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Investigations of an "ice plain" in the mouth of Pine Island Glacier, Antarctica

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ABSTRACT. We present newly acquired airborne radar data showing ice thickness and surface elevation for Pine Island Glacier, Antarctica. These data, when combined with earlier measurements, suggest the presence of a lightly grounded area immediately above the grounding line of Pine Island Glacier. We identify this region as an "ice plain". It lies close to the centre line of the glacier, has an elevation above buoyancy of <50 m and extends inland for >28 km. The upstream edge of the ice plain is defined by a "coupling line". The configuration of the ice plain implies that nearby thinning of the ice stream would result in substantial grounding-line retreat. We suggest that the grounding-line retreat of Pine Island Glacier, observed between 1992 and 1996, probably commenced sometime after 1981.

1. INTRODUCTION

Pine Island Glacier drains an area of approximately $165\,000 \pm 700\ \mathrm{km}^2$ (Vaughan and others, in press), almost one-quarter of the West Antarctic ice sheet (WAIS) (Fig. 1). Extensive regions of its catchment area, as with most of the WAIS, rest on a bed which is considerably below sea level and is thus known as a marine ice sheet. Pine Island Glacier is unusual amongst WAIS glaciers in that it terminates in only a short ice shelf with <70 km between the grounding line and ice front. Of the ice discharged across the grounding line, only half is accounted for by calving, and the remainder is removed by basal melting of the ice shelf at a mean rate of $12\ \mathrm{m\,a}^{-1}$ (Jenkins and others, 1997). Interest in the area has increased recently because of the detection both of grounding-line retreat (Rignot, 1998) and of changes in surface elevation in its and neighbouring basins (Wingham and others, 1998).

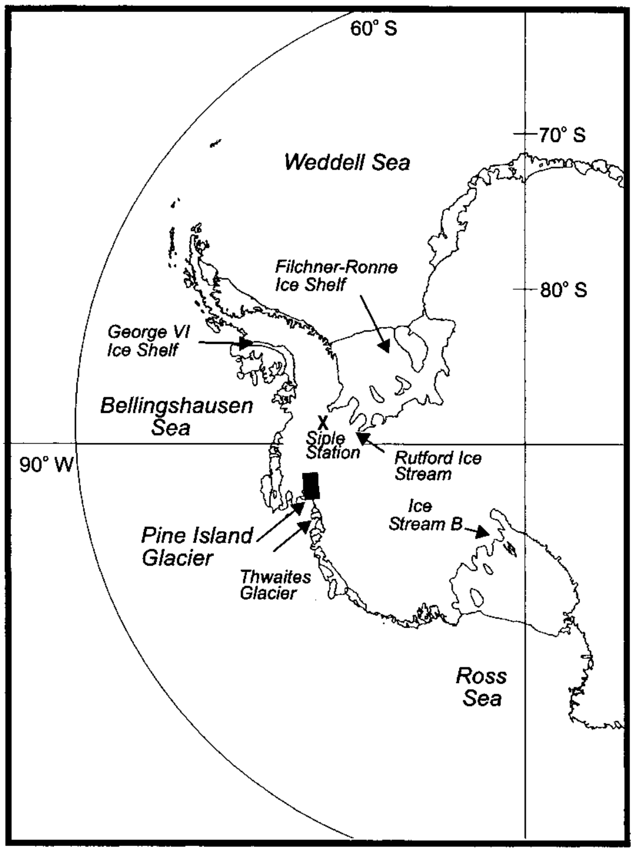

Fig. 1. Map of West Antarctica showing the location of Pine Island Glacier and other locations referred to in the text. The solid black rectangle is the area detailed in Figure 2.

The concept that a marine ice sheet is inherently unstable was developed by Weertman (1974) who showed that a grounding line retreating over bedrock that sloped downward towards the centre of the ice sheet could continue retreating. As a consequence of Pine Island Glacier's short ice shelf and fore-deepened bed profile, it was identified as an ice-stream/ice-shelf system that is potentially unstable (Hughes, 1975, 1977; Thomas, 1979). Indeed, Hughes (1981) suggested that this area represented the "weak underbelly" of the WAIS. More recently, Hindmarsh (1993) pointed out that having an ice stream between an ice sheet and an ice shelf allows the possibility of the system reaching neutral equilibrium, suggesting stability rather than instability. This suggests that ice shelves do not "buttress marine ice sheets against decay" and may be unimportant in maintaining the stability of the inland ice sheet. Although a definitive numerical model for the dynamics of marine ice sheets and ice streams has yet to be produced, the likelihood of a collapse of the WAIS in the next century or two has been discounted by Bentley (1998) in a review of recent models.

In this paper we focus our attention on the grounding-line region of Pine Island Glacier. The grounding line was first located from airborne radar measurements by Crabtree and Doake (1982). Its location was subsequently queried and suggested to be further downstream; firstly by Thomas (1984), based on a stability argument, then by Lucchitta and others (1995), using satellite images. Rignot (1998), using interferometric synthetic aperture radar (InSAR) measurements, identified the flexing limit about 30 km downstream from the grounding-line position identified by Crabtree and Doake (1982).

There is a striking similarity between the interpretation history of airborne radar data collected along Ice Stream B and our reinterpretation of the data collected by Crabtree and Doake (1982) along Pine Island Glacier. A grounding-line position for Ice Stream B was first determined by Rose (1979) from airborne radar measurements. Then Shabtaie and Bentley (1987), using extensive airborne radar soundings tied to accurately located ground stations, demonstrated that the true grounding-line position of Ice Stream B actually lay some 100 km further downstream. They also showed that the intervening area between the true grounding-line position and that identified by Rose, a region now commonly termed an "ice plain", was lightly grounded, with elevations above buoyancy of only 30–40 m.

We amalgamate previous datasets and observations with new airborne measurements and conclude, in agreement with earlier speculations on the position of the grounding line by Thomas (1984) and Lucchitta and others (1995), that Pine Island Glacier has an ice plain above its grounding line. We suggest that the grounding line identified by Crabtree and Doake (1982) for Pine Island Glacier is actually a "coupling line", like the supposed grounding line identified by Rose (1979) on Ice Stream B. A coupling line, as distinct from a grounding line, is a "dynamic boundary between grounded and very slightly grounded ice, rather than between grounded and floating ice" (Bindschadler, 1993).

2. PINE ISLAND GLACIER GROUNDING-LINE MEASUREMENTS

Pine Island Glacier is an area with few direct geophysical measurements. The reason for this lack of data is principally its location, being distant from any national Antarctic station. Ground-based traverses along Pine Island Glacier ice shelf and ice stream would have to negotiate a route through severely crevassed regions. In addition, Pine Island Bay is annually beset by sea ice, hindering ship access to the ice front. It is therefore no surprise to find that geophysical knowledge of the region has been largely gleaned from airborne and satellite measurements, which we will now review.

The first airborne ice-sounding radar measurements of Pine Island Glacier were made in February 1981 by Crabtree and Doake (1982). Ice thickness and surface elevation were measured on two flights along and across the glacier, with an uncertainty of ± 30 m for ice thickness and of ± 10 m for surface elevations derived from pressure altimetry. Crabtree and Doake (1982) located the grounding line via a two-stage process. First, they calculated the thickness of ice that when floating would have a surface elevation equal to that measured. Second, they selected a grounding line at the point where a significant divergence between measured and derived ice thickness began. The coincidence of a notable break in surface slope with the start of the deviation between measured and derived ice thicknesses corroborated their choice for the location of the grounding line. One of the flight tracks (9 February 1981) and the supposed location of the grounding line are marked in Figure 2.

Using an ice-dynamics model, based on Weertman's (1974) concept of balancing the creep thinning of an ice shelf with the advection of grounded ice into it, Thomas (1979) estimated that the depth of the grounding-line sill, for the ice sheet to be in equilibrium, had to be 520–610 m. Citing these estimates, Thomas (1984) used force-balance arguments to dispute the location of the grounding line identified by Crabtree and Doake (1982). Thomas speculated that the region between the grounding line recognized by Crabtree and Doake and the one he favoured (~29 km downstream) was occupied by a partially grounded glacier sliding over a lubricated bed. This is the feature we identify as an ice plain.

The placement of the grounding line by Crabtree and Doake (1982) was also queried by Lucchitta and others (1995), who, by an examination of a pair of 1992 ERS-1 SAR images, noted that the location was in a region of rolling topography. Furthermore, by tracking the movement of crevasses, Lucchitta and others (1995) showed the ice-stream velocity continues to increase for 20 km downstream of the grounding line identified by Crabtree and Doake (1982). The inferred topography and the measured velocities led Lucchitta and others (1995) to speculate that Crabtree and Doake (1982) had identified an "ice plain" of the type described by Bindschadler (1993).

The technique of satellite InSAR was successfully employed by Rignot (1998) to locate the limit of flexing for Pine Island Glacier. Generally, the limit of flexing was shown by Stephenson (1984) and Smith (1991) to lie over grounded ice, upstream of where the ice column first reaches hydrostatic equilibrium (Fig. 3). The InSAR process does not require precise geolocation of individual SAR images, but rather precise relative registration. In transferring Rignot's results on to an absolute polar stereographic projection (Fig. 2), so that we can compare datasets, we estimate a positional uncertainty of ± 1 km. Rignot (1998) removed the effects of topography and ice velocity from interferograms, leaving vertical displacements in response to ocean tides. An elastic-beam model of the ice shelf was then fitted to the tidal displacements, and the limit of flexing located. The interferometric flexing limit for 1992 was about 30 km downstream from the position Crabtree and Doake (1982) gave as the grounding line, in about the same position as the grounding line suggested by Thomas (1984).

Rignot (1998) calculated that in the centre of the ice stream the limit of flexing retreated by 1.2 ± 0.3 km a^{-1} between 1992 and 1996 (Fig. 2). He attributed the retreat to an increase in basal melting causing the ice shelf to thin by 3.5 ± 0.9 m a^{-1}.

On 13 February 1998 a British Antarctic Survey aircraft (de Havilland DHC-6 Twin-Otter) completed a return flight from the abandoned Siple station (75°54′ S, 84°54′ W) to the front of Pine Island Glacier (A′–A–C–C′ in Fig. 2). The survey aircraft measured the ice thickness and the magnetic signature of the lithosphere. Post-processing of differential global positioning system (GPS) data has provided accurate navigational and altimetric data for the aircraft. Ice thicknesses were measured everywhere except for a 15 km long segment on A′–A starting about 17 km upstream from the limit of flexing, where the bed echoes were obscured by heavy surface crevassing. The lowest bed elevation measured was approximately 1575 m below sea level, 170 km from the ice divide between Pine Island Glacier and Rutford Ice Stream drainage basins. The radar returns from the ice shelf of Pine Island Glacier are characterized by strong reflections from basal crevasses, whilst over the grounded ice the returns are generally weaker and exhibit rapid changes in the power of the echo. There is a gradual change in the character of the radar returns in the region of the grounding-line area, and

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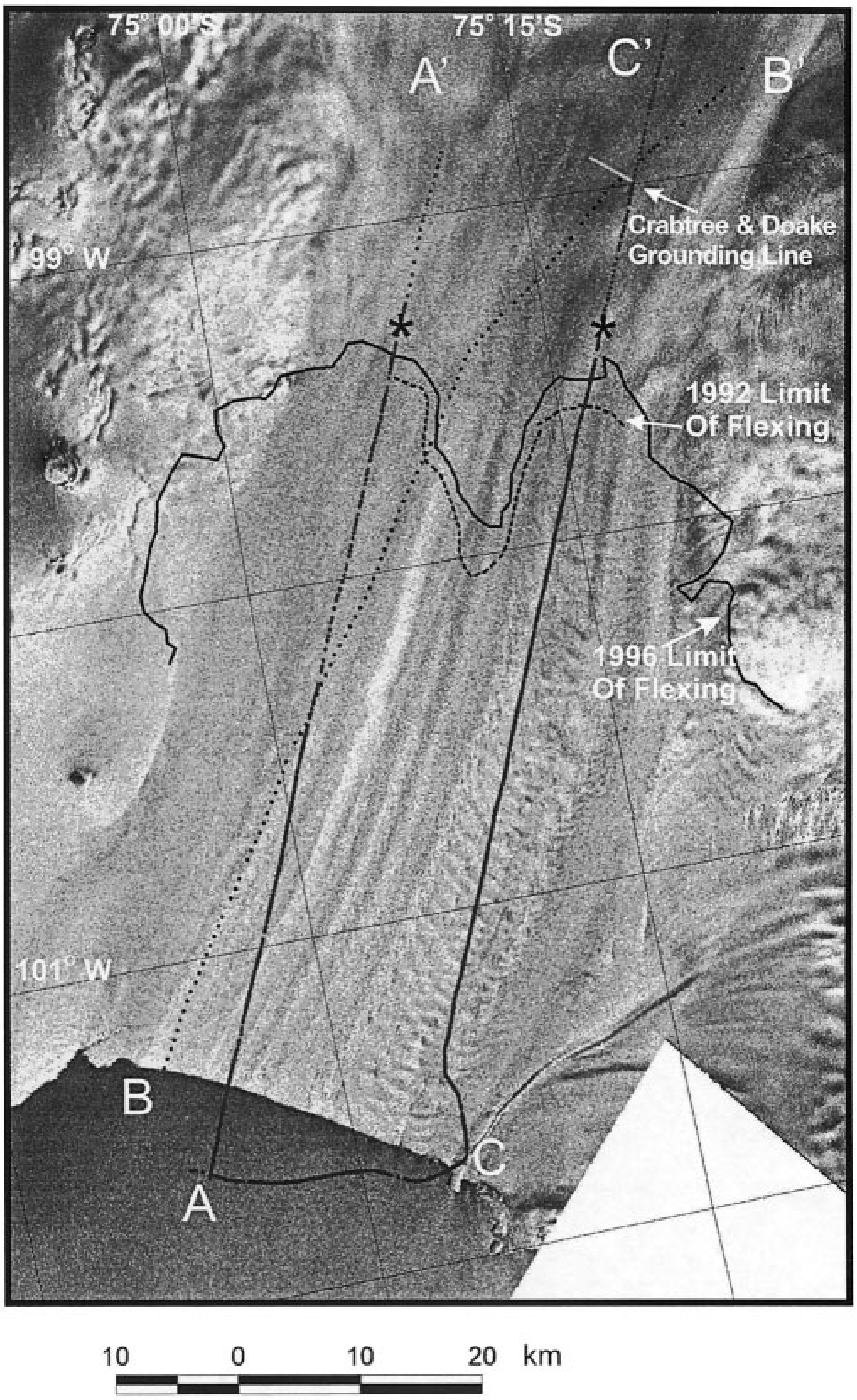

Fig. 2. Mosaic of SAR images (orbit 3174, frames 5193 and 5211) of the grounding-line region and ice shelf, Pine Island Glacier. Marked on the image are the 1992 and 1996 limits of flexing (Rignot, 1998), grounding-line location initially deduced by Crabtree and Doake (1982) and the 1998 calculated hydrostatic positions (marked by stars). The February 1981 flight track is along the line B–B′. The February 1998 flight track is along A′–A–C–C′.

not a distinct sharp change. Thus ice-sounding radar cannot, alone, identify the grounding-line location accurately.

3. DETERMINATION OF HYDROSTATIC SURFACE

For an ice shelf of known thickness floating in hydrostatic equilibrium we can estimate the elevation of the surface above sea level. The difference between the measured and the calculated surface elevation is called the *hydrostatic anomaly*. The position where the hydrostatic anomaly starts to show a significant increase, coinciding with an associated break in surface slope, we identify as the *coupling line* (Fig. 3). In this section we formulate a relationship, using data from 1998, to calculate a hydrostatic surface from the ice thickness.

A consequence of using GPS for navigation is that the sur-

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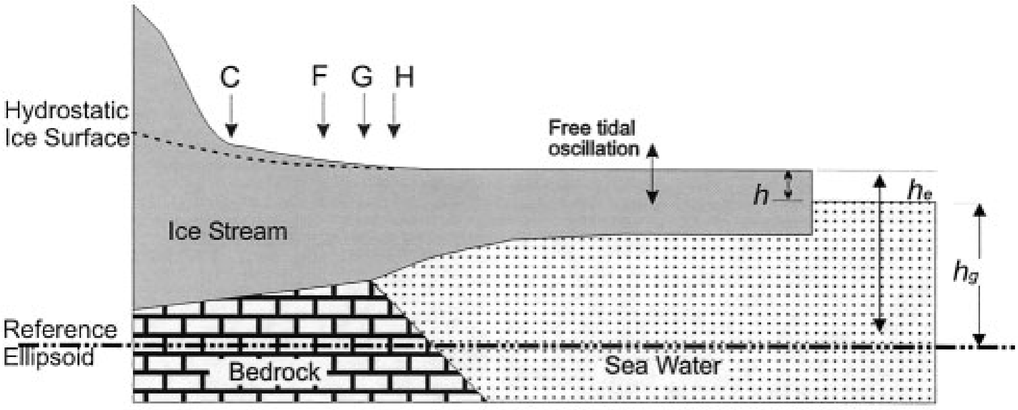

Fig. 3. Schematic of grounding line (after Smith, 1991), showing the expected juxtaposition of the coupling line (C), the limit of flexing (F), the grounding line (G) and the hydrostatic point (H). The parameters used to derive a hydrostatic surface in Equation (1) are also shown.

face and bed elevation measurements of Figure 4a and c are referenced to an ellipsoid (WGS84) and not sea level. Only where the aircraft crossed the ice front and flew over open sea are the height of the ice surface above sea level, h, and above the reference ellipsoid, h_e, both known. The difference between h and h_e gives h_g, the ellipsoid–geoid separation at the ice front. Expanding the hydrostatic relationship for calculating the height, h, of ice above sea level for an ice shelf of thickness H gives:

$$h = h_e - h_g = \left(1 - \frac{\rho_i}{\rho_w}\right) H + \frac{\rho_i}{\rho_w} d\,, \tag{1}$$

where d is the air gap to be subtracted from H to reduce it to an equivalent solid ice thickness (see Shabtaie and Bentley, 1987, for an elaboration of Equation (1)). The density of solid ice, ρ_i, is taken as 920 kg m^{-3}, and the mean sea-water density, ρ_w, is taken as 1030 ± 3 kg m^{-3} (Jenkins and Doake, 1991). Although it is not known how h_g varies along the flight tracks from the ice front to the grounding line, a distance of 70 km, geoid models such as OSU91A (Rapp and others, 1991) suggest there will be only minor differences. Making h_g a first-order function of distance, x, from the ice front,

$$h_g = Mx + C\,, \tag{2}$$

and substituting from Equation (1) gives:

$$h_e(x) - \left(1 - \frac{\rho_i}{\rho_w}\right) H(x) = Mx + C + \frac{\rho_i}{\rho_w} d\,, \tag{3}$$

where C is a constant that includes any systematic errors in the measurements. We estimate the resolution in the measurements of ice thickness, H, and surface height above ellipsoid, h_e, to be ± 6 and ± 0.7 m, respectively. The formulation allows a linear regression of the lefthand side of Equation (3) against x that gives the gradient of the line, $M = (0.0 \pm 8.0) \times 10^{-6}$, and intercept, $[C + (\rho_i/\rho_w)d] = -3.6 \pm 3.4$ m. This suggests that the variation in the ellipsoid–geoid separation, h_g, is negligible unless it is exactly offset by a counteracting change in the density profile of the ice shelf. We can therefore calculate a hydrostatic surface, $h_e(x)_{\text{fitted}}$ (relative to the reference ellipsoid), from the ice thickness.

$$h_e(x)_{\text{fitted}} = 0.107 H(x) - 3.6\,. \tag{4}$$

The possible range of mean sea-water densities (± 3 kg m^{-3}) produces only minor variations in the regression values. The total uncertainty on the derived ice surface, $h_e(x)_{\text{fitted}}$, is therefore ± 3.5 m.

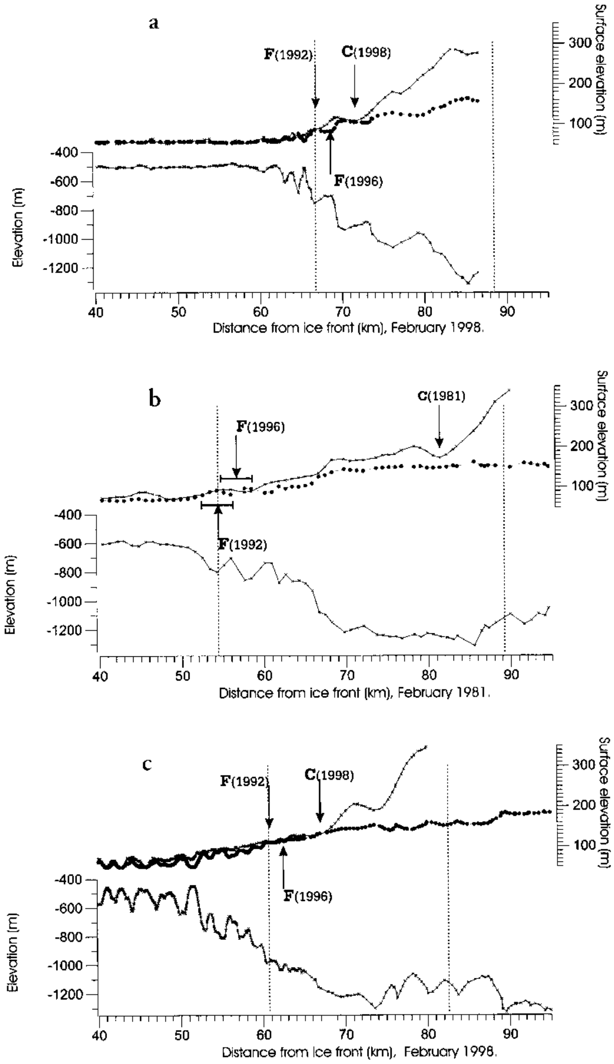

Fig. 4. Ice-base elevation and measured (crosses) and derived (circles) surface elevations, relative to the World Geodetic System (WGS84) ellipsoid, along sections of the Crabtree and Doake (1982) flight-line and the 1998 flight-lines. F marks the limit of flexing position and C marks the coupling line. (a) Profiles along a section of line A–A′ on the northern side of the glacier. (b) Profiles along a section of line B–B′ near the centre of the glacier. The horizontal bars at the limit of flexing positions represent the ±2 km uncertainty in their placement along the 1981 data. (c) Profiles along a section of line C–C′ on the southern side of the glacier. Sections of the profiles between the vertical dotted lines are detailed in Figure 5.

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Figure 4a and c are plots of the ice-bottom elevation and the measured and derived hydrostatic surfaces along the northern and southern sides of Pine Island Glacier. The coupling points that fulfil our criteria are also shown on Figure 2. The small error on the derived ice surfaces has negligible influence on our positioning of the coupling points.

Application of Equation (4) permits an accurate hydrostatic surface to be fitted to the ice-thickness data collected by Crabtree and Doake (1982). Also, the measurement of h_g at the ice front allows the surface elevation data from 1981 to be transferred to our reference frame (WGS84). Navigation for this flight was by a Doppler-Omega navigation system. Unless this system is updated with a reference position it can accumulate a positional uncertainty of >1 km per flying hour. We estimate the uncertainty in the aircraft positioning on 9 February 1981 (Fig. 2) to be ± 2 km, controlled by observing in the radar records when the aircraft crossed the ice front. We then estimated the 1981 ice-front location based on its position in a 1982 Landsat image, assumed there was no calving of the ice front during the intervening year and used an ice-front velocity of 2.6 km a^{-1} (Jenkins and others, 1997). An independent confirmation of this level of error is provided by comparing the surface elevations at the crossing point of lines B–B′ and C–C′ (Fig. 2). The gradient of the surface slope in this grounded region allows an accurate estimate to be made of the likely error between the 1981 and 1998 flight positions, and is within 2 km, even after allowing for any reasonable change in surface elevation in the interval (Wingham and others, 1998). Unfortunately, the precision of the navigation for the 1981 data precludes the accurate measurement of possible changes in ice thickness where the 1981 and 1998 flight tracks cross over the ice shelf (lines A–A′ and B–B′ in Fig. 2).

Our confidence in the precision of navigation for the flight on 9 February 1981 is in contrast to the main level of uncertainties for an earlier flight on 6 February 1981. Jenkins and others (1997) showed, by identifying the glacier margins in the radar returns, that the along-track positional uncertainty could be as high as 6 km. This earlier flight probably crossed Pine Island Glacier grounding line, but, given the uncertainty in along-track position and probable similar uncertainty in cross-track position, we have had to discount using those data.

Figure 4b shows the ice-base elevation and the measured and derived ice surfaces for the 1981 data. Although there is an uncertainty of ± 10 m associated with the measured ice surface, there is a clear and unambiguous position for the location of a coupling point. The dramatic break in surface slope means that, even with our more accurate calculation of a hydrostatic surface, the position of the coupling point agrees, within the error bounds, with the location Crabtree and Doake (1982) ascribed to the grounding line.

4. DISCUSSION

Although Bentley (1987) has shown that the physical parameters that describe one glacier may not be appropriate for another, it is instructive to compare and contrast the ice-plain regions of Pine Island Glacier and Ice Stream B. For parity, a 25 km long section of ice plain above the grounding line of Ice Stream B is chosen for the comparison (data from Bindschadler and others, 1987). The physical parameter that distinguishes the ice-plain region of Ice Stream B is its very low driving-stress value of 4.8 kPa, a value more typical of an ice shelf. The driving stress on the ice plain of Pine Island Glacier is an order of magnitude higher, at 38 kPa. Bindschadler and others (1987) showed that the ice-plain region on Ice Stream B had negligible basal shear stress, the resistance to flow coming from side shear and longitudinal stress gradients. Measurements made through boreholes drilled on Ice Stream B (Engelhardt and others, 1990) confirmed that the ice base was at its melting point, that the glacier was close to flotation and that the glacier bed consisted of unconsolidated sediments. The basal water and sediments provide the mechanisms for lubricating ice-stream motion, either by basal sliding or by subglacial till deformation.

When Thomas (1984) speculated that Pine Island Glacier was sliding over a lightly grounded area, he guessed a basal shear stress of order 10 kPa. He supposed that the bed might be lubricated by meltwater produced by a region directly upstream of the coupling line where the surface slope is steep, the driving stress is 100–150 kPa and ice-stream motion is due to shear deformation rather than sliding. But perhaps the most notable difference between the two ice-plain regions is that the bed beneath Pine Island Glacier slopes up towards the grounding line, even over the ice plain. To apportion the resistance to the driving force acting on Pine Island Glacier into its individual components would require more ice-thickness data combined with accurate velocity and strain-rate data.

For each of the three profiles in Figure 4, the hydrostatic anomalies have been plotted in Figure 5 as a function of distance from the February 1992 limit of flexing position. Because the ice-thickness profiles were not taken at the same time as the interferometry, we cannot be certain that the flexing limits can be identified with features on the elevation profiles. However, at the highest reported thinning rates of 3.5 ± 0.9 m a^{-1} (Rignot, 1998), we would not expect a significant difference to the actual thickness profiles (Fig. 4), although the hydrostatic anomalies could have changed (Fig. 5). It can be seen from Figure 5b that in 1981 an area along the centre of Pine Island Glacier, now identified as an ice plain, had elevations above buoyancy of <50 m. The region of lightly grounded ice extended some 26 km inland from the 1992 limit of flexing along line B–B′. The distance between the coupling line and the limit of flexing along line A–A′ on the northern side of Pine Island Glacier is 4 km (Fig. 5a), whilst along line C–C′ on the southern side it is 5.5 km (Fig. 5c).

We assume that any ice-stream thinning that removes the hydrostatic anomaly (i.e. the basal overburden pressure) will be accompanied by a retreat of the grounding line. The sensitivity of the ice plain of Pine Island Glacier is clearly shown in Figure 5b, where a 30 m thinning of the ice stream would cause the limit of flexing to retreat inland by almost 20 km. At the thinning rates estimated by Rignot (1998) the ice plain which existed in 1981 would have been consumed in <10 years and the limit of flexing would now lie upstream of the then coupling line. This sensitivity to thinning suggests that the retreat of the flexing limit is a recent event, initiated sometime after 1981 and possibly only a few years prior to the 1992 observations.

It is possible that the recession of the hinge zone has continued since 1996. On Figure 5a and c the 1998 coupling line appears to lie at about the same elevation as the 1996 flexing limit, or even lower. It seems likely that the grounding line would continue to retreat until it was close to the coupling-line position. We do not know by how much the 1981 profile may have changed in the intervening years, but the extended

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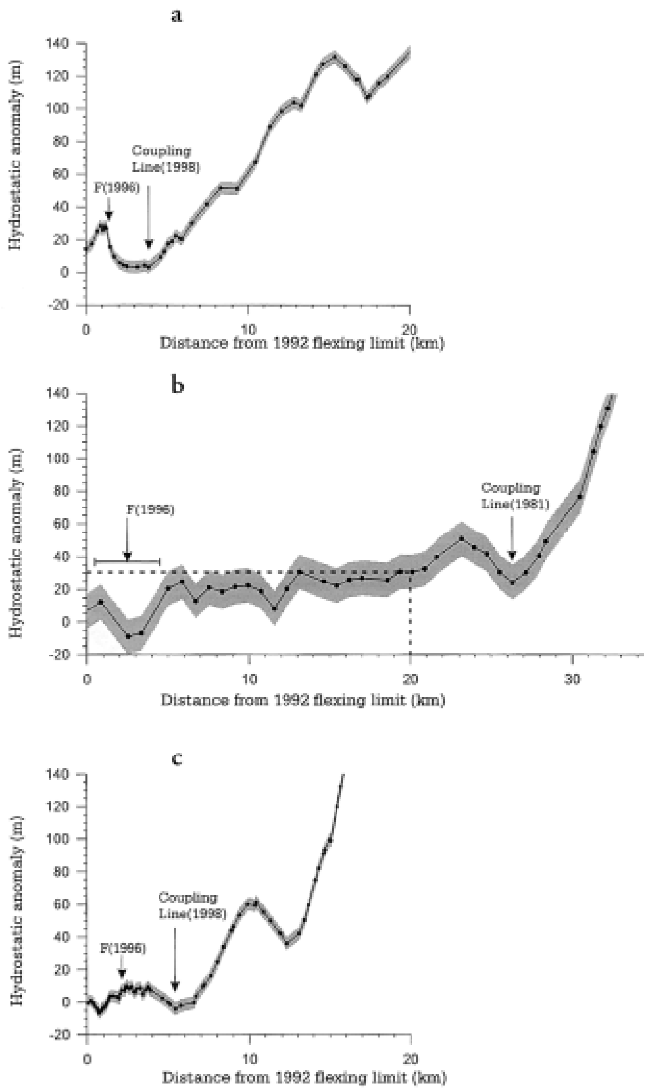

Fig. 5. The hydrostatic anomaly obtained by subtracting the derived hydrostatic profile from the observed surface profile (shown in Fig. 4); the shaded region represents the resultant uncertainty. (a) The thickness in excess of flotation on the northern side of the glacier, uncertainty ±3.6 m. (b) The thickness in excess of flotation along the Crabtree and Doake (1982) flight-line close to the centre of the glacier, uncertainty ±10.6 m. The uncertainty on the placement of the limit of flexing position on the 1981 data is ±2 km. (c) The thickness in excess of flotation on the southern side of the glacier, uncertainty ±3.6 m. For each of the three profiles the 1992 flexing limit has been chosen as the reference position.

area of near-zero hydrostatic anomaly shown in Figure 5b suggests that once a threshold of thinning has been exceeded, bringing the flexing limit back to just upstream of its 1996 position, there could be rapid retreat of the grounding line back to the coupling line. All three profiles in Figure 5 demonstrate, by definition, the very large positive hydrostatic anomalies that lie upstream of the coupling line. We therefore expect the position of the coupling line to be fairly static and relatively insensitive to small changes in ice thickness.

The retreat rate of the limit of flexing was measured by InSAR to be 1.2 km a^{-1}, and attributed to an increase in the basal melt rate of the ice shelf (Rignot, 1998). An alternative cause of the glacier retreat is suggested in the recent InSAR observations that reveal Pine Island Glacier to be an amalgamation of around 10 separate tributaries (Stenoien, 1998), which merge some 130 km upstream of the grounding line. Any one of the tributary ice streams could exhibit non-steady behaviour that would then propagate down-glacier. Satellite altimeter measurements between 1992 and 1996 show a falling surface elevation in the catchment areas of both Pine Island Glacier and its neighbour, Thwaites Glacier (Wingham and others, 1998). This may imply a long-term re-adjustment to changing accumulation which is manifested as a thinning of the ice stream at the grounding line. There have been many attempts to assess whether Pine Island Glacier is in a state of balance, by computing the input from snow accumulation minus the output by ice flow (for a summary of these see Vaughan and others, 2001. However, the error budget from velocity, ice thickness, accumulation and glacier geometry measurements means that the results are as yet inconclusive.

5. CONCLUSIONS

The advent of GPS has permitted the formulation of an accurate relationship between the ice-shelf thickness and its surface elevation, relative to the WGS84 ellipsoid. This relationship has allowed us to incorporate the ice-thickness data collected in 1981 and 1998 onto a common reference frame and construct a hydrostatic surface across the terminus of Pine Island Glacier. Points that have positive hydrostatic anomalies, that are coincident with a break in surface slope and that are upstream of the limit of flexing are locations on the glacier's coupling line. Given only the hydrostatic analysis, we would agree with the Crabtree and Doake (1982) placement of the grounding line. However, the InSAR-derived limit of flexing lies seaward of our coupling line and has a maximum separation of >28 km along the centre of the glacier. The region between the limit of flexing and the coupling line is the ice plain. Our accurate construction of a hydrostatic surface shows that the ice plain has elevations above buoyancy of <50 m. It is intriguing that this value of 50 m above flotation thickness is that found by Van der Veen (1996) to be the minimum necessary for the terminus of Columbia Glacier to remain steady. Perhaps there is a connection between the dynamics of tidewater glaciers and the behaviour of Antarctic grounding lines.

Rignot (1998) reported no increase in ice-stream velocity associated with the retreat of the limit of flexing inland and reasoned that the retreat was therefore well established before the 1992 interferogram. If our data interpretations are correct, we conclude that given the vulnerability of the ice plain to a small thickness change, the retreat must be a recent phenomenon, otherwise the present grounding line would lie upstream of the coupling line. The derived hydrostatic surface for the 1981 data invites speculation as to the location of the 1981 limit of flexing. The profile in Figure 4b suggests that in 1981 the limit of flexing would not have been far removed from its 1992 position. We therefore conclude that the retreat, observed by Rignot (1998), must have begun sometime after 1981 and probably only a few years prior to 1992.

Clearly Pine Island Glacier is a system in the process of change, and although we have reasoned that the retreat of the flexing limit could have begun within the last decade, we do not yet know if this is a short-term perturbation or a forerunner of more substantial retreat. It may even be part of a continuing adjustment to natural Holocene climate change. We echo the recommendation of all who have studied the region, that direct geophysical measurements

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are urgently required to understand and predict the dynamic behaviour of Pine Island Glacier, a region of potentially global significance.

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