Using inverse methods constrained by recent satellite observations, we have produced a comprehensive estimate of the basal shear stress beneath the Filchner-Ronne ice streams. The inversions indicate that a weak bed (approx. 4–20 kPa) underlies much of these ice streams. Compared to the Ross ice streams, the distribution of weak subglacial till is more heterogeneous, with ‘sticky spots’ providing much of the resistance to flow. A weak bed beneath Recovery ice stream extends several hundred kilometres inland with flow. Along this ice stream, discrepancies between thickness measurements and flux estimates suggest the existence of a deep (>1400 m) trough not resolved by existing maps of subglacial topography. We hypothesize that the presence of this and other deep troughs is a major influence on this sector of the ice sheet that is not fully incorporated in current models of ice-sheet evolution.

Keywords: ice streams; glaciology; Antarctica; ice sheets

1. Introduction

Each year Antarctic ice streams discharge to the ocean a volume of ice that approximately balances the annual accumulation from snowfall over the ice sheet (Paterson 1994). The floating Filchner-Ronne ice shelf, which is comparable in

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area to the state of Texas, is sustained by inflow from eight large ice streams (figure 1), the catchments of which comprise approximately 22% of the ice sheet’s grounded area and include area from both East and West Antarctica. Because only a handful of ice streams regulate discharge from this large area, the dynamics and stability of these ice streams are important for determining Antarctica’s current and future mass balance.

Satellite altimetry data indicate that the Antarctic ice sheet thickened at an average rate of 1.4 cm yr\(^{-1}\) from 1992 to 2003 (Davis et al. 2005), with East Antarctic thickening offsetting West Antarctic thinning. One of the largest imbalances is the rapid thinning along the Amundsen Coast (Rignot & Thomas 2002; Shepherd et al. 2002), which may represent a dynamic response of grounded ice to increased ice shelf melt from a warming ocean (Payne et al. 2004). Conversely, the catchment feeding the Ross ice streams is thickening, which is largely due to ice stream C’s stagnation roughly 150 years ago (Retzlaff & Bentley 1993; Joughin & Tulaczyk 2002).

The catchment supplying the Filchner-Ronne ice shelf is thickening (Joughin & Bamber 2005; Davis et al. 2005). Unlike some other areas with imbalances, this thickening is more evenly distributed over the catchment area, without any strong thinning or thickening on the fast moving regions to indicate recent

Figure 1. Ice stream speed (colour and thin black 100 m yr\(^{-1}\) contours) over shaded DEM surface. Thick black lines show catchment boundaries, and thick grey lines show 1000 m elevation contours. The white box on inset shows the figure location and the black box shows the location of the region shown in figure 2.
changes in ice stream flow. This thickening may be the result of some combination of a twentieth century accumulation increase and a continuing response to the approximately 50% accumulation increase that began in the early Holocene (Joughin & Bamber 2005).

Concerns about marine ice-sheet instability have driven numerous studies of the Ross ice streams, which are relatively accessible because of their proximity to McMurdo Station. After more than two decades of research, we now know that despite low driving stresses, these ice streams flow rapidly over a weak ($O(10)$ kPa) water-saturated till bed (Alley & Bindschadler 2001). The presence of weak till was initially inferred from reflection seismic data from beneath Whillans ice stream (Blankenship et al. 1986) and subsequently confirmed by till samples from boreholes drilled through several of the Ross ice streams (Kamb 1991, 2001). Control-method inversions of surface velocity, elevation and thickness data (MacAyeal et al. 1995; Joughin et al. 2004) indicate this till is homogeneously distributed beneath most of the fast moving ice stream area, with only a few isolated sticky spots (Alley 1993). These inversions also indicate there are both weak- and strong-bedded areas beneath many of the slower moving tributaries.

Seismic data collected on Rutford, Carlson and Evans ice streams indicate areas of weak dilatant till with properties similar to that of the till beneath the Ross ice streams (Smith 1997b; Vaughan et al. 2003). Many of the seismic lines, however, also indicate the presence of stronger regions, suggesting a less uniform till distribution than beneath the Ross ice streams. Little is known about the basal shear stress and the distribution of till beneath the other Filchner-Ronne ice streams, where no seismic data have been collected. The goal of this paper is to use control-method inversions and remote sensing observations to constrain the basal drag beneath these ice streams.

2. Inverse methods

Our results are based on a finite-element ice-stream model that was developed for application to West Antarctic ice streams (MacAyeal 1989; Hulbe & MacAyeal 1999). In brief, this depth-averaged model determines the velocity and stress fields as functions of ice thickness, surface elevation and basal shear stress and various boundary conditions, including the location of seaward ice fronts, ice shelf grounding lines and connections with non-streaming inland ice. The model’s governing stress-balance equations include both basal drag and various deviatoric stresses acting in the horizontal plain as a means to balance the driving stress. These equations are most applicable when the ice flow is dominated by basal sliding so that the vertical shear is small.

Here, we invert the forward model to find the basal shear stress, $\tau_b$, that minimizes the misfit between model-determined and observed velocity fields. This inversion relies on ‘adjoint-trajectory methods’ (MacAyeal 1992, 1993), which are related to optimal control theory. With this method, an ‘adjoint-trajectory’ version of the model is created to develop an automatic and objective means to evaluate the ‘cost function’ (i.e. least-squares misfits between the model and observations) as it varies with changes in the undetermined parameters ($\tau_b$ for the results described here). As with many models, the ice
stream equations are ‘self-adjoint,’ so that little modification of the forward model code is needed to solve the inverse problem. Similar methods have been used successfully to study the basal friction field of the Ross ice streams (MacAyeal et al. 1995; Joughin et al. 2004) and the Northeast Greenland ice stream (Joughin et al. 2001).

A linearly viscous deforming bed model (Alley et al. 1986) was assumed in the original development of the inverse technique. More recent results suggest that till behaves more as a plastic material with a distinct yield stress (Kamb 1991, 2001; Tulaczyk et al. 2000a). Thus, the forward model and inversion have been adapted to accommodate a plastic bed (Joughin et al. 2004). Force balance is achieved regardless of the assumed bed rheology. Thus, the inversion results alone cannot distinguish which is the most appropriate bed model. For the work described here, we use a solution that solves for both horizontal components of the basal shear stress. We then assume that the true basal shear stress is aligned with the velocity vector, and thus discard the cross flow component of resistance, which may have served to improve the model-data misfit by compensating for errors in the data.

The outcome of our inversions depends on the flow-law parameters. In all experiments, we make the common assumption that the flow-law exponent, \( n = 3 \) (Paterson 1994). The flow-law parameter, \( A \), is strongly sensitive to ice temperature. As a result, we used a simple one-dimensional model of heat transfer to estimate temperature in the ice column (Zotikov 1986) from which we determined the temperature-dependent values of \( A \) (Paterson 1994). Our standard set of inversions does not include any flow-law enhancement factor to account for strain softening and/or viscous heating (Budd & Jacka 1989). To examine the sensitivity of the inversions to these effects, we performed an additional set of inversions using a flow-law enhancement factor of 3 as described below.

The inverse techniques we used have been tested extensively with synthetic data (Joughin et al. 2004), which indicate that the inversions are fairly robust with respect to the typical levels of error encountered with surface velocity and bed topography data. The inversions are more sensitive to surface–slope errors so that elevation errors of greater than 5 m can substantially degrade the inversion results. The inversions have difficulty accurately resolving \( \tau_b \) at length-scales of less than a few ice thicknesses, particularly when the fluctuations are relatively weak. The initial ‘guess’ for \( \tau_b \) used in the inversion can also substantially affect the results at these length-scales, with initial estimates based on the structure of the driving stress seeming to provide the closest fit (Joughin et al. 2004). While the inversions can have trouble resolving subtle details in the basal shear stress distribution, experiments with synthetic data demonstrate that the results are reliable when averaged over distances greater than a few ice thicknesses. For this reason, we centre our interpretation on values that are averaged over areas with dimensions of several ice thicknesses.

It is important to note that the model was developed for ice streams with little vertical deformation (MacAyeal 1989). Thus, the model is not strictly applicable to regions where the bed is relatively ‘strong’ (i.e. \( \tau_b > 40 \) kPa) and the basal shear stress is comparable to the driving stress. Analysis of the model by D. R. MacAyeal (1997, unpublished work) indicates that where there is substantial deformation due to vertical shear, the model correctly predicts the locations of sticky spots (Alley 1993), but tends to underestimate the basal shear

stress. Qualitatively, however, the model tends to correctly identify regions where the bed is strong and supports most or all of the driving stress. Thus, our analysis focuses more qualitatively on regions where the bed is strong and more quantitatively on regions where the bed is ‘weak’ (i.e. \( \tau_b < 40 \text{ kPa} \)).

### 3. Data

Our inversions use measurements of ice-flow velocity, surface topography, and ice thickness. In 1997 and 2000, RADARSAT was used to collect extensive interferometric synthetic-aperture radar coverage over Antarctica (Jezek 2002, 2003). We have used these data to estimate the velocity of the ice streams feeding the Filchner-Ronne ice shelf (see figure 1), using both conventional interferometry and speckle tracking (Joughin 2002; Joughin & Padman 2003). Error estimates for the velocities over most of the region are less than 5 m yr\(^{-1}\). In general, the errors are uncorrelated over distances less than an ice thickness, and simulations show the inversions are relatively insensitive to these errors (Joughin et al. 2004).

For surface elevation we used a new, 1-km resolution Antarctic digital elevation model (DEM) (Bamber 2005, unpublished data), which was created using both European remote sensing (ERS) radar altimeter and ICESAT laser altimeter data. The use of ICESAT data greatly improved DEM accuracy for the area south of 81.6\(^\circ\) S, which is the southern extent of the ERS altimeter coverage. While the accuracy of this DEM is generally good (less than 5 m), there are a few areas where ice stream boundaries (e.g. Rutford) lie near sloped areas that degrade the accuracy of the ERS altimeter data and impact our results.

We used the BEDMAP compilation of ice thickness in our inversions (Lythe & Vaughan 2001). These data are posted at 5 km intervals, although actual resolution depends on the density of the source data. The inversions are relatively insensitive to errors in ice thickness of a few hundred metres, with errors becoming a significant problem only when they approach a significant fraction of the ice thickness. Below, we describe regions in the BEDMAP compilation where the errors can be this large.

The MODIS imagery shown in many of our figures was extracted from the Mosaic of Antarctica (MOA; Bohlander et al. 2004). This mosaic is a composite of 260 manually cloud-cleared MODIS Band 1 (red visible light) scenes, which were assembled using an image-stacking method that enhanced resolution and radiometric detail. The mosaic has a grid scale of 125 m and is composed of images acquired between 0530 and 1330 UT so that the mean illumination is always towards the upper right in the subscene, but not limited to a single azimuth. Images also were selected to maintain a moderately low sun elevation across the continent. As a result, the composite image provides a rendition of the snow surface topography with significant enhancement of subtle features (e.g. flows stripes and surface undulations).

### 4. Ice stream inversions

We applied the inversion procedure to the eight largest Filchner-Ronne ice streams. To speed convergence, we divided each of the larger ice streams into several sub-regions over which the data were inverted individually. Application
of the inversion procedure to smaller areas in this manner also yields better results in terms of model-data mismatch (Joughin et al. 2004). After determining $\tau_b$ for each sub-region, we combined the results to produce the $\tau_b$-map shown in figure 2. This figure also shows the locations of the boxes over which we computed average basal shear stresses (table 1). These results are described throughout the remainder of §4.

(a) **Bailey ice stream**

Bailey ice stream (figure 1), which is the smallest of the ice streams we examined, discharges 4.7 Gtons yr$^{-1}$ of ice to the Filchner ice shelf. The basal shear-stress inversions (figure 2) indicate that there is a sticky spot near the grounding line in an area where the surrounding bed is relatively weak. With this sticky spot, the average basal shear stress (table 1) for the box shown in figure 2 is 33 kPa, which is equivalent to about 70% of the driving stress (figure 3).

(b) **Slessor ice stream**

Slessor ice stream discharges 24.8 Gtons year$^{-1}$ across its grounding line from a basin that is thickening at approximately 20% of the accumulation rate
(Joughin & Bamber 2005; Davis et al. 2005). The inversions show that a relatively weak bed extends roughly 85 km inland from the grounding line. The MOA imagery (figure 4) indicates that the surface topography is relatively smooth over this region, which is consistent with a uniformly weak bed. The basal shear stress of 9 kPa averaged over the box shown in figure 2 balances nearly half the driving stress. This suggests that lateral shear stresses may support much of the remaining driving stress (approx. 8 kPa) near the grounding line. The width and thickness of this part of the ice stream are comparable to the upper branches of Whillans ice stream (e.g. UpB), where lateral shear stresses range from about 10 to 13 kPa (Whillans & Van der Veen 1997; Joughin et al. 2004).

Upstream from the weak region, there is a gap where we have no velocity data and, consequently, could not perform an inversion. Features in the MOA image suggest that there is a mixture of strong and weak spots at the bed in this region. Just upstream of this gap, the inversions suggest the bed is weak, but it then abruptly transitions to an area where $\tau_b$ is much larger.

Recovery ice stream

With a catchment covering about 8% of the grounded ice sheet’s area, Recovery ice stream has the second largest flux (35 Gtons yr\(^{-1}\)) in our study area. Unlike several of the other ice streams that have weak beds at their respective grounding lines, the bed is strong for a region that extends about 40 km upstream from the Recovery grounding line. Despite this stronger bed, this ice stream discharges ice across its grounding line much faster (greater than 900 m yr\(^{-1}\)) than any of the other Filchner-Ronne ice streams (less than 650 m yr\(^{-1}\)).

Over the 50 km just upstream of the strong-bedded region, the fast ice flow narrows in the down-flow direction as its centre-line speed increases from about 300 to 600 m yr\(^{-1}\). The bed beneath this area appears to consist of a mixture of weak and strong regions before giving way to a region where the bed is consistently weak for the next 180 km upstream (Box A in figure 2). Over this extended weak area, the ice stream flows at about 200–300 m yr\(^{-1}\) while maintaining a nearly constant width (approx. 50 km). This is one of the weakest regions we sampled (\(\tau_b = 6\) kPa), and it represents by far the largest expanse of uniformly weak bed in our study area. As described below, however, errors in the bed topography likely cause the inversion to underestimate \(\tau_b\) in this region.

A large strong spot terminates the large expanse of weak area, just upstream of the area where a tributary from the south merges with the main body of the ice stream. This strong spot correlates well with surface slopes visible in the MOA image (figure 5), which indicate a region of locally high-driving stress. Further upstream there is a gap in the velocity data where similar features visible in the MOA data indicate that there are likely additional strong spots at the bed. A similar set of strong spots is located nearly 100 km further upstream. In between these stronger regions, the bed is relatively weak (14 kPa) over the region enclosed by Box B.

Table 1. Values averaged over boxes shown in figure 2.

<table>
<thead>
<tr>
<th>ice stream box</th>
<th>area (km(^2))</th>
<th>thickness (m)</th>
<th>(\tau_a) (kPa)</th>
<th>(\tau_b) (kPa)</th>
<th>(E=1)</th>
<th>(E=3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bailey</td>
<td>867</td>
<td>982</td>
<td>47</td>
<td>33</td>
<td>36</td>
<td></td>
</tr>
<tr>
<td>Slessor</td>
<td>1870</td>
<td>521</td>
<td>17</td>
<td>9</td>
<td>10</td>
<td></td>
</tr>
<tr>
<td>Recovery-A</td>
<td>6894</td>
<td>786</td>
<td>20</td>
<td>6</td>
<td>7</td>
<td></td>
</tr>
<tr>
<td>Recovery-B</td>
<td>4474</td>
<td>1926</td>
<td>43</td>
<td>14</td>
<td>17</td>
<td></td>
</tr>
<tr>
<td>Support Force</td>
<td>1706</td>
<td>1528</td>
<td>57</td>
<td>26</td>
<td>34</td>
<td></td>
</tr>
<tr>
<td>Foundation-A</td>
<td>680</td>
<td>2212</td>
<td>45</td>
<td>12</td>
<td>18</td>
<td></td>
</tr>
<tr>
<td>Foundation-B</td>
<td>1792</td>
<td>577</td>
<td>22</td>
<td>10</td>
<td>13</td>
<td></td>
</tr>
<tr>
<td>Foundation-C</td>
<td>3900</td>
<td>2030</td>
<td>43</td>
<td>18</td>
<td>18</td>
<td></td>
</tr>
<tr>
<td>Institute-A</td>
<td>2348</td>
<td>1332</td>
<td>26</td>
<td>11</td>
<td>14</td>
<td></td>
</tr>
<tr>
<td>Institute-B</td>
<td>534</td>
<td>1038</td>
<td>10</td>
<td>4</td>
<td>3</td>
<td></td>
</tr>
<tr>
<td>Rutford</td>
<td>492</td>
<td>1742</td>
<td>45</td>
<td>6</td>
<td>12</td>
<td></td>
</tr>
<tr>
<td>Evans-A</td>
<td>1289</td>
<td>1279</td>
<td>32</td>
<td>8</td>
<td>12</td>
<td></td>
</tr>
<tr>
<td>Evans-B</td>
<td>1014</td>
<td>1282</td>
<td>30</td>
<td>7</td>
<td>12</td>
<td></td>
</tr>
<tr>
<td>Evans-C</td>
<td>855</td>
<td>1698</td>
<td>29</td>
<td>8</td>
<td>11</td>
<td></td>
</tr>
</tbody>
</table>

(c) Recovery ice stream

With a catchment covering about 8% of the grounded ice sheet’s area, Recovery ice stream has the second largest flux (35 Gtons yr\(^{-1}\)) in our study area. Unlike several of the other ice streams that have weak beds at their respective grounding lines, the bed is strong for a region that extends about 40 km upstream from the Recovery grounding line. Despite this stronger bed, this ice stream discharges ice across its grounding line much faster (greater than 900 m yr\(^{-1}\)) than any of the other Filchner-Ronne ice streams (less than 650 m yr\(^{-1}\)).

Over the 50 km just upstream of the strong-bedded region, the fast ice flow narrows in the down-flow direction as its centre-line speed increases from about 300 to 600 m yr\(^{-1}\). The bed beneath this area appears to consist of a mixture of weak and strong regions before giving way to a region where the bed is consistently weak for the next 180 km upstream (Box A in figure 2). Over this extended weak area, the ice stream flows at about 200–300 m yr\(^{-1}\) while maintaining a nearly constant width (approx. 50 km). This is one of the weakest regions we sampled (\(\tau_b = 6\) kPa), and it represents by far the largest expanse of uniformly weak bed in our study area. As described below, however, errors in the bed topography likely cause the inversion to underestimate \(\tau_b\) in this region.

A large strong spot terminates the large expanse of weak area, just upstream of the area where a tributary from the south merges with the main body of the ice stream. This strong spot correlates well with surface slopes visible in the MOA image (figure 5), which indicate a region of locally high-driving stress. Further upstream there is a gap in the velocity data where similar features visible in the MOA data indicate that there are likely additional strong spots at the bed. A similar set of strong spots is located nearly 100 km further upstream. In between these stronger regions, the bed is relatively weak (14 kPa) over the region enclosed by Box B.

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(d) Support Force ice stream

We only have velocity data for the upstream region of Support Force ice stream. Beneath much of this section of the ice stream, the inversions indicate the bed is significantly weaker than in the surrounding regions. Basal shear stress (26 kPa) supports about half of the driving stress (56 kPa), which is similar to the case for tributaries with similar widths and speeds feeding the Ross ice streams (Joughin et al. 2002).

(e) Foundation ice stream

Foundation ice stream discharges 33 Gtons yr\(^{-1}\) from a catchment that includes area from both East and West Antarctica. Its two major tributaries coalesce roughly 70 km inland of the grounding line to form the ice stream’s main trunk (figure 1). Beneath this trunk (Box A), basal shear stress (12 kPa) resists just over 25% of the driving stress (45 kPa). A weak bed also is indicated by the presence of flow stripes that are visible in the MOA data (Gudmundsson et al. 1998).

The low-\(t_b\) region continues inland along the ice stream’s western tributary, averaging 10 kPa over Box B (figure 2), before strengthening upstream of the 250 m yr\(^{-1}\) contour. Further upstream along this tributary, the flow slows and widens over an area where the inversions indicate a weaker bed (Box C) and the MOA imagery indicates a smoother surface. The average \(\tau_d\) over this region is 18 kPa, with a single strong spot providing much of this resistance. In the MOA imagery for the upstream end of this region (figure 6), surface undulations are visible that correspond well with the locations of strong spots determined by the inversion.

The bed topography is poorly resolved beneath parts of Foundation ice stream, probably because of surface clutter from crevasses on the active ice stream (Lythe & Vaughan 2001). In particular, Boxes A and C have mean thicknesses roughly four times as large as Box B, which lies between them. This means that it is likely that that both \(\tau_b\) and \(\tau_d\) for Box B are larger than is indicated in table 1 since the ice should be much thicker than indicated by the BEDMAP data. Since the ice stream is relatively narrow (approx. 20 km) in this region, however, the thicker ice means that the shear margins may also provide substantial resistance. In §6, we describe an example for Recovery ice stream that also supports this conclusion. Consequently, while the \(\tau_b\) estimate may be quantitatively incorrect, we conclude that the inference of a weaker bed beneath Box B is qualitatively correct.

Foundation ice stream’s eastern tributary originates in an area of enhanced flow that also supplies Support Force ice stream. Basal shear stress along this tributary remains high all the way to the intersection with the ice stream’s main trunk.

(f) Institute ice stream

Institute ice stream discharges 21.6 Gtons yr\(^{-1}\) to the centre of the Ronne ice shelf. The main trunk is partially fed by a long tributary flowing from west to east, which has a relatively strong bed. The basal shear stress beneath the ice stream’s main trunk (Box A) is 11 kPa, which balances 43% of the driving stress. The regions of relatively small basal shear stress on the ice stream’s trunk corresponds well with the region downstream from where Scambos et al. (2004)
inferred a rapid transition to low basal shear stresses. A grounded area (Box B) juts out into the ice shelf along the ice stream’s eastern margin. The average basal shear stress for this region is 3.5 kPa, which is comparable to weak till beneath the Ross ice streams (Tulaczyk et al. 2000a; Kamb 2001). At the downstream end of Box A and just inland of the grounding line, the bed is substantially stronger (approx. 20 kPa) than the average over the trunk, which may play a role in determining the grounding line’s present location.

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Rutford ice stream

With a discharge of 17.8 Gtons yr\(^{-1}\), Rutford is the third smallest ice stream in our study area. The inversions show a mixture of strong and weak spots beneath Rutford ice stream, which extends more than 100 km inland from the grounding line. Several of these stronger spots agree well with bumps visible in the MOA image shown in figure 7. In at least one case (near the right edge of figure 7), an inferred sticky spot seems to be displaced by several kilometres (down and to the right) from a prominent bump visible in the imagery. This may be an artefact in the DEM, since the increase in \(\tau_d\) associated with the bump extends over a larger area than indicated by the MOA imagery.

The area enclosed by the box in figure 2 indicates that bed strength can be relatively weak (6 kPa) in places, even though the driving stresses are relatively large (45 kPa). These weak areas, as well as the abrupt transitions between weak and strong regions, are consistent with seismic data collected at several locations on the ice stream that show strong porosity variations over short distances (less than 10 km; Smith 1997a,b; Vaughan et al. 2003). In particular, the seismic data indicate that the weak regions have porosities (greater than 0.4) similar to those of till samples from beneath Whillans ice stream where basal shear stresses typically are less than 4 kPa (Tulaczyk et al. 2000a; Kamb 2001).

Evans ice stream

Several tributaries merge near an irregularly shaped grounding line to form Evans ice stream (see figure 8). Despite having one of the smaller catchments
in our study area, this ice stream produces the largest annual discharge (35.7 Gtons yr$^{-1}$) because the accumulation rate is nearly an order of magnitude higher for this catchment than for the larger East Antarctic catchments (e.g. Recovery).

The bed is weak (8 kPa) beneath much of the eastern half of the ice stream’s main trunk (Boxes A and C), with higher basal shear stresses beneath the western portion. Most of the tributaries have relatively high basal shear stresses above the regions where they merge with the main ice stream. A notable exception is the easternmost tributary, where a weak bed (7 kPa) extends nearly 50 km upstream (Box B) from the ice stream’s main trunk.

Several flow stripes and an otherwise smooth surface (figure 8) indicate a weak bed (Gudmundsson et al. 1998) in roughly the same areas where the inversions show the basal shear stresses are small. In contrast, the surface topography appears much rougher in areas that the inversions show have larger basal shear stresses.

Seismic data were collected along a 7 km line on a fast-flowing section of Evans ice stream (see light-blue triangle in figure 8; Vaughan et al. 2003). The inferred porosities along this line are 0.4 and greater suggesting the presence of weak dilated till, which is in good agreement with the low basal shear stress determined by our inversions.

5. Inversions with flow-law enhancement

It is often assumed that ice stream margins are weakened through strain softening and/or viscous heating (Budd & Jacka 1989; Echelmeyer et al. 1994; Scambos et al. 1994). This effect is often taken into account using a scalar multiplier, $E$, in the flow-law: $E > 1$ implies enhancement. For Whillans ice stream, evidence of weak margins comes from a profile across the margin where a one-dimensional model requires strong enhancement ($E = 10$) to match observed velocities (Echelmeyer et al. 1994). More recent velocity measurements for several profiles across the margin, however, agree well with their theoretical counterparts, suggesting little or no margin enhancement (Joughin et al. 2004). Laboratory tests of ice samples extracted from the margin of Whillans ice stream also suggest little or no margin enhancement (Jackson & Kamb 1997). For these reasons, we performed our standard set of inversions using no flow-law enhancement ($E = 1$).

It is important to note that many of the data used for determining the temperature dependence of the flow-law at temperatures less than $-10 \, ^{\circ}C$ were obtained from observational data on the spreading of ice shelves (Paterson 1994). The resulting parameterization for the flow-law yields values of the flow-law parameter, $A$, up to 60% greater than suggested by some laboratory measurements (Paterson 1994). Also, by ignoring horizontal advection, our temperature model may overestimate temperatures (Joughin et al. 2003). In addition, without horizontal advection the temperature model we used may overestimate ice stream temperatures. If this is the case, then some degree of enhancement is included since the warmer temperatures estimates yield larger values for $A$. Thus, even with a value of $E = 1$ some degree of enhancement may be implicitly included in the flow-law parameterization used for the inversions.

The degree to which margin enhancement occurs is still an open issue. If there is margin enhancement, then our inversions with no enhancement will tend to
under-predict basal shear stress by over-predicting lateral resistance. To evaluate the sensitivity of this effect, we performed an additional set of inversions with $E=3$, which is toward the high range of what can be expected for a strong single-maximum fabric (Paterson 1994). It is also important to note that in very strong shearing, as in the sides of the ice streams, one does not really expect (nor did Jackson & Kamb (1997) observe) a single-maximum fabric, because recrystallization should actively produce a fabric that is not grossly different in enhancement from random (Alley 1992). Thus, we believe that the values of $E=1$ and 3 used in the inversions bracket the range of likely values.

Table 1 provides a comparison of the inversion results with and without enhancement. In general, the inversions with enhancement yield values of $\tau_b$ that are 1–8 kPa larger than the cases with no enhancement, largely because ‘softer’ ice can provide less lateral resistance. A similar increase was found for inversions on the Ross ice streams (Joughin et al. 2004). Even with enhancement, the areas enclosed by the boxes in figure 2 have relatively small basal shear stresses that, in general, support less than half the driving stress. Thus, the qualitative picture of the bed comprising a mixture of weak and strong spots does not change dramatically when the inversions include flow-law enhancement.

6. Discussion

For much of the area beneath the Ross ice streams, basal shear stresses are less than 4 kPa (Tulaczyk et al. 2000a; Kamb 2001; Joughin et al. 2004). In contrast, basal shear stresses average 12 kPa over the boxes shown in figure 2, indicating a substantially stronger, although still relatively weak, bed. Over nearly 27% of this area, basal shear stress is less than 4 kPa, which indicates that there may be areas of dilatant till beneath these ice streams with characteristics similar to the till beneath the Ross ice streams. This conclusion is consistent with seismic data that suggest the presence of high-porosity (greater than 0.4) dilatant till beneath Rutford and Evans ice streams (Smith 1997a,b; Vaughan et al. 2003).

Although it appears that weak dilatant till is present beneath some areas of the Filchner-Ronne ice streams, it does not appear to be as evenly distributed as beneath the Ross ice streams. This suggests that numerous sticky spots and, in some cases, ‘sticky regions’ contribute to higher basal shear stresses (Alley 1993). This hypothesis is consistent with seismic results showing high- and low-porosity regions in close proximity to each other (Smith 1997a,b; Vaughan et al. 2003). Even with sticky spots, however, many of the weak regions do not fully support the driving stress. Thus, both sticky spots and marginal shear stresses are important controls on these ice streams.

Most of the differences between the Ross and Filchner-Ronne ice streams may be attributable to the regional geology. The Ross ice streams flow through a relatively low-relief continental shelf that contains abundant, erodable, Tertiary sediments, which provide a suitable substratum for the development of widely distributed weak till (Tulaczyk et al. 1998; Studinger et al. 2001; Winberry & Anandakrishnan 2003). Conversely, the Filchner-Ronne ice streams drain across the Transantarctic mountains, and the pattern of alternating weak and strong bed patches may correspond to ice moving over alternating sedimentary basins and mountain ranges.

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The mean thicknesses given in table 1 indicate that the Filchner-Ronne ice streams are thicker than the Ross ice streams by several hundred metres. Assuming similar geothermal heat fluxes and surface temperatures, this thicker ice implies less conductive heat loss from the bed. The larger mean basal shear stresses also imply greater basal shear heating. These factors combined likely mean that there is significantly more water at the beds of these ice streams than there is beneath the Ross ice streams (Raymond 2000). This makes it unlikely that the regions of higher basal shear stress can be attributed to a lack of lubricating water, which suggests that bed rock bumps or other till free areas contribute to a stronger bed (Alley 1993). This is consistent with the geologic setting of these ice streams, which flow across the Transantarctic mountains.

Recovery ice stream exists as a well defined flow feature over a distance of several hundred kilometres inland from its grounding line. Mountains along its northern margin (figure 5) and, to a lesser extent, along its southern margin indicate a strong degree of topographic control. The BEDMAP data indicate a mean bed elevation of 260 m for the ice stream’s fast-flowing section situated between these mountains (Box A). If this elevation is correct, then it is unlikely that the weak bed can be attributed to marine sediments, although non-marine sediments might still serve to weaken the bed. Unfortunately, there is almost a complete absence of ice thickness measurements for this sector of the ice sheet, so the BEDMAP DEM may contain large errors (Lythe & Vaughan 2001).

The mean BEDMAP thickness for Box A on Recovery ice stream is 786 m, and the mean velocity is 244 m yr$^{-1}$. With an approximate width of 50 km, this yields an estimated ice flux of 9 Gtons yr$^{-1}$. This section of the ice stream, however, drains about 80 to 90% of the catchment area, which should yield a flux of roughly 28–32 Gtons yr$^{-1}$ based on the flux at the grounding line (Joughin & Bamber 2005). Since the ice-stream velocity and width are well constrained, this suggests that the mean thickness for Box A is too small by a factor of about 3.2–3.6. Scaling the BEDMAP thickness by a factor of 3.2 yields a mean bed elevation of $-1460$ m for Box A. This suggests that Recovery flows through a deep, well-defined trough in the subglacial topography, which extends several hundred kilometres inland from the grounding line. The depth of this

Figure 7. Basal shear stress (colour) estimate for Rutford ice stream (a) and corresponding MOA image (b). Flow speed is shown with 100 m yr$^{-1}$ contours (black lines).
trough is such that it would remain well below sea level even after rebound during deglaciated periods, increasing the likelihood that marine sediments may weaken the bed. Recent results suggest that there is an approximately 3 km-thick layer of marine sediments in a trough at the upper end of neighbouring Slessor ice stream (Bamber et al. 2006).

If our hypothesis that Recovery ice stream flows through a deep trough is correct, then the thickness data poorly constrain our \( \tau_b \) estimates. To investigate the impact of this error, we performed new inversions after re-scaling the ice thickness by a factor of 3.2 and adjusting the basal topography accordingly. This rescaling increased the mean driving stress estimate for Box A, which is proportional to thickness, to 62 kPa. This inversion yielded a basal shear stress estimate of 18 kPa, which while greater than the thin-ice estimate, still suggests a relatively weak bed beneath much of the ice stream. Marginal resistance and the driving stress are proportional to ice thickness, which may explain why the basal shear stress for both the ‘thin’ and ‘thick’ inversions supports approximately 30% of the driving stress.

Vaughan & Bamber (1998) note several areas where the geometry of the Antarctic ice sheet appears to be ‘drawn down’ relative to an idealized ice-sheet model. The most apparent anomaly is the ‘concave-up’ profile of the catchment

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**Figure 8.** Basal shear stress estimate (colour) for Evans ice stream (a) and corresponding MOA image (b). Flow speed is shown with 100 m yr\(^{-1}\) contours (black lines). Blue triangle shows the approximate location of seismic surveys (Vaughan et al. 2003) and the red line indicates the grounding line position.
drained by the Ross ice streams. This departure from the typical ‘concave-down’ ice-sheet profile has been attributed to the weak beds of these ice streams, which cannot support high driving stresses (Whillans & Van der Veen 1997). In East Antarctica the largest departure from an idealized ice-sheet profile is in Recovery ice stream’s catchment (Vaughan & Bamber 1998), where several processes related to a large subglacial trough would have a major influence on the ice sheet’s geometry. As with the Ross ice streams, a weak bed would limit the driving stress that the bed can support, which in turn would influence the slope and shape of the ice sheet. Thicker ice in the trough would maintain high driving stresses but with lower slopes. Perhaps a more important effect is the influence that a 1400 m deep trough would have on basal temperature. For a given set of boundary conditions (geothermal heat flux, surface accumulation rate and temperature) and steady state conditions, ice thickness determines whether there is melting or freezing at the bed (Paterson 1994). This means that at the base of the trough, temperatures can reach the pressure melting point equivalent to that for an ice sheet with a flat bed and a surface elevation approximately 1500 m lower than observed. Once the ice sheet thickens enough to allow basal melting in the trough, enhanced basal motion would increase advection of the ice from the interior to counteract further thickening, resulting in a low profile ice sheet. Although the data suggest sliding, a similar effect, albeit at slower flow speeds, might be achieved without basal melting since the thick trough ice would promote enhanced deformation through higher driving stresses and warmer basal ice (Hulbe et al. 2000).

Slessor ice stream also is a well-defined feature that extends deep into its catchment, which indicates that its flow is guided by a subglacial trough. Recent radio-echo sounding results have revealed that its three tributaries flow through deep basal troughs (Rippin et al. 2003). Areas of the bed beneath the northernmost tributary lie up to 800 m below sea level and would remain below sea level even after isostatic compensation (Rippin et al. 2003). As in the case of Recovery ice stream, continuity suggests that this trough extends all the way to the grounding line at a depth well below sea level.

In addition to Recovery and Slessor ice streams, the BEDMAP data indicate that Support Force, Foundation, Rutford and Evans ice streams all flow through deep subglacial troughs, though in many cases these features are poorly resolved by the sparsely sampled data. The catchments of these ice streams are all drawn down relative to an idealized ice-sheet profile (Vaughan & Bamber 1998). Just as for Recovery ice stream, it is likely that these troughs have a strong influence on the shape of ice sheet, resulting in a significantly lower profile relative to a flat-bedded ice sheet.

Deep subglacial troughs should have a strong influence on the ice sheet’s geometry and evolution. The fact that these troughs are poorly resolved or missing altogether (e.g. Recovery) indicates that ice-sheet models for determining the evolution of the Filchner-Ronne ice streams are poorly constrained. As an example, full ice-sheet models predict substantial thinning for this region as part of the ice sheet’s long term response to environmental changes since the Last Glacial Maximum (Huybrechts et al. 2004). Although recent and long-term changes in accumulation offset some of this thinning, the model still predicts strong (up to several centimetres) present-day net thinning. Recent observations, however, suggest this region is thickening (Joughin & Bamber 2005; Davis et al. 2005). This may be because fast moving
ice streams flowing through deep troughs yield faster response times than those predicted by models based on sheet flow over a nearly flat bed.

Unlike the other Filchner-Ronne ice streams, the main trunk of Institute ice stream is not constrained by so deep or well defined a trough (Scambos et al. 2004). It also differs from these ice streams in that it has a wider and more expansive ice plain above the grounding line, which more closely resembles the ice plains of Whillans and Kamb ice streams. Flow lines in the ice shelf downstream of the Institute grounding line also indicate some evidence of flow variability approximately 360 years ago (Scambos et al. 2004). These are characteristics that are similar to those of Whillans and Kamb ice plains. Institute ice stream, however, is several hundred metres thicker than these other ice plains, which combined with its higher basal shear stresses, should yield warmer basal temperatures and maintain melting. Thus, while freezing beneath the other ice plains may dewater and strengthen till (Tulaczyk et al. 2000b; Bougamont et al. 2003a, b), stagnation of Institute ice stream appears unlikely as long as it is able to maintain its thicker ice. On the whole, Institute ice stream appears to represent a combination of characteristics of both the Filchner-Ronne and Ross ice streams.

7. Summary

Seismic observations have suggested relatively heterogeneous conditions beneath some of the Filcher-Ronne ice streams (Smith 1997a, b; Vaughan et al. 2003). Our results confirm these observations and provide a comprehensive picture of basal conditions beneath the other ice streams feeding the Filchner-Ronne ice shelf. The most notable feature of these inversions is the extensive area of weak bed beneath Recovery ice stream, which extends deep into the ice sheet’s interior. While limitations in knowledge of the bed topography make our results in this area more qualitative than quantitative, they suggest the presence of marine sediments beneath Recovery ice stream, similar to those that may lie beneath Slessor ice stream (Bamber et al. 2006). Furthermore, our results indicate that flow in this large catchment is dominated by a large, unmapped subglacial trough. Failure to account for this and other large subglacial troughs is likely to have a substantial effect on model predictions of ice-sheet evolution. Thus, the significance of these troughs demonstrates the need for further radio-echo sounding campaigns to better constrain subglacial topography, particularly in the large expanses where little or no data currently exist.

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