Journal of the Royal Society of New Zealand

ISSN: 0303-6758 (Print) 1175-8899 (Online) Journal homepage: https://www.tandfonline.com/loi/tnzr20

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To cite this article: C. R. Bentley & K. C. Jezek (1981) RISS, RISP and RIGGS: Post-IGY glaciological investigations of the Ross Ice Shelfin the U.S. programme, Journal of the Royal Society of New Zealand, 11:4, 355-372, DOI: 10.1080/03036758.1981.10423327

To link to this article: https://doi.org/10.1080/03036758.1981.10423327

Published online: 09 Feb 2012.

© *Journal of the Royal Society of New Zealand*,
Volume 11, Number 4, pp. 355-372, 12 figs.

RISS, RISP and RIGGS: Post-IGY glaciological investigations of the Ross Ice Shelf in the U.S. programme

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Geophysical and glaciological data compiled since the IGY have led to a new picture of ice shelf dynamics, so that the ice shelf is now viewed as a dynamic system that responds strongly to changes in ice input and grounding line fluctuations, and that may be crucial to the existence of the inland ice sheet.

Measurements over the last twenty years have revealed the following characteristics about the shelf. Sonic logging in a drill hole in the ice demonstrates a striking anomaly in wave velocity that may be associated with the passage of the ice over the grounding line. Seismic shear wave velocities show anisotropic characteristics that may reflect either crystal anisotropy or stress anisotropy, or both. The apparent dielectric constant in the ice is significantly less than expected from laboratory studies and may exhibit real regional variations. The mass output from the ice shelf is only about half as great as the glacier input through the Transantarctic Mountains plus the surface accumulation on the shelf and its West Antarctic drainage basin, yet measurements on the ice shelf indicate that the mass balance is near zero. This could be consistent with recovery from a past surge of the West Antarctic inland ice. The submarine topography is dominated by broad ridges and valleys extending in an unbroken pattern from the open Ross Sea past the Ross Ice Shelf to the grid eastern part of the West Antarctic inland ice. Convolutions in the ice thickness suggest turbulent flow. Detailed examination of thickness variations, of present-day ice movement rates, of the record of past flow lines revealed by radar soundings, and of isostatic gravity anomalies has led to a picture of Holocene retreat and fluctuations in grounding line positions during the last 1500 years.

INTRODUCTION

During the International Geophysical Year (IGY) and the International Geophysical Cooperation (IGC) that immediately followed, the first general geophysical and glaciological survey of the Ross Ice Shelf was carried out (Crary *et al.*, 1962). This pioneering oversnow traverse (Fig. 1) produced a great volume of information about the ice shelf and its regimen and yielded a broad outline of the characteristics of the ice shelf. We will be referring to various aspects of that work as we examine the way in which the outline has been filled in during the succeeding years.

In 1960-62 Swithinbank (1963) carried out the first precise measurements of ice movement rates related to the Ross Ice Shelf when he measured the velocities of seven outlet glaciers flowing through the Transantarctic Mountains. Ice thicknesses were calculated from gravity observations on the grounded glaciers and from elevation measurements on those that are floating, leading to an estimate of the mass flux into the ice shelf (Giovinetto *et al.*, 1966). Additional input to the shelf comes from the regions in between the main outlet glaciers; estimating that contribution led to a calculated total mass flux of $48 \pm 15 \times 10^{12}$ kg/yr. This figure was only half of the mass input to the surface of the East Antarctic drainage system that feeds those glaciers, as estimated by Giovinetto (1964). There were many uncertainties in the estimates; nevertheless the indicated positive net mass flux for the drainage system was nearly twice the estimated standard error.

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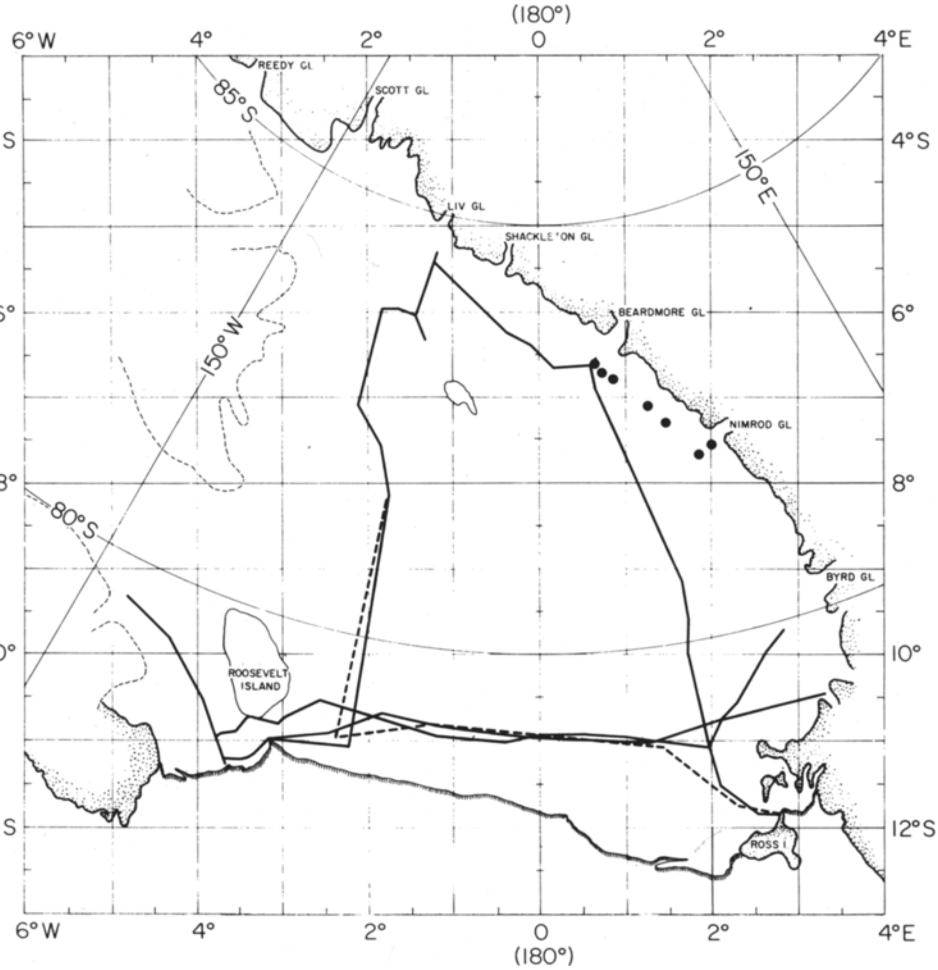

Fig. 1—Pre-RIGGS traverses. Solid line: IGY traverses (Crary *et al.,* 1962); dashed line: Ross Ice Shelf Survey (Dorrer *et al.,* 1969); solid circles: additional movement measurements (Swithinbank, pers. comm., 1979). This map and all succeeding maps show "grid" coordinates: a rectangular system with meridians parallel to the Greenwich meridian (north towards Greenwich) and its origin ($L = 0°$, $\lambda = 0°$) at the South Pole.

The next important programme was the Ross Ice Shelf Survey (RISS), which comprised measurements of the velocity vectors and snow accumulation rates along a trail from Ross Island nearly to Roosevelt Island, thence grid northward (for explanation of grid coordinates see caption to Figure 1) for about 300 km. Markers were set out for the movement measurements in the summer of 1962-63, and the remeasurement was carried out in 1965-66. As part of the 1962-63 survey, snow accumulation was measured at 1800 bamboo poles along the Dawson Trail between Little America V and McMurdo stations, the heights of which had previously been measured during 1959-60. The velocity measurements showed that the central part of the ice shelf moves nearly uniformly with a velocity between 800 and 935 mm/yr, the velocity diminishing fairly rapidly to the grid east within 100 km of Ross Island and more gradually grid westward within 200 km or so of Roosevelt Island. The surface accumulation rates showed a very regular pattern also, mostly differing little from 120 to 150 mm/yr of water equivalent, increasing rather rapidly to around 200 mm/yr only at each end of the trail where the trail approached the barrier. These values are about 25% less than the accumulation rates reported by Crary *et al.* (1962). It is possible that the discrepancy represents a real difference between the different time periods, but it is more likely, in light of subsequent studies, that the smaller figure represents the better long-term average.

With the velocity measurements it was then possible to calculate the outward mass flux through a section near the barrier and to compare it with the calculated mass flux into the ice shelf through the Transantarctic Mountains, combined with the estimated net mass flux on the surfaces of the ice shelf and the West Antarctic drainage system that feeds its grid western part. That comparison led to an estimate of the net mass flux for the drainage system that was again positive (Giovinetto and Zumberge, 1968), and again almost twice the estimated error. If the figure for the net mass flux is correct, and *if the ice is assumed to be in steady state,* then the implication would be that the mean rate of bottom melting under the entire ice shelf is more than 200 mm/yr, a very large melt rate; alternatively, the implication could be that the ice sheet is growing.

Also during the 1960s, a glaciological and geophysical programme was carried out on Roosevelt Island. The field programme was inaugurated in 1961-62 to determine the mass balance, strain rates, velocities, and thickness of the ice dome, and to measure the physical characteristics of the ice-bedrock interface and the underlying rock. The first survey was completed in the 1962-63 season, and the resurvey was carried out in 1967-68. The ice dome on Roosevelt Island was found to have a transverse elevation profile that is closely in accord with a Nye equilibrium profile, even though the accumulation rates on the grid southwest side of the island, and the corresponding velocities, are 2½ times as great as those on the grid northeast side (Clapp, 1965, and unpublished report). A recent, more detailed, analysis by Thomas *et al.* (1980) has shown that the Roosevelt Island strain rates can be satisfied by a Glen's flow law with an exponent of 3, provided there is a relatively high geothermal flux and a strain rate enhancement due to recrystallization that increases outward from the center of the ice rise.

During the 1960s early studies of electromagnetic wave propagation in the ice were begun. A. H. Waite continued his pioneering radar work of the IGY in 1961-62 and the next two seasons, when he made the first airborne ice thickness sounding surveys over, among other places, the marginal parts of the Ross Ice Shelf. Detailed studies on the McMurdo Ice Shelf, Skelton Glacier, and Roosevelt Island were undertaken in 1964-65 using Waite's equipment (Jiracek and Bentley, 1971). This work gave several accurate determinations of the wave velocity in shelf ice and showed that most of the wave velocity measurements were consistent with laboratory studies of the dielectric constant. Near Roosevelt Island, however, both on the ice shelf and on the ice rise itself, the wave velocity was found to be unexpectedly high. Jiracek (1967) also was the first to report evidence for anisotropy within the Antarctic ice sheet. That discovery, on Skelton Glacier, was pursued by Clough (1974), who showed a remarkable change in the depolarization patterns on Skelton Glacier, the strong anisotropy disappearing in a distance of only a few kilometres downstream.

In 1967-68 there began a long and fruitful series of radar sounding flights carried out in a joint U.S.-U.K. programme. Many of these sounding flights crossed the Ross Ice Shelf, leading even before RIGGS to a vastly improved map of the ice shelf thickness. Data were collected by personnel of the Scott Polar Research Institute (SPRI) using SPRI equipment mounted on a U.S. C-130 Hercules aircraft (Robin, this volume).

In 1969, J. W. Brodie suggested a multi-disciplinary study centered around a drill hole through the Ross Ice Shelf so it would be possible to study not only the ice but the ocean and ocean floor beneath the ice shelf. This suggestion was enthusiastically received and planning for the Ross Ice Shelf Project (RISP) began, both nationally in the United States and internationally through SCAR in 1970. It was apparent from the outset that a programme to measure ice thickness and water depths below the ice would be necessary in order to find an optimum site for the drill hole and that the maximum benefit would be obtained if it were viewed as a comprehensive geophysical and glaciological programme for study of the whole Ross Ice Shelf and the solid earth beneath, rather than simply a site

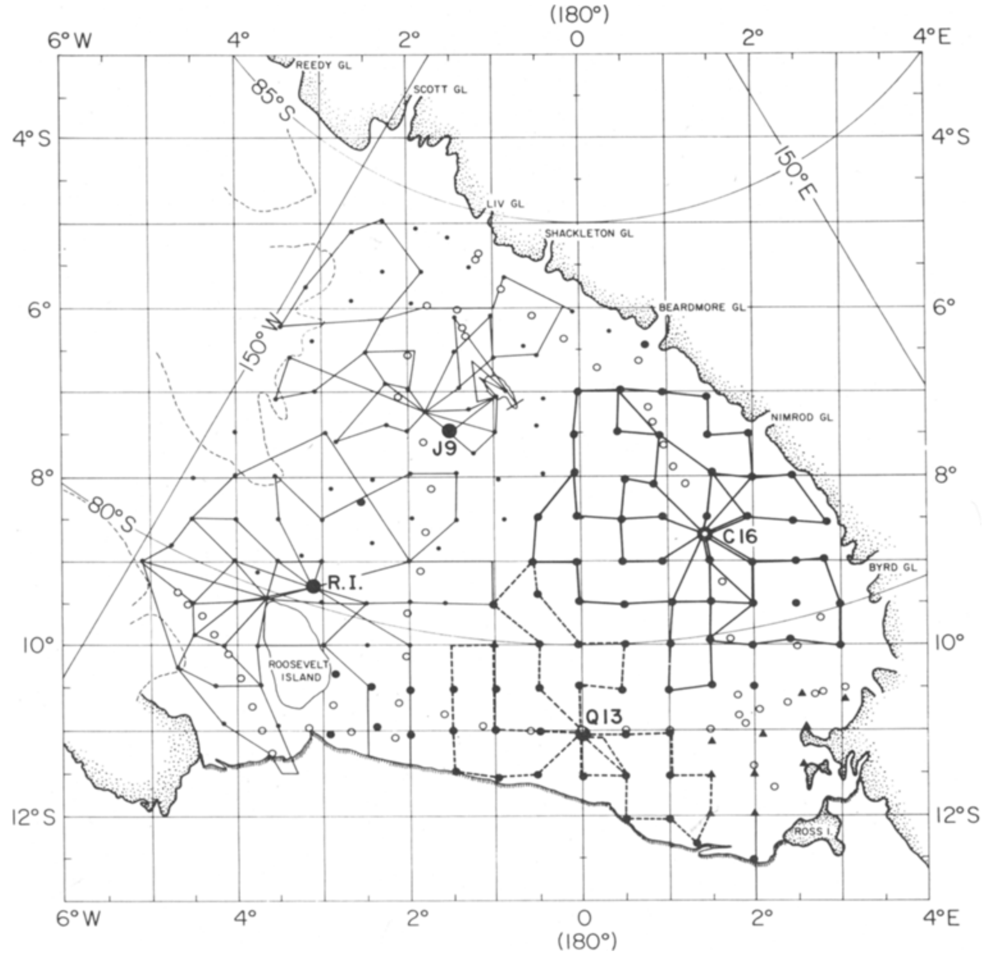

Fig. 2—Location map for RIGGS showing surface stations (solid circles and triangles) and airborne radar flight lines (continuous and dashed lines). The open circles were stations occupied during IGY (Crary *et al.*, 1962). (Different symbols and line widths refer to different field seasons.)

selection project. Consequently, in 1973-74 the Ross Ice Shelf Geophysical and Glaciological Survey (RIGGS) commenced.

The RIGGS programme was a cooperative effort involving many different institutions, including the Geophysical and Polar Research Center at the University of Wisconsin-Madison (radar sounding, seismic reflection and refraction shooting, resistivity soundings, and gravity survey experiments); the University of Maine at Orono (strain rates, 10-m temperatures, and surface accumulation); the U.S. Geological Survey (absolute movement of ice); the University of Nevada Desert Research Institute (near-surface snow studies); State University of New York at Buffalo (shallow core drilling); and the University of Copenhagen (accumulation rates).

Each season's work involved setting up one or two base camps around which detailed, local, surveys were conducted, and from which remote stations, positioned roughly on a 55-km grid (Fig. 2), were occupied via Twin Otter aircraft. At most of the grid sites, stake networks were laid out using Tellurometers; the distances between stakes were remeasured after a year to determine the spreading rate of the ice shelf. Satellite positioning was carried out at 116 sites; 80 of them were reoccupied after one year to determine the velocity of the ice shelf. At a majority of the sites core samples to a depth of 10 m were collected for the determination of accumulation rates by identifying dated radioactive fallout horizons, and for the examination of oxygen isotope ratios; temperatures were usually measured in the 10-m holes. At about 15 stations scattered over the ice shelf, a more extensive geophysical programme, comprising principally seismic and electromagnetic studies of the internal physical properties of the ice, was carried out. Surface samples were collected for studies of precipitation chemistry at a number of stations, particularly in the grid eastern part of the ice shelf, and widely spaced 100-m holes were drilled for core. In addition, radar antennas were fitted beneath the wings of the Twin Otter aircraft and a series of radar flights was conducted from each base camp at the end of each season. These flights (Fig. 2) supplemented those of the joint U.S.-U.K. programme previously mentioned. The field work took five years to complete (including one season of inactivity due to a long C-130 casualty list) and has led to measurements and analyses over the whole ice shelf of ice thickness and internal properties, water depth, subglacial structure, strain rates, velocities, surface balance rates, and englacial temperatures. The analysis of this great mass of data is still continuing.

Meanwhile, after many difficulties, the RISP drilling team successfully completed an access hole to the ocean beneath the ice shelf in December 1977 and also obtained ice cores to a depth of 155 m (Clough and Hansen, 1979). A year later, an ice core through the entire thickness of the ice shelf was recovered (Zotikov *et al.*, 1980).

In the remainder of this paper we review the principal glaciological results of RISP and RIGGS; results more directed towards marine geophysics are discussed by Davey (this volume).

ICE PHYSICS

Correctly interpreting data collected using remote sensing techniques requires a knowledge of the physical properties of the ice. Although there is a large literature on laboratory measurements of the electrical and mechanical properties of ice, there is no guarantee that these properties are the same in an ice sheet. Consequently, during RIGGS, experiments to measure these properties in situ were conducted, in some cases yielding unexpected results at variance with laboratory measurements.

Seismic compressional-wave (P-wave) speeds can be calculated as functions of depth within the ice shelf from seismic short refraction data. Because P-wave velocities increase with increasing density and decrease with increasing temperature, the temperature and density profile of the ice shelf serves to create a velocity maximum at about 100 m depth. There is little variation in the maximum velocity across the ice shelf (about 3800 m/sec; Robertson, 1975), but this velocity is significantly lower than that found in inland ice sheets (about 3850 m/sec; Kohnen, 1974); the latter, in turn, is closely in agreement with both laboratory measurements and theoretical expectations. P-wave speeds determined by ultrasonic measurements on ice cores to a depth of 155 m in the RISP drill hole are even higher (3880 m/sec; Kohnen and Bentley, 1977). The mystery of this unexplained discrepancy was heightened by measurements made during the 1978-79 field season. A sonic logger, which measured the average P-wave speed with an accuracy of about 0.1% over a vertical distance of a few metres, revealed a maximum of 3800 m/sec in a borehole at station J9, completely consistent with the refraction shooting and at variance with the ultrasonic measurements on ice cores from the same location. Another interesting feature of the sonic logging profile (Fig. 3) is a marked second-order discontinuity in P-wave speed at a depth of 100 m, approximately (but perhaps coincidentally) the depth at which the source of the snow that forms the ice changes from accumulation on the ice shelf to accumulation on the West Antarctic ice sheet. We have as yet no physical explanation for these peculiarities in wave speed.

P-wave velocities can be converted into densities in the ice shelf down to a depth of about 100 m with an accuracy of a few percent. Lateral variations were found to occur over a distance as short as 2 km near the RISP drill site (Kirchner *et al.*, 1979). We

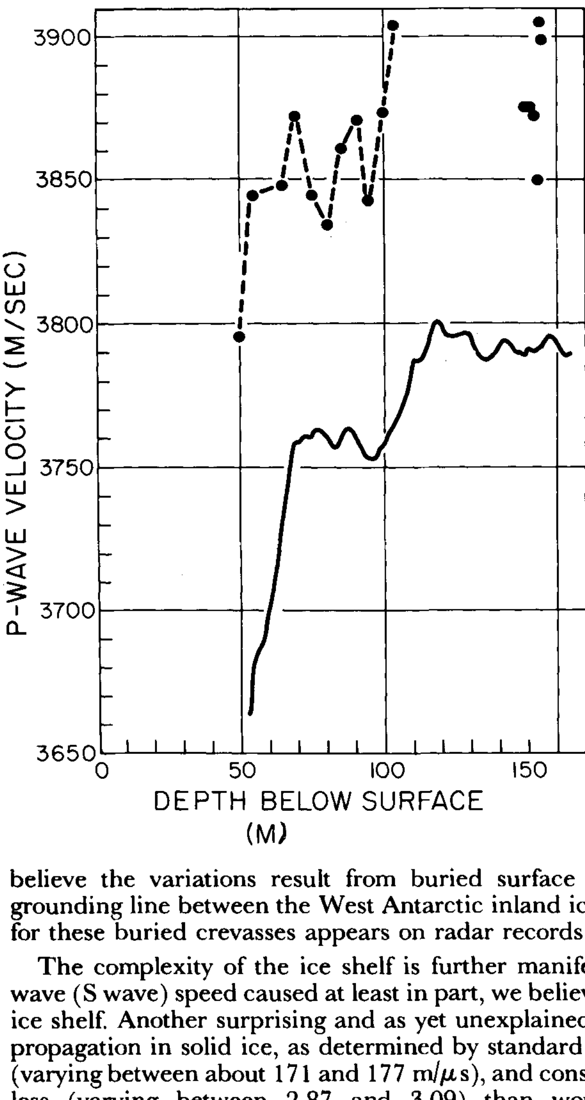

Fig. 3—Sonic log (solid line) made in the 168 m hole drilled at J9 in 1978. Solid circles show velocities measured by ultrasonic techniques on core samples taken from the same hole (Kohnen and Bentley, 1978).

believe the variations result from buried surface crevasses which formed near the grounding line between the West Antarctic inland ice and the ice shelf. Direct evidence for these buried crevasses appears on radar records (Jezek, 1980).

The complexity of the ice shelf is further manifested in variations of seismic shear wave (S wave) speed caused at least in part, we believe, by anisotropic stresses within the ice shelf. Another surprising and as yet unexplained result is that speeds of radio wave propagation in solid ice, as determined by standard geophysical techniques, are higher (varying between about 171 and 177 m/μs), and consequently the electrical permitivity is less (varying between 2.87 and 3.09) than would be expected from laboratory measurements (Jiracek and Bentley, 1971; Jezek *et al.,* 1978). Furthermore, there are significant variations from station to station that appear to form a coherent pattern.

Hochstein (1967), Bentley (1977; 1979), and Shabtaie and Bentley (1979) have used electrical resistivity measurements to examine the thermal regime of the ice shelf. Values for the activation energy, which determines the temperature dependence of the resistivity, and the basal mass balance, which strongly influences the temperature profile in the ice column, were chosen to model the data. Indicated activation energies ranged from 0.25 to 0.15 eV (Bentley, 1979); after allowance for a decrease in ionic impurity concentration in the lower part of the ice shelf (Herron and Langway, 1979), the suggested best value was 0.25 eV, in agreement with laboratory studies. Except for one station where the temperature distribution may not have been in equilibrium, the calculations indicated a near-zero basal mass balance. The only direct measurement of bottom mass balance is at the RISP drill hole; Zotikov *et al.* (3980) have reported the existence of 6 m of saline ice accreted to the base of the shelf. If the saline ice has been accumulating since the ice sheet became afloat, then a basal accumulation rate of about 20 mm/yr is indicated, not measurably different from zero by the electrical technique.

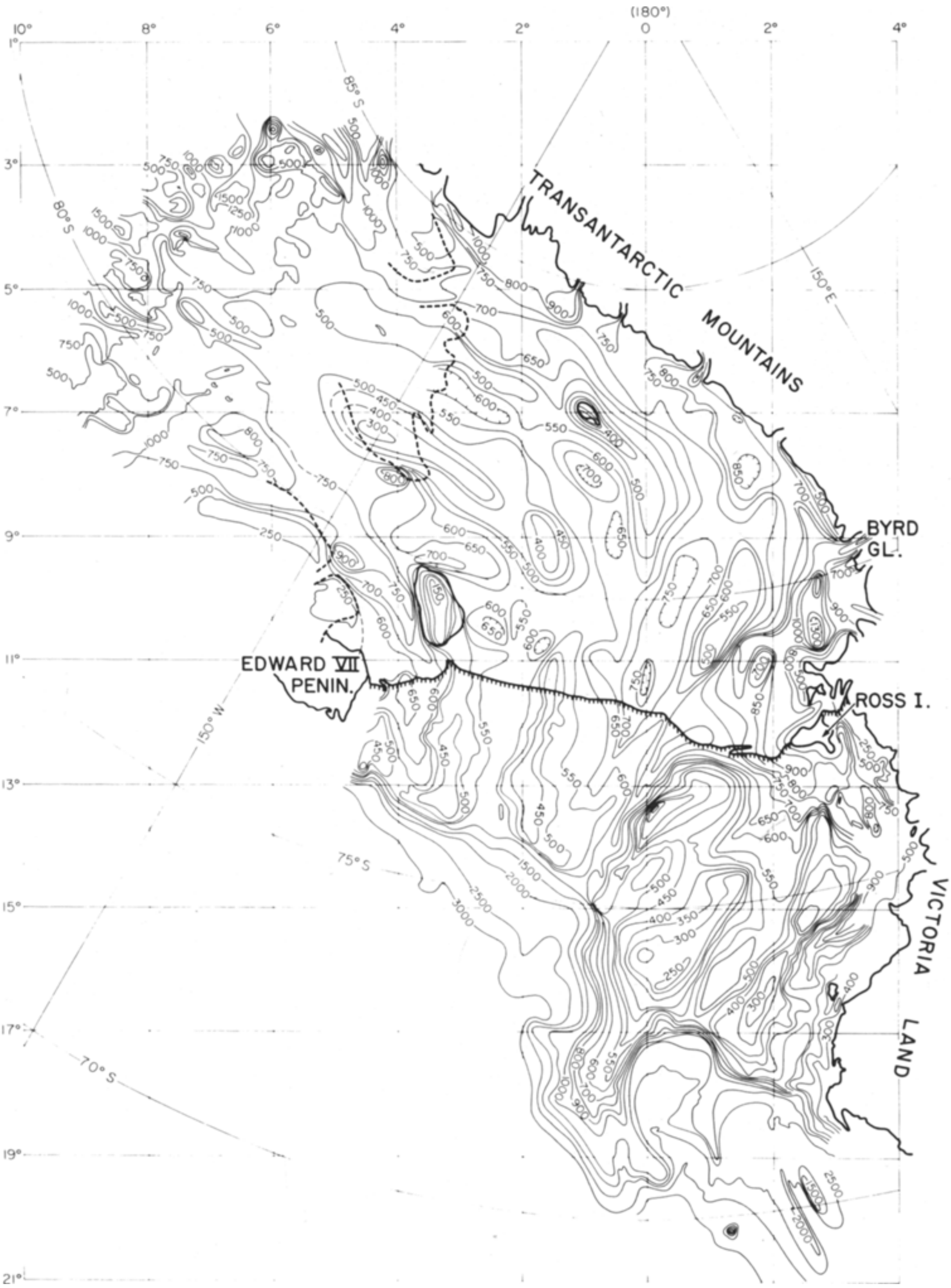

Fig. 4—Submarine topography of the Ross Embayment. The contour interval changes from 50 m in the Ross Sea and under the Ross Ice Shelf to 250 m beneath the West Antarctic inland ice (Albert *et al.,* 1978).

THE ROSS EMBAYMENT

The RIGGS survey, together with airborne radar sounding in West Antarctica (Rose, 1979) and shipboard echo sounding in the Ross Sea (Hayes and Davey, 1975), has made it possible to draw a subglacial and submarine topographic map of the entire "Ross

Embayment" (Albert *et al.*, 1978; Robertson *et al.*, in press). Beneath the ice shelf itself, the sea bottom topography (Fig. 4) is characterized by a series of troughs and ridges crossing the ocean floor nearly parallel to the Transantarctic Mountains and extending for about 250 km under the grounded ice sheet before terminating in the more rugged and deeper Byrd Subglacial Basin. The unbroken continuity of the subglacial and submarine topography across the West Antarctic ice sheet grounding line shows that the position of the grounding line is largely determined by ice sheet dynamics and the heights of sea level, and is, therefore, easily subject to change in time. In fact, we believe that its position has varied greatly in the past, and that it may in fact be moving now.

Gravity anomalies (Robertson *et al.*, in press; L. L. Greischar, pers. comm. 1981) on the ice shelf are closely associated with the ridge-and-trough submarine topography. The anomalies are too large to be explained by a process of glacial scouring and partial filling of the trough with glacial marine sediments; furthermore, they are not correlated with sediment thickness where the latter has been measured. This suggests that tectonic structure, and not glaciation, is the fundamental determinant of the sea bottom topography beneath the grid western portion of the shelf, and thus the positions of ice streams are tectonically controlled.

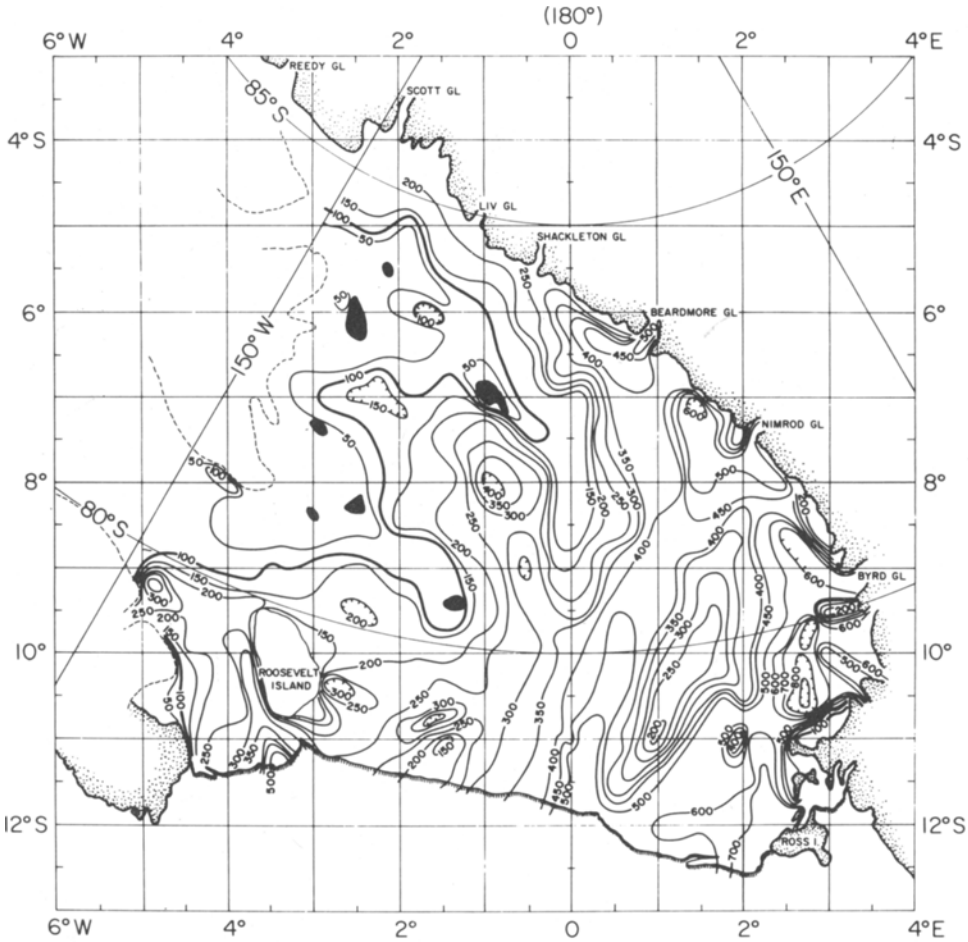

Fig. 5—Water layer thickness contours and the locations of ice rises (solid black areas) from RIGGS data. The contour interval is 50 m.

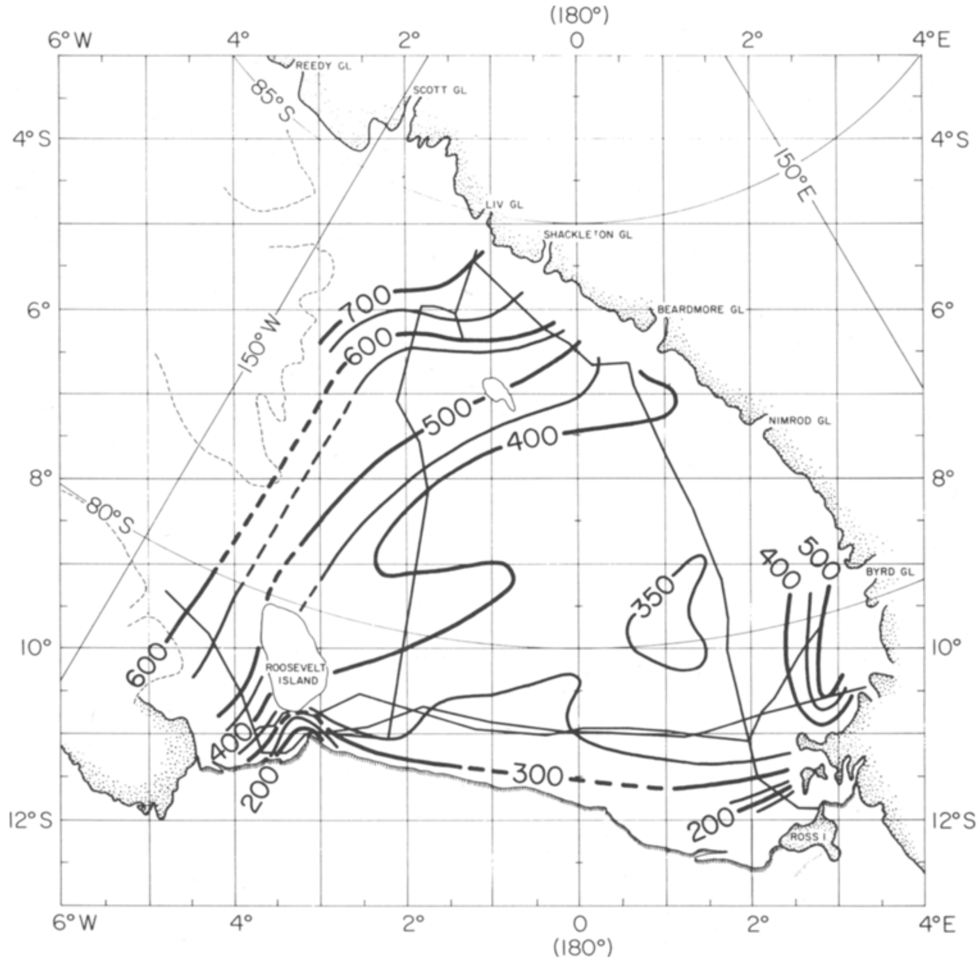

Fig. 6—Ice thickness map (in metres) produced from seismic reflection data collected during IGY (Crary *et al.,* 1962). The traverse route is represented by the thin line. The contour interval is 50 m.

The submarine topography beneath the southeastern part of the Ross Embayment, both beneath the ice shelf and in the open sea, exhibits a contrasting pattern of deeps and shoals. The topographic variations beneath the ice shelf, exceeding 800 m in height, show very subdued isostatic anomalies, suggesting that the topography here, in contrast with that beneath the more westerly part of the ice shelf, does not reflect a deeper structural control. It is likely that the features have been eroded by ice associated with an active Byrd Glacier at an earlier time when the ice in this region was grounded.

Because depths to ocean bottom were measured on a 55 km grid, small, isolated rises in the ocean floor easily could be missed. After combining evidence from the distribution of bottom crevasses found by analysis of radar data, from ice and water layer thicknesses, and from surface crevassing, Jezek (1980) concluded that there are six additional ice rises (besides Roosevelt Island and Crary Ice Rise) on the ice shelf — all in the grid western sector and generally associated with areas of shallow water (Fig. 5). It is widely believed that ice rises may play a central role in stabilizing the ice sheet by acting as pinning points in the ice shelf.

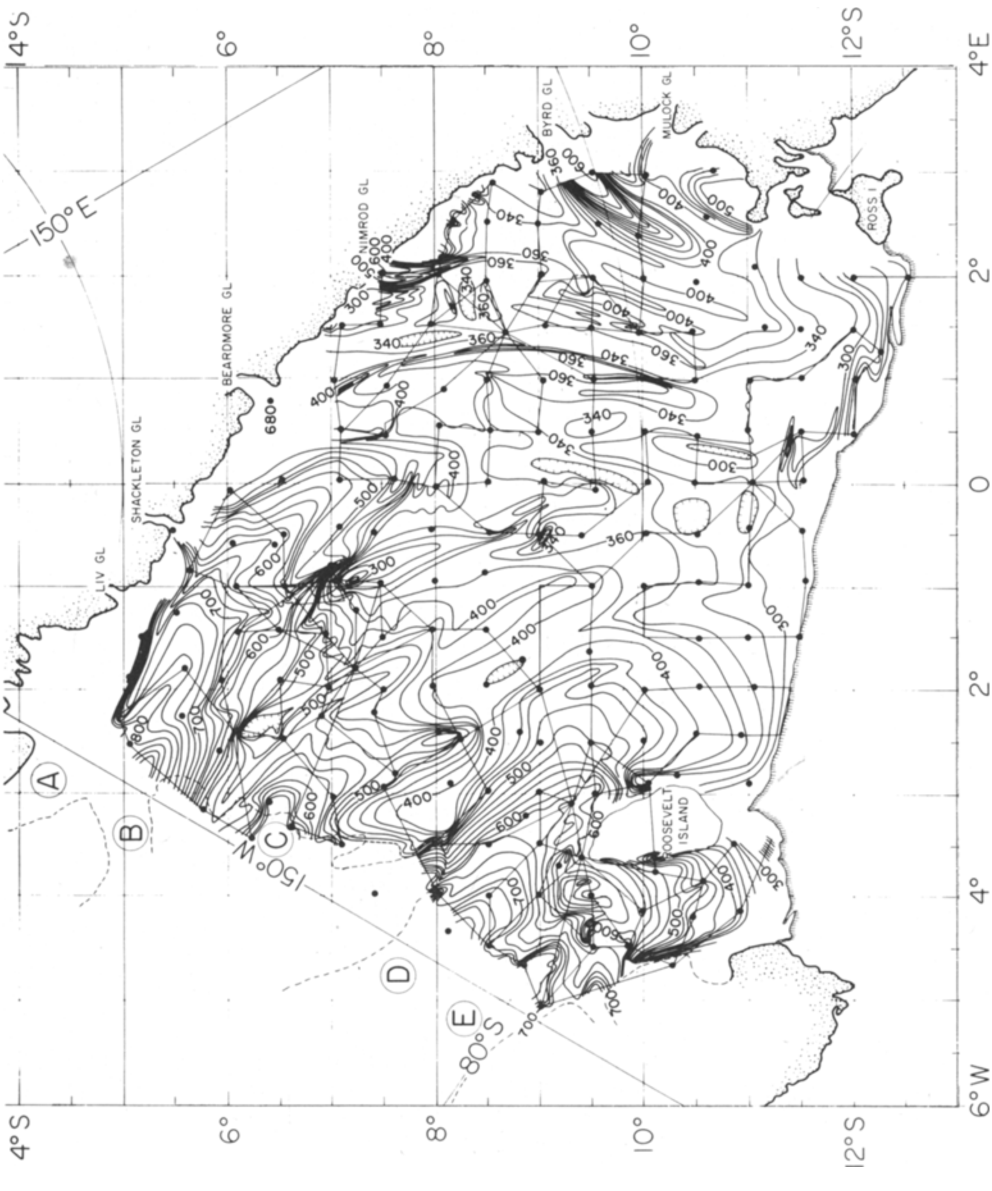

Fig. 7—Ice thickness map produced from radar sounding data collected during RIGGS. Flight-lines and surface stations (solid circles) are shown (Bentley *et al.*, 1979). The contour interval is 20 m.

THE ROSS ICE SHELF — ICE THICKNESS

The measurement and interpretation of the thickness of the Ross Ice Shelf have undergone dramatic changes since the IGY. The thickness map compiled by Crary *et al.* (1962) shows the shelf gradually thinning from about 700 m thick in the grid northwest corner of the shelf to about 200 m thick near the barrier (Fig. 6). This map can be contrasted with the map compiled using the RIGGS data (Bentley *et al.,* 1979) (Fig. 7). The general trends of Crary's map are preserved, but superimposed on these trends is an unexpected level of complexity.

Most of the shelf is characterized by broad intrusions of thick ice that reflect the tremendous volume of ice entering the shelf from the ice streams and outlet glaciers. Particularly striking are the lobes associated with the Ice Streams B and D and with Byrd Glacier. Crary Ice Rise, centrally located on the ice shelf, forms an obstruction to ice flow that causes severe changes in ice thickness. The grid eastern flank of the ice rise is wrapped by a thick finger of ice which may result from the damming effect the ice rise has on upstream ice, but which also may be interpreted as evidence for a more grid easterly position of the grounding line on this flank of the ice rise some time in the recent past. The damming effect is also manifest in the pronounced minimum in ice thickness in the lee of the ice rise, i.e. directly downstream. Contours upstream of Roosevelt Island are convoluted and contain anomalous closed minima suggesting considerable turbulence as Ice Streams D and E are deflected around the island. There are several other relative maxima and minima. The entire grid eastern part of the shelf is characterized by convoluted ice thickness patterns. Finally, an unusual ridge-trough structure — perhaps resulting from basal melting and freezing — trails off downstream of Beardmore Glacier.

We believe that some of the complexities arise from disequilibrium conditions in the ice shelf. This impression is supported by the presence of remarkable fault-like steps in ice thickness that occur at around grid position 9° S, 0° W (Fig. 8). There is as much as 80 m of displacement on the steps and their south central location on the ice shelf makes their existence all the more remarkable since this area would be the least likely to be experiencing unusual stresses.

ICE FLOW AND MASS BALANCE

Repeated position location by navigational satellite, carried out by U.S. Geological Survey personnel and compiled by Thomas *et al.* (in press) yields a straightforward map of the present-day velocity vectors on the ice shelf (Fig. 9). The agreement with the RISS data is excellent, and the association with ice-stream and outlet-glacier input is well defined, except for Ice Stream C, where the coverage is sparse and the evidence

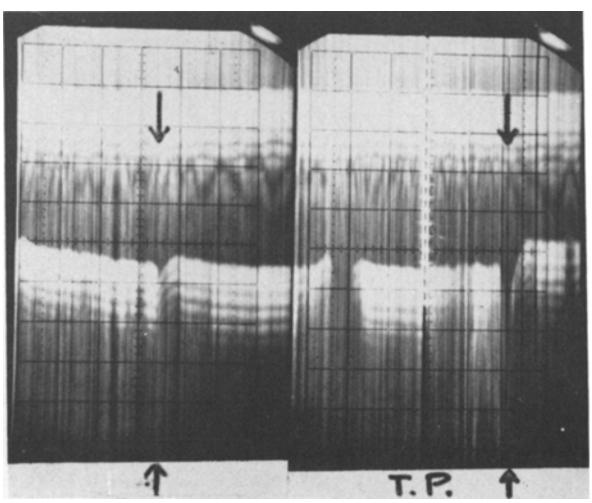

Fig. 8—Airborne radar profile from RIGGS, showing a remarkable step-like change (indicated by arrows) in ice thickness occurring at about 9° S,0° W. The reflection from the bottom of the ice shelf is shown by the white band just below the center of the figure. The grid-line interval represents 1 μs of travel time vertically, and about 4 km horizontally (Bentley *et al.,* 1979).

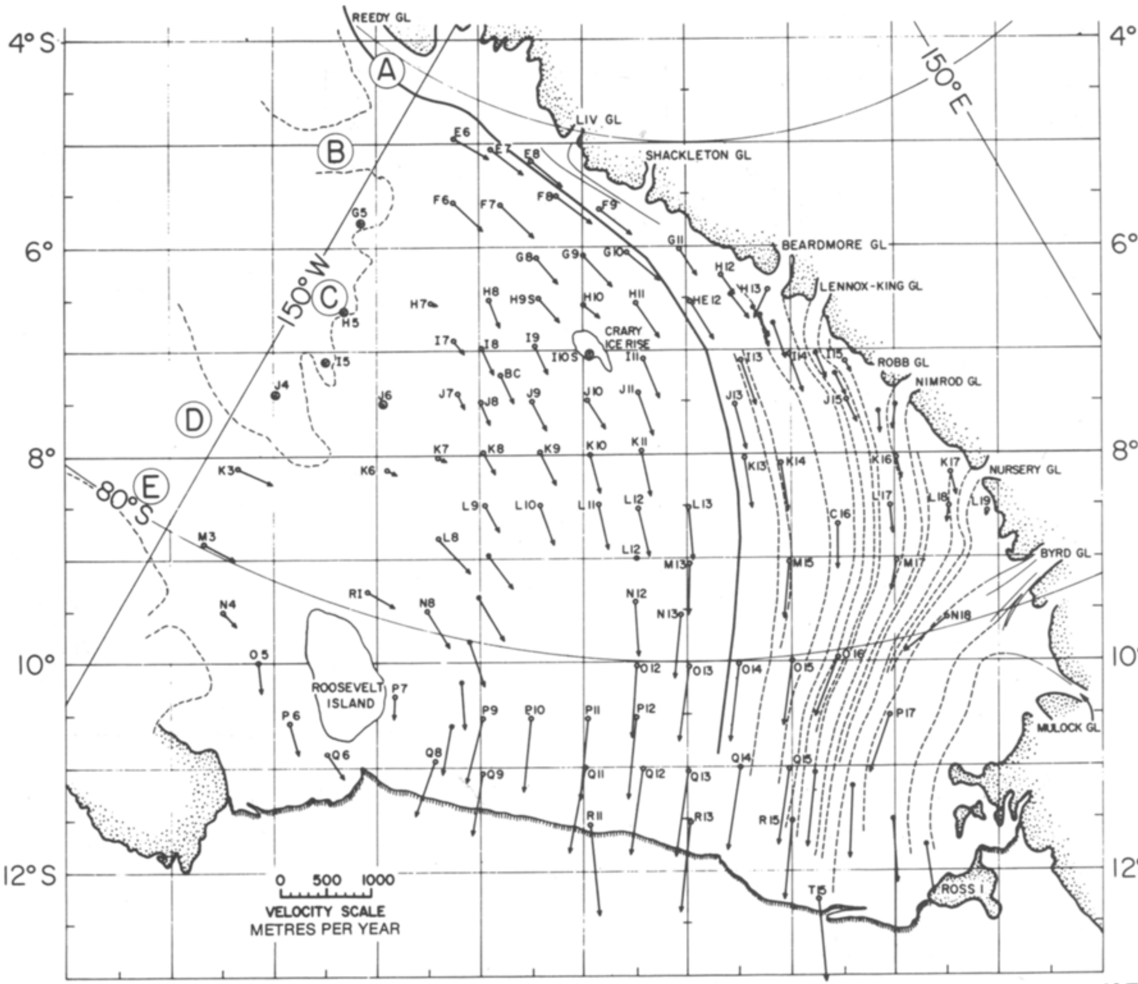

Fig. 9—Velocity vectors determined from Doppler satellite tracking data (arrows) and flow lines based on variations in radar echo strength (dashed lines). The heavy line in the central part of the map represents the boundary between East and West Antarctic ice.

inconclusive. Some glaciologists believe that Ice Stream C is now inactive but that it flowed much more actively in the past.

Flow lines have been mapped in the grid eastern sector of the ice shelf by Bentley *et al.* (1979), using the RIGGS airborne data (Fig. 9) and by Neal (1979) who worked with the airborne radar data from the U.S.-U.K. programme. They have drawn flow lines by correlating changes in the amplitude of bottom reflections between flight lines — for instance, the ice emerging from East Antarctic glaciers generally produces a strong reflection, whereas the reflections off the ice between outlet glaciers is poor. Neal suggests that these reduced signal strengths are associated with brine percolated into the firn from upwardly penetrating bottom crevasses; however, there is also evidence to suggest that there is absorption and scattering taking place in the deeper ice as well. The pattern of bright and dark bands is easily correlated between flight lines and can be used to infer the direction of ice flow since there is good correspondence between the radar reflection bands and velocity vector data in the eastern sector of the ice shelf.

Mapping flow lines on the basis of correlatable radar reflectors is not as straightforward in the western sector. Although there are some features which can be traced down the ice shelf (Neal, 1979; Jezek, 1980). Jezek has shown that they do not correspond to modern

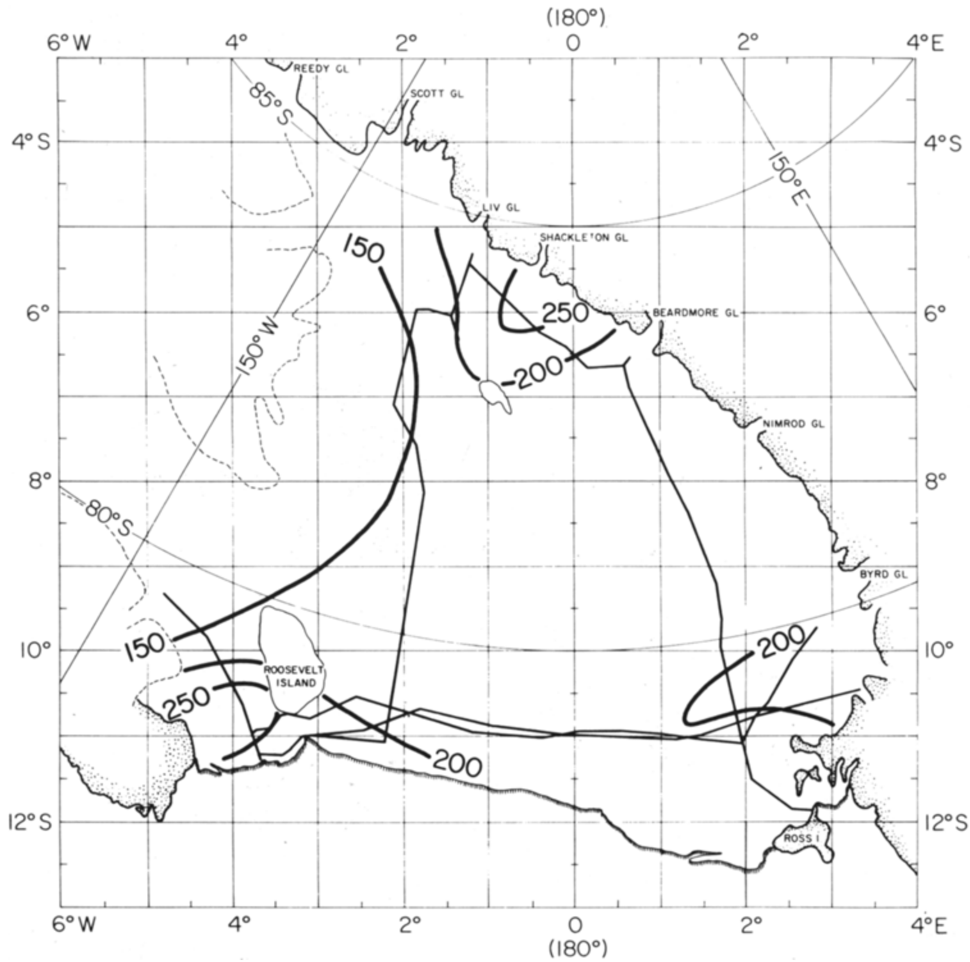

Fig. 10—Contours of surface mass balance rate (heavier lines) from IGY traverse pit studies (in mm/yr of water equivalent) (Crary *et al.,* 1962). Thinner lines denote traverse routes. The contour interval is 50 mm/yr.

flow lines determined from measured velocities. Instead, there is indication that these tracks are records of ice flow changes, a point that is discussed further below.

The stability of an ice sheet may depend on the state of any boundary ice shelves (Thomas, 1973). As a result, a knowledge of the present condition of the Ross Ice Shelf may indicate the state of the West Antarctic ice sheet. Using measured velocities, strain rates and ice thicknesses, Thomas and Bentley (1978a) calculated the mass balance in three flow bands located in the grid western half of the ice shelf. A positive sum of ice shelf thickening and bottom melting (it is not possible from the available data to distinguish between the two) of 0.15 ± 0.06 m of ice per year was found near Roosevelt Island — an area which may experience significant basal melting. Near the RISP drill site, the figure was approximately 0, and upstream from Crary Ice Rise, 0.34 ± 0.15 m of ice per year. Since bottom balance rates on the latter two flow bands are probably small, these data suggest that the ice shelf is near equilibrium near the drill site but is thickening upstream of Crary Ice Rise. Ice shelf thickening upstream of Crary Ice Rise is supported by calculations on the temperature profile at J9 (MacAyeal and Thomas, 1979). However, the mass balance calculations upstream of Crary Ice Rise must be reconsidered because one of the newly discovered ice rises falls within the flow band. Jezek (1980)

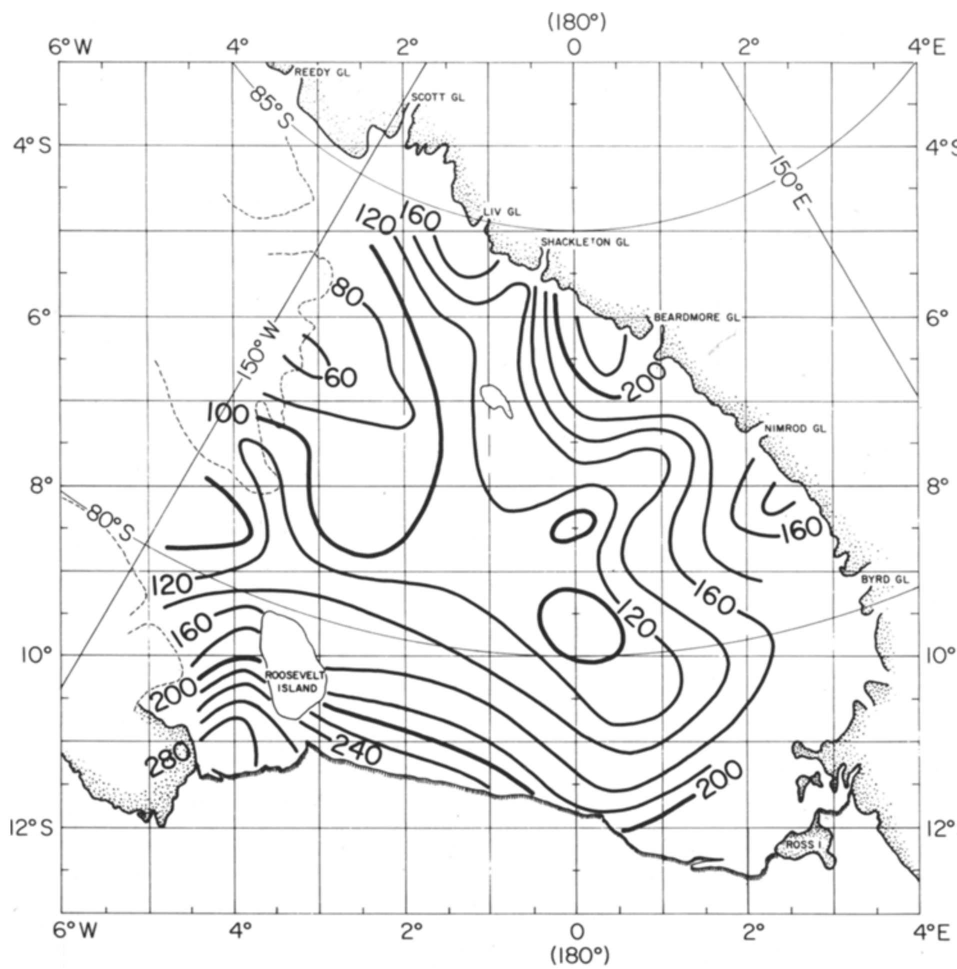

Fig. 11—Surface mass balance rates (in mm/yr of water equivalent) from RIGGS, based on depths to dated radioactive fallout horizons (Clausen *et al.*, 1979) and measured accumulation at stakes (Thomas *et al.*, in press). The contour interval is 20 mm/yr.

reports that two new calculations in this area, one taking the ice rise into account and the other avoiding it, indicate a mass balance near zero.

Data collected during RIGGS have permitted separate balance estimates corresponding to ice input from East and West Antarctica and have substantially reduced the errors in mass balance below those in previous estimates. Using the technique of measuring the depth to dated radioactive fallout horizons, Clausen *et al.* (1979) have determined, with substantially improved accuracy, the surface mass balance over most of the ice shelf. Again, the improvement over what was known at the end of IGY can be seen by comparing Figure 11 with Figure 10, the equivalent map as presented by Crary *et al.* (1962) based on pit studies. These data, together with the flow line map and new knowledge about the thickness of Byrd Glacier (P. Calkin, cited by Hughes, 1975), enable us to calculate 0.13 ± 0.05 m/yr for the sum of ice shelf thickening and bottom melting in the grid eastern sector of the shelf, fed from East Antarctica. Since there is no direct measure of solid ice flow into the grid western sector of the ice shelf, the calculation is less accurate, because it must include an estimate of surface accumulation rates throughout the West Antarctic catchment area. But even assuming the low value of 100 kg m^{-2} yr^{-1} throughout the entire central region in which there are no measurements, the indicated net balance is still positive (0.20 ± 0.13 m/yr) although not significant at the 95% confidence level. It should be noted that a positive mass balance in the grid eastern sector would mean directly that the ice shelf is thickening, whereas in the grid western part a positive balance would imply an average thickening of the ice in the entire drainage system, but not necessarily a thickening of the Ross Ice Shelf itself. Since the studies previously mentioned imply indeed that the ice shelf is not thickening, the implication of a build-up of the West Antarctic inland ice is strong, and the suggestion of a recovery phase following a former surge of the inland ice is unavoidable.

TEMPORAL VARIATIONS OF THE ICE SHELF

Thomas and Bentley (1978b) have modelled the retreat of the West Antarctic ice sheet after the Holocene maximum. Incorporating the effects of sea level rise, which tends to force the grounding line to retreat, and shear stress along the margins of the ice shelf, which tend to inhibit retreat, the modelling showed that retreat of the ice sheet from the edge of the continental shelf began about 15,000 years ago, and that the present-day Ross Ice Shelf was largely established by about 7,000 yr B.P. An important result of these calculations was the fact that buttressing of the inland ice sheet by the bounded ice shelf probably prevented the complete collapse of the West Antarctic ice sheet. Interpretation of the isostatic gravity anomalies in the Ross Embayment (Greischar and Bentley, 1980) supports the history of ice sheet retreat deduced from the modelling studies. Long-wavelength free-air gravity anomalies show a clear trend of increasingly negative values from about −15 mgal near Ross Island to −30 mgal near the Siple Coast that is consistent with incomplete rebound from a former crustal depression resulting from glacial loading, and also a relative maximum over the part of the central ice shelf, indicating earlier crustal unloading, where, according to the model results, the grounding line first retreated. Greischar and Bentley (1980) calculate that, in the absence of changes in the ice sheet from other causes, isostatic uplift will cause a readvance of the grounding line at rates of 25-200 m/yr, eventually reaching a position running roughly from Roosevelt Island to Beardmore Glacier.

On a shorter time scale, there is evidence suggesting that the state of the ice shelf varies considerably over periods of several hundred years. Bentley (in press) has observed both transverse and longitudinal variations (i.e. variations along a flow line) in basal reflectivity and signal strengths from internal layers as recorded on radar sounding of a section of the ice shelf comprising ice flowing from a group of valley glaciers between Beardmore and Nimrod Glaciers. Correlation between individual glacier bands was good enough to yield a striking pattern of large variations in reflection characteristics with time for some 1500 years. Although he proposes no specific causal mechanism for these variations, Bentley believes they are related to fluctuations in the grounding line near the valley glaciers induced by climatic changes. This idea is supported by a rather close correlation between the glacial variations with time and oxygen isotope ratios in ice cores from Byrd Station (Johnsen *et al.*, 1972) and Dome C (Lorius *et al.*, 1979).

Further evidence of recent changes in the ice shelf comes from an analysis by Jezek (1980) of reflectors in the ice shelf downstream of Crary Ice Rise. Although the nature of the reflectors has not been determined, they can be traced downstream across the ice shelf. These tracks were found to cut across the flow lines drawn on the basis of present-day velocities (Fig. 12). Jezek (1980) interprets this lateral migration of the reflector tracks relative to the flow lines as evidence for a shift in the origin point of the reflector, which he believes to be the grid eastern margin of Crary Ice Rise, in the sense of an areally shrinking ice rise, with the most rapid change occurring about 400 years ago. That the fluctuations in the ice shelf are regionally controlled is suggested by the additional observation of a discrepancy between a reflector track near the Siple Coast and the associated velocity vectors.

Evidence of recent climatic change in the Ross Ice Shelf comes from an analysis of temperatures by Thomas (1976). A 1° increase in the mean annual temperature (i.e., the temperature at a depth of 10 m in the ice shelf) has been observed to occur between 1958 and 1974 at stations on the ice shelf as well as at McMurdo and Byrd Stations.

SUMMARY AND PROSPECT

Knowledge of the Ross Ice Shelf has expanded dramatically in the last 20 years. In particular, studies have shown that the ice shelf is more susceptible to rapid fluctuations than was previously suspected. Nevertheless, critical questions still remain. Can the ice shelf be expected to expand or shrink in the future? How vulnerable is it to climatic change? How does the presence or absence of the ice shelf affect (or reflect) the dynamics of the West Antarctic inland ice? We hope these questions will be a prime focus of the U.S. glaciological programme in the years to come.

Fig. 12—Debris tracks (heavy lines) and a flow line (dashed line) based on the velocity vectors downstream of Crary Ice Rise.

ACKNOWLEDGEMENTS

This work was supported by National Science Foundation grant number DPP79-20736.

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University of Wisconsin-Madison, Geophysical and Polar Research Center, Contribution #388.

Received, 9 April 1981