillhole site J9. The grounding lines of Ice Streams A, B and C are shown by the hatched line (Bindschadler and others, 1987; Shabtaie and Bentley, 1987). The lighter dashed line indicates the grounding line (as mapped by Rose (1979)); this line is now believed to be the "coupling" line of the ice streams (Alley and others, 1987). In the rectangular-grid coordinate system shown, 0° grid longitude lies along the Greenwich and 180° meridians with grid north toward Greenwich, and the grid equator passes through the geographic South Pole. (Modified from Shabtaie and Bentley (1987).) Squares have sides of length equal to 1° of latitude along the 180° meridian.

from interfaces many hundreds of meters deep in the bed (Rooney and others, 1991). A good example of a strong sub-bottom reflection at a depth of about 10 wavelengths beneath the base of Ice Stream C can be seen in Figure 7. Consequently, we will henceforth consider the phases of the reflections to be real, i.e. either 0° or 180°.

To determine whether there is a phase reversal in a particular reflection, we compare its initial deflection on the seismogram with that of the direct wave through the ice. Although this comparison, in principle, should not be necessary for uniform explosive sources, it is a precaution against anomalous source conditions (such as, perhaps, energy blow-out into a crevasse) and accidental reversal of polarity in connecting seismometers to the seismic cables. We will show later that possible phase shifts along the direct-wave propagation path do not affect the results.

FIELD WORK

At UpB (Fig. 2), seismic-reflection data were collected along the SA, SB and SC lines in 1983–84 (Blankenship

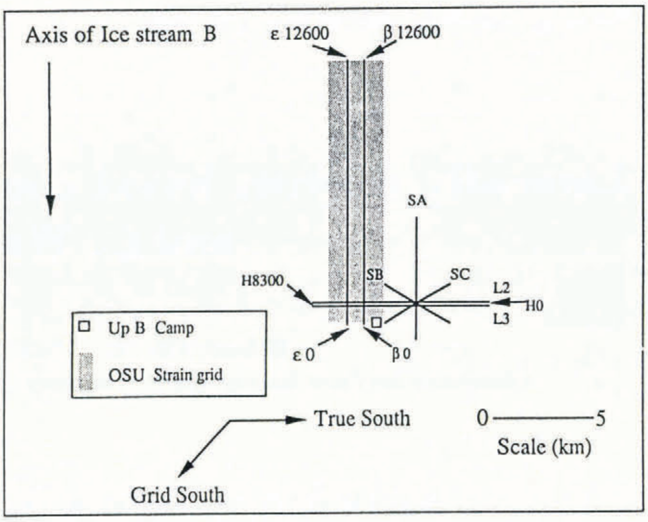

Fig. 2. Lay-out of the seismic reflection experiments on Ice Stream B.

and others, 1987; Blankenship, 1989), along lines L2 and L3 in 1984–85 (Rooney and others, 1987), and along lines ϵ, β, 1800, 7920, 9720 and 11520 in 1985–86 (Rooney, 1988).

On Ice Stream C and Ridge BC during the 1988–89 field season (Fig. 3), nearly 60 km of high-resolution P-wave reflection profile data were collected; 36 km on a line (X line) profiled across the ice stream, 7 km on a parallel (Y line), and the rest along 7 km and 3.5 km lines parallel to the axis of the ice stream (U, V, W and Z lines) (Atre, 1990). Five wide-angle experiments were also conducted at X3600, Z3600/X18000 (two perpendicular profiles), X34200 and X59400.

For all experiments the receiving spread was 690 m long, with 24 geophones spaced 30 m apart. Sources were 0.15 kg or 0.45 kg explosive charges detonated in boreholes 14–18 m deep. Amplifiers had a flat frequency response without phase shift within the frequency band employed (50–450 Hz), which extended well beyond the signal frequencies (200–300 Hz) at both ends. Detailed descriptions of the data collection and reduction can be

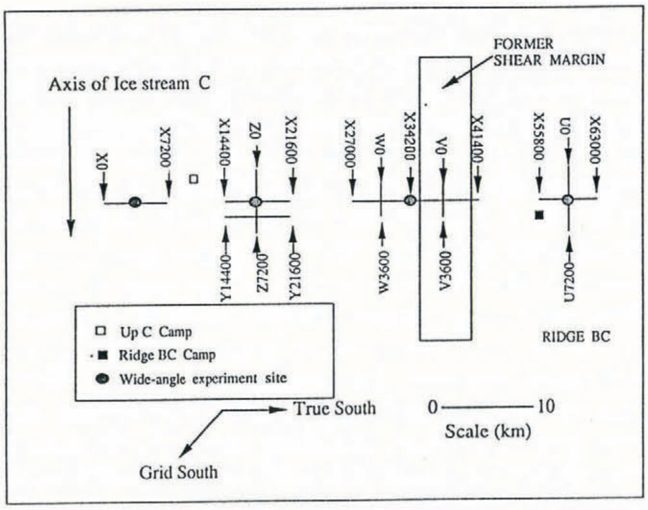

Fig. 3. Lay-out of the seismic reflection experiments on Ice Stream C and Ridge BC.

found in Blankenship and others (1987), Rooney and others (1987), Rooney (1988) and Atre (1990). For this study of phases, we have not carried out any stacking, move-out corrections or any other processing that could conceivably distort the phase information. Instead, we rely on visual inspection of individual seismograms.

REFLECTION PHASES

Ice Stream C

The seismograms shown in Figure 4 provide typical examples of the relationship between direct and reflected P-wave arrivals that is observed on all records between X0 and X7200 of the X line (Fig. 3). The basal reflection has the same phase as the direct P wave, which implies that the acoustic impedance of the bed in this region is higher than that of the ice (we will call this a high-impedance bed). In contrast, the seismic records in Figure 5, which are typical of the X16200–X21600 section of the X line, all show reflections that are reversed in phase, which means that the acoustic impedance of the bed is less than that of the ice (a low-impedance bed). Basal reflections on the Y and Z lines, which respectively intersect and run parallel to this section of the X line, also show phase reversals.

Some records in this region show reflections of very low amplitude (Fig. 6), presumably where the acoustic impedance of the bed is nearly equal to that of the ice. In both halves of Figure 6, which show seismograms from

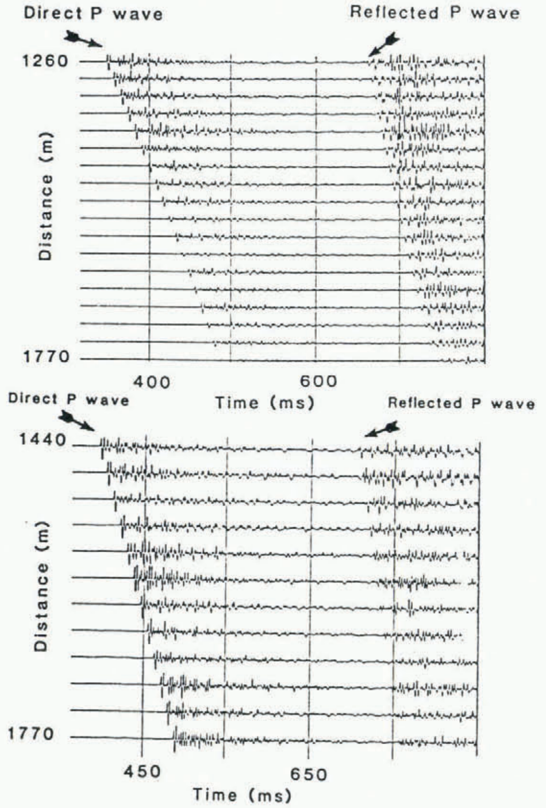

Fig. 4. Parts of seismic records from locations X3960 and X6120. The direct P waves through the ice and the P waves reflected from the ice bottom are marked.

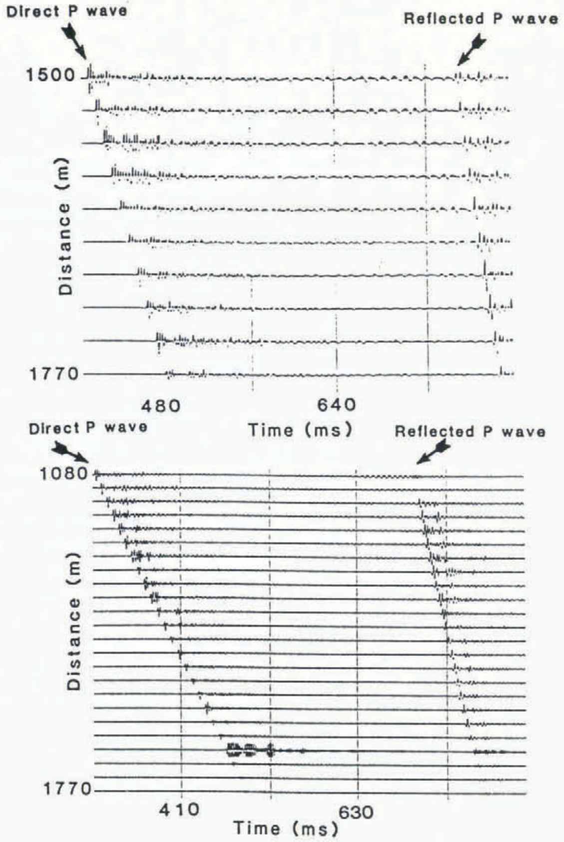

Fig. 5. Parts of seismic records from locations X18720 and X19080. The direct P waves through the ice and the P waves reflected from the ice bottom are marked.

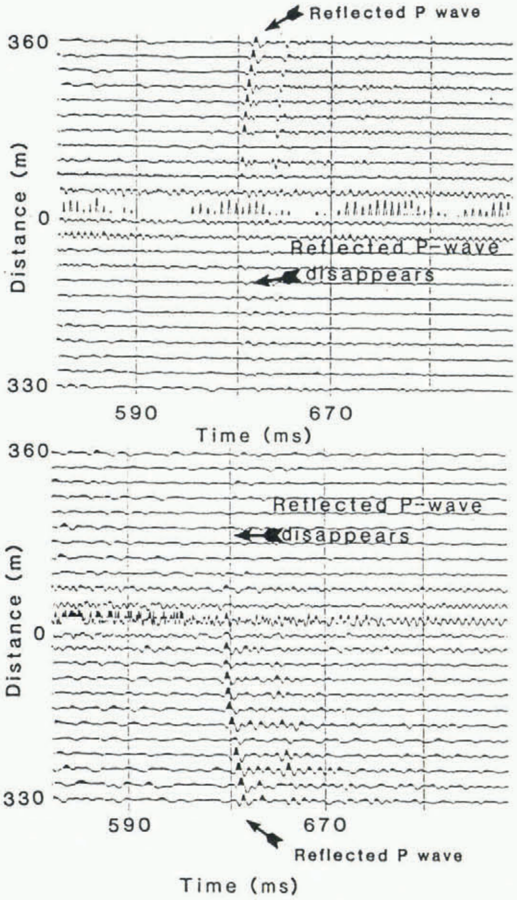

Fig. 6. Parts of seismic records from locations X20880 and X21240. The P waves reflected from the ice bottom are marked. Direct arrivals do not appear on these center-shot records.

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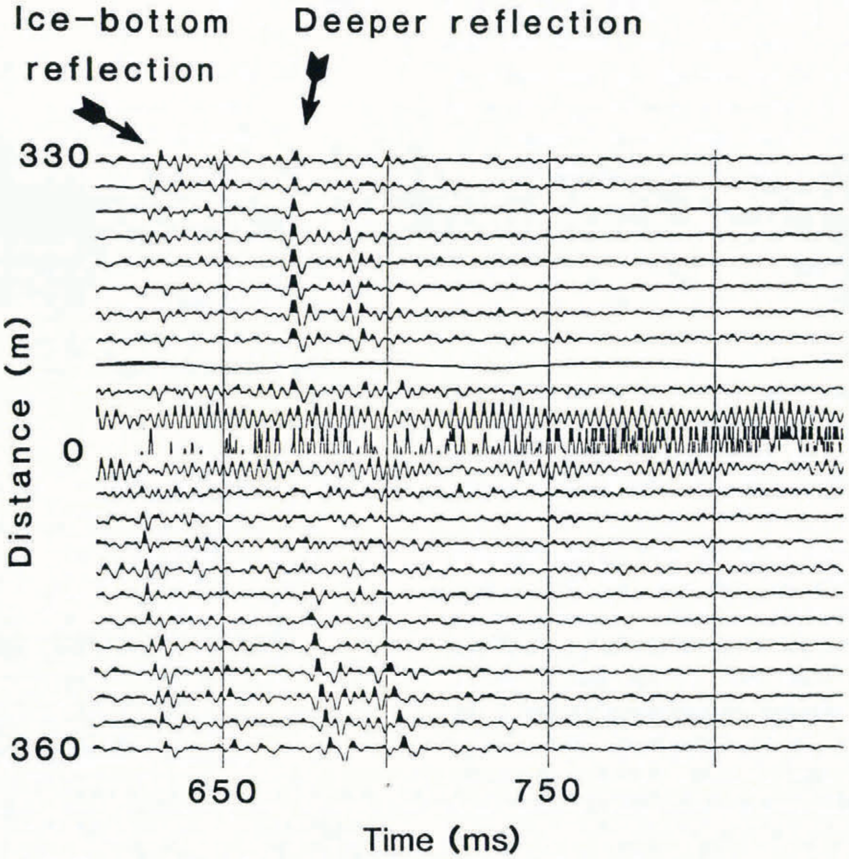

Fig. 7. Part of a seismic record from a center shot at location X31680. The P waves reflected from the ice bottom and from a reflector below the ice bottom are marked. Direct arrivals do not appear on the record.

adjacent shots, the basal reflection disappears within a spread length. (These disappearances also show on shots at greater offsets.) The move-out of the reflection on both seismograms shows that the basal interface is flat, so the disappearance strongly suggests that there is a significant lateral change in the acoustic impedance of the bed over a distance of the order of 100 m.

A low-impedance bed generally characterizes the next section of the X line (X2700–X35280) and the intersecting W line, although there are a few isolated areas, 50–100 m wide, where the amplitude of the basal reflection is low and its phase unclear (e.g. Fig. 7; also on offset shots).

On Ridge BC, all the basal reflections, including those from the U and V lines, are unreversed, indicating a high-

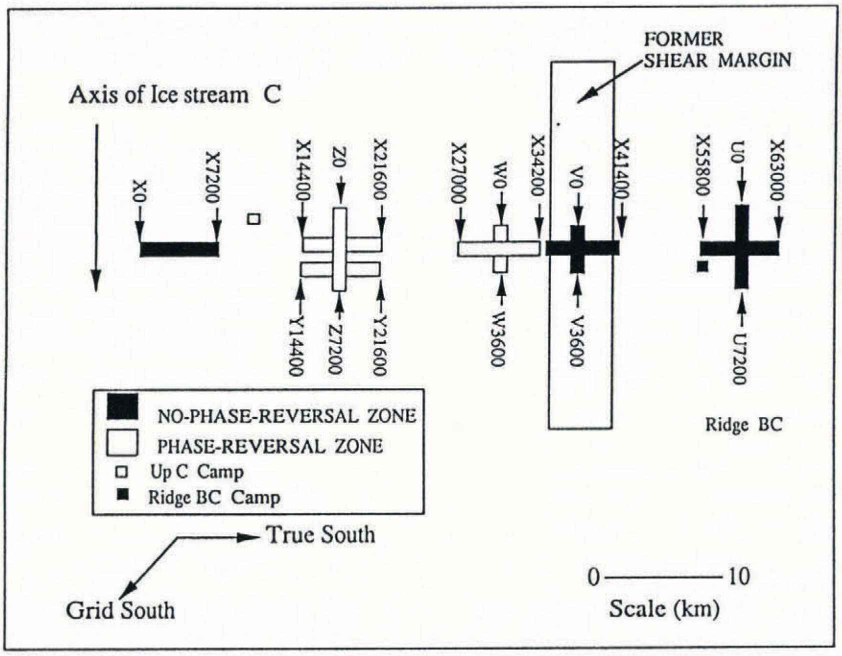

Fig. 8. A cartoon summarizing the phase of the P wave reflected from the ice bottom relative to that of the direct P wave on Ice Stream C and Ridge BC.

impedance bed. The change-over from a low-impedance bed occurs between X34920 and X35280, a distance of 360 m. This transition is within a few hundred meters of the inner boundary of the former margin as located by Shabtaie and others (1987) from radar observations.

In summary, the surveyed sections of the beds of Ice Stream C and Ridge BC can be divided into three zones: two zones of high-impedance (no-phase-reversal) beds beneath the central ice stream and beneath Ridge BC, including the former shear margin, and an intervening low-impedance (phase-reversed) zone more than 20 km wide (Fig. 8). Seismic-reflection data strongly suggest that there is a layer of low-wave-speed sediments everywhere beneath the seismic line on Ice Stream C (Munson and Bentley, 1992), so it is highly unlikely that the high-impedance bed is solid bedrock.

Ice Stream B

Lines SA, SB and SC, near UpB camp (Fig. 2) were the ones along which Blankenship and others (1987) first detected a low-velocity basal layer. Subsequent analyses indicated that this layer probably existed in the whole area of study (Rooney and others, 1987). On seismic records from these three lines (e.g. Fig. 9), the phase of the

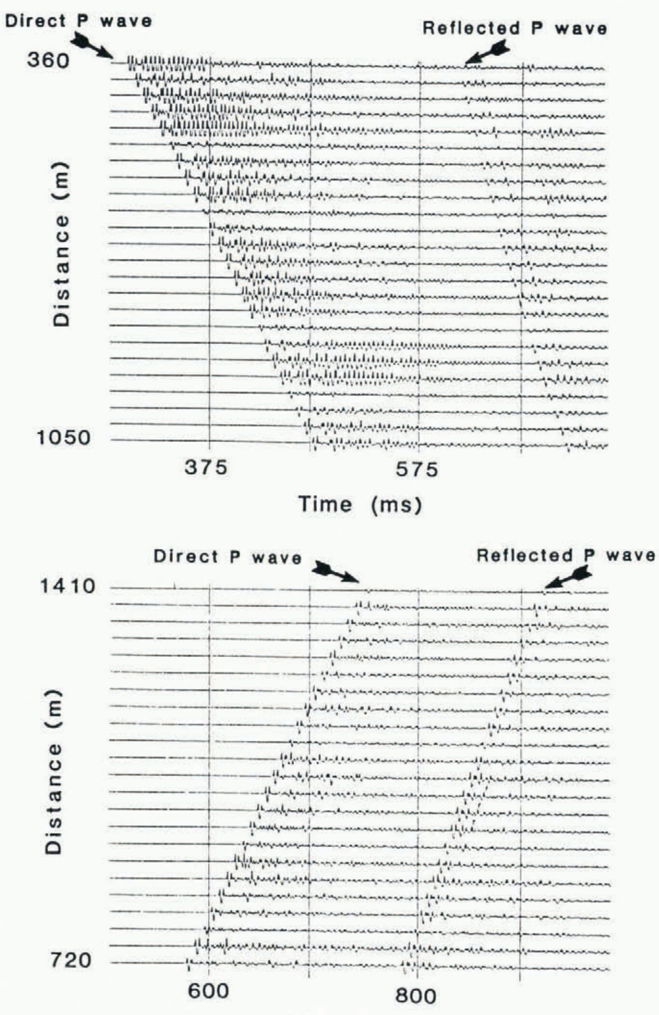

Fig. 9. Parts of seismic records from seismic profile SA. The direct P waves and P waves reflected from the ice bottom are marked.

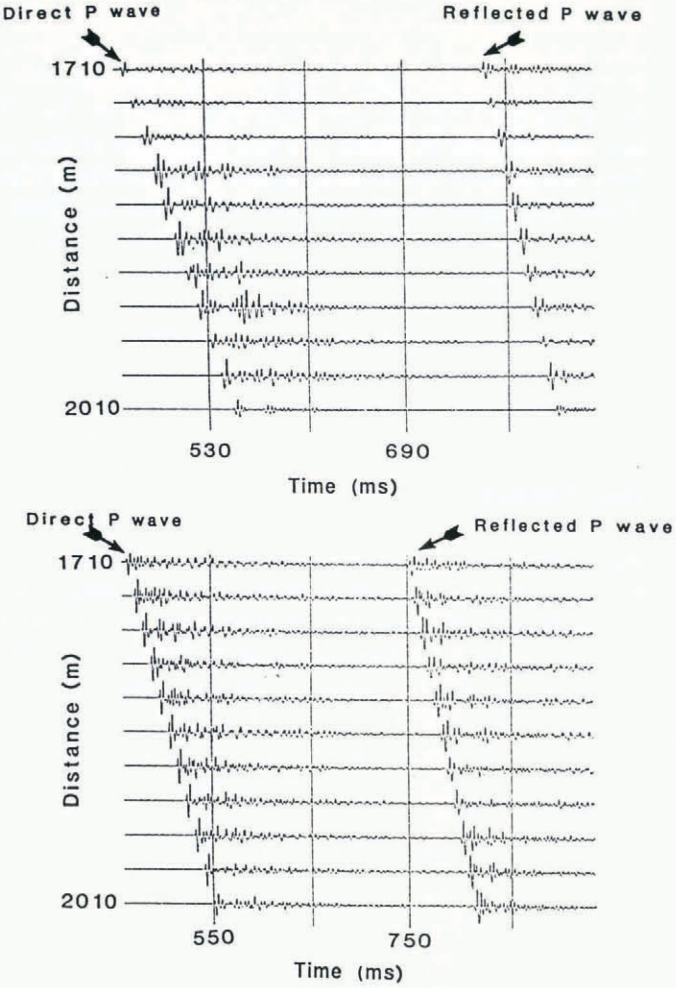

Fig. 10. Parts of seismic records from locations L3H0 and L3H360. The direct P waves through the ice and the P waves reflected from the ice bottom are marked.

basal reflection is the same as the phase of the direct arrival, which implies that the acoustic impedance of the bed there is higher than that of the ice.

On lines L2 and L3 (Fig. 2) there is a boundary, near the true southern end of the lines, across which there is a

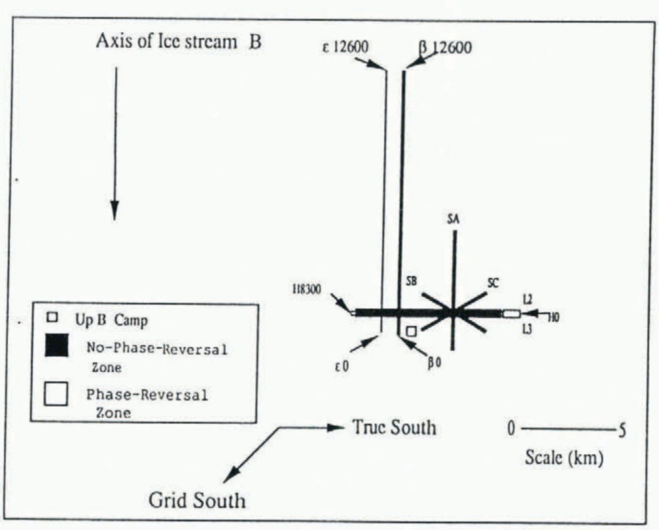

Fig. 11. A cartoon summarizing the phase of the P wave reflected from the ice bottom relative to that of the direct P wave on Ice Stream B.

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lateral change from the phase reversal on the south (the side nearer the margin of the ice stream) to no phase reversal on the north. Basal reflections from H0 and H360 on the L3 line (Fig. 10) and H720 and H1080 on the L2 line are reversed in phase. On the remainder of the L lines, the phase of the basal reflections is not reversed. The β line also shows basal reflections unreversed in phase. Seismic records from the ϵ line are generally too noisy to determine whether there is a phase reversal or not.

Thus, almost all of the seismically sampled part of the bed of Ice Stream B shows no phase reversal (Fig. 11), which means it has an acoustic impedance greater than in the ice.

DISCUSSION

Before proceeding to interpret the phase information in terms of the characteristics of the basal interface, we should examine the possibility that the phase of the direct P wave is not the same everywhere. This conceivably could be true for the following reasons. When the ray path for a P wave has its lowest point (turning point) in a velocity gradient, it theoretically undergoes a phase change of $\pi/2$ (Aki and Richards, 1980, p. 418), whereas a head wave (the "refraction arrival") does not. Because of the large stresses in Ice Stream B, the firn increases in density with depth much more rapidly than on Ice Stream C (Retzlaff and Bentley, 1993). This means that for some range of distances, far enough for the ray paths to reach the velocity maximum in Ice Stream B but not in Ice Stream C, the direct waves might differ in phase by $\pi/2$ between the two ice streams. Since the phase on the reflection is picked to be either 0 or π, that shift might lead to an apparent phase reversal between the two ice streams for reflections of the same phase. However, both at very short distances, where rays in both ice streams have their turning points within the velocity gradient, and at very large distances, where rays in both ice streams reach the velocity maximum, this consideration would not lead to a phase difference between ice streams.

This concept can best be checked by consideration of the seismograms themselves. A phase shift of $\pi/2$ in the direct wave should manifest itself as a diminution of the initial upward motion of the ground, which, from an explosive source, would be strong in the absence of any phase shift. Examination of many seismograms reveals no such effect on Ice Stream C (e.g. Figs 4 and 5) but shows that it does occur commonly on Ice Stream B (e.g. Fig. 10). This is just the opposite of the expectation from the arguments of the previous paragraph. More to the point, however, is the fact that even where the first deflection is weak, it is still upwards, so the upwards first motion of the reflection still is taken to be no change in phase. Thus, the seismograms themselves refute the suggestion that a phase change at the turning point might be affecting the judgement as to the phase of the basal reflection.

Critical to any discussion of the reflection coefficient at the base of the ice streams is the evaluation of the acoustic impedances in the ice and subglacial bed. The impedance in the ice is difficult to ascertain precisely because of the uncertainty in the value of the P-wave speed, v_p, in ice at, or very near, the pressure-melting point. In his review monograph, Röthlisberger (1972) stated that for v_p in temperate ice "a value of 3600–3620 m s^{-1} ... is probably the best figure now recommendable, but higher values (around 3660 m s^{-1}) ... are equally justified". We have been unable to find any better information published since that report, so we will assume $v_p = 3630 \pm 30$ m s^{-1}.

The next question is to what part of the glacier that low wave speed applies. Robin (1958) found that the decrease in wave speed as the melting point (0°C) is approached starts to become appreciable at a temperature of –0.3°C and is pronounced at –0.2°C. At UpB, the bed must be at the pressure-melting point (because this is a fast-sliding glacier); a temperature of –0.2°C colder than that is reached 5 m above the bed (Engelhardt and others, 1990). We therefore assign a wave speed of 3630 ± 30 m s^{-1} to a 5 m thick basal layer. The wave speed just above that layer is taken to be 3800 m s^{-1}, in accordance with the formula of Kohnen (1974).

The frequency of the reflected signals is 200–300 Hz, which corresponds to a wavelength of 16 ± 3 m. This is substantially larger than the thickness of the low-speed basal layer, so the wave impinging on the base of the ice will be traveling at a speed that is some average of the wave speeds in and above the layer. We assume that the contribution to the effective value of v_p of the ice at a distance z from the boundary is proportional to e^{-kz}, where k is the wave number. Thus, we take as a weighted average value for v_p

$$\overline{v_p} = \frac{\int_0^H v_p e^{-kz} dz}{\int_0^H e^{-kz} dz}$$

where H is the ice thickness. Carrying out the integration of this expression with the numerical values given above yields $\overline{v_p} = 3650 \pm 40$ m s^{-1} (the uncertainty combines the uncertainties in v_p and the wavelength). This, finally, leads (for ice with density 0.912 Mg m^{-3}) to an acoustic impedance $z_{ice} = 3.33 \pm 0.04 \times 10^6$ kg m^{-2} s^{-1}.

In this analysis, we have assumed that the ice is vertically isotropic. Blankenship and others (1987) found an averaged v_p in the ice sheet of 3831 m s^{-1}, just about what one would expect for an isotropic ice sheet. This is consistent with the fact that the ice sheet is probably anisotropic with a horizontal girdle fabric (Blankenship, 1989). Furthermore, Blankenship (1989) showed that a layer at the base of the ice with a fabric tightly clustered about the vertical, which might be expected if basal stresses have had an important effect and which would produce a high vertical v_p, is probably thin or non-existent at UpB. We therefore do not consider anisotropy further.

At UpB, the wave speed and the density in the subglacial till have been measured: $v_p = 1550 \pm 150$ m s^{-1}; density = 1.99 Mg m^{-3} (the uncertainty in the density at UpB is negligible compared to the uncertainty in v_p). However, the cited value for v_p should not be taken to imply that the most probable value is the central one, because the field measurements by Blankenship and others (1987) actually provide only an upper limit to v_p (1700 m s^{-1}); the lower limit is simply the sound speed through water. Values closer to 1700 m s^{-1} than to 1400 m s^{-1} are much more likely, according to the range of wave speeds given by Hamilton (1970), Morgan (1969)

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and Anderson (1974) for shallow-water marine and lacustrine sediments. Converted to acoustic impedances (Z_{bed}), that range is shown in Figure 12 as the band labelled "till". (Corrections have been applied to their measured wave speeds to adjust for the different sound speeds in water of different temperature and salinity, as discussed by Blankenship and others (1987). Densities were calculated from porosities assuming a rock-particle density of 2.67 $\mathrm{Mg\,m^{-3}}$.) Only the upper part of the error bar that represents the measurement at UpB (Fig. 12) overlaps the expected range.

Both the measured and the expected values overlap the acoustic impedance in the ice (horizontal band in Figure 12). This implies that, at the measured porosity, the acoustic-impedance ratio could be either greater or less than 1, depending on the particular character of the local sediment or ice. But it is clear from Figure 12 that, if the impedance of the bed Z_{bed} is to be higher than Z_{ice}, it must be near to the top of its possible range, regardless of what value we choose for the ice.

We should distinguish here between the *uncertainty* in the knowledge of the impedance, which limits our ability to calculate the reflection coefficient but which does not change from place to place, and the *variability* in the impedance, which reflects physical changes in the media and hence can change laterally. The principal source of both uncertainty and variability is v_p. The uncertainty in v_p is not much greater for the subglacial material than for the ice; 45 $\mathrm{m\,s^{-1}}$ (3%) (Blankenship and others, 1987) vs 40 $\mathrm{m\,s^{-1}}$ (1%). However, there is a large difference in the variability of v_p between the ice and the bed. The range in acoustic impedance for ice shown by the band in Figure 12 arises from the uncertainties in the wave speed as a function of temperature and in the thickness of the low-speed basal-ice layer. Neither the wave speed nor the layer thickness is likely to change much laterally, so the

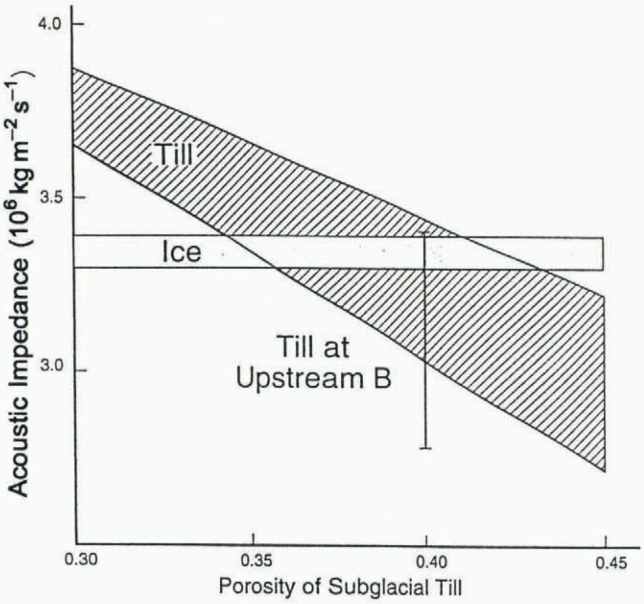

Fig. 12. Acoustic impedance of till as a function of porosity. The band covered by diagonal shading represents the range of reported laboratory measurements. The vertical error bar gives the allowable range in the subglacial till at UpB. The allowable acoustic-impedance range for ice is indicated by the shaded horizontal band.

variability of Z_{ice} is small. In the bed, on the other hand, lateral variations in v_p, hence in the acoustic impedance, could occur easily as a result of changes in the subglacial material, so the variability in Z_{bed} is substantial. Since the impedances are so nearly matched between the ice and the bed, the reflection coefficient could easily change sign as a result of only a small change in either the porosity, the sediment type or both.

For Ice Stream B, beneath which we expect (from the uniformly fast movement of ice) that the till is everywhere dilated, we conclude that there must be a lateral change in the nature of the till, perhaps in the geologic character of the rock fragments that it comprises, near the beginning of the L lines where the reflection phase changes. There may also be a slightly different character under the ϵ line where, as noted above, the reflections exhibit a low signal : noise ratio.

The situation beneath Ice Stream C is less clear. We can assert that the bed in the 20 km wide low-impedance zone must be dilated, but the phase information is not enough to distinguish, for the two high-impedance zones — the central ice stream and Ridge BC — between a dilated and an undilated basal till. Work is in progress to ascertain whether the reflection amplitudes can answer that question.

CONCLUSION

Reflection phases cannot be used simply to distinguish between a dilated and an undilated subglacial bed because a dilated bed may have an acoustic impedance that is either greater than or less than that in the basal ice. This fact is borne out by a lateral change in phase on Ice Stream B where the bed is presumably everywhere dilated. A dilated bed is also indicated for a 20 km wide zone on Ice Stream C that shows subtle flowlines in Landsat images. The state of the bed beneath the central part of Ice Stream C and on Ridge BC cannot be determined from phase data alone.

ACKNOWLEDGEMENTS

We thank S. Anandakrishnan, N. Lord, C. G. Munson, A. N. Novick and R. Retzlaff for their help with the field work on Ice Stream C, and personnel of the Polar Ice Coring Office, University of Alaska, for their efficient preparation of the seismic shot holes. We are grateful to D. R. MacAyeal and two anonymous referees for comments that have led to substantial improvements in the paper. This research was supported by U.S. National Science Foundation grant DPP86-14011. This is Contribution 529 of the Geophysical and Polar Research Center, University of Wisconsin–Madison.

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MS received 4 September 1992 and in revised form 1 March 1993