

SEISMIC SHORT-REFRACTION STUDIES ON THE ROSS ICE SHELF, ANTARCTICA*

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ABSTRACT. Seismic short-refraction studies were carried out at five stations on the Ross Ice Shelf during the 1976–77 summer season as part of the comprehensive Ross Ice Shelf Geophysical and Glaciological Survey. Measurements of the velocities of compressional waves were made at each location. Compressional wave velocities were measured along more than one azimuth at three sites, and shear wave velocities (both components) at two. Travel-time curves were fitted to an exponential expression by means of a non-linear least-squares regression technique. The errors in the apparent velocities are estimated to be about $\pm 50 \text{ m s}^{-1}$ at short distances, diminishing to about $\pm 10 \text{ m s}^{-1}$ near the ends of the profiles. Compressional-wave velocities show only slight variations with azimuth and only over certain depth intervals, showing that constant-velocity surfaces are essentially horizontal. Shear-wave velocities, however, exhibit large variations according to azimuth and polarization, indicating that transverse isotropy is violated at least in the upper 30–40 m of the ice shelf. It is believed that the anisotropy is caused by structural details in the firn perhaps modified by preferred crystal orientation and that it may arise at least partly from anisotropic stresses in the ice shelf.

RÉSUMÉ. *Études des réfractions sismiques courtes sur le Ross Ice Shelf dans l'Antarctique.* On a exécuté des études de réfraction sismiques courtes en cinq stations du Ross Ice Shelf au cours de la saison d'été 1976–77 dans le cadre des expéditions glaciologiques et géophysiques. En chaque point on a mesuré les vitesses des ondes de compression. On a mesuré ces vitesses des ondes de compression selon plusieurs azimuts en trois sites et les vitesses des ondes de cisaillement (deux composantes) en deux sites. Les courbes distance parcourue-temps ont été ramenées à une expression exponentielle par une technique de régression aux moindres carrés non linéaire. Les erreurs dans les vitesses apparentes sont estimées à $\pm 50 \text{ m s}^{-1}$ à de courtes distances, diminuant jusqu'à $\pm 10 \text{ m s}^{-1}$ près de l'extrémité des profils. Les vitesses des ondes de compression ne montrent que de faibles variations avec l'azimuth et seulement pour certains intervalles de profondeurs, ce qui montre que les surfaces de vitesse constante sont essentiellement horizontales. Les vitesses des ondes de cisaillement cependant, montrent de larges variations selon l'azimuth et la polarisation, ce qui indique que l'isotropie transversale n'existe pas au moins dans les 30 ou 40 m supérieurs de la calotte. On pense que cette anisotropie provient de détails de structure dans le névé qui peuvent être modifiés par une orientation préférentielle des cristaux et que ceci peut se produire au moins partiellement à cause d'efforts anisotropes dans la calotte glaciaire.

ZUSAMMENFASSUNG. *Seismische Refraktionsstudien über kurze Distanzen auf dem Ross Ice Shelf, Antarktika.* Während der Sommerkampagne 1976–77 des umfassenden Ross Ice Shelf Geophysical and Glaciological Survey wurden seismische Refraktionsstudien auf kurze Distanzen an fünf Stationen durchgeführt. An jedem Punkt wurde die Geschwindigkeit der Kompressionswellen gemessen. Auf drei Stationen wurden diese Geschwindigkeiten in mehr als einem Azimut bestimmt, die Geschwindigkeit der Scherwellen in beiden Komponenten auf zwei Stationen. Die Laufzeitkurven wurden mit Hilfe einer nichtlinearen Regressions-technik nach der Methode der kleinsten Quadratsumme einer Exponentialfunktion angepasst. Die Genauigkeit der scheinbaren Geschwindigkeiten kann bei kurzen Distanzen auf etwa $\pm 50 \text{ m s}^{-1}$ abgeschätzt werden; sie steigt nahe den Profilen auf ca. $\pm 10 \text{ m s}^{-1}$. Die Geschwindigkeit der Kompressionswellen schwankt mit dem Azimut nur geringfügig und nur bei gewissen Tiefenintervallen, womit sich erweist, dass die Flächen konstanter Geschwindigkeit im wesentlichen horizontal sind. Dagegen zeigen die Geschwindigkeiten der Schwellen beträchtliche Schwankungen je nach Azimut und Polarisation, was auf eine transversale Anisotropie zumindest in den oberen 30–40 m des Schelfeises hinweist. Sie wird auf strukturelle Einzelheiten im Firn, vielleicht in Form von bevorzugten Kristallorientierungen, und zumindest teilweise auf anisotropen Druck im Schelfeis zurückgeführt.

INTRODUCTION

Seismic short-refraction experiments were carried out at five stations on the Ross Ice Shelf during the 1976–77 season as part of the Ross Ice Shelf Geophysical and Glaciological Survey (RIGGS). At all of the locations (Fig. 1), compressional wave (P-wave) velocities were measured. At the three base camps, C-16, J-9, and Q-13, shear waves (S-waves) of both

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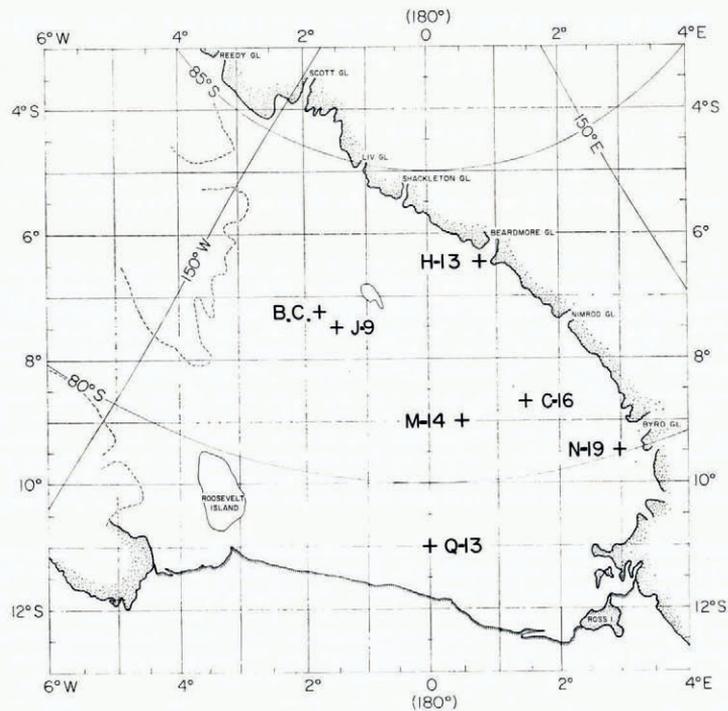


Fig. 1. Locations of short-refraction stations on the Ross Ice Shelf occupied during RIGGS 1976-77.

TABLE I. RIGGS 1976-77 SHORT REFRACTION PROFILE SPECIFICATIONS

Station	Line No.	Wave type	Maximum shot-spread separation m	Spread orientation (magnetic)
J-9	1	P	1 242	N.-S.
	1	SV	346	N.-S.
	1	SH	346	N.-S.
	2	P	1 150	N. 60° W.
	3	P	1 012	N. 120° W.
	1A	SV	396	E.-W.
	1A	SH	446	E.-W.
C-16	1	P	246	N.-S.
	1	SV	250	N.-S.
	1	SH	300	N.-S.
	2	P	1 256	N. 120° E.
	3	P	1 754	N. 60° E.
Q-13	1	P	1 012	N.-S.
	1	SV	297	N.-S.
	1	SH	500	N.-S.
	2	P	1 012	N. 45° W.
	2	SV	368	N. 45° W.
	2	SH	322	N. 45° W.
H-13		P	320	N.-S.
M-14		P	322	N.-S.
N-19		P	325	N.-S.

horizontal (SH) and vertical (SV) polarization were also measured. Profiles for both P- and S-waves were made along more than one geographical azimuth at J-9 and Q-13, and for P-waves only at C-16. All spreads were in-line, and recordings were made on photosensitive paper using standard exploration-type seismographs. Table I gives the specifications of the 1976-77 short refraction profiles. Further details are given by Kirchner (unpublished).

DATA ANALYSIS

The seismograms were read to the nearest 0.1 ms with an estimated error of ± 0.3 ms. By fitting the following exponential expression to the travel-time curves, an unbiased, reproducible estimate of the velocity (i.e. inverse slope) as a function of distance was obtained (Kirchner, unpublished):

$$t = a_1\{1 - \exp(-a_2x)\} + a_3\{1 - \exp(-a_4x)\} + a_5x, \quad (1)$$

where x is the shot-detector distance, t is the predicted travel time, and the a_i are constants that were determined by a computer-assisted non-linear regression analysis (see Kirchner (unpublished) for a more detailed discussion).

Velocities as a function of distance were converted to velocities as a function of depth using the standard integral formula of Wiechert, Herglotz, and Bateman. An estimate of the errors in velocity was obtained by separating the travel-time points into two groups—those falling above the best-fit travel-time curve, and those falling below it. Equation (1) was then fitted to each set of points and the analysis leading to the velocity-depth curves was repeated. This method should produce an approximate probable error as a function of depth. The probable errors in both P-wave and S-wave velocities were estimated in this way to be about ± 50 m s⁻¹ at short distances, falling to ± 10 m s⁻¹ near the ends of the profiles.

P-WAVE RESULTS

The P-wave velocities for the five stations are very similar to each other (Kirchner, unpublished). Furthermore, only at C-16 among the three base camps was there any clear difference in v_P along different azimuths, and even there the differences was small: a maximum of about 4% in the 10 to 25 m depth range and negligible at other depths (Fig. 2). At J-9 and Q-13 the difference was less than 1% at all depths. The azimuthal invariance shows that the "layers" (i.e. the constant-velocity surfaces) are very nearly horizontal, a 1% velocity difference would correspond to a dip of about 0.2%.

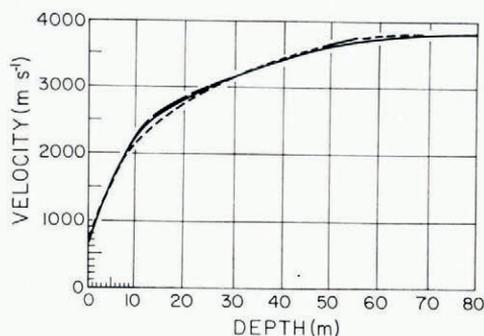


Fig. 2. P-wave velocity plotted against depth at station C-16. The three profiles were oriented at 60° to each other. Two of the curves are indistinguishable; the third (dashed line) shows a maximum deviation at a depth of 15 m.

Profiles at the three base camps were extended to sufficiently large distances (>1 km) to record maximum P-wave velocities (corresponding to solid ice). These velocities all were within ± 5 m s⁻¹ of 3 800 m s⁻¹, thus comparing reasonably well with values for the grid western part of the ice shelf calculated by three different methods (Robertson, unpublished): 3 810 \pm 18 m s⁻¹, 3 812 \pm 5 m s⁻¹, and 3 824 \pm 14 m s⁻¹. However, Robertson (unpublished) did find evidence at some stations for sloping layers (see also Kirchner and others (1979) for further evidence of anisotropy).

P-wave velocities obtained from refraction shooting on the Ross Ice Shelf and on other ice shelves, are lower than those from refraction shooting on grounded ice sheets and laboratory measurements. Other reported values on ice shelves include: 3 750 \pm 20 m s⁻¹ at Maudheim (Robin, 1958), a mean of 3 789 \pm 7 m s⁻¹ on the Ross Ice Shelf (Crary and others, 1962), and 3 810 m s⁻¹ (no error given) at Ellsworth on the Filchner Ice Shelf (Thiel and Behrendt, 1959).

The refracted-wave velocities correspond to horizontal propagation. A larger value, 3 840 m s⁻¹, was found for the P-wave traveling vertically from seismic bore-hole logging at Little America V (Thiel and Ostenso, 1961), a value in apparent agreement with those measured on grounded ice at the same temperature ($\approx -25^\circ\text{C}$) (Kohnen, 1974), although, unfortunately, Thiel and Ostenso (1961) gave no evidence of the accuracy of their measurement. The difference in maximum refracted P-wave velocities measured on ice shelves and grounded ice has been attributed by several investigators (e.g. Thiel and Ostenso, 1961) to dissimilar temperature and density structures, related to the different basal boundary condition in the two cases. However, no quantitatively satisfactory explanation has been found.

Kohnen and Bentley (1977) have analyzed ultrasonic (2 MHz) measurements on cores from station J-9. The horizontal (i.e. transverse to the core) velocity is consistently lower than the vertical (longitudinal) velocity down to the greatest depth sampled by refraction shooting (about 100 m), but is still significantly larger than those obtained by refraction shooting. In the upper 100 m, the mean difference between the two is 30 \pm 9 m s⁻¹, a difference in agreement with that between the velocities from logging and refraction shooting. This suggests that there may be a general preferred crystal orientation or structural fabric in the ice shelf, favorable to higher vertical velocities, to explain the low refracted-wave velocities. However, this does not explain why the refracted-wave velocities are less than the horizontal ultrasonic velocities on the ice cores, nor is there any apparent reason for such a fabric to develop in the upper 100 m of ice shelves but not in grounded ice sheets. The problem still requires further study.

S-WAVE RESULTS

The S-wave velocity–depth results for stations C-16, J-9, and Q-13 show large variations. Velocities of the horizontally-polarized (SH) wave (v_{SH}) along different azimuths show differences that are several times the error over extensive depth ranges (e.g. Fig. 3, between 15 and 30 m). Differences in the velocities of the vertically-polarized (SV) wave (v_{SV}) along different azimuths are smaller but still substantial, reaching a maximum of about 8% in the 5 to 35 m depth range at Q-13. Large differences between v_{SH} with v_{SV} along the same line (e.g. Figs 4 and 5) are also evident.

As already discussed, the P-wave results show that the firn layers are essentially horizontal. It follows that the S-wave velocities reveal anisotropy, to which they are more sensitive than P-wave velocities (Bentley, 1975). More specifically, the results indicate that the firn in the upper 30 or 40 m of the ice shelf is not only anisotropic (from the differences between SV and SH), but not even transversely isotropic (as shown by the azimuthal variation). Surface-wave dispersion analysis also indicates anisotropy (personal communication from D. Albert, 1978).

Comparison between SH and SV along a single line shows either a cross-over pattern (Fig. 4), such as that previously reported at Roosevelt Island base camp by Robertson (unpublished), or a curve in which SV predominates over most of the depth range sampled (Fig. 5). The cross-overs, where they do occur, are found at similar depths at each station (Table II). (The error in picking the depths to the cross-over points is about one meter.) Perhaps coincidentally, the cross-over points also approximate the depths to changes in the

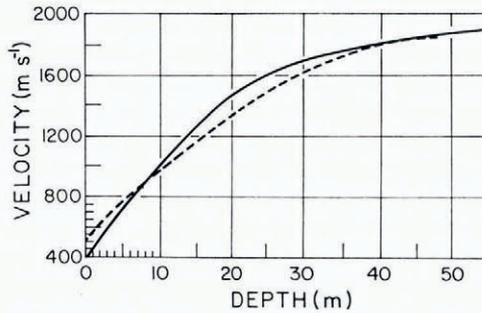


Fig. 3. SH-wave velocities plotted against depth at station J-9 for two perpendicular profiles.

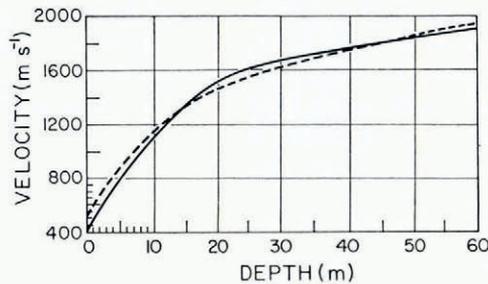


Fig. 4. SH and SV velocities plotted against depth at station C-16, showing a typical double-cross-over pattern.

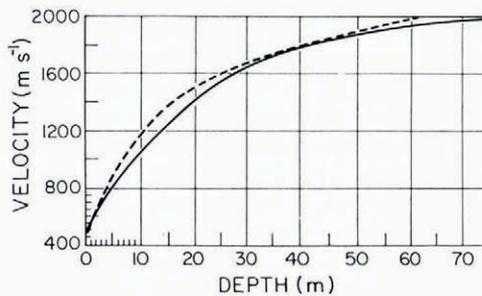


Fig. 5. SH and SV velocities plotted against depth at station Q-13, showing greater SV velocities at all depths.

TABLE II. DEPTH TO CROSS-OVERS IN SHEAR WAVE VELOCITY-DEPTH RESULTS

	C-16 No. 1	Q-13 No. 2	J-9 No. 1	Roosevelt Island*
	m	m	m	m
$V_{SV} > V_{SH}$ to $V_{SV} < V_{SH}$	14	8	13	12
$V_{SV} < V_{SH}$ to $V_{SV} > V_{SH}$	41	35	43	43

* From Robertson (unpublished).

P-wave velocity gradients (Robertson, unpublished), which in turn presumably correspond to changes in metamorphic mechanisms (Robertson and Bentley, 1975).

The higher near-surface SV velocities (compared with the SH velocities) could be due to vertical structures in the firn, such as vertically-elongated air pockets or vertical ice glands. Ramseier (1966) found anisotropy between vertical and horizontal directions in some measurements of unconfined compressive strength and air permeability on oriented polar snow samples, and he suggested a favorable vertical direction for larger bonds and a higher vertical density of bonding between snow grains. Such vertical features should become less important as the density increases with the depth. The occurrence of greater SH-wave velocities in the intermediate depth ranges might be explained by horizontal structures, such as the annual winter and summer layers or ice lenses, that became predominant as the vertical structures fade.

These models, however, provide no explanation for anisotropy in the horizontal plane; for that we must consider a non-vertical axis of symmetry. Bentley (1971) used a single-crystal model as an approximation to a random distribution of *c*-axis directions within a cone. The same general velocity pattern might be expected of any structural fabric that is axially symmetric. We have therefore attempted to fit the J-9 and Q-13 velocity distributions to the single-crystal model utilizing the velocities calculated as a function of the inclination of the *c*-axis (axis of symmetry) and the azimuth and angle of incidence of the seismic ray, by Bentley (1971, tables 1 and 2). (The angle of incidence in the present case is 90° .) No fit to the J-9 data (four lines at different azimuths (Table I)) was possible for an angle of inclination $I < 30^\circ$; the best fits occurred for inclinations between 75° and 90° . The Q-13 data have several possible fits (there are only two lines). Still, none was found for $I < 20^\circ$ and the best fits again were in the range $I \geq 75^\circ$, although several possible fits were found for inclinations between 20° and 90° . The single-crystal model therefore strongly implies a nearly-horizontal axis of symmetry, a configuration that does not correspond to observations of either crystalline or structural fabric anywhere on polar ice sheets. We conclude that the axially-symmetric model is not acceptable.

There are several reasons why this might be so:

1. A single-pole model may not be appropriate for the crystalline fabric. A two-pole fabric, for example, would be strongly anisotropic around any axis. Thus crystal orientation could contribute to the velocity anisotropy in the horizontal plane even with a fabric that is laterally symmetric about a vertical axis. The Little America V petrofabrics (Gow and Williamson, 1976) do show multiple maxima, albeit at greater depths.
2. There may be a rapid change with depth in the number of maxima and their orientations, again shown by the Little America V cores (Gow and Williamson, 1976), in which case the velocity may represent an average of several orientations.
3. Structural effects (such as ice lenses, subsurface fold structures, buried sastrugi, etc.), probably predominate over crystal anisotropy as the cause of the observed velocity distribution, particularly in the upper 30–40 m of the ice shelf where the largest velocity anisotropy occurs, since the little evidence available from core data indicates that strong crystal orientation is not present. (100-m holes were cored at stations C-16 and Q-13 during the 1977–78 season, but the results are not yet available.) The “single-crystal” model may not be a particularly good one for such structural phenomena. Nevertheless, the transverse anisotropy of the S-wave velocities requires that there be some characteristics in the firn that are directionally dependent, perhaps due to the wind, but also perhaps partly due to the stress distribution in the ice.

A detailed map of ice thickness of the Ross Ice Shelf compiled from airborne radar soundings and supplemented by seismic reflection and radar spot measurements made at survey stations over much of the shelf, reveals a complex pattern of ice-thickness variations

(Bentley and others, 1979). There is a clear implication that simple stress boundary conditions are not appropriate to the Ross Ice Shelf. Stresses needed to cause these thickness variations, and to tilt layers within the shelf (Robertson, unpublished; Kirchner and others, 1979), may also contribute substantially to horizontal anisotropy of near-surface velocities.

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