A review of ice-sheet dynamics in the Pine Island Glacier basin, West Antarctica: hypotheses of instability vs. observations of change.

David G. Vaughan, Hugh F. J. Corr, Andrew M. Smith, Adrian Jenkins
British Antarctic Survey, Natural Environment Research Council

Charles R. Bentley, Mark D. Stenoien
University of Wisconsin

Stanley S. Jacobs
Lamont-Doherty Earth Observatory of Columbia University

Thomas B. Kellogg
University of Maine

Eric Rignot
Jet Propulsion Laboratories, National Aeronautical and Space Administration

Baerbel K. Lucchitta
U.S. Geological Survey

Abstract

The Pine Island Glacier ice-drainage basin has often been cited as the part of the West Antarctic ice sheet most prone to substantial retreat on human time-scales. Here we review the literature and present new analyses showing that this ice-drainage basin is glaciologically unusual, in particular; due to high precipitation rates near the coast Pine Island Glacier basin has the second highest balance flux of any extant ice stream or glacier; tributary ice streams flow at intermediate velocities through the interior of the basin and have no clear onset regions; the tributaries coalesce to form Pine Island Glacier which has characteristics of outlet glaciers (e.g. high driving stress) and of ice streams (e.g. shear margins bordering slow-moving ice); the glacier flows across a complex grounding zone into an ice shelf coming into contact with warm Circumpolar Deep Water which fuels the highest basal melt-rates yet measured beneath an ice shelf; the ice front position may have retreated within the past few millennia but during the last few decades it appears to have shifted around a mean position. Mass balance calculations of the ice-drainage basin as a whole show that there is currently no measurable imbalance, although there is evidence within the basin some specific areas are significantly out of balance; and finally, the grounding line has been shown to be retreating. The Pine Island Glacier basin is clearly important in the context of the future evolution of WAIS, because theoretically this basin seems to have a high potential for change, and because observations already show change occurring here. There is, however, no clear evidence to indicate that the ice sheet in the Pine Island Glacier basin has begun a phase of ongoing retreat or collapse.

1. Introduction

The West Antarctic ice sheet (WAIS, Figure 1) drains into the Southern Ocean by three main routes; through the ice streams on the Siple Coast into the Ross Ice Shelf; through the glaciers and ice streams feeding Ronne Ice Shelf, and through the glaciers which debouch, either directly or through small ice shelves, into the Bellingshausen and Amundsen seas. While the dynamics of the Siple Coast ice streams have been studied under the West Antarctic ice sheet Initiative, and those feeding Ronne Ice Shelf have been studied under the auspices of the Filchner-Ronne Ice Shelf Programme (FRISP), there has been no coordinated effort to understand the dynamics of glaciers feeding the Bellingshausen and Amundsen seas. Consequently, this area is seldom visited and its glaciology is poorly understood.

The largest glaciers in this sector are Pine Island Glacier and Thwaites Glacier, both transport ice from the interior of WAIS into the Amundsen Sea. *Vaughan and Bamber* [1998] calculated that, in terms of the mass of snow accumulating in their catchment basins, Pine Island and Thwaites glaciers are, respectively, the second and fifth most active basins in Antarctica, with Pine Island Glacier alone accounting for around 4% of the outflow from the entire Antarctic Ice Sheet. The ice-drainage basins that feed these glaciers rest on beds as much as 2500 m below sea level, perhaps the deepest in Antarctica, and some authors have suggested that this in itself implies a great potential for rapid collapse [e.g., *Hughes*, 1981; Fastook,1984; Thomas, 1984].

Together, Pine Island and Thwaites glaciers may be key to the future evolution of WAIS, but in this review, we concentrate on the ice-dynamics of the Pine Island Glacier basin, largely because in addition to theories of instability there is a growing body of observations of change and unsteady flow here. After some introductory notes we will consider each of the component parts of the basin in turn and then its interaction with sea into which it flows. In conclusion, we discuss how models which predicted that Pine Island Glacier might be particularly prone to collapse are currently being reviewed and if they are supported by recent observations of change in the ice sheet.

1.1 Introductory notes on nomenclature

1.1.1 Glaciers, ice streams and outlet glaciers

The terms *glacier*, *ice stream* and *outlet glacier* are often loosely applied in the scientific literature as well as the non-specialist press. Here we will use widely accepted definitions; *ice streams*, being areas of fast-moving ice sheet bounded by slower moving ice; *outlet glaciers*, being bounded by nunataks or mountain ranges [*Bentley*, 1987; *Swithinbank*, 1954]; and *glaciers*, being a generic term for a "mass of snow and ice continuously moving from higher to lower ground, or if afloat, spreading continuously..." [*Armstrong et al.*, 1973]. The distinction between ice streams and outlet glaciers "becomes rather hazy in practice" [*Bentley*, 1987], and is particularly acute in this case as Pine Island and Thwaites *glaciers* share most of the dynamical characteristics of pure ice streams and need not be considered as inherently different.

1.1.2 Floating portion of Pine Island Glacier

Hughes [1981] stated that neither Pine Island Glacier nor Thwaites Glacier were "buttressed by a confined and pinned ice shelf" and Stuiver et al. [1981], that "they are unimpeded by an ice shelf". At that time, the position of the grounding line of Pine Island Glacier was poorly mapped, and Pine Island Glacier was assumed to calve directly into Pine Island Bay, but airborne radar sounding soon revealed that the seaward -80-km of the glacier was indeed floating *[Crabtree and Doake*,1982]. The original idea that these glaciers are dynamically different from others in WAIS has, however, persisted and subsequent authors support the original notion, that Pine Island Glacier does not debouch through "an ice shelf" [*Kellogg & Kellogg*, 1987], "a substantial ice shelf" [*Jenkins et al.*, 1997], or "a large ice shelf" [*Rignot*, 1998]. Taking the widely accepted definition of an ice shelf, *A* "floating ice sheet of considerable thickness attached to a coast..." [*Armstrong et al.*, 1973] it is clear that the floating portion of Pine Island Glacier, together with the adjacent floating areas, do constitute an *ice shelf*. Whether the ice shelf is a significant dynamic control on the glacier is, however, still an open question.

1.1.3 Pine Island Bay

Strictly, *Pine Island Bay* is the bay (around 75 x 55 km) into which flows the ice from Pine Island Glacier [*Alberts*, 1981], although the term is sometimes applied to a rather wider area.

1.1.4 West Antarctic ice sheet

The West Antarctic ice sheet (WAIS) is not an officially-recognised placename. We follow the accepted usage, meaning the term to refer to the ice sheet that covers West Antarctica, but excluding the Antarctic Peninsula.

1.2 Introductory note on meteorology

Since the 1960s it has been widely recognised that the coastal portions of WAIS bordering on the Amundsen and Bellingshausen seas experience high precipitation rates compared with the rest of Antarctica [*Shimizu*, 1964], matched only on the Antarctic Peninsula and around the coast of Wilkes Land [*Giovinetto*, 1964]. A recent compilation of net surface mass balance derived from *in situ* measurements and satellite data [*Vaughan et al.*, 1999] is shown in Figure 2. It agrees broadly agrees with earlier estimates [eg. *Giovinetto & Bentley*, 1985], but shows an increased level of detail which may eventually be confirmed by *in situ* measurements.

Meteorologically, the high precipitation rate in this sector results from synoptic-scale cyclones which travel around the Antarctic in the circumpolar trough. The trough is deepest over the Amundsen Sea, and many synoptic-scale cyclones come ashore here, producing considerable precipitation in the coastal zone. The effect is seen in simple moisture transport calculations [*Bromwich*, 1988] and in precipitation fields derived from sophisticated General Circulation Models [*e.g., Connolley and King*,1996]. Precipitation may be particularly high during winter months when the circumpolar trough moves south and more cyclones track across the coastal region [*Jones & Simmonds*, 1993].

There are no meteorological stations in the Bellingshausen / Amundsen sea sector of WAIS, and consequently, direct measurements of decadal climate change (or stasis) have yet to be reported from either the Pine Island Glacier or Thwaites Glacier basins, although it is possible that surface elevation changes (see Section 2.7) do reflect recent anomalous precipitation rates [*Wingham et al.*, 1998]. Offshore, a reduction in sea-ice extent in both the Amundsen and Bellingshausen seas has been noted in all seasons over the two decades prior to 1995 [*Jacobs & Comiso,* 1997] and this perhaps reflects a change in surface temperatures.

2. The interior ice-drainage basin

The interior of ice-drainage basins are sometimes viewed as cisterns, which passively accumulate ice and then supply it to the glacier (or ice stream) at whatever rate the glacier can transport it away. In this section, however, we present evidence that flow in the Pine Island Glacier basin is far from homogeneous, there is no clear distinction between ice-sheet and glacier flow, and that flow in the basin may have a strong influence on the overall configuration and glacier activity.

2.1 Delineation of the ice-drainage basin

Field-measurements of surface elevation in the Pine Island Glacier basin are few, but altimetry from the ERS-1 satellite is available to 81.9ES, which includes the entire basin. This altimetry has been used to create several high-resolution Digital Elevation Models (DEMs) of Antarctica [e.g., *Bamber & Bindschadler*, 1997; *Legresy & Remy*, 1997, *Stenoien*, 1998; *Liu et al.*, in press].

Here we have used a 5-km-resolution ERS-1-derived DEM [*Bamber & Bindschadler*, 1997] (Figure 3) to delineate the Pine Island Glacier basin and its neighbours, and for comparison, a 200-m resolution DEM [*Liu et al.*, in press] to delineate the Pine Island Glacier basin alone (Figure 4). The methods were described in detail by *Vaughan et al.* [1999], but in summary, we identified segments of grounding line and then delineated the basins that feed them by tracing the line of steepest ascent inland as far as the ice divide. The procedure was limited to the grounded ice sheet as it assumes that ice-flow is parallel to the surface slope.

Table 1 shows the area of the Pine Island Glacier drainage basin measured from the above delineation using an equal area projection, together with earlier estimates shown for comparison. There was considerable disagreement between the early estimates and while the more recent ones that rely on ERS-1 data have reduced the uncertainty, there remains some residual uncertainty, presumably, resulting from the different methods of analysis. An average of the three most recent estimates probably constitutes the best estimate of the basin area and uncertainty (165 000 ± 7 000 km^2).

2.2 Shape of the catchment basin

The shape of the basin delineated in Section 2.1 (Figure 4) has broadly the same shape as the earlier delineations and this is worthy of discussion. It consists of two lobes, one immediately upstream of Pine Island Glacier and another, the *southern lobe*, feeding the first through a <100-km wide neck. A delineation of the catchment basins of the 70 largest glaciers in Antarctica, by a similar method, showed this configuration is most unusual [*Vaughan and Bamber*, 1999]. Generally, ice-drainage basins which drain through glaciers or ice streams, are uniformly convergent; only the Ice Stream C basin has a similar “necked” basin although in that basin the DEM was based on non-satellite data and so maybe less reliable. It is possible that *southern lobe* of the Pine Island Glacier basin indicates unsteady conditions in the basin, with this particular portion currently being transferred between catchment basins, or it may simply reflect unusual bed morphology (See Section 2.5).

2.3 Mass input and balance flux

Overlaying the basins for Pine Island Glacier derived in Section 2.2, on a grid representing the mean surface balance over Antarctica [*Vaughan et al.*, 1999], we have estimated the total rate of snow accumulation in the Pine Island Glacier basin (Table 1). This is the amount of ice that must leave the basin for mass balance to be maintained and so is termed the *balance flux*. The aggregate of the best three estimates gives a balance flux for Pine Island Glacier of (66 ± 4) Gt a^{-1}, where the uncertainty is derived from the spread of the results, but is consistent with that in area (± 4%) and accumulation (± 5%) [*Vaughan et al.*, 1999].

A similar process has been used to find the balance fluxes of the other major glaciers of Antarctica, and while many glaciers are fed by basins with larger areas, only one balance flux exceeds that of Pine Island Glacier; Totten Glacier, East Antarctica has a balance flux of around 75 Gt a^{-1} [*Vaughan and Bamber*, 1998]. Outside Antarctica, the most active glacier is Jacobshavns Isbrae, Greenland which supports about half this flux [*Bindschadler*, 1984].

2.4 Glacier tributaries

Two techniques employing satellite data and yielding wide coverage now allow us to identify areas where ice-flow is concentrated within the interior of the ice-drainage basin. The pattern that emerges is one of great complexity, with many glacier-tributaries coalescing to form the main glacier.

2.4.1 Interferometric SAR

Goldstein et al. [1993] used Synthetic Aperture Radar (SAR) data from the ERS-1 satellite to construct interferometric SAR (InSAR) images of ice flow in Antarctica. These showed only ice movement along the line of sight to the satellite, but the method has since been refined, to produce a 2-dimensional velocity-field for parts of the interior of the Pine Island Glacier basin [*Stenoien*,1998]. These velocity fields show a system of tributary ice streams which merge about 100 km above the grounding line to form a single unit of flow, which is Pine Island Glacier. These tributaries are identifiable several hundreds of km inland, much further inland than the set of arcuate crevasses at the point where the ice enters a more confined channel

[*Luchitta et al.*,1995; *Luchitta et al.*, 1994], which were previously thought to mark the onset of streaming flow.

Stenoien [1998] suggested that similar patterns of tributary ice streams have not been seen elsewhere, and that they are perhaps unique to Pine Island Glacier. There are, however, several pieces of evidence that suggest complex systems of tributaries do exist in other basins: bright margins in ERS-1 SAR data have shown that Evans Ice Stream, Antarctica forms at the confluence at least five tributaries [work reported in, *Jonas and Vaughan*, 1996], Radarsat data shows that Recovery Glacier, Antarctica has two tributaries which extend hundreds of km inland [*Jezek, et al.*, 1998], and flowlines in Landsat imagery show that Institute Ice Stream, Antarctica also has several tributaries [*Mantripp et al.*, 1996].

This presence of several tributaries in the interior drainage system may imply that Pine Island Glacier is unlikely to respond dramatically to changes in one locality. For example, if “water-piracy” [*Alley* et al., 1994] or a reduced supply of basal till were to shut-off one of the tributaries the others would probably be unaffected and the flux in Pine Island Glacier would remain fairly constant. Perhaps more significant is Stenoien’s observation that none of the Pine Island Glacier tributaries shows a clear onset region, but rather there is a gradual increase in ice speed down the length of each tributary. The lack of a clear onset region, suggests notions of a bi-stable state of glacier-flow i.e. fast or slow may be unrealistic, but rather that a progressive response to changing boundary conditions is possible.

2.4.2 Satellite Altimetry

We can also derive some understanding of the distribution of ice flow on the interior of the basin using the DEMs discussed in Section 2.1. It yields an understanding that is less quantitative than that from InSAR, but it does give complete coverage of the Pine Island Glacier basin. The method is to calculate the aspect (flow-direction) for each cell, then to assign a numerical value corresponding to the number of cells upstream of it, and whose accumulation will eventually flow through it. The technique is known as *flow-accumulation* and a grey-scale representation of this *flow-accumulation* grid gives an indication of where flow is localised within the basin (Figure 5).

Figure 5 shows almost the same set of tributaries to that drawn by Stenoien and in Figure 5 they are numbered using Stenoien's designation. One tributary (5) appears to drain much of the southern lobe of the catchment basin described in Section 2.2 and its presence is perhaps enough to explain the existence of the southern lobe.

The flow-accumulation analysis is similar to that employed by *Budd and Warner* [1996], and a close inspection of their velocity field (their Figure 1) indicates the presence of the three main flow streams upstream in the Pine Island Glacier basin.

Taken together, InSAR and flow-accumulation show that the interior of the Pine Island Glacier basin is complex with around 10 tributary ice streams coalescing to form a single glacier. None of these tributaries appears to have a well-defined onset zone and what controls their location and longevity remains to be determined, although radar data presented in Section 2.6 suggest that the control is through basal conditions. At present even sophisticated thermomechanical models of the area fail to reproduce them [e.g. *Payne*, 1999].

2.5 Subglacial topography

The bed topography of WAIS was first mapped in detail using a combination of traverse data, airborne sounding data and TWERLE balloon altimetry [*Jankowski and Drewry*, 1981]. While this study clearly delineated the major subglacial features of the area, the availability of new data prompts us to repeat the exercise.

Figure 6 shows a new compilation of bed topography beneath the grounded portion of the Pine Island Glacier basin. To create a grid of ice thickness we used traverse data [*Bentley* and

Ostenso, 1964; Behrendt, 1964; *Bentley and Chang*, 1971], airborne radar data [*Jankowski and Drewry*, 1981; unpublished data collected by British Antarctic Survey] and rock outcrop polygons which were used as an isopleth of zero ice thickness (Figure 6a). The resulting grid of ice thickness was subtracted from the ERS-1 derived DEM of surface elevation described in Section 2.1 to give bed topography Figure 6b.

Figure 6b. shows clearly the main features identified in earlier compilations; the Bentley Subglacial Trench; the Byrd Subglacial Basin which here reaches almost 2000 m below sea level, and between these depressions, the "sinuous ridge" described by *Jankowski and Drewry* [1981]. The Ellsworth Subglacial Highlands, are also well-defined. Despite significantly improved data coverage in the Pine Island Glacier basin, Figure 6b shows no new substantial features except a trough 1000 m below sea level, in which the Pine Island Glacier and its main tributaries flow.

2.6 Driving stress

The *driving stress* in an ice sheet is generally calculated from the surface slope and ice thickness according to a simple relation [*Paterson*, 1994; page 241]. Here we have calculated the driving stress for the region using the ERS-1-derived DEM and the ice thickness grid described above (Figure 7). The calculated driving stress is negligible near the ice divide where the surface slopes are low; intermediate (50-110 kPa) on the (presumably) slow-moving areas between the ice divides and the tributary glaciers, low (<50 kPa) on the tributary glaciers, but rises to >110 kPa along the main trunk of Pine Island Glacier.

The driving stresses shown in Figure 7 compare well with data collected during an airborne sortie in 1998 from the disused Siple Station (75E 54' S 84E 30' W) to the ice front of Pine Island Glacier (the flight-track is shown in Figure 6a). Ice-penetrating radar data from this sortie show that the margin of the main tributary to Pine Island Glacier is marked by a downwards step in the bed elevation and a change to a smoother ice-base reflection which has as "ice-shelf-like" character (Figure 8). The driving stress calculated from the along-track topography and smoothed over a 10-km window (Figure 8), has four distinct zones; the interior of the basin where driving stress is 50-75 kPa, the main tributary glacier where it is around 30 kPa and on the main trunk of Pine Island Glacier where the driving stress rises significantly to over 100 kPa, once floating the driving stress falls to less than 10 kPa. The marked change in roughness over the tributary presumably indicates that the location of the tributary is controlled by underlying geological constraints, and the low driving stress, that it flows over a well-lubricated bed.

Consideration of the pattern of driving stresses suggest that the Pine Island Glacier basin is dynamically different from the idealised ice-stream basin. Here we find a cold-based ice sheet, which feeds a set of wet-based, lubricated tributaries which have driving stresses similar to some ice streams, these merge to form Pine Island Glacier whose flux and basal conditions are such that high driving stress (>100 kPa) must be generated to maintain the balance flux. Thus, in terms of driving stress, Pine Island Glacier may be more akin to East Antarctic outlet glaciers than to West Antarctic ice streams [*Bentley*, 1987], while its tributaries maybe dynamically quite similar to ice streams.

2.7 Surface elevation change

ERS-1 Satellite altimetry for the period 1992-1996 were analysed for the evidence of surface elevation change [*Wingham et al.*,1998]. These data, which covered most of the interior of the Antarctic Ice Sheet north 82ES; showed only one spatially-coherent surface elevation change during this period, a fall (-11.7 ± 1.0 cm per year) in the Pine Island Glacier-Thwaites Glacier basin. Wingham et al. indicated that the change was centred and most significant over the Thwaites Glacier basin (see their Figure 2), rather than the Pine Island Glacier basin, but the trend did appear to extend across both. The simplest interpretation is that the surface lowering resulted from a change in surface mass balance, but a change in the glacier flux due to glacier surging or grounding line retreat might also be the cause. However, in either case being obtained from only 4 years of data, the result gives little indication of future behaviour and it is hoped that more detailed analysis of the ERS-1 altimetry will refine the pattern of change. In 2001 the launch of NASA's Geoscience Laser Altimeter System (GLAS) will allow similar measurements to be made even in the coastal margin of Antarctica.

3. The Glacier

Until the discovery of the network of ice- stream tributaries in the basin [*Stenoien*, 1998; Section 2.4], Pine Island Glacier was generally considered to extent only around 70 km above the grounding line to where the ice is first channelled into parallel flow (Figure 9). Perhaps this is still a useful definition, since it draws some distinction between the tributaries and the main trunk of the glacier, approximately marking the increase in driving stress mentioned in Section 2.6. Thus defined Pine Island Glacier is bounded to the north by nunataks in the Hudson Mountains, and in the south by ice sheet without nunataks but which is presumably slow-moving.

3.1 Surface features

Surface features on Pine Island Glacier revealed by Landsat and SAR imagery have been shown by various authors and are reproduced in Figure 9. Flowlines of the type discussed by *Whillans and Merry* [1993] show considerable convergence at the head of the glacier, around the zone of arcuate "crevasses" revealed by ERS-SAR images (shown in Figure 9 and in greater detail by *Luchitta et al.* [1995]) which presumably mark a zone of longitudinal extension. Luchitta et al. noted that these "crevasses" had not been previously described and are not shown by visible images, and concluded that they may be covered by a layer of snow.

Below the zone of convergence and arcuate crevasses, flowlines on the glacier are roughly parallel. The surface of the glacier has smooth undulations that are typical of fast-moving glaciers.

3.2 Velocity

Luchitta and co-authors [*Luchitta et al.*, 1995; *Luchitta et al.*, 1994; *Ferrigno et al.*, 1993] have measured the ice velocity on Pine Island Glacier using sequential SAR images acquired by the ERS-1 satellite. They found that on the main trunk of the grounded portion the centre-line speed ranged from about 1 km a^{-1} near the arcuate crevasses, to 1.5 km a^{-1} at the grounding line identified by *Crabtree and Doake* [1982]. The flow speed then rose rapidly to 2.5 km a^{-1} between that grounding line and the one identified by *Rignot* [1998], and then remained approximately constant to the ice front. Their velocity measurements were generally higher than earlier estimates [*Kellogg & Kellogg*, 1987; *Lindstrom & Tyler*, 1984; *Crabtree and Doake,* 1982; *Williams et al.*, 1982] although the data are not adequate to determine if there has been any aceleration.

3.3 Grounding line

The grounding line of Pine Island Glacier is not easily discernable either on the Landsat or ERS-1 SAR imagery in Figure 9, but its position was determined by hydrostatic calculations based on airborne data [*Crabtree and Doake*, 1982] (Figure 9). This hydrostatic condition downstream of this grounding line was, however not entirely clear and *Thomas* [1984] argued that there a zone of partial grounding might exist for 30 km downstream. This downstream position has now been confirmed, using InSAR to detect the limit of tidal flexing [*Rignot*, 1998]; this analysis also showed that the limit of flexure extends seaward (around 15 km) near the centre of the glacier and is re-entrant on either side (Figure 9). This pattern is possibly an indication of a bedrock obstruction similar to that known to underlie Rutford Ice Stream [*Stephenson*, 1984] or it might be solely due to the glacier being thicker close to its centre line.

The position of the limit of flexure was measured at five epochs between 1992 and 1996, and showed that it moved during this period [*Rignot*, 1998]. The simplest interpretation is that between 1992 and 1994 there was a retreat in the position of the grounding line in the centre of the glacier that averaged 1.2 ± 0.3 km a^{-1}. The pattern of change within the re-entrant parts is not so clear. Rignot calculated that this might be caused by a thinning rate of 3.5 ± 0.9 m a^{-1}, which is a small fraction of the basal meltrates in this area (see Section 4.2) and thus could result from a relatively small change in the oceanographic conditions, but it could also be caused by a thinning of the glacier upstream of the grounding line.

4. The Ice Shelf

As noted above the Pine Island Glacier debouches into an ice shelf comprising the -80-km floating portion of Pine Island Glacier and the slow-moving floating ice sheet that surrounds it. The *floating portion of Pine Island Glacier* is easily identified by plentiful flowlines generated near at the grounding line which continue to the ice front and show little divergence.

4.1 Surface features

Kellogg et al. [1985] found that dense (650 kg m^{-3}) “well-sintered” firn predominated at the surface of the floating portion of Pine Island Glacier close to the ice front. They interpreted this as an indication that strong katabatic winds producing net sublimation from the ice surface. The extent, persistence or magnitude of this negative net surface mass balance is, however, not well-established.

Landsat and SAR images of the floating portion of Pine Island Glacier (Figure 9) show crevasses formed at the grounding line, moving in plumes to the ice front. Much of the southern side of the floating portion of the glacier is covered by periodic transverse surface undulations visible on the Landsat images. These also form in plumes emanating from the grounding line and dissipating towards the ice front. Airborne radar sounding data show that these undulations have an amplitude of around 20 m, a horizontal wavelength of around 2.5 km and are hydrostatically compensated by ice thickness changes of -190 m. The ice flow velocity measured on this section of ice shelf was 2.3 - 2.6 km a^{-1} [*Luchitta et al.*, 1997] which suggests that one undulation is produced each year, although the mechanism that causes them is uncertain.

A former feature of the floating portion of Pine Island Glacier, not previously discussed but clearly visible in the image shown in Figure 9, was a raft of thicker ice embedded in the ice shelf. In later images, the raft was seen to have been advected downstream at approximately the same speed as the ice shelf flow. The origin and significance of this raft is unclear and it is no longer available for investigation as it would have calved from the ice shelf in the late-1980s. Similar features have been noted in grounded ice streams [e.g., *Whillans et al.*, 1993] and possibly in floating ice shelves [*Cassasa et al.*, 1991].

4.2 Basal melting

In early 1994, the research ship *Nathaniel B. Palmer* entered Pine Island Bay and conducted an oceanographic survey of the area which has led to three studies revealing a most unusual oceanographic regime:

- *Jacobs et al.* [1996] used oceanographic measurements and a “salt-box” calculation to infer that the mean basal melt rate beneath the floating portion of Pine Island Glacier was 10-12 m a^{-1}, 5 times the highest rate previously measured on George VI Ice Shelf [*Bishop and Walton*, 1981]. They concluded that these high melt-rates were fuelled by relatively warm Circumpolar Deep Water (CDW) flooding this portion of the continental shelf.
- *Jenkins et al.* [1997] determined that the extent of the ice shelf showed no persistent trend in the period 1973-1994 (in Figure 10 we extend this series to 1966-1998). Combining this with a flux calculation across the grounding line and with the assumption that it is a steady-state system, they calculated a mean basal melt rate over the ice shelf of 12 ± 3 m a^{-1}.
- *Hellmer et al.* [1998] used analyses of dissolved oxygen and oxygen isotopes to confirm the strong melt-water signal in the outflow and applied a thermohaline model,

which suggested that in some areas the basal melt-rate is twice the mean value and that temporal variations in the temperature of the inflowing CDW could cause substantial changes in the basal melt rate.

Rapid melting from beneath the floating portion of Pine Island Glacier was later confirmed using satellite measurements of mass balance [*Rignot*, 1998] giving a mean melt-rate of (24±4) m a^{-1} increasing to (50±10) m a^{-1} close to the grounding line. These melt-rates are larger than those calculated by *Jenkins et al.* [1997] due to a new position for the grounding line constrained by InSAR observations of tidal flexing - using this grounding line position we should increase the estimate of mean melt-rate produced by Jenkins et al. to around 17 m a^{-1}, and that from Jacobs et al. by a similar amount.

These observations of a thick ice shelf coming in contact with relatively warm Circumpolar Deep Water has led to speculation that ice - ocean interactions in Pine Island Bay may be similar to those prevalent during the Last Glacial Maximum (LGM) when much of WAIS probably extended out to the edge of the continental shelf and came into contact with similarly warm water [e.g. *Jenkins et al.*, 1997]. In Pine Island Bay we now know that such conditions generate high sub-ice shelf meltrates, and we may infer that these may have also been prevalent during glacial periods, severely limiting the size of ice shelves surrounding the ice sheet. In addition, the results highlight the importance of oceanographic conditions as a significant control on the present configuration of WAIS.

4.3 Ice front stability

The icefront position of Pine Island Glacier has been reconstructed by several authors using several sources of data and covering various periods. Figure 10 shows the position of the icefront sporadically since 1966. Taken at face value the pattern suggests a retreat of around 10 km between 1966 and 1973, followed by a period of general stability, and readvance in recent years. This simple interpretation should, however, be qualified on two counts. a) The presence of an iceberg just off the icefront on the sketch map drawn from 1966 aerial photography, indicates that a significant calving had recently occurred, and that prior to this the icefront was even further advanced than the most extreme icefront shown. The same is true for the 1973 image. b) We should bear in mind that the rate of icefront advance (>2.5 km a^{-1}) allows for fluctuations within the intervals between the observations large enough to exceed the extreme positions shown in Figure 10.

Kellogg and Kellogg [1987] have suggested that the Thwaites Iceberg Tongue was not actually formed from calving of Thwaites Glacier Tongue but might have calved from Pine Island Glacier. This is, however, unlikely, as a profusion of surface transverse lineations on the iceberg tongue matched well with the transverse lineations on Thwaites Glacier, but did not match the longitudinal lineations that predominate on Pine Island Glacier [*Ferrigno et al.*, 1993].

In summary, there are insufficient data available to discern a decadal trend in the ice front position of Pine Island Glacier, although it is probable that some retreat (- 10 km) has occurred in the last 30 years. (Longer-term retreat of the ice shelf front is discussed in Section 5).

5. The Marine environment

5.1 Retreat of ice in Pine Island Bay

Seabed sediments provide a valuable source data on glacier retreat in Pine Island Bay. Results from four cores from Pine Island Bay and 19 from the outer continental shelf and eastern Amundsen Sea have been presented [*Anderson & Myers*, 1981; *Kellogg & Kellogg,* 1987; *Kellogg et al.*, 1987], but their value is limited because they contain very little material suitable for radiocarbon dating. The radiocarbon dates that do exist, only poorly constrain the retreat of the ice sheet in Pine Island Bay, to any time during the last few millennia. Prior to this the ice sheet may have occupied the entire Pine Island Bay, and perhaps butted against the Thwaites Glacier Tongue to form an extensive ice shelf or ice sheet.

Kellogg and Kellogg [1987] suggested the retreat was very recent, with Pine Island Bay being filled with grounded ice only 100 years ago, but this interpretation relied heavily on an ongoing retreat rate of - 0.8 km per year inferred from aerial photography acquired in 1966 and Landsat imagery acquired in 1973. The variable position of the ice front shown in Figure 10 now casts doubt on such an extrapolation.

5.2 Ice Extent at the Last Glacial Maximum

Sediment cores have been collected on the outer continental shelf, near 110°W [*Anderson & Myers,* 1981], and north of Thurston Island between 100°W and 102°W [*Kellogg & Kellogg* 1987; *Kellogg et al.*, 1987]. The cores show a thin (0-15 cm) upper layer of sandy mud, probably of Holocene age, containing common planktonic and calcareous benthic Foraminifera. Diatoms are relatively rare in this layer despite high abundances in the surface water. A compact, poorly-sorted diamicton underlies the sandy mud. This diamicton is generally more than 2.3 m thick, and probably represents deposition beneath grounded ice that extended to the continental shelf break.

If we assume that the diamicton is a remnant of subglacial basal till, then grounded ice probably remained over most of the Amundsen Sea continental shelf until relatively late in the Holocene. The postglacial sediments on the outer shelf are much thinner in Pine Island Bay than on the outer shelf in the Ross Sea [*Kellogg et al.*, 1979; *Domack et al.*, in press; *Licht et al.,* 1999; *Shipp et al.*, in press] which, given similar deposition rates, might suggest an earlier deglaciation in the Ross Sea than in the Amundsen Sea. Finally, if we assume the post-glacial deposition rates are similar to current measured rates e.g., 10-35 cm a^{-1} beyond coastal Alaskan glaciers [*Molnia & Carlson*, 1999] and ~10 cm a^{-1} in Antarctic Peninsula fjords [*Domack and McClennen,* 1996], then we can conclude that deglaciation occurred at most a few thousand years ago.

6. Mass balance

A major preoccupation of glaciologists, particularly those considering Pine Island Glacier has been comparing accumulation and output fluxes to establish the overall mass balance of the ice-drainage basin. Comparisons of balance flux (Table 1) and grounding line flux (Table 2) for Pine Island Glacier show a progression toward reduced uncertainty but significant uncertainty remains. The best estimate of overall mass balance, (-2.4 ± 4 Gt a^{-1}), which is not significantly different from zero, probably results from the integrated basin accumulation rate (balance flux) calculated in this study and the grounding line flux calculated by *Rignot* [1998]. Given a catchment basin area of -170 000 km^2, this is equivalent to a lowering of surface elevation in the range 1.5 - 3 cm a^{-1}, depending on the density of the layers being lost.

Rignot [1998] , however, determined the ice thickness by inverting the ice surface elevation at the limit of flexing using a hydrostatic condition and there is evidence that the limit of flexing is often thousands of m upstream of the hydrostatic limit. One example is on Rutford Ice Stream where along the centre line the limit of flexing is 2 km upstream of the hydrostatic point and the surface is 50 m above the hydrostatic condition [*Vaughan*, 1994; *Smith*, 1991]. If a similar situation applies on Pine Island Glacier then the ice flux then the grounding line flux may have been over-estimated by Rignot.

The mass balance within two subdivisions of the basin has also been calculated [*Stenoien*, 1998], the north-eastern slopes of the main basin apparently having a positive mass balance of (6.4 ± 3.7) Gt a^{-1} and the south-western slope a negative mass balance of (-7.7 ± 4.7) Gt a^{-1}. These values were calculated using a hand-drawn map of mean surface mass balance, but no significant difference in his result was found when we repeated the calculation using an updated map of surface mass balance [*Vaughan et al.*,1999]. Spread evenly across the areas these imbalances represent changes in surface elevation of (23 ± 14) cm a^{-1} and (-51 ± 31) cm a^{-1} , respectively.

There is thus a contradiction between mass balance calculations and measurements of change in surface elevation [*Wingham et al.*,1998; see Section 2.7] which seems to imply one of three possibilities; a) substantial changes in the density-depth relation in the snow in the period 1992-1996, b) unusually low precipitation accumulation in the period 1992-1996 or, c) one of the analyses is substantially in error.

7. Discussion

Weertman [1974] developed a simple model of the behaviour of the stability of the junction between ice sheet and ice shelf (i.e., the grounding line). He concluded that where the bed sloped down inland the grounding line would not be stable, except when it was at the continental shelf edge. This led to the widely cited hypothesis that the current configuration of West Antarctic Ice Sheet (WAIS) was inherently unstable and might collapse, especially if the restraint exerted by ice shelves was removed. Some authors have even suggested that climate change might begin the process by removing the ice shelves [*e.g., Mercer*, 1978], although this hypothesis was never very widely supported within the glaciological community. *Hughes* [1980] suggested that Pine Island Glacier was the "weak underbelly of the West Antarctic Ice Sheet" and was most vulnerable to this process, being grounded on rock well below sea level and being bounded by the lowest ice divides in WAIS and these results were later confirmed by modelling studies [*Fastook*, 1984].

On theoretical grounds, Pine Island Glacier has thus occupied a central position in the arguments regarding the stability of WAIS, but recently, the theories of marine ice-sheet instability have been questioned. Although it seemed self-evident, Weertman's assumption that ice thickness must be continuous at the grounding line is now seen as simplistic. Ice streams and outlet glaciers, may act as intermediates, buffers, between grounded ice sheets and ice shelves, allowing the strict boundary conditions that result from a sharp transition to be relaxed [*Hindmarsh*, 1993], (see [*Bentley*, 1998] for a summary of this argument). This allows that an infinite number of grounding line positions are compatible with *neutral equilibria* in the ice sheet and the grounding line transition maybe relatively unimportant in controlling the configuration of the ice sheet. Furthermore, other theories suggest that ice sheet collapse might occur, but as a result of instabilities within the ice-drainage basins, rather than from a grounding line instability, and only as an asynchronous response to millennial climate forcing [*MacAyeal*, 1992].

Nevertheless, the hypotheses of marine ice-sheet instability have led to many studies seeking direct observational evidence of change in WAIS and such observations of change are now established in the literature. We have highlighted several observations of change in this study from the Pine Island Glacier drainage basin but none make a strong case for ongoing basin-scale ice sheet change or readjustment, and certainly, none suggest that the ice sheet in this area has entered a phase of significant collapse or retreat;

- the grounding line has retreated a few km along the centre of the glacier over a period of a few years [*Rignot*, 1998], but we really do not understand if this has significantly changed the force-balance of the glacier,
- the surface elevation in the ice-drainage basin dropped over four years [*Wingham et al.*, 1998], but we are uncertain if this is the result of changing precipitation or changing ice flow,
- locally the ice sheet may be out of balance [*Stenoien*, 1998] but these measurements do not seem to concur with changes in surface elevation,
- the ice sheet or ice shelf in Pine Island Bay may have retreated significantly in the last few millennia [*Kellogg and Kellogg*, 1987], but more radio-carbon dating and better discrimination of ice-berg-rafted, sub-ice-shelf and sub-ice-sheet sediments we cannot be confident,
- the ice – ocean interaction appears to be unusually dynamic in Pine Island Bay due to intrusion of CDW onto the continental shelf in this area, we cannot, however, be sure of the past or future longevity of this intrusion,
- and the shape of the ice-drainage basin is unusual and perhaps surprising but we can only hypothesize that its shape is changing.

At a time when the paradigm of marine ice sheet instability is being questioned, our observations of change, perhaps, appear ambiguous and inconclusive -- interpreting as precursors of collapse is certainly inadequate. This shortcoming suggests that we should prepare field and remote sensing experiments that will allow us to determine the causes of changes in the ice sheet, as only an understanding the root cause of change will lead us to a sound foundation for predicting future behaviour.

Acknowledgements

References

Alberts, F.G., *Geographic names of the Antarctic*. National Science Foundation, Washington. 1981

Alley, R. B., S. Anandakrishnan, C.R. Bentley and N. Lord. A water-piracy hypothesis for the stagnation of Ice Stream C, Antarctica. *Ann. Glaciol.,* 20, 187-194, 1994.

Anderson, J.B. and M.C. Myers. *USGSC Glacier Deep Freeze 81 Expedition to the Amundsen Sea and Bransfield Strait*. Antarc. J. U.S., 16, 5, 1981

Armstrong, T., B. Roberts and C.W.M. Swithinbank. *Illustrated glossary of snow and ice.* Scott Polar Research Institute, Cambridge, 1973.

Bamber, J. L. and R. A. Bindschadler, An improved elevation dataset for climate and ice-sheet modelling: validation with satellite imagery, *Ann. Glaciol.,* 25, 1997.

Bentley, C. R., Antarctic ice streams: a review, *J. Geophys. Res.,* 92, 8843-8858, 1987.

Bentley, C. R. and M. B. Giovinetto, Mass balance of Antarctica and sea level change, *Geophysical Institute and Centre For Global Change and Arctic System Science, University of Alaska.,* 481-488, 1991.

Bentley, C. R., Rapid sea-level rise from a West Antarctic ice-sheet collapse: a short-term perspective, *J. Glaciol.,* 44, 157-163, 1998.

Bentley, C.R. and N.A. Ostenso, Glacial and subglacial topography of West Antarctica. *J. Glaciol.*, 3, 29, 882-911. 1961.

Bentley, C.R. and F.-K. Chang, Geophysical exploration in Marie Byrd Land, Antarctica. Ant. Res. Ser., 16. 1-38, 1971.

Behrendt, J.C., Distribution of narrow-width magnetic anomalies in Antarctica. *Science*, 144, 3621, 993-994, 1964.

Bindschadler, R., Jakobshavn Glacier drainage basin: a balance assessment, *J. Geophys. Res.,* 89, 2066-2072, 1984.

Bishop, J.F. and J.L.W. Walton, Bottom melting under George VI Ice Sheet, Antarctica. *J. Glaciol.*, 27, 429-447, 1981.

Bromwich, D. Snowfall in high southern latitudes. *Rev. Geophys.*, 26, 149-168, 1988.

Budd, W. F. and R. C. Warner, A computer scheme for rapid calculations of balance-flux distributions, Ann. Glaciol., 23, 21-27, 1996.

Cassasa, G., K.C. Jezek,, J. Turner and I.M. Whillans. Relict flow stripes on the Ross Ice Shelf. *Ann. Glaciol.*, 15, 132-138, 1991

Connolley, W. M. and J. C. King, A modeling and observational study of East Antarctic surface mass balance, *J. Geophys. Res.,* 101, 1335-1343, 1996.

Crabtree, R. D. and C. S. M. Doake, Pine Island Glacier and its drainage basin: results from radio echo- sounding, *Ann. Glaciol.,* 3, 65-70, 1982.

Domack, E. W. and C. E. McClennen, Accumulation of glacial marine sediments in fjords of the Antarctic Peninsula and their use as late Holocene palaeoenvironmental indicators,*Antarc. Res. Ser.,* 70, 135-154, 1996.

Domack, E.W., E.A. Jacobson, S.A. Shipp, et al. Sedimentologic and stratigraphic signature of the late Pliestocene/Holocene fluctuation of the West Antarctic Ice Sheet in the Ross Sea: A new perspective part 2. *Bulletin, Geological Society of America*. in press

Fastook, J.L., West Antarctica, the sea-level controlled marine instability: past and future. In: J.E. Hansen and T. Takahashi (Eds), Climate Processes and Climate Sensitivity. Geophys. Mono. AGU, 29, 275-287, 1984.

Ferrigno, J. G., B. K. Luchitta, K. F. Mullins, A. L. Allison, R. J. Allen, and W. G. Gould, Velocity measurements and changes in the position of Thwaites Glacier/iceberg tongue from aerial photography, Landsat images and NOAA AVHRR data, *Ann. Glaciol.,* 17, 239-244, 1993.

Giovinetto, M. The drainage systems of Antarctica: Accumulation. Antarct. Res. Ser., 2, 127-155, 1964.

Giovinetto, M. B. and C. R. Bentley, Surface balance in ice drainage systems of Antarctica, *Antarc.J.US.,* 20, 6-13, 1985.

Goldstein, M., H. Engelhardt, B. Kamb, and R. M. Frolich, Satellite radar interferometry for monitoring ice sheet motion: application to an Antarctic ice stream, *Science,* 262, 1525-1530, 1993.

Hellmer, H. H., S. S. Jacobs, and A. Jenkins, Ocean erosion of a floating Antarctic Glacier in the Amundsen Sea, *Antarc. Res. Ser.,* 75, 83-100, 1998.

Hindmarsh, R. C. A., Qualitative dynamics of marine ice sheets, *NATO ASI.Series.,* I 12, 68-99, 1993.

Hughes, T. J., The weak underbelly of the West Antarctic Ice Sheet, *J. Glaciol.,* 27, 97, 518-525, 1980.

Jacobs, S. S., H. H. Hellmer, and A. Jenkins, Antarctic ice sheet melting in the Southeast Pacific, *Geophys. Res. Lett.,* 23, 957-960, 1996.

Jacobs, S. S. and J. C. Comiso, Climate variability in the Amundsen and Bellingshausen Seas, *J. Clim.,* 10, 697-709, 1997.

Jankowski, E. J. and D. J. Drewry, The structure of West Antarctica from geophysical studies, *Nature,* 291, 17-21, 1981.

Jenkins, A., D. G. Vaughan, S. S. Jacobs, H. H. Hellmer, and J. R. Keys, Glaciological and oceanographic evidence of high melt rates beneath Pine island glacier, west, *J. Glaciol.,* 43, 114-121, 1997.

Jezek, K.C., H.G. Sohn and K.F. Noltimier. The Radarsat Antarctic Mapping Project. IGARSS '98 - 1998 International Geoscience and Remote Sensing Symposium, Proceedings, Vol 1-5, Ch.888, 2462-2464, 1998

Jonas, M. and D. G. Vaughan, ERS-1 SAR mosaic of Filchner-Ronne-Schelfeis, *Filchner Ronne Ice Shelf Programme Reports,* 10, 47-49, 1996.

Jones D. A. and I. Simmonds, A climatology of Southern Hemisphere extratropical cyclones, *Climate Dynamics*, 9, 135-145, 1993.

Kellogg, T. B., R. S. Truesdale, and L. E. Osterman, Late quaternary extent of the West Antarctic Ice Sheet: new evidence from Ross Sea cores, *Geology,* 7, 249-253, 1979.

Kellogg, T. B., D. E. Kellogg, and T. J. Hughes, Amundsen Sea sediment coring, *Antarc. J. the US,* 20, 79-81, 1985.

Kellogg, T. B. and D. E. Kellogg, Recent glacial history and rapid ice stream retreat in the Amundsen Sea., *J. Geophys. Res.,* 92, 8859-8864, 1987.

Kellogg T. B., D. E. Kellogg, E. D. Waddington, and J. S. Walder., Late Quaternary deglaciation of the Amundsen Sea: implications for ice sheet modelling, *IAHS Pub.*, The physical basis of ice sheet modelling. Proceedings of an International Symposium held during the 19th General Assembly of the IUGG at Vancouver, British Columbia, Canada. 9-22 August 1987., 349-357, 1987.

Legresy, B. and F. Remy, Altimetric observations of surface characteristics of the Antarctic Ice Sheet, *J. Glaciol.,* 43, 265-275, 1997.

Licht, K. J., N. W. Dunbar, J. T. Andrews, and A. E. Jennings, Distinguishing subglacial till and glacial marine diamictons in the western Ross Sea, Antarctica: implications for a last glacial maximum grounding line, *Bulletin of the Geological Society of America,* 111, 91-103, 1999.

Lindstrom, D. and T. J. Hughes, Downdraw of the Pine Island Bay drainage basins of the west antarctic ice sheet, *Antarc. J. US.,* 19, 56-58, 1984.

Lindstrom, D. and D. Tyler, Preliminary results of Pine Island and Thwaites Glaciers Study, *Antarctic Journal of the United States,* 19, 53-55, 1984.

Liu, H., K. Jezek, and B. Li, Development of Antarctic DEM by integrating cartographic and remotely sensed data: A GIS-based approach., *J. Geophys. Res.,* in press.

Lingle, C.S. The West Antarctic ice sheet may be necessarily characterized by discharge through fast-flowing ice streams. Paper presented at the Chapman Conference on Fast Glacier Flow: Ice Streams, Surging and Tidewater Glaciers, AGU, Vancouver, B.C. May 5-8, 1986.

Luchitta, B., C. Rosanova, and K. Mullins, Velocities of Pine Island and Thwaites Glacier, West Antarctica, from ERS-1 SAR images, *Ann. Glaciol.,* 21, 277-283, 1995.

Luchitta B. K., C.E. Smith, J. Bowell, and K.F. Mullins, Velocities and mass balance of Pine Island Glacier, West Antarctica, derived from ERS1-SAR. ESA Publication SP-361, 2nd ERS-1 Symposium, Space at the Service of Our Environment, Hamburg, Germany, 11-14

October 1993 Proceedings, 147-151, 1994.

Luchitta B.K., and C.E. Rosanova, Velocities of Pine Island and Thwaites Glaciers, West Antarctica, from ERS1-SAR images. ESA Publication SP-414, 3rd ERS-1 Symposium, Space at the Service of Our Environment, Florence, Italy, 17-21 March, 1997. 819-824, 1997.

MacAyeal, D. R., Irregular oscillations on the West Antarctic Ice Sheet, *Nature,* 359, 29-32, 1992.

Mantripp, D.R., J. Sievers, H. Bennat, C.S.M. Doake, K. Heidland, J. Idhe, M. Jonas, B. Reidel, A.V. Robinson, R. Scharroo, H.W. Shenke, U. Shirmer, F. Stefani, D.G. Vaughan and D. J. Wingham, 1996. Topographic map (satellite image map), Filchner-Ronne-Shelfeis (2nd Edition). Map at 1 : 2 000 000 Published by Institut für Angewandte Geodäsie, Frankfurt am Main, Germany. 1996.

Mercer, West Antarctic ice sheet and CO2 greenhouse effect: a threat of disaster, *Nature,* 271, 321-325, 1978.

Molnia, B. F. and P. R. Carlson, Surface sedimentary units of northern Gulf of Alaska continental shelf., *Bulletin of the American Association of Petroleum Geologists,* 62, 633-643, 1999.

Paterson, W.S.B., The physics of glaciers. Elsevier Science, Oxford. 480 pages. 1994.

Payne, A.J., A thermomechanical model of ice flow in West Antarctica. *Climate Dynamics*, 15, 115-125, 1999.

Rignot, E. J., Fast recession of a West Antarctic Glacier. *Science,* 281, 549-551, 1998.

SCAR, 1993. Antarctic digital database user's guide and reference manual. Scientific Committee on Antarctic Research, Cambridge. Xi+156pp. (This manual accompanies a CD-ROM.)

Shimizu, H. Glaciological studies in West Antarctica. Antarct. Res. Ser., 2, 37-64, 1964.

Shipp, S.A., J.B. Anderson and E.W. Domack. Seismic signature of the late Pleistocene fluctuation of the West Antarctic Ice Sheet system in the Ross Sea: A new perspective, Part 1. *Bulletin, Geological Society of America*, in press.

Smith, A.M., The use of tiltmeters to study the dynamics of Antarctic ice-shelf grounding lines. *J. Glaciol.*, 37, 125,51-58, 1991.

SPRITE Group and C. G. Boyer, The southern rim of the Pacific Ocean: preliminary geologic report of the Amundsen Sea - Bellingshausen Sea cruise of the Polar Sea 12 February - 21 March 1992, *Ant. J. US,* 27, 11-14, 1992.

Stenoien M. D., Interferometric SAR observations of the Pine Island Glacier catchment area, Thesis for the Degree of Ph.D., University of Wisconsin - Madison. 1-127, 1998.

Stuiver, M., G. H. Denton, T.J. Hughes and J.L. Fastook, History of the marine ice sheet in West Antarctica during the last glaciation: a working hypothesis. In The Last Great Ice Sheets. Ed G.H. Denton and T.J. Hughes. Wiley-Interscience, New York. 1981.

Stephenson, S. N., Glacier flexure and th position of grounding lines: measurements by tiltmeter on Rutford Ice Stream, Antarctica, Ann. Glaciol., 5, 165-169, 1984.

Swithinbank C. W. M., Ice streams, *Polar Record*, 7, 148, 185-186, 1954.

Thomas, R. H., Ice sheet margins and ice shelves. In: J.E. Hansen and T. Takahashi (Eds), Climate Processes and Climate Sensitivity. Geophys. Mono. AGU, 29, 265-274, 1984.

USGS, Antarctica sketch map, Thurston Island - Jones Mountains, 1:500,000, 1993

Vaughan, D. G., Investigating tidal flexure on an ice shelf using kinematic GPS. *Ann. Glaciol*, 20, 372-376, 1994.

Vaughan, D. G., J. L. Bamber, M. Giovinetto, J. Russell, and A. P. R. Cooper, Reassessment of net surface mass balance in Antarctica, *J. Clim.*, 12 (4), 933-946,1999.

Vaughan D. G. and J. L. Bamber., Drainage basin analysis and improved calculation of balance fluxes for West Antarctic ice streams and glaciers, Abstract for AGU Chapman Conference on the West Antarctic Ice Sheet, Orono, Sept. 1998.

Weertman, J., Stability of the junction of an ice sheet and an ice shelf, *J. Glaciol.,* 13, 3-11, 1974.

Whillans, I. M. and C.J. Merry, Ice-flow features in the Ice Stream-B, Antarctica, revealed by SPOT HRV imagery. *J. Glaciol.*, 39, 515-527, 1993.

Whillans, I. M., M. Jackson and Y-H Tseng, Velocity pattern in a transect across Ice Stream B, Antarctica, *J. Glaciol.*, 39, 562-572, 1993.

Williams, R. S., J. G. Ferrigno, T. M. Kent, and J. W. Schoonmaker, Landsat images and mosaics of Antarctica for mapping and glaciological studies, *Ann. Glaciol.,* 3, 321-326, 1982.

Wingham, D. J., A. J. Ridout, R. Scharroo, R. J. Arthern, and C. K. Schum, Antarctic elevation change from 1992 to 1996, *Science,* 282, 456-458, 1998.

Figures

Figure 1 Location map showing area covered by Figures 2-7.

Figure 2 Map of the area around the ice-drainage basin of Pine Island Glacier showing sites of measured surface mass balance (black dots), and interpreted grid of surface mass balance derived from field measurements and passive microwave satellite data from *Vaughan et al.* [1999].

Figure 3 Map of the area around Pine Island Glacier showing surface elevation contours derived from Geodetic Mission of the ERS-1 satellite on a 5-km grid [*Bamber and Bindschadler*, 1997]. The boundary of the Pine Island Glacier ice-drainage basin is shown by the dotted line and the ice shelves are shaded.

Figure 4 Map of ice-drainage basins in the vicinity of Pine Island Glacier. Major glaciers are numbered and their direction of flow indicated by arrows; 1. Pine Island Glacier, 2. Thwaites Glacier, 3. Evans Ice Stream, 4. Carlson Inlet, 5. Rutford Ice Stream, 6. Institute Ice Stream. Rock outcrop and ice shelves are hatched with the mountain ranges numbered (7. Hudson Mountains, 8. Jones Mountains, 9. Ellsworth Mountains). The solid lines show basin-boundaries derived from *Bamber and Bindschadler* [1997], and the dotted line the Pine Island Glacier basin derived from the *Liu et al.* [in press]. The coastline is derived from the Antarctic Digital Database [SCAR, 1993]and the ice shelves are shaded.

Figure 5 Map of area around surrounding the basin drained by Pine Island Glacier showing flow-accumulation derived from 5-km resolution surface elevation grid. Grid cells are shaded such that, cells fed by many others are darker than those fed by few others. The darker areas thus represent areas into which the flow is channelled. The numbered features are the tributaries ice streams as identified by *Stenoien* [1998]. Stenoien's tributaries 1 & 10 are not resolved on this representation.

Figure 6a Map of the area around the ice-drainage basin of Pine Island Glacier showing measurements of ice thickness from airborne survey and oversnow traverses. Grey lines indicate unsuccessful sounding by airborne survey.

6b Map of bed elevations. A grid of ice thickness was calculated using the ice thickness measurements shown in Figure 6a.; this grid was subtracted from a surface elevation grid [*Bamber and Bindschadler*, 1997] to give a grid of bed elevation.

Figure 7 Map of driving stress for the grounded ice sheet in Pine Island Glacier and neighbouring ice-drainage basins derived from surface elevation (Figure 3) and bed elevation (Figure 6b). Light-grey shading denotes driving stresses in the range 0-50 kPa, mid-grey 50-110 kPa, and dark-grey greater than 110 kPa.

Figure 8 Driving stress, topography and radar section collected during airborne survey flight-track shown in Figure 6a. a) Driving stress calculated using 10-km smoothing along the flight-track, the driving stress is calculated to be positive in the direction of Pine Island Glacier ice front and is negative where the ice flows into a different basin-, b) Ice surface and bottom topography along the flight-track with features ice flow features marked, c) section of radar data, showing the difference in the character of the ice-bottom return between the interior drainage basin (rough) and the tributary glacier (smooth). This difference probably reflects the difference between a frozen bed and one that is a melting.

Figure 9 (a) Landsat 1 sub-scene of Pine Island Glacier acquired on January 24, 1973 (path 246,row 114). The "R" marks the position of the raft of ice discussed in Section 4.1. "U" indicates plumes of ice thickness undulations formed close to the grounding line and dissipating towards the ice front discussed in Section 4.1). (b) Mosaic of two ERS-1 SAR images (orbit 3174, frames 5193 and 5211) acquired December 4, 1992 showing the same area. "A" marks the set of arcuate crevasses identified by *Luchitta et al.* [1995]. The dotted line marks the grounding line identified by *Crabtree and Doake* [1982] and the red line indicates the limit of tidal flexing determined by *Rignot* [1998] for 21 January 1996.

Figure 10 Map of selected icefront positions for Pine Island Glacier, between 1966 and 1996 overlaid on an excerpt of a sketch-map drawn from aerial photography collected in 1966 (USGS, 1993). Sources: 1966, Aerial photography (USGS, 1993); January 24,1973, Landsat 1 (path 246, row 114); January 15, 1982, Landsat image; February 9, 1992, ERS-1 SAR image; December 4, 1992, ERS-1 SAR images (orbit 3174, frames 5193 and 5211); 15 March 1994, ERS-1 SAR image; February, 1996, ERS-1 SAR image; February 13, 1998, flight-track of BAS airborne survey aircraft which flew along the ice front.

Tables

Table 1. Estimates of ice-drainage basin area and balance flux for the Pine Island Glacier basin.

Basin Area / 1000 km^2	Balance Flux / Gt a^{-1}	Source
159	63.4	From DEM by *Liu et al.* [1999] and surface balance compilation of *Vaughan et al.* [1999]
175	69	From DEM by *Bamber and Bindschadler* [1997] and surface balance compilation of *Vaughan et al.* [1999]
159 ± 1	63.9 ± 6	[*Rignot*, 1998]
	76	[*Bentley & Giovinetto*, 1991]
		(Arithmetic mean of estimates from *Crabtree & Doake* [1982] and *Lindstrom & Hughes* [1984])
214 ± 20	86± 30	[*Crabtree & Doake*, 1982]
182	65.9± 5	[*Lindstrom & Hughes*, 1984]

Table 2. Estimates of grounding line flux for Pine Island Glacier.

Mass / Gt a^{-1}	Source
68.4 ± 2	[*Rignot*, 1998]
> 56 ± 6	[*Jenkins et al.*, 1997]
70	[*Luchitta et al.*, 1995]
25.5 ± 5	[*Lindstrom & Hughes*, 1984]

Notes

AGU Antarc. Res. Series has around 500 words per column
This paper is 8800 (inc. refs) i.e. around 9 printed pages + diagrams
I think that it should be well within the allocated 19 pages.