Widespread Complex Flow in the Interior of the Antarctic Ice Sheet

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It has been suggested that as much as 90% of the discharge from the Antarctic Ice Sheet is drained through a small number of fast-moving ice streams and outlet glaciers fed by relatively stable and inactive catchment areas. Here, evidence obtained from balance velocity estimates suggests that each major drainage basin is fed by complex systems of tributaries that penetrate up to 1000 kilometers from the grounding line into the interior of the ice sheet. This finding has important consequences for the modeled or estimated dynamic response time of past and present ice sheets to climate forcing.

The generally accepted view of discharge from an ice sheet is that it is controlled largely by the dynamics of fast-flowing outlet glaciers and ice streams, here jointly defined as fast-flow features, and that these are separated from their drainage basins by onset areas. As a consequence of their assumed importance, field observations have focused on these fast-flow features (*1–3*) as have numerical-modeling investigations (*4–6*). The implicit assumption has been that each basin contains a fast-flow feature that draws ice down from the inland ice sheet—a region of predominantly cold-based ice. There, the ice flow is controlled by internal deformation, which we define here as inland flow, but has also been termed sheet flow (*7*). Thus, based on this view, the basin acts simply as a passive source for the fast-flow feature. Recent evidence from synthetic aperture radar (SAR) interferometry suggests that this view of ice flow is not appropriate for the catchment area feeding the Siple Coast ice streams in West Antarctica (*8*). There have, however, been no large-scale observations of the velocity field for most of the interior of the 13 million km^2 Antarctic Ice Sheet that can test this view of ice-sheet dynamics. In particular, the East Antarctic Ice Sheet, which encompasses about 90% of the ice in Antarctica, remains extremely poorly sampled.

Here, we have a series of recently published data sets, derived from satellite remote sensing and in situ measurements, to calculate a quantity known as the balance velocity (U_b). This quantity represents the depth-averaged velocity required at any point to maintain the ice sheet in a state of balance, given a specified distribution of net surface mass flux. We used a two-dimensional finite-difference scheme (*9*) to estimate the balance velocities for the grounded portions of the Antarctic Ice Sheet. Ice shelves were excluded from the analysis because the assumption that flow is parallel to the local surface slope is inappropriate for floating ice. Three input data sets are required to perform this calculation: surface slope, ice thickness, and mean net surface mass balance. A digital elevation model (DEM), containing information on ice surface slope, was derived from ERS-1 satellite radar altimetry and terrestrial data (*10*). The DEM was smoothed to remove short-wavelength undulations over a spatially variable distance of 20 times the ice thickness (*11*, *12*). The ice thickness grid was obtained from a recently undertaken reanalysis of airborne radar data that had been collected during the 1970s (*13*). The grid of mean net surface mass balance was obtained from a recent compilation that combined in situ and passive microwave satellite measurements (*14*).

In broad terms, the surface slope controls the spatial pattern of the balance velocities, which are scaled by the surface mass balance and ice thickness. Uncertainty in the surface mass balance is estimated to be 10% (*14*). An average uncertainty of 20% is estimated for the ice thickness grid, although there are substantial regional variations because of the variable density of coverage of airborne radar data (*13*) and, consequently, in some areas the error is greater. The ERS-1 satellite-derived surface elevation data only extends as far south as 81.5°S. Beyond this latitudinal limit, the quality and coverage of elevation data is highly variable. On the basis of these uncertainties, we estimate that the error in U_b north of 81.5°S is about 25% because of errors in the input data sets. The U_b value will only match the true depth-averaged velocity if the ice sheet is in balance. Although regional variations in balance have been observed, they appear to be concentrated near the ice sheet margins (*15*).

The balance velocities were compared with recently obtained surface velocity data derived from RADARSAT imagery (*8*) and Landsat feature tracking (*16*) that cover an extensive region of the Siple Coast, both north and south of 81.5°S (Fig. 1). There is

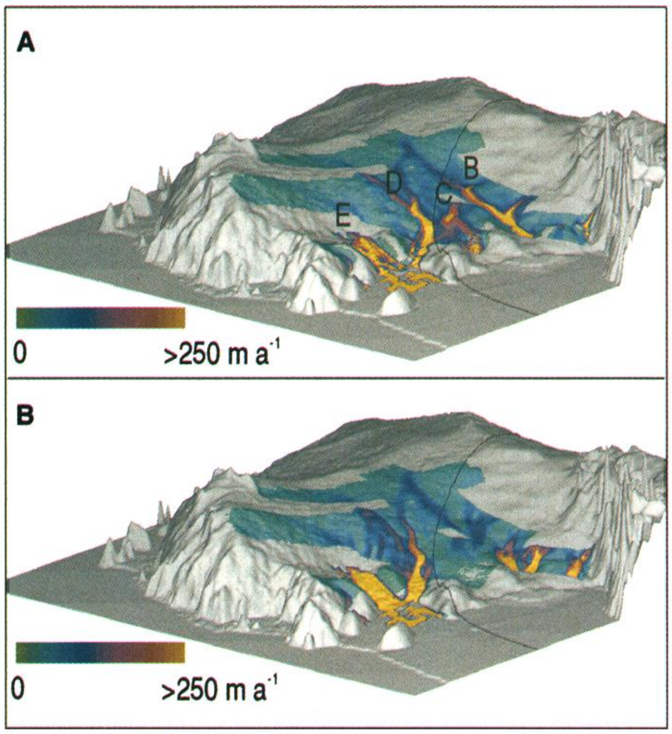

Fig. 1. (**A**) Balance velocities for part of the Siple Coast covering parts of Ice Streams B, C, D, and E. The area to the left of the black line marks the limit of the ERS-1 satellite-derived elevation data. (**B**) Surface velocities derived from feature tracking (*16*) and RADARSAT SAR images with the use of a combination of interferometric and "speckle tracking" techniques (*8*). m a^{-1}, m year^{-1}.

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good agreement in the pattern of the velocity fields north of 81.5°S, but the agreement deteriorates beyond the ERS-1 satellite coverage. In particular, there is an area of relatively high balance velocities (~150 m year^{-1}) just south of 81.5° that is not observed in the RADARSAT data. We believe that this is because of the large imbalance observed in this area (*8*) after the shutdown of Ice Stream C about 140 years ago.

The calculated values of U_b for the grounded ice sheet in Antarctica are shown in Fig. 2. The complexity of flow in the interior of the ice sheet (Fig. 2) has been confirmed qualitatively by several other observations of complex flow in the interior of drainage basins. For example, the pattern of tributaries shown in Fig. 2A in the Pine Island Glacier basin is similar to that obtained from SAR interferometry (*17*), and the pattern of five tributaries feeding the Evans Ice Stream has been observed in Landsat and ERS-1 SAR images (*18*). We believe, therefore, that the balance velocities presented in Fig. 2 provide a good description of the spatial pattern of horizontal ice flow in the interior of Antarctica north of 81.5°S. The spatial coverage of terrestrial elevation data south of 81.5° is shown in Fig. 3. Although much of West Antarctica and south of the Transantarctic Mountains has reasonable coverage, a substantial region of East Antarctica is relatively poorly sampled. It is, consequently, not possible to draw any conclusions about the pattern of flow in this region.

Complex flow is observed throughout the continent. Byrd Glacier has two substantial tributaries, which bifurcate and extend 800 km inland from the coast, although these features lie south of the satellite surface elevation coverage. Slessor Glacier also demonstrates a complex flow pattern that appears to be linked to the inflow to Bailey Ice Stream. A remarkable finger-like structure of four tributaries can be seen feeding into Institute Ice Stream that flows into the Ronne Ice Shelf. These tributaries have balance velocities in the range of 80 to 100 m year^{-1}, which is substantially slower than typical active ice stream velocities (*1*, *2*) but faster than the surrounding ice that is moving at a rate of about 30 m year^{-1}. They fall into the accepted definition of neither ice stream nor inland flow. Active ice streams typically have velocities in the range of 100 to 2000 m year^{-1} (*1*, *2*), and high strain rates are required to develop shear margins that allow well-defined ice stream flow.

Our observations, however, show many tributaries of slower but nevertheless distinctly channeled flow in the interior of drainage basins. These tributaries do not possess clear onset regions and do not have well-developed shear margins for much of their length. It is not for several hundred kilometers downstream that a substantial acceleration in ice motion takes place and velocities reach those associated with ice stream flow. Figure 4 shows balance velocity profiles down two tributaries that feed fast-flow features on the ice sheet. The balance velocities increase steadily down each feature within 100 to 200 km from the grounding line. There is then a rapid increase in U_b that reflects a change in flow regime, where the driving stress decreases with increasing flux. Profiles running across the fast-flow features indicate that the flow rates within a feature are about 5 to 10 times greater than the surrounding ice at these locations (Fig. 4). Small perturbation theory suggests that the dynamic response time of an ice mass is inversely proportional to its velocity (*19*), implying that parts of the interior of Antarctica and probably former ice sheets can respond more rapidly to climate forcing than model simulations might suggest (*20*).

The distinction between the slow-moving interior ice sheet and fast-moving outlet glaciers and ice streams is not as clear as has been previously believed. It appears that a continuous gradation between these two extremes represents an as yet undefined flow regime that is neither ice stream nor inland flow. Ice streams

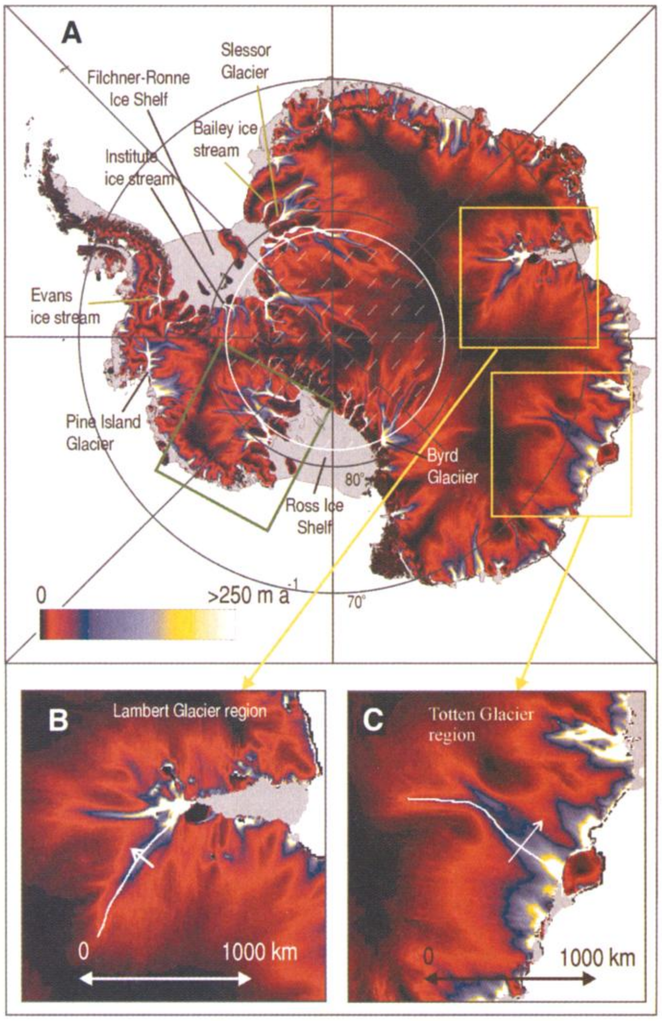

Fig. 2. (**A**) Balance velocities calculated for the grounded part of the Antarctic Ice Sheet. Ice shelves and floating ice tongues are gray. The coastline and rock outcrops are black. The locations of (**B**) and (**C**) are shown by the yellow boxes (upper right and bottom right, respectively). The area covered in Fig. 1 is shown by the green box. The quality and coverage of elevation data is substantially poorer south of 81.5°S. This region is indicated by the hatched area inside the white circle (center). (B) Balance velocities for the Lambert Glacier region showing the two large flow-features that feed the Amery Ice Shelf. The arrow across the flowline indicates the direction of the profile as plotted in Fig. 4. (C) Balance velocities centered on Totten Glacier with Northcliffe, Denman, and Scott Glaciers forming the fast-flow feature at top right. The arrow across the flowline indicates the direction of the profiles as plotted in Fig. 4B.

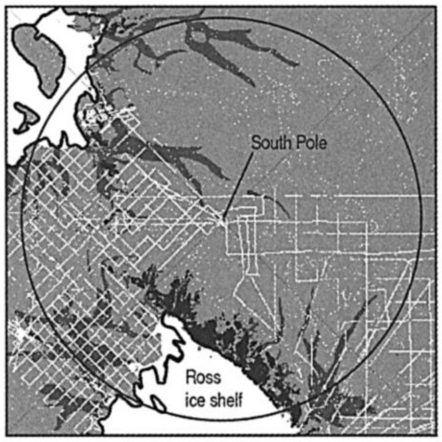

Fig. 3. A plot locating the elevation data used in the DEM for the area south of 81.5°. The data were collected during the late 1960s and early 1970s from airborne and ground traverses over the snow (*22*). The latitudinal limit of ERS-1 radar altimeter coverage is shown by the area outside the solid black circle. Areas of faster flow are shown by dark shading to indicate their location with respect to the elevation data.

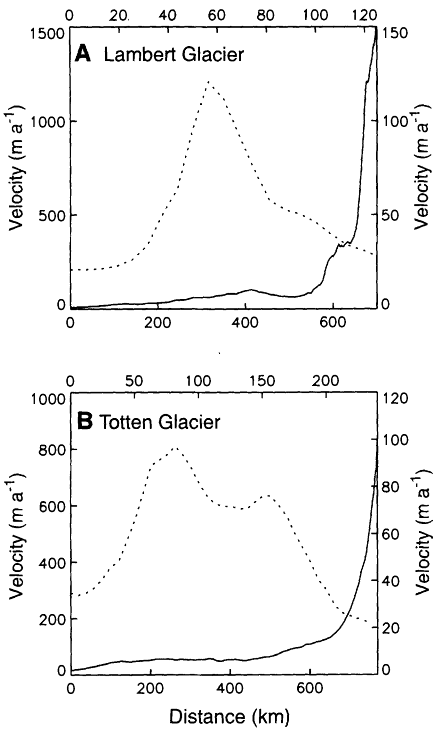

Fig. 4. Balance velocity profiles down two fast-flow features (**A** and **B**) in East Antarctica, the locations of which are shown by the white lines in Fig. 2, B and C. Transects running across each flow feature (dashed lines) are plotted with scales displayed on the right-hand ordinate and the upper abscissa. To avoid including floating ice, the profiles stop short of the estimated grounding line positions (*23*).

and inland flow represent, therefore, end members of a continuum of flow states within an ice sheet. This continuum of flow states is observed extensively in East and West Antarctica. It is also interesting to note that this type of behavior is not seen solely in Antarctica. A recently discovered ice stream in northeast Greenland (*21*) shows characteristics similar to those described above, and tributary-type flow is also present (*12*). It is likely, therefore, that former large ice masses, such as the Laurentide Ice Sheet, also possessed these flow characteristics.

This evidence challenges the view that the Antarctic plateau is a slow-moving and homogeneous region. Numerical models of the ice sheet have yet to reproduce the pattern and the complexity of flow shown in Fig. 2, therefore suggesting a limitation of their ability to simulate the present-day dynamics and future changes in the behavior of the largest ice mass on Earth.

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23. The assumptions about ice flow used to derive the balance velocities break down when the ice begins to float in the vicinity of the grounding line. Velocities close to the grounding line are, therefore, unreliable. Here, a simple test for whether the ice thickness was close to the value predicted by hydrostatic equilibrium was used to identify floating ice in Fig. 2.
24. We thank the European Space Agency for provision of the data used to produce the DEM and J. Dowdeswell for useful discussions on the text.

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Density of Phonon States in Iron at High Pressure

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The lattice dynamics of the hexagonal close-packed (hcp) phase of iron was studied with nuclear inelastic absorption of synchrotron radiation at pressures from 20 to 42 gigapascals. A variety of thermodynamic parameters were derived from the measured density of phonon states for hcp iron, such as Debye temperatures, Grüneisen parameter, mean sound velocities, and the lattice contribution to entropy and specific heat. The results are of geophysical interest, because hcp iron is considered to be a major component of Earth's inner core.

The inner core of Earth is composed almost entirely of Fe or Fe-rich alloys. This geophysical aspect is one of the many motivations to study the phase diagram of Fe with its various isomorphs in a wide pressure and temperature range (*1*). The hcp high-pressure phase of Fe (ε-Fe) is considered the most relevant phase for the inner core (*2*). At ambient pressure and temperature, Fe is magnetic and crystallizes in the body-centered cubic (bcc) structure (α-Fe). The phase transformation of α-Fe to nonmagnetic ε-Fe at around 13 gigapascals (13 GPa) occurs over a broad pressure range depending on the sample environment (*3*). Although a considerable amount of crystallographic data on the pressure and temperature phase diagram of Fe is available (*1*), there is less experimental information available on the lattice dynamics of ε-Fe. The sound velocities in the ε-phase at high pressure and temperature are of particular interest, because seismic wave experiments indicate that the sound velocities in Earth's inner core are anisotropic (*4*).

The lattice dynamics of solids is described by phonons, which are characterized by their dispersion relations and by their density of states (DOS). Dispersion relations are conventionally measured with single-crystal samples using coherent inelastic neutron scattering (*5*). The phonon DOS can be obtained after adjusting the dispersion curves with a Born–von Kármán model (*6*). In special cases, incoherent neutron scattering can be used to determine the phonon DOS of polycrystalline samples (*5*). Alternatively, the density of phonon states of polycrystalline samples can be determined with the recently established technique of resonant nuclear inelastic absorption of synchrotron radiation (SR) (*7, 8*), which employed the 14.413-keV resonance of ^{57}Fe. Measurements of the phonon DOS in α-Fe (*9, 10*), Fe clusters (*11*), thin Fe layers (*12*), and Fe-containing quasicrystals (*13*) demonstrated the high sensitivity of this method. We used nuclear inelastic absorption to study a tiny (0.5 μg) Fe sample pressurized in a diamond anvil cell (DAC) and to measure the phonon DOS of ε-Fe.

The experiments were performed at the nuclear resonance station ID22N at the European Synchrotron Radiation Facility (ESRF),

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