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Overview of areal changes of the ice shelves on the Antarctic Peninsula over the past 50 years

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Abstract

In recent decades, seven out of twelve ice shelves around the Antarctic Peninsula (AP) have either retreated significantly or have been almost entirely lost. At least some of these retreats have been shown to be unusual within the context of the Holocene and have been widely attributed to recent atmospheric and oceanic changes. To date, measurements of the area of ice shelves on the AP have either been approximated, or calculated for individual shelves over dissimilar time intervals. Here we present a new dataset containing up-to-date and consistent area calculations for each of the twelve ice shelves on the AP over the past five decades. The results reveal an overall reduction in total ice-shelf area by over 28 000 km^2 since the beginning of the period. Individual ice shelves show different rates of retreat, ranging from slow but progressive retreat to abrupt collapse. We discuss the pertinent features of each ice shelf and also broad spatial and temporal patterns in the timing and rate of retreat. We believe that an understanding of this diversity and what it implies about the baseline dynamics and control will provide the best foundation for developing a reliable real predictive skill.

1 Introduction

The changing position of the margin of the Antarctic ice sheet, both floating and grounded, is being mapped as part of the USGS Coastal-change and Glaciological Maps of Antarctica programme (Williams and Ferrigno, 1998). As part of this programme, a comprehensive time-series of ice front changes around the Antarctic Peninsula was compiled from sources dating from 1940 to 2002 (Cook et al., 2005). The time-series data further reveals changes in glacier, ice shelf and other ice fronts and is published as hardcopy maps with detailed accompanying reports (Ferrigno et al., 2006, 2008) and digital data (Scientific Committee on Antarctic Research, 2005). The main trends observed in the fronts of marine and tidewater glaciers have already been discussed elsewhere (Cook et al., 2005), but the changes in ice shelf fronts were excluded from that study and are brought up to date and described here.

The retreat of ice shelves on the Antarctic Peninsula over the past century has been widely documented and attributed to atmospheric warming (e.g. Vaughan and Doake, 1996; Doake and Vaughan, 1991a; Mercer, 1978; Rott et al., 1998; Skvarca et al., 1999) and several specific mechanisms have been suggested to explain particular phases of retreat and collapse (Scambos et al., 2000; Doake et al., 1998; MacAyeal et al., 2003; Vieli et al., 2006). This is, however, a view that is not universally held, with some evidence that ice-shelf thinning may, in part, be the result of oceanographic change which may "pre-condition" ice shelves to retreat (Shepherd et al., 2003). This raises the question about the degree to which ice shelves in general, and specific ice shelves, are currently responding (and importantly will respond in future) to oceanographic and atmospheric drivers of change.

It has also been shown that the loss of floating ice shelves and ice in fjords can cause profound acceleration and thinning of the tributary glaciers flowing from the Antarctic Peninsula plateau (Rignot et al., 2004, 2005; Rott et al., 2002; Scambos et al., 2004; Pritchard and Vaughan, 2007; De Angelis and Skvarca, 2003; Hulbe et al., 2008). This acceleration implies, and will almost certainly continue to imply, a contribution to sea-level rise. So while the effects of ice shelf retreat on the tributary glaciers will not be discussed directly in this paper, the capability to predict the contribution to sea-level rise that could arise from the Antarctic Peninsula must begin with an understanding of the relevant processes influencing the ice shelves, and this is the subject of the current study.

Individual ice shelves have been monitored and documented (Table 1; see Fig. 1a for locations), but until now there has been no consistent approach to measuring changes for all the ice shelves on a regular time-series. We present a discussion of ice shelf areas measured from the USGS coastal-change dataset, which has been updated for this study using Envisat ASAR images from 2008/2009 (acquired for the IPY Polar View programme [1]). The data has been collected using a consistent approach to give

[1] http://www.polarview.org/

reliable figures for the areas of each ice shelf over the past five decades. We present an analysis of the changes and discuss the trends observed.

2 Method

The methods used in compiling the coastal-change database are described in detail elsewhere (Cook et al., 2005). In summary, historical data sources, including early maps, aerial photographs and satellite images, were used to map ice fronts onto a common reference, using ArcGIS. The quantity of available source material varied considerably across the AP. In some regions, photography coverage and satellite images allowed more than one coastline position to be mapped within each decade, whilst in other regions there was little or no coverage in some decades, and implied large gaps between recorded positions. The quality of the source material also varied, so for this reason, each coastline digitised was assigned a reliability rating (Ferrigno et al., 2006). For the purpose of measuring how the ice shelves have changed in area over the last 50 years, one ice front position was chosen for each decade. This coastline was chosen according to the following scheme:

- As close to mid-decade as possible.
- Where this was not possible, a coastline at either the start or end of the decade, depending on which has the most complete, reliable ice front or most representative line.
- Where there was no coastline available within the decade, a coastline at the end of the previous decade as an approximate position.
- Where none of the above information was available, the decade remained blank.

It should be noted that the coastline chosen could have occurred prior to or following a major ice shelf collapse event and it has not been selected specifically to reflect these events.

The resulting areas therefore represent the area of the ice shelf on a particular year, and do not represent an average position for the decade. Accuracies cannot be assigned to each area value as the variation within each decade is unquantifiable. The reliability rating attached to each coastline indicates the likely errors in the positioning of the ice front and should be taken into account when assessing the accuracy of the areas (Table 2). Since 87% of all coastlines are reliable to within 150 m, and because area accuracies depend not only on variations in the width but also the length of the coastline, the areas are left at 1 km^2 precision.

Using ArcGIS, the coastlines were joined to the grounding lines describing each ice shelf (obtained from the Antarctic Digital Database: SCAR, 2005) to create a single polygon for each ice shelf, in each decade. The area of each polygon was then measured, within the Lambert Azimuthal Equal-Area projection.

3 Results

On the AP, the total area of ice shelf lost (from the earliest available records to the present) is 28 117 km^2 (Table 3). This is a total of 18% of the original area of floating ice (where the original area is based on the earliest available areas for each ice shelf). Despite the episodic nature of the retreat of individual ice shelves, on the decadal timescale there has been a steady decline in total area of the Antarctic Peninsula ice shelves that began in the 1970s and continued to the present (Fig. 2).

The largest change in ice lost from all ice shelves combined, occurred between the 1970s and 1980s (from 1621 km^2 ice gain in the previous decade, to ~9823 km^2 ice loss), with a slight decline in ice loss since the 1980s from 7640 km^2 to 5461 km^2 lost within the present decade (Fig. 3).

Vaughan and Doake (1996) suggested that an ice shelf that is no longer viable is one that will suffer progressive retreat, perhaps over a period of many years, with no major readvance. Of the twelve ice shelves, three have shown significant retreat (Larsen B, Müller and Wilkins) and four of them have shown total collapse (Jones, Wordie, Prince

Gustav, Larsen A) within the last half-century (Table 2). The definition we give here of significant retreat is where less than 70% of the original ice shelf area remains; total collapse is where 10% or less of the original area remains.

Five of the ice shelves (George VI, Bach, Stange, Larsen C and Larsen D) have more than 90% of their original area remaining and are showing no steady retreat. Larsen D is the only ice shelf to have shown overall advance over this time period. The largest ice shelf, Larsen C, is still the largest contributor to the overall area, and is currently at approximately 90% of its original area and showing no significant change.

It has been proposed that the thermal limit of viability for ice shelves on the AP is north of the -9°C isotherm (Vaughan and Doake, 1996; Morris and Vaughan, 2003) and these latest results largely corroborate this theory as all five unchanging ice shelves lie south of the -9°C isotherm (in 2000) (see Fig. 1b). As the following descriptions indicate, the ice shelves that have retreated have generally followed a pattern that suggests a southerly migration of a climatic limit. The Wilkins Ice Shelf is the latest and southernmost ice shelf to begin collapse, suggesting that the thermal boundary has continued to move southwards and may soon threaten the currently stable ice shelves on the southwest AP.

4 Description of change on individual ice shelves

Despite the overall trend in retreat of ice shelves around the AP, individual ice shelves have shown considerable differences in the timing and course of their retreat. Our expectation is that these variations result from the particular characteristics of both the local forcing experienced by individual ice shelves and their structure and balance, or their particular mass balance terms. Indeed, the ice shelves of the AP are diverse and have many peculiarities and individual characteristics. Ice shelf dynamics such as thinning, increase in tributary flow and rheological weakening must be considered alongside ice front retreat in order to interpret the type of collapse shown by individual ice shelves.

TCD
3, 579–630, 2009

So while the changes seen in each ice shelf may be broadly interpreted through discussion of these factors, as is undertaken below, it is clear that a formal sensitivity analysis using models that can take account of differences in mass balance sensitivity and ice rheology, similar to those performed for Larsen B (Vieli et al., 2007), are the most powerful tool available in understanding the detailed aspects of the course of change. Similarly, the "structural glaciology" approach based on a detailed examination of individual retreat events (Glasser and Scambos, 2008; Braun and Humbert, 2009) can provide significantly greater insight into particular retreat events than can be presented here.

4.1 West coast of the Antarctic Peninsula, North to South

4.1.1 Müller Ice Shelf (Fig. 4a)

Müller Ice Shelf (67°14′ S and 66°52′ W) is one of the smallest remaining ice shelves, currently covering 40 km^2. It is the northernmost ice shelf on the western side of the AP and drains an area of the Arrowsmith Peninsula. Two glaciers feed the ice shelf (Bruckner and Antevs), which is pinned on Humphreys Island. Aerial photography from 1947 revealed a significant advance beyond the island until 1956 (Ward, 1995), after which retreat began and by 1963 it was almost back to its 1947 position. A second period of ice front advance of 4 km^2 took place between 1974 and 1986. This was followed by a fast retreat until 1996, after which it has been relatively stable. It is currently at 50% of its size in 1957, although a lack of ground surveying makes interpretation of the grounding line difficult.

A longer historic context for Müller Ice Shelf, obtained from marine sediment cores, shows that advance took place c. 400 years ago (Domack et al., 1995). Although this correlates with the onset of the Little Ice Age (LIA), climatic conditions across Antarctica were variable and this period was one of warmer temperatures along the AP. Domack et al. (1995) concluded that the advance at this time was probably linked to colder circumpolar deep water (CDW) filling the fjord than at present. The warm

TCD
3, 579–630, 2009

CDW that is currently within the fjord may be contributing to the rapid undermelt of the ice shelf system observed in recent years. However, with 50% of the ice shelf remaining and little change in area since the 1990s, the question is why does so much of Müller Ice Shelf remain today, when neighbouring Jones Ice Shelf (Fig. 1a) is entirely gone? Swithinbank (1986) suggested that without the protection afforded by the fjord embayment Müller Ice Shelf probably would not exist under present climatic conditions. This could offer some of the answer, but it might also be wise to question how much of the remaining portion of Müller Ice Shelf is actually floating. The photograph shown in Fig. 4a seems to show that only the western ice front as a clear calving front and it is possible that east of Humphreys Island, the ice is more ice plain than ice shelf. However, with water depths immediately adjacent to the ice shelf reaching over 600 m (Domack et al., 1995) this would seem to imply a surprising bed topography.

4.1.2 Jones Ice Shelf (Fig. 4b)

For a description of the full coastal-change time series for Jones Ice Shelf see Ferrigno et al. (2006, p. 8); Fox and Vaughan (2005).

Jones Ice Shelf (67°30′ S and 66°55′ W) filled Jones Channel between Bigourdan Fjord and Bourgeois Fjord, and was fed mainly by the 10-km long Heim Glacier on Arrowsmith Peninsula. Although substantially smaller than most (29 km^2 in 1947), it has shown a similar pattern of retreat to other ice shelves on the Antarctic Peninsula. Including records prior to the 1950s, the imagery reveals there was a slight increase in area between 1947 and 1978, since when it began a rapid retreat, culminating in complete disappearance by 2003. Blaicklock Island and the Arrowsmith Peninsula are now separated by open water in Jones Channel.

However, despite being less than ten kilometres apart and under similar climate forcing, the east and west portions of the ice shelf behaved differently, with most of the early advance and subsequent retreat occurring on the eastern ice front. The western front was stable until the early 1990s, after which it too retreated, disconnecting from Blaicklock Island sometime between December 1998 and January 2001. The difference in response between the two ice fronts was examined by Fox and Vaughan (2005). Their conclusion was that the response of the east and west fronts of the ice shelf were controlled by the configuration of the fjord in which the ice shelf lay. The western and eastern portions of Jones Ice Shelf lie in convergent and divergent embayments respectively, and therefore have different strain rate regimes, which affected the retreat rates. This supports modelling studies (Doake et al., 1998) which compared contrasting ice flow regimes in the Larsen B Ice Shelf prior to its collapse. The course of retreat therefore, is controlled not only by climatic events but also, in this case, the geometry of the embayment and the location of pinning points (Fox and Vaughan, 2005).

Ice shelf thickness does not appear to have had a strong influence on the progress of the retreat. There is, however, insufficient evidence to determine whether surface-melting may be related to specific retreat events (Scambos et al., 2000).

4.1.3 Wordie Ice Shelf (Fig. 4c)

For a description of the full coastal-change time series for Wordie Ice Shelf see Ferrigno et al. (2006, p. 8).

Wordie Ice Shelf (69°10′ S, 67°30′ W), the northernmost large (>1000 km^2) shelf on the western AP, disintegrated in a series of events during the 1970s and 1980s, and by 1992 there was little more than a few disconnected and retreating glacier tongues remaining. It was the most dramatic retreat of all ice shelves on the AP up until that time. Understanding the cause and mechanism of collapse has given important clues as to the behaviour of ice shelves as they undergo progressive ice front retreat.

During the British Graham Land Expedition (1934–1937), the early explorers identified approximately ten ice rises or ice rumples. The ice shelf was then interpreted as a relict feature and because of poor surface conditions it was suggested that it would not survive for long (Fleming, 1940).

Doake and Vaughan (1991b) suggested that the break-up was triggered by the warming trend in mean annual air temperatures in this region since the early 1970s, which increased ablation and the amount of melt-water and hence the rate of fracture;

an assertion that has never really been supported by strong evidence, but which remains a reasonable hypothesis. The major ice rises, however, appear to have played a significant role in controlling the behaviour of Wordie Ice Shelf. The earlier ice front retreat was punctuated by periods of stasis when the ice was pinned on the ice rises (Vaughan, 1993), and the retreat between periods of stasis was rapid. The ice rises also appear to have acted as wedges, contributing to weakening of the ice shelf and hastening the break-up. As the shelf ice flowed past them, the ice rises caused splitting of the ice shelf upstream, and created broken wakes downstream, introducing weakness (see oblique aerial photograph in Fig. 4c). The increasing stresses caused by the separation around the ice rise as the ice shelf accelerated eventually became sufficient for fracture. Fracture, in the form of surface crevasses or rifts extending to the bottom of the ice shelf, appeared prior to retreat of the ice front, but whether this was a cause or effect of the collapse remains unclear.

Wordie Ice Shelf was fed by a number of tributary glaciers which join to make three main input units. As the fracture processes weakened the interior of the ice shelf, and after the ice front had retreated past a critical point, the outlet glaciers appeared to punch through the remaining ice shelf because of the reduced restraint on them (Vaughan et al., 1993). However, the break-up of the ice shelf was not initiated by surging activity on the input glaciers.

Vaughan (1993) also concluded that there were no changes in the relative dominance of the tributary glaciers (between 8 to 20 km upstream) following the ice shelf collapse and therefore the ice shelf provided no significant restraint on the inland glaciers. Given recent work, however, this conclusion is now in doubt. The impact of the loss of the ice shelf has been observed upstream on Fleming Glacier (Rignot et al., 2005). Satellite radar interferometry data from 1995 to 2004 and airborne ice thickness data from 2002 reveal that 50 km from its ice front, Fleming Glacier flows 50 percent faster than it did in 1974, prior to the main collapse of the Wordie Ice Shelf. Rignot et al. (2005) conclude that the glaciers accelerated following ice shelf removal and have been thinning and losing mass to the ocean over the last decade.

4.1.4 George VI Ice Shelf (Fig. 4d)

George VI Ice Shelf is the largest ice shelf on the west coast of the AP, covering a total area of ~24 000 km^2. It is unusual in two respects, as discussed in Smith et al. (2007). Firstly, it has two ice fronts that are almost 500 km apart and experience quite different climatic conditions. Secondly, it is an ice shelf that is constrained within a narrow channel (25 km wide in the north, widening to over 70 km in the south) and loses most of its mass to melting (both surface melt-pools and basal melting), rather than changes in ice-dynamics brought about by a changing calving rate (Roberts et al., 2005).This suggests that this ice shelf may be more sensitive to changes in ocean conditions, and less sensitive than its neighbours to atmospheric change (Doake and Vaughan, 1991b).

Both the north and south ice fronts have shown retreat in recent decades but the overall area has not shown significant change, currently at 92% of its original size (in 1947). Although the relative change is small, it has been argued that it may be in its first stage of disintegration (Luchitta and Rosanova, 1998).

The northern front (at 70° S, 68°45′ W) showed a short period of advance between 1947 and 1960, followed by a major phase of ice shelf retreat until 1980, then slower but continual retreat through to the present day. A total of 610 km^2 of ice has been lost since 1947. It is thought that calving occurs along rifts and fissures which penetrate the entire depth of the ice shelf (approximately 100 m at the northern ice front) (Potter and Paren, 1985). Conversely, it is affected by the occurrence of perennial and on occasion, semi-permanent sea-ice, which prevents break-up of ice when cracks form near the edge, and holds large icebergs within the channel. Doake (1982) commented that it was difficult to locate the true ice shelf front in images because of the fragmented shelf. Indeed, interpretations of the ice front from imagery in 1974 in this study differ from those by Luchitta (1998).

The southern front (at 73° S, 72°10′ W) has shown a regular rate of retreat since 1947, although retreat has slowed since the 1990s. It has lost a total of 1330 km^2 since

1947, more than twice the area of that lost from the northern front. It also appears to have begun retreating earlier, without any significant re-advance. Here the ice shelf is significantly thicker (a maximum of 600 m) and wider (over 70 km) than at the northern front (Smith et al., 2007). There is little sea-ice build-up within the Ronne Entrance which also differs from the northern end of George VI Sound.

An unusual abundance of annual melt ponds on George VI ice shelf has been known, at least since the 1960s (Wager, 1972). In recent decades, every summer much of southern part of the ice shelf has been covered in melt-ponds that tend to recur on the same area of ice each year. This is clear evidence that the presence of persistent melt does not necessarily condemn an ice shelf to rapid retreat, and cannot be considered the sole factor in causing ice shelf retreat. There may, however, be a single factor that is responsible both for the persistence of melt ponding on George VI ice shelf and its continued survival. Most of the ice in George VI Ice Shelf derives from glaciers feeding it from the Antarctic Peninsula in the east. This ice flows largely across the ice shelf towards Alexander Island in the west (Pearson and Rose, 1983). The rates along the axis of the ice shelf are thus very low, but the westerly flow of ice against Alexander Island, creates strong horizontal uniaxial compression over large portions of the ice shelf. As a result, much of the ice shelf surface is covered by pressure rollers, which control the shape and pattern to the lakes. Pearson and Rose likened the flow to that of a subducting continental plate, and noted that most of the ice loss from this ice shelf is by basal melting rather than iceberg calving. Being in compression means that there are few crevasses in the ice shelf, and these, and any other routes by which water can drain down out of the ice shelf, are quickly closed. Melt water thus tends to remain at the surface of this ice shelf much longer than is normal.

The overall pattern of retreat of both ice fronts coincides with the well-documented atmospheric warming across the AP, but since the southern ice front retreat preceded retreat at the northern ice front, it cannot be interpreted simply as a result of the southern migration of a key mean annual isotherm (Smith et al., 2007). It is likely that the southern ice front is also responding to other forcing mechanisms and feedbacks (e.g. changes in sea-ice extent or oceanography). One theory is that the mass balance of the ice shelf is strongly influenced by high basal melt rates associated with intrusions of warm Upper Circumpolar Deep Water (UCDW) (Shepherd et al., 2003; Payne et al., 2004) and therefore its stability is sensitive to changes in ocean circulation. The thinning of floating glaciers in the West Antarctic Ice Sheet (WAIS) region has been attributed to heat contained within the CDW flowing towards the continent via a submarine glacial trough and driving basal melting (Walker et al., 2007). In order to assess whether George VI Ice Shelf has entered a phase of rapid retreat and whether this is largely driven by atmospheric warming, basal melting, or both, requires further work (Smith et al., 2007).

4.1.5 Wilkins Ice Shelf (Fig. 4e)

Wilkins Ice Shelf (70°20′ S, 72°20′ W) is the largest ice shelf on West Antarctica currently undergoing significant break-up. It was a stable and slowly evolving ice shelf between 1947 and mid-1980s, but beginning in 1986 this pattern changed and a series of retreats on the northern edge occurred (Scambos et al., 2000). As with the Larsen A and Larsen B Ice Shelves (see Sect. 4.2.2), the retreat events were characterized by episodic calving of many small elongated bergs. A "disintegration" event was observed in 1998 (Scambos et al., 2000) but much of the ice remained until March 2008, when dramatic calving removed 800 km^2 of ice, followed by a further 1450 km^2 in July 2008. The ice shelf was then only supported by a single strip of ice just 1 km wide, strung between Latady and Charcot islands. This strip gave way in April 2009, reducing the area by a further 340 km^2. More than one third of the original area has been lost with the remaining area currently standing at 5434 km^2.

Wilkins Ice Shelf has an unusual mass balance regime and differs in many ways from its neighbours, Wordie Ice Shelf and George VI Ice Shelf, which are fed by glaciers. It is unusually stagnant with flow rates of 30–90 ma^{-1}, compared to Wordie (200–2000 ma^{-1}) and George VI (200 ma^{-1}) (Vaughan et al., 1993). The catchment area is 16 900 km^2, which is an unusually low proportion of grounded ice sustaining the ice shelf, compared with Wordie Ice Shelf and George VI Ice Shelf. In this respect it may be similar to Larsen A Ice Shelf in being sustained largely by in situ accumulation. Vaughan et al. (1993) also found that in the middle of the ice shelf the surface was low, but there is a topographic bulge along the shelf's western margins, where most of the accumulation appeared to occur. Abundant melt-ponding and the observation that the ice shelf was unusually thin and parts may be brine-soaked to sea-level (Vaughan et al., 1993; Swithinbank and Luchitta, 1986), suggested that the shelf may be subject to rapid break-up (Lucchitta and Rosanova, 1998). Vaughan et al. (1993) however, stated that Wilkins Ice Shelf front was likely to decay by normal calving processes and unlikely to undergo sudden disintegration (as seen in the Wordie Ice Shelf), at least until the calving front decoupled from the fringing islands. They speculated that if the climatic limit continued to move south at the same rate, within 30 years the Wilkins Ice Shelf may no longer be viable.

Analysis of the recent break-up suggests that surface or near-surface melt-water is the main pre-condition for disintegration, and that hydro-fracture is the main mechanism (Scambos et al., 2009). The break-ups were characterized by repeated rapid fracturing that created narrow parallel blocks at the ice-edge. The ice-front bending stresses induced by buoyancy forces were responsible for fracture formation. Humbert and Braun (2008) showed that stresses in Wilkins Ice Shelf changed with each major disintegration event, resulting in lengthening of rifts immediately following the events. These changes in stress distribution, together with widespread brine infiltration, led to the May 2008 disintegration event (Scambos et al., 2009). It has also been inferred that compressive stress along the narrow ice bridge between Charcot and Latady Islands increased as this region narrowed. Near-surface melt-water is effective at driving fracture extension and it has been concluded that most disintegration events will occur during or just after an extensive melt season. The two break-ups that occurred during Autumn/Winter seasons, however, showed that melt ponds played no role in these particular break-up processes (Braun and Humbert, 2009). The fact that Wilkins Ice Shelf experienced two break-up events under two widely contrasting surface conditions reveals that there is more than one mechanism at play in its disintegration. Prediction of where and when the next step in the break-up might occur is impossible due to this highly complex pattern of fracturing, and close observation of subsequent events by satellite is key to examining the causes and triggers of break-up events (Humbert and Braun, 2008). Latest research (carried out in February 2009) suggests that about 3100 km^2 at the northern Wilkins Ice Shelf are endangered, but that there is no visible signature that the remaining 8000 km^2 are at risk (Braun et al., 2009).

In addition to climatic forcing (i.e. the migration southwards of the thermal limit of viability), the Wilkins Ice Shelf is subject to the same oceanic forcing mechanism as George VI Ice Shelf as it receives the same warm-water influx of the UCDW current. The major decrease in sea-ice extent observed in the Bellinghausen Sea from 1988 through 1991 may also have affected the ice shelf margin, although this remains to be explored (Lucchitta and Rosanova, 1998).

4.1.6 Other ice shelves on west Antarctic Peninsula (Fig. 4f and g)

There are two other substantial ice shelves located on the western side of the Antarctic Peninsula: Bach Ice Shelf, located between Beethoven Peninsula and Monteverdi Peninsula, at the southern end of Alexander Island, and Stange Ice Shelf, west of Spaatz Island. There are also a number of small un-named ice shelves close to the Wilkins Ice Shelf, on Beethoven Peninsula, which are not discussed in this report.

Although Bach Ice Shelf and Stange Ice Shelf are named and are considerably larger than other ice shelves on the AP (namely: Jones, Müller, Wordie, Prince Gustav and Larsen A), there has to date been very little discussion about either. There has been much greater interest paid to their neighbours, Wilkins Ice Shelf and George VI Ice Shelf.

They should be considered, however in studies of ice shelf behaviour in this region, as they both display relative stability in an area that may be subject to both atmospheric and oceanic forcing. Bach Ice Shelf (72° S, 72°20′ W) lies between Wilkins Ice Shelf and George VI Ice Shelf southern ice front, both of which have shown significant retreat in recent years. Its area was 4800 km^2 in 1947 and since then it has lost only 300 km^2, showing stable growth and calving during this period. Stange Ice Shelf (73° S, 76°40′ W), situated to the west of George VI Ice Shelf, also shows no signs of retreat. At 8300 km^2 in 1970s (there are no reliable records of the ice front position prior to 1973), it is currently at 8000 km^2, or 96% of its original size. It can be concluded that both Stange and Bach Ice Shelves show stable ice shelf behaviour. They both currently lie to the south of the climatic limit of viability, but with retreat of the Wilkins Ice Shelf suggesting that this limit may be moving south, Bach Ice Shelf may be the next ice shelf under threat. The impact of the UCDW current, which is speculated to be contributing to the ice loss at the George VI Ice Shelf southern front, is yet to be examined.

4.2 East coast of the Antarctic Peninsula, North to South

4.2.1 Prince Gustav Ice Shelf (Fig. 4h)

For a description of the full coastal-change time series for Prince Gustav Ice Shelf see Ferrigno et al. (2006, p. 6–7).

The Prince Gustav Ice Shelf (64°20′ S, 58°30′ W) was the most northerly ice shelf and was the first to show signs of retreat. Although 1600 km^2 in 1957, this ice shelf retreated progressively through the second half of the 20th century. In 1995, it finally collapsed, leaving open water between James Ross Island and the main Antarctic Peninsula. The remnants close to the coast have now also completely melted.

There is some evidence that Prince Gustav Ice Shelf was already retreating prior to 1957, the first position recorded for this study. Historical accounts as far back as 1843 suggest that the ice shelf had been retreating for most of this period (Cooper, 1997). A study by Reece (1949), based on expedition reports, showed that in 1945 the southern ice front stretched from James Ross Island as far south as Robertson Island. He concluded that at this time the ice was continuous with Larsen Ice Shelf. He discussed the difference in the types of ice observed in the Prince Gustav Channel: “[The ice to the north of Sjögren Glacier] is land-fast sea ice, which persists for more than one season due to the presence of islands; that in the south is shelf ice of the same type and origin as the Larsen Ice Shelf but perhaps less thick. This is mainly due to the presence of shoals although islands are also there." There is evidence to suggest that the ice shelf became separated from Larsen Ice Shelf sometime in the late 1940s (Francis, 1947), retreating to Cape Longing. Since then there was a rapid retreat from 1957 to 1961, followed by a more steady retreat until the collapse in 1995.

As with Jones Ice Shelf, the response of the ice fronts at each side of the channel differed considerably. The northern ice front was relatively stable until the early 1990s, but the southern ice front has shown a slow progressive retreat since the earliest records. The mechanisms underlying the retreat were discussed by Cooper (1997). The ice of the Prince Gustav Channel has two components: glacier tongues (flowing into the northern part of Prince Gustav Ice Shelf) and ice derived from in situ accumulation (south of Sjogren Glacier). The different geometries may also explain the differing stability of the northern and southern margins: in the north, the land margins are convergent and the ice margin is pinned at the narrows between Cape Obelisk and the opposite shore; in the south the land margins are broadly divergent, allowing the ice to retreat steadily along a featureless coast (Cooper, 1997). The geometry of the southern coast has changed due to the significant retreat of Sjogren Glacier, making it unlikely that the ice shelf will reform.

One study of sub-ice-shelf sediments has demonstrated that Prince Gustav Ice Shelf varied in extent in response to natural Holocene climate forcing (Pudsey and Evans, 2001). This study concluded that ice shelves in this region are sensitive indicators of regional climate change, but without necessarily implying that recent decay is due to human-induced climate change (Pudsey and Evans, 2001).

4.2.2 Larsen A and Larsen B Ice Shelves[2] (Fig. 4i and j)

For a full time series of retreat for Larsen Inlet and Larsen A see Ferrigno et al. (2006, p. 7) and Skvarca (1999). For Larsen B see Ferrigno et al. (2008, p. 8). For a detailed analysis of Larsen Ice Shelf changes in extent between 1840 and 1960, see Karstkarel (2005).

Close to the Prince Gustav Ice Shelf, the Larsen A Ice Shelf (64°40′ S, 60° W) has also now disappeared. Larsen A Ice Shelf extended from Cape Longing to Robertson Island and merged with Larsen B Ice Shelf to the south at Seal Nunataks. At 4000 km^2 in 1961, Larsen A remained relatively stable until the 1980s, when it began to retreat in steps, culminating in January and February 1995 in a major collapse. During this period, almost 2000 km^2 of ice shelf were lost within a few weeks, breaking up into thousands of icebergs producing a plume 200 km into the Weddell Sea. This breakup was the most spectacular event within a 50-year period of ice shelf retreat along both coasts of the Antarctic Peninsula, and was the first collapse that was properly observed. The ice within Larsen Inlet (between Cape Longing and Sobral Peninsula) also continued to retreat after it became disconnected from Larsen A Ice Shelf, suffering major retreat during the period 1987 to 1989, and very little ice remains. What does remain is a small promontory held together by Seal Nunataks, which may or may not be floating.

The same form of collapse occurred on an even larger scale in 2002 with the break-up of Larsen B Ice Shelf. Larsen B (65°30′ S, 61° W) is the ice shelf south of Larsen A, extending from Robertson Island in the north to Jason Peninsula in the south. In 1963, it was 12 000 km^2 but now has only 20% of its area remaining (2400 km^2). Most of the remaining ice is held in Scar Inlet, between Cape Disappointment and Jason Peninsula. Unlike Larsen A, Larsen B continued to advance until the early 1990s. A large tabular iceberg (~1700 km^2) and smaller bergs calved in 1995, after which it followed

[2]There are four glaciological distinct sections of the feature that is formally named "Larsen Ice Shelf", these have been termed Larsen "A", Larsen "B", Larsen "C" and Larsen "D" following nomenclature proposed by Vaughan and Doake, 1996. See Fig. 1a.

a similar pattern of retreat to its neighbour Larsen A (Scambos et al., 2000). It culminated in a dramatic fragmentation of approximately 3250 km^2 within only a few days in February 2002. This was the first ice shelf for which good data was collected prior to retreat, assisting with understanding the controls of the dynamics of ice shelves.

The potential mechanisms leading to the disintegration of these ice shelves have been widely discussed. The underlying cause appears to be rising atmospheric temperatures in the region (Rott et al., 2002, 1998; Vaughan et al., 2001; Skvarca et al., 1998; Domack et al., 2005), but while this may be the principal driver, it does not explain the mechanism behind dramatic collapse rather than normal progressive retreat.

Both ice shelves were observed to have significant melt-water accumulation, both immediately prior to and in preceding melt-seasons. One theory is that the build-up of surface melt-water causes initial fragmentation of the ice shelf by downward propagation of water through crevasses (Scambos et al., 2000; van der Veen, 2007; Scambos et al., 2008; van den Broeke, 2005). The rapid disintegration of the rifted shelves into thousands of small fragments could be explained by a secondary process proposed by MacAyeal et al. (2003). This involves energy being released when narrow ice-shelf fragments capsize in a coherent manner, which then initiate further fragmentation in a "domino" effect. This mechanism could also help explain the time delay between previous seasons of greatest melt-water build-up and the season of rapid collapse. The pre-break-up surface texture and ice thickness, which are related to the glacier source areas, are also critical in determining the speed of ice shelf disintegration (MacAyeal et al., 2003).

Another theory suggests that the geometry of the ice front could affect disintegration. The mechanism is similar to that which keeps the arch of a stone bridge stable, with the bottom stone held in compression, until too many stones fall away from the underside and the bridge collapses. For an ice shelf, so long as the second principal stress (perpendicular to the ice front) is compressive the ice front can remain stable, but when calving causes the geometry to bow inwards and begins to remove the ice in this compressive arch (Doake et al., 1998; Rack et al., 2000), the ice shelf could become potentially unstable. Once the ice front breaks back through the critical arch then an irreversible retreat will occur, possibly catastrophically (Doake et al., 1998). Large rifts, almost parallel to the ice front, formed lines of weakness along which calving occurred in the years prior to collapse (Skvarca, 1993). These transverse compression fractures differ from the straight-line fractures that appeared in Wilkins Ice Shelf prior to collapse, suggesting there is a difference in calving mechanisms.

Another hypothesis has been used to suggest that the ice shelves may have become susceptible to crevasse-fracture through a sustained ice thinning (Shepherd et al., 2003). Satellite altimetry has shown that Larsen B Ice Shelf lowered by up to $0.17 \pm 0.11\,\mathrm{m\,a^{-1}}$ between 1992 and 2001, which could be explained not only by increased summer melt-water changing the near surface density structure, but the loss of basal ice through melting (Shepherd et al., 2003). This suggests that enhanced ocean-driven melting could provide a link between regional climate warming and successive disintegration of these sections of the Larsen Ice Shelf. Oceanographic observations in this region are limited, however, and another study has shown that regional ocean warming may not have had a major impact on basal melt rates (Nicholls et al., 2004). See Sect. 4.2.3 for a fuller explanation.

Other changes in the ice shelf dynamics prior to their collapse have also been detected, such as an acceleration in ice shelf flow of up to 50%, revealed through observations using satellite radar interferometry for the Larsen B Ice Shelf (Rignot et al., 2004; Vieli et al., 2006; Skvarca et al., 2003). It has been speculated that the ice shelf acceleration is explained by the retreat of the ice front but also by further rheological weakening of the shear margins, and not by the observed thinning or the acceleration of tributary glaciers (which appear to be an effect, not a cause of ice shelf acceleration) (Vieli et al., 2007). Ice surface features were examined by Glasser and Scambos (2008), who concluded that structural glaciological discontinuities played a large part in the break-up of the Larsen B IS because they rendered the ice shelf mechanically weak prior to its collapse. Large open-rift systems, which were not present more than ~20 years ago, became more pronounced in the years preceding break-up, possibly as a result of ice-shelf front retreat during 1998–2000, which removed the strongest portions of shelf ice (Glasser and Scambos, 2008).

These changes indicate that the Larsen A and Larsen B ice shelves reacted to the changing climate significantly before their collapse. The causes of these pre-collapse changes and their link to the changing climate, however, are still poorly understood. It is accepted that rapid regional warming has been occurring on the Antarctic Peninsula over the last 50 years, at a rate of 3.7±1.6°C per century (Vaughan et al., 2003). The steady retreat and collapse of the northern Larsen Ice Shelf coincided with this warming trend on the east coast (Skvarca et al., 1998). In 2006, the first direct evidence linking human activity to the collapse of Larsen ice shelves was reported (Marshall et al., 2006). That study showed that the Southern Hemisphere Annular Mode has been enhanced by greenhouse gas emissions, and has had a profound effect on AP summer temperatures by causing warm Föhn winds to blow off the Antarctic Peninsula. If warming continues, the Larsen C may soon display a similar pattern of retreat.

4.2.3 Larsen C Ice Shelf (Fig. 4k)

Larsen C Ice Shelf (67°30′ S, 62°30′ W) is the largest ice shelf on the AP, currently at just under 51 000 km^2. It extends from Jason Peninsula in the north to Gipps Ice Rise in the south, where it merges with Larsen D Ice Shelf. To date, this ice shelf has not shown evidence of climate driven retreat, as the variations in the ice front follow normal ice shelf fluctuation patterns.

There has been relatively little discussion about Larsen C Ice Shelf compared with its neighbours, Larsen A and B. Skvarca (1994) described changes in the ice shelf margins and surface-ice features along the Larsen Ice Shelf from Cape Longing as far south as Ewing Island. He noted advance along the area ∼70 km southeast of Jason Peninsula since 1975, observing that the ice shelf is sustained by snow accumulation as there is no significant glacier input from Jason Peninsula. For the majority of the Larsen C ice front, to Gipps Ice Rise, there was major change between 1975 and 1988. The ice shelf is heavily rifted and crevassed, strongly influenced by the inflow of land ice, mainly through Cabinet, Mill, Whirlwind and Mobiloil Inlets. A section broke off the northernmost part, shortly after being hit by an iceberg in 1976, but the major change occurred in 1986 by the calving of two giant icebergs. The larger berg was ~6520 km^2 in area and the smaller one ~1260 km^2 (Skvarca, 1994). In total the area of ice lost from Larsen C between 1975 and 1988 was 7800 km^2. Since then there have been no further large calving events and the ice front has been slowly advancing. It is currently at 90% of its original size (in 1963) and appears to remain stable.

Prior to the calving event in 1986, the ice shelf floated free to the north of Gipps Ice Rise. The rise acted as a constraining point causing chasms and inlets to form parallel to the ice front, which formed lines of weakness where the calving took place (Skvarca, 1994). It is thought that this was not climate-driven retreat but part of the natural process of ice shelf mass balance regulation. If the same atmospheric and oceanic forces are at play which caused its neighbours to collapse, however, Larsen C Ice Shelf is also expected to retreat. The timescale and possible mechanism are as yet uncertain.

Using satellite radar measurements from 1992 and 2001, Shepherd et al. (2003) found that the northernmost sections of Larsen C displayed the greatest decrease in surface elevation, $0.27\pm0.11\,\mathrm{m\,a^{-1}}$. Towards the southern tip however, the ice shelf thickened where ice is discharged from Palmer Land. The mean rate of elevation change of Larsen C was $0.08\pm0.04\,\mathrm{m\,a^{-1}}$. There is evidence that nearby Weddell Deep Waters (WDW) have warmed (Robertson et al., 2002), and in 2002 oceanographic measurements showed large quantities of Modified Weddell Deep Waters (MWDW) present in front of the northern Larsen C (Nicholls et al., 2004). The study by Shepherd et al. (2003) stated that although basal melt rates are uncertain, enhanced ocean melting has been and is thinning the Larsen Ice Shelf. They predicted that if basal melting continues at the estimated rate of $0.78\,\mathrm{m\,a^{-1}}$, Larsen C will approach the thickness of Larsen B at the time of its collapse in around 100 years, or more rapidly with a warming ocean (Shepherd et al., 2003).

Conversely, other research indicates that it is unlikely that temperature changes of the deep Weddell Sea are having a major impact on melt rates at the base of Larsen C Ice Shelf (Nicholls et al., 2004). Observations show that although MWDW is present at the ice front, the Ice Shelf Water is not derived from this directly, but from MWDW pre-conditioned by winter cooling and by salinification from sea-ice production. This suggests that there is no clear source of warm water for increased basal melting, since the pre-conditioned MWDW is cooled to the surface freezing temperature. Recent modelling suggests there are regions of basal freezing (Holland et al., 2009) and this supports the prevalence of cold MWDW beneath Larsen Ice Shelf. There is an urgent need for more data before further conclusions can be drawn (Nicholls et al., 2004).

4.2.4 Larsen D (Fig. 4l)

Larsen D Ice Shelf is a unique ice shelf on the AP, as it has advanced over the past 50 years, although this has been small relative to the retreats of the other ice shelves. The ice shelf extends from Gipps Ice Rise (68°45′ S, 60°45′ W) southwards to Smith Peninsula (74°30′ S, 61° W). The first ice front position for this study was mapped from oblique aerial photographs taken in 1966 and the area calculated at 21 700 km^2. The most recent calculation was 22 600 km^2. Within that time there have been small fluctuations but it can be concluded that Larsen D is generally stable. This ice shelf is the furthest from the thermal limit of viability and so it appears unlikely that it will show signs of change until this limit is driven southwards. The Larsen D has been barely mentioned in reports of AP ice shelves. Skvarca (1994) observed ice margin changes between Gipps Ice Rise and Ewing Island and noted a steady advance between 1975 and 1988. In the same period, the surrounding area of fast-ice receded. Fast ice exists along the whole front of the Larsen D Ice Shelf. It makes ice front interpretation difficult as the semi-permanent sea-ice is of varying thicknesses and almost indistinguishable from shelf ice on some satellite images. For this study, it was important to select a consistent ice front between images to accurately represent the real position for each decade. Although more difficult to interpret than other ice shelves, the results accurately describe the small variations that occurred in this area.

5 Discussion

Atmospheric warming on the Antarctic Peninsula over the past 50 years has been widely documented (King and Harangozo, 1998; King, 1994; Skvarca et al., 1998; Vaughan et al., 2003; Morris and Vaughan, 2003; Marshall et al., 2006). At a rate of 3.7±1.6°C (century)$^{-1}$ this is significantly higher than the global mean (0.6±0.2°C during the 20th Century) and also the rest of the Antarctic continent (Vaughan et al., 2003). The ultimate cause of the atmospheric changes remains to be determined but there is emerging evidence that it can be linked to human activity (Marshall et al., 2006).

A map showing interpolated mean annual temperature across the AP shows a contrast (3–5°C) between the east and west coast and also reveals that the limit of ice shelves closely follows the −9°C isotherm (Morris and Vaughan, 2003) (see Fig. 1b). The limit of ice shelves that have retreated in the last 100 years is bounded by the −9°C and −5°C isotherms, suggesting that the retreat of ice shelves in the AP region is consistent with a warming of around ~4°C (Morris and Vaughan, 2003). The retreat of ice shelves on the Antarctic Peninsula over the past century is now widely attributed to atmospheric warming (e.g. Vaughan and Doake, 1996; Doake and Vaughan, 1991a; Mercer, 1978; Rott et al., 1998; Skvarca et al., 1999). Furthermore, although across the AP warming is strongest in the winter months, warming in the summer is statistically significant and has been sufficient to cause a substantial increase in surface melt-water production on several ice shelves (Vaughan, 2006).

The behaviour of the ice shelves in their response to the warming climate has varied considerably between the twelve AP ice shelves, from slow, steady retreat by calving to rapid disintegration. Ice shelf dynamics and other forcing mechanisms have been suggested to explain particular phases of retreat and collapse. Geometry of the embayment can have a key effect in controlling the response of different sides of an ice shelf within a channel (e.g. Jones Ice Shelf and Prince Gustav Ice Shelf). The geometry

TCD
3, 579–630, 2009

of the ice front itself could affect disintegration, when normal calving produces a compressive arch close to the ice front and causes strain sufficient to lead to rapid retreat (e.g. Larsen B Ice Shelf). The location of pinning points can affect the course of retreat, as seen in the Wordie Ice Shelf, where major ice rises acted as wedges and contributed to weakening, hastening break-up.

The mass balance regime strongly affects the behaviour of the ice shelf. Where shelf ice is not fed by grounded ice, it relies on in-situ accumulation and is more susceptible to rapid breakup through surface melting and thinning (e.g. Wilkins Ice Shelf). Surface melt-water can be both an instigator and indicator of imminent ice shelf collapse (e.g. Larsen A, Larsen B and Wilkins Ice Shelves), although not in the case of George VI Ice shelf which is under uniaxial compression. Only the currently retreating ice shelves (Larsen B and Wilkins) and the northernmost portion of Larsen C Ice Shelf, have the firn characteristics and melt season length associated with imminent break-up (Scambos et al., 2003). The melting may be caused by atmospheric warming, but the mechanism for break-up is described as being a vertical propagation of water through crevasses causing initial fragmentation of the ice shelf (van der Veen, 2007; Scambos et al., 2008). A "domino effect" may also occur as the rifted ice blocks topple, causing mass disintegration (e.g. Larsen B Ice Shelf) (MacAyeal et al., 2003). The melt pools observed on the surface of the Larsen C Ice Shelf currently lie to the north of the position of the -9°C isotherm and may be the precursor to collapse in this region, but without knowing the rate of southwards migration of the limit of viability the timescale is difficult to predict.

The surface texture, ice thickness and ice velocity also play an important part in determining the behaviour of an ice shelf undergoing retreat. It is also possible that rheological weakening of the shear margins can cause an increase in the ice flow rates (e.g. Larsen B), and the development of rifts and the rupturing of sutures between flow units can precondition an ice shelf to collapse (Glasser and Scambos, 2008).

A further aspect of ice shelf change is thinning caused by ocean-driven basal melting, which could pre-condition ice shelves to retreat. This could be due to warm intrusions of Upper Circumpolar Deep Water currents (e.g. southern George VI ice front and Müller Ice Shelf) and Modified Weddell Deep Waters (e.g. Larsen B and Larsen C Ice Shelves). Recent modelling suggests that the basal melt of ice shelves would increase quadratically as the ocean offshore of the ice front steadily warms (Holland et al., 2008), but as yet there is not enough data available to be able to interpret the warming trends of the Southern Ocean. Further oceanographic observations are needed to determine the extent to which ocean temperature change may impact on the ice shelves. It is also still unclear if a critical minimum ice shelf thickness is a factor in ice-shelf disintegration (Shepherd et al., 2003). In addition, it is possible that in a warmer climate, the reduction in rates of sea-ice formation could cause a reduction in the flux of High Salinity Shelf Water (HSSW) beneath the ice shelf, in turn reducing basal melting (Nicholls, 1997). This has been observed in oceanographic measurements from beneath the Filchner-Ronne Ice Shelf, in the southern Weddell Sea. Sea-ice must therefore also be taken into account as a factor affecting ice shelf changes.

One ice front does not appear to fit within the pattern of ice shelf retreat being directly related to atmospheric warming. Prior to this study, the southern ice front of George VI Ice Shelf has not been discussed in the literature. Although the ice shelf has changed little overall, the southern front has shown retreat since the 1950s and yet it lies south of the predicted limit of viability. Because few ice-front positions were available, this ice shelf was not included in earlier studies that sought to establish a connection between the known pattern of atmospheric temperatures with ice shelf changes through discussion of a southerly-migrating “limit of viability” for ice shelves (Vaughan and Doake, 1996; Morris and Vaughan, 2003). The map presented by Morris and Vaughan was derived by a multivariate regression technique in which the mean residuals were approximately two degrees. It shows that the mean annual temperature in the area around the southern ice front of George VI ice shelf is -10°C (Fig. 1b). This is slightly cooler than the -9°C mean annual temperature that was identified as the limit of viability for ice

TCD
3, 579–630, 2009

A. J. Cook and D. G. Vaughan

shelves, which lies to the north. Certainly, this new information requires a reassessment, or at least adjustment, of the hypothesised −9°C isotherm as the limit of viability. The gradient of isotherms is shallower on the west coast and therefore the isotherm locations could be less precise in this region. To investigate whether the concept of a thermal limit of viability of ice shelves can survive the introduction of this new data, we need to either: determine the distribution of mean annual air temperature in greater detail; allow that the limit can only be identified in approximate terms (existing in the region of −9 to −10°C); or investigate whether a more refined description of the limit of viability is required.

6 Conclusions

Ice shelf retreat may be an icon of climate change science, but not all ice-shelf retreats are the same. There are a variety of responses: protracted retreat, rapid collapse and possibly staged-collapse. The rate of retreat does not simply appear to be related to the rate of change of climate, but it is modulated by the ice shelf configuration and conditions of mass balance. It is not clear exactly which processes dominate in each case or why. There are speculations, but very little definitive interpretation because most ice shelves can be subject to several different forces. The foregoing discussion suggests that the theory of a thermal air temperature limit of viability seems to work in broad terms but may only be sufficient to provide a explain or predict a general pattern of change, but will be insufficient to give a more precise predictive capacity for any individual ice shelf. The prediction by Vaughan et al. (1993) that Wilkins Ice Shelf would collapse in thirty years was arguably 50% wrong, in that much of the ice shelf has already been lost. However, it is questionable whether that ad hoc prediction could be much improved today. The basis for prediction continues to be based on observing the broad patterns and expecting them to continue, rather than on fundamental understanding or improvements in modelling. The ice shelf dynamics we have discussed above (including ice shelf geometry, pinning points and pressure zones; surface melt and texture; mass balance, ice source, thickness and velocity) and external forces such as atmospheric and ocean temperature and sea-ice extent, impact the timing and style of collapse/retreat. We must undertake many more studies and sensitivity tests such as those undertaken by Vieli et al. (2007), before we can build an effective predictive model. Perhaps progress is not best served by over-focussing on a few examples when there are many to consider. A better understanding of the many distinct cases of ice shelf retreat around the AP could also improve our modelling and projections of future changes in West Antarctic Ice Sheet ice shelves.

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TCD
3, 579–630, 2009

Ice shelf changes on the Antarctic Peninsula

A. J. Cook and D. G. Vaughan

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Table 1. Summary of changes observed in ten ice shelves located on the Antarctic Peninsula. The figures were obtained from references that recorded the measured area of a particular ice shelf on both the earliest and most recent dates available.

Ice Shelf	First recorded date	Last recorded date	Area on first recorded date (Km^2)	Area on last recorded date (Km^2)	Change (Km^2) remaining	% of original area	Reference
Müller	1956	1993	80	49	−31	61	(Ward, 1995)
Jones	1947	2003	25	0	−25	0	(Fox and Vaughan, 2005)
Wordie	1966	1989	2000	700	−1300	35	(Doake and Vaughan, 1991a)
George VI	1974	1995	~26 000	~25 000	−993	96	(Luchitta and Rosanova, 1998)
Wilkins	1990	1995	~17 400	~16 000	−1360	92	(Luchitta and Rosanova, 1998)
	1995	1998			−1098	85	(Scambos et al., 2000)
Prince Gustav	1945	1995	2100	~100	−2000	5	(Cooper, 1997)
	1995	2000		47		2	(Rott et al., 2002)
Larsen Inlet	1986	1989	407	0	−407	0	(Rott et al., 2002)
Larsen A	1986	1995	2488	320	−2168	13	(Rott et al., 1996)
Larsen B	1986	2000	11500	6831	−4669	59	(Rott et al., 2002)
	2000	2002		3631	−3200	32	(Scambos et al., 2004)
Larsen C	1976	1986	~60 000	~50 000	−9200	82	(Skvarca, 1994; Vaughan and Doake, 1996)

TCD
3, 579–630, 2009

Table 2. Coastlines used to compile ice shelf fronts came from different source material. These are from specific years and have different reliability ratings, as defined below (reliability shown in brackets). See Ferrigno et al. (2006) for a description of the source material contributing to each reliability rating.

Ice shelf	1950s	1960s	1970s	1980s	1990s	2000s	2008/2009
Müller	1957[1]	1969[1]	1974[1]	1986[1]	1996[1]	2001[1]	2008[1]
Jones	1957[1]	1969[1]	1978[1]	1986[1]	1996[1]	2001[1]	2008[1]
Wordie	1947[2]	1966[2]	1974[2]	1989[1]	1997[1]	2001[1]	2008[1]
Wilkins	1947[3]			1990[1]	1997[3]	2001[1]	2008/2009[1]
George VI (N)	1947[1]	1966[3]	1974[3]	1989[1]	1997[2]	2001[1]	2008[1]
George VI (S)	1947[3]	1968[2]	1973[2]	1986[1]	1997[2]		2008[1]
Bach	1947[3]	1968[3]	1973[2]	1986[1]	1997[2]	2001[1]	2008[1]
Stange			1973[2]	1986[1]	1997[2]	2001[1]	2008[1]
Prince Gustav (N)	1957[1]	1957[1]	1977[3]	1988[1]	1993[2]	1997[2]	2008[1]
Prince Gustav (S)	1957[5]	1961[3]	1961[3]	1989[2]	1993[2]	1997[2]	2008[2]
Larsen A	1961[3]	1963[1]	1975[2]	1986[1]	1995[2]	2000[1]	2008[1]
Larsen B		1963[1]	1975[2]	1986[1]	1999[2]	2002[1]	2008[1]
Larsen C		1963[2]	1975[2]	1988[1]	1997[2]	2001[2]	2008[1]
Larsen D		1966[2]		1988[1]	1997[2]	2000[1]	2008/2009[1]

[1] Within 60 m [2] Within 150 m [3] Within 300 m [4] Within 600 m [5] Within 1 km

TCD

3, 579–630, 2009

Ice shelf changes on the Antarctic Peninsula

A. J. Cook and D. G. Vaughan

Title Page
Abstract
Introduction
Conclusions
References
Tables
Figures
Back
Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

Table 3. Changes in area of ice shelves on the Antarctic Peninsula. The total loss for each ice shelf is calculated from the earliest recorded area to the most recently measured area in 2008/2009. The % of ice remaining is also based on the earliest recorded areas. N. B. The decade "2000s" consists of ice front positions between 2000–2002, with the latest position "2008/2009" given separately.

Ice shelf	1950s	1960s	1970s	1980s	1990s	2000s	2008/2009	Total change (km^2)	% remaining
	Areas in km^2								
Müller	78	69	60	64	45	44	40	−38	51
Jones	29	31	36	26	21	10	0	−29	0
Wordie	1420	1917	1538	827	906	312	139	−1281	10
Wilkins	16 577			15 986	14 694	13 663	11 144	−5434	67
George VI	25 984	25 806	25 249	24 707	24 260		24 045	−1939	93
Bach	4798	4721	4825	4685	4582	4562	4487	−311	94
Stange			8286	8148	8030	7949	8022	−264	97
Prince Gustav	1632	1299	1328	1019	665	276	11	−1621	1
Larsen A	4021	3736	3873	3394	926	638	397	−3624	10
Larsen B		11 573	11 958	12 190	8299	4429	2407	−9166	21
Larsen C		56131	58036	50241	51246	51593	50837	−5295	91
Larsen D		21 716		22 372	22 345	21 851	22 602	886	104
Total Area	152 246	151 862	153 483	143 661	136 020	129 589	124 128		
Total Change		−384	1621	−9823	−7640	−6432	−5461	−28117	82

TCD

3, 579–630, 2009

Ice shelf changes on the Antarctic Peninsula

A. J. Cook and D. G. Vaughan

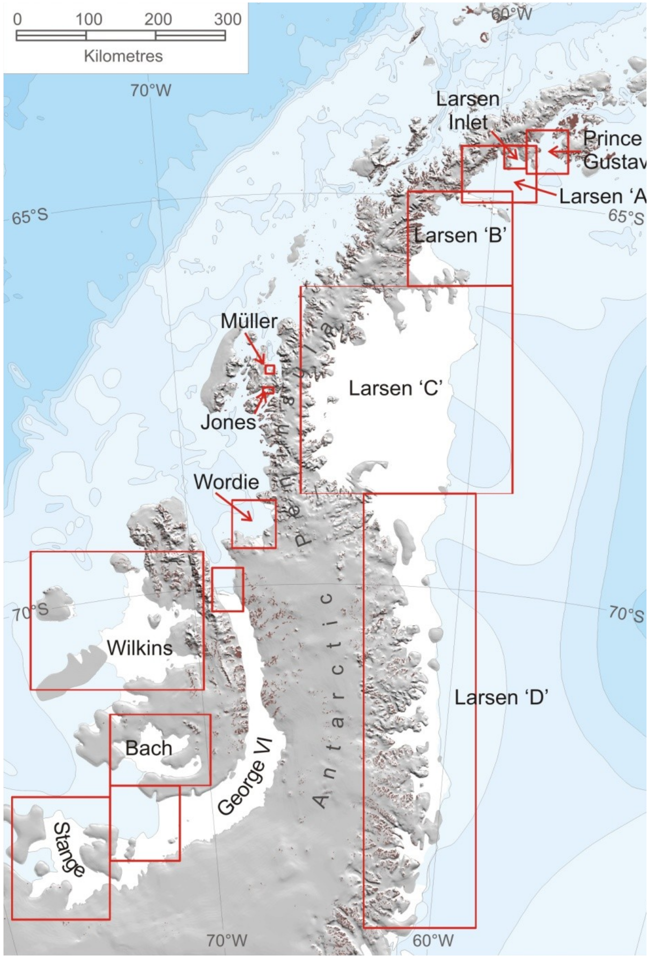


Fig. 1a. Location of ice shelves on the Antarctic Peninsula.

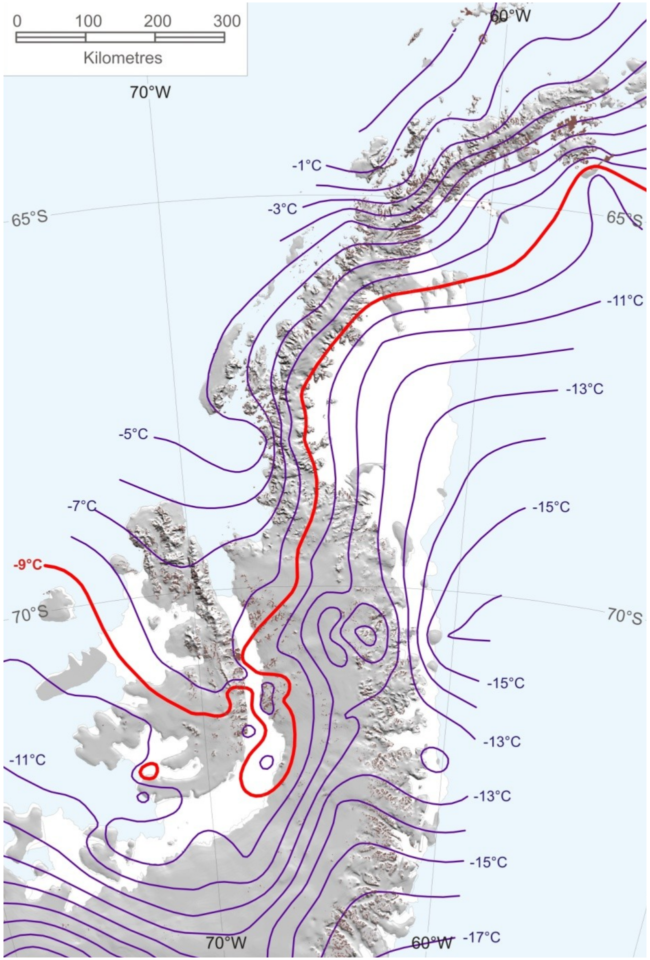


Fig. 1b. Contours of interpolated mean annual temperature in 2000 A.D., as compiled by Morris and Vaughan (2003).

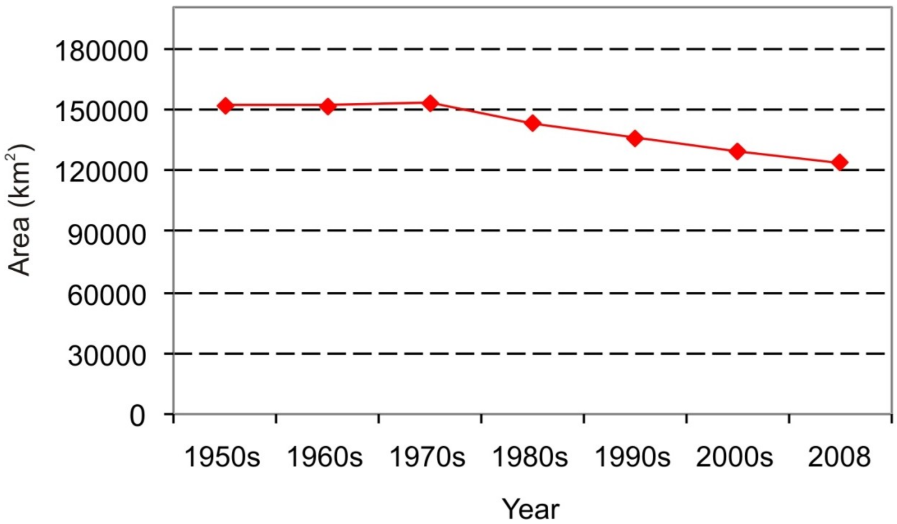


Fig. 2. Total area of floating ice on the Antarctic Peninsula over the past 5 decades.

TCD
3, 579–630, 2009

Ice shelf changes on the Antarctic Peninsula

A. J. Cook and
D. G. Vaughan

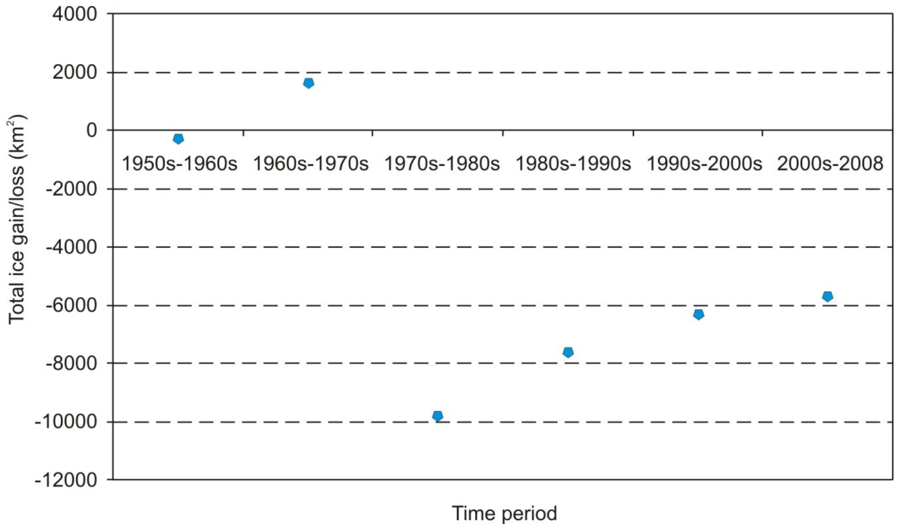


Fig. 3. Total amount of floating ice lost between each decade.

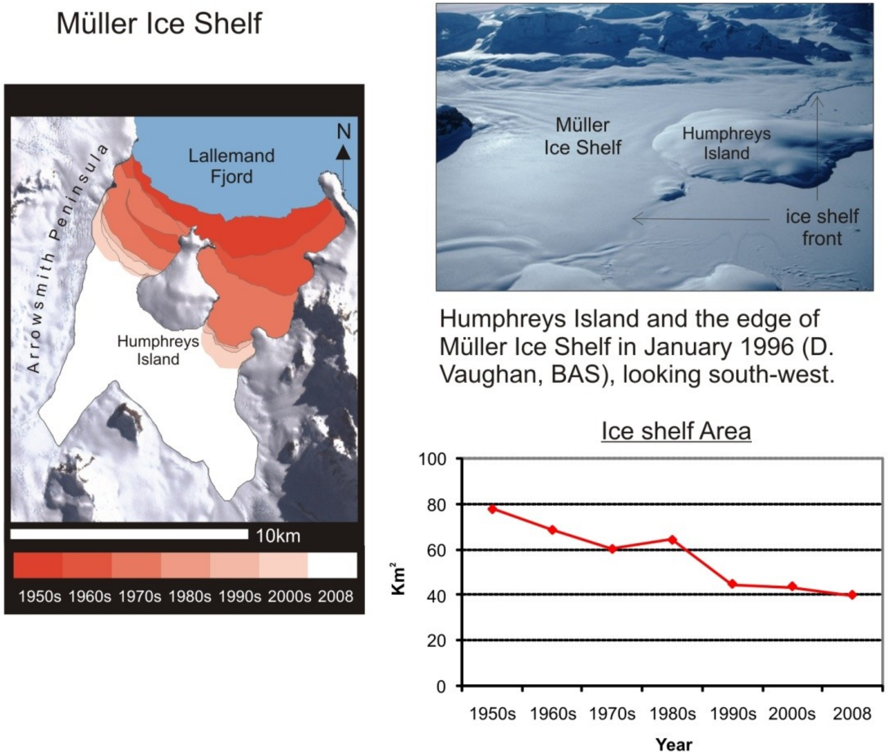
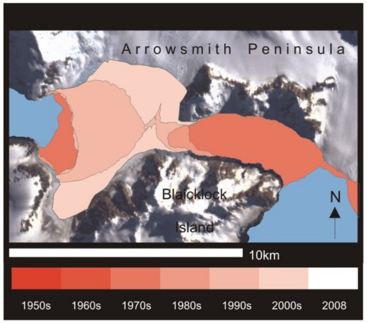


Fig. 4a. Müller Ice Shelf

TCD
3, 579–630, 2009

Jones Ice Shelf



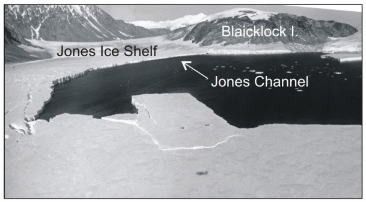


Image of Jones Ice Shelf taken during its retreat in 1998 (photographer unknown), looking east. Two depots of fuel drums are visible on the calving iceberg in the centre of the frame.

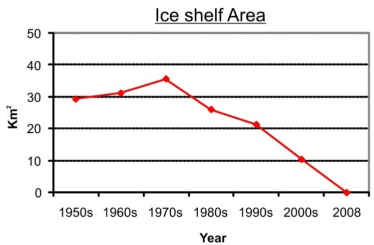


Fig. 4b. Jones Ice Shelf

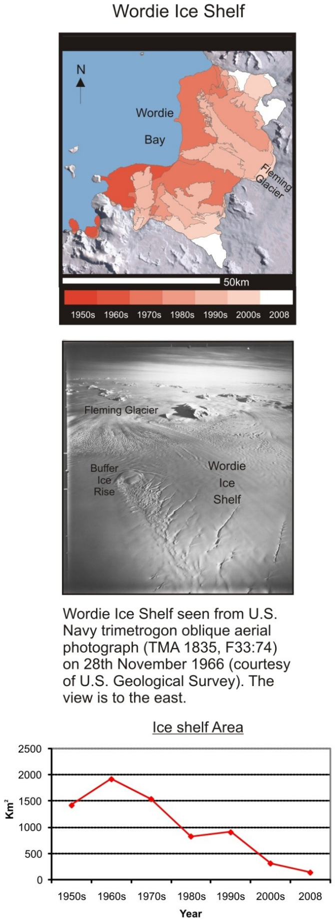


Fig. 4c. Wordie Ice Shelf

TCD

3, 579–630, 2009

Ice shelf changes on the Antarctic Peninsula

A. J. Cook and D. G. Vaughan

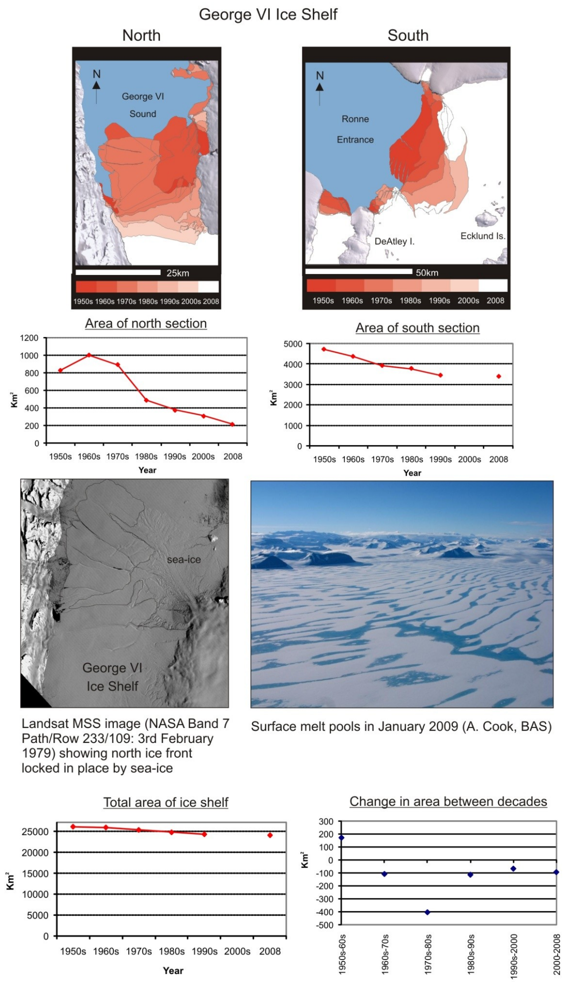


Fig. 4d. George VI Ice Shelf

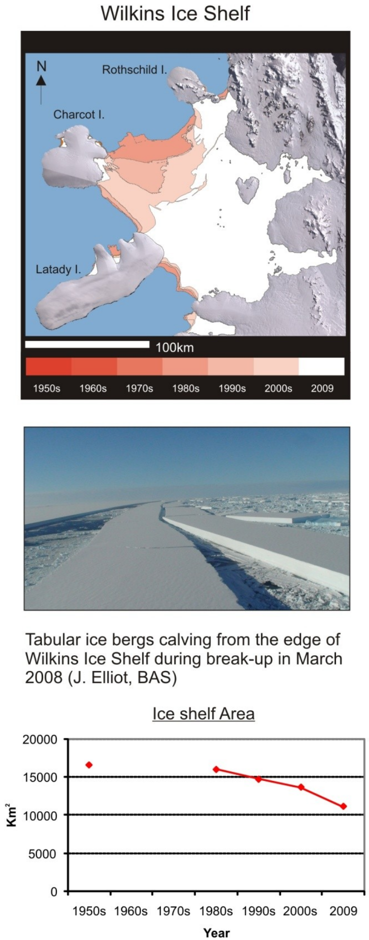


Fig. 4e. Wilkins Ice Shelf

TCD

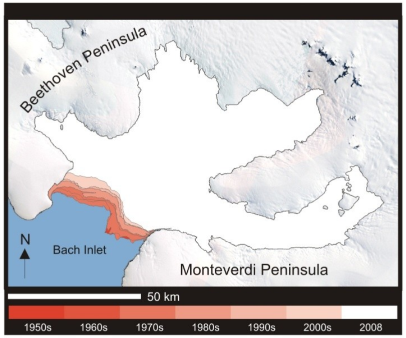
3, 579–630, 2009

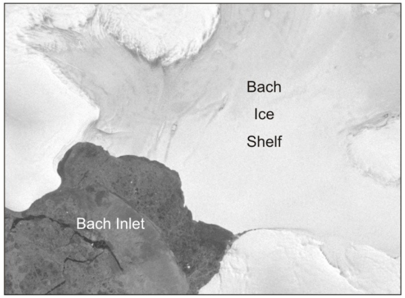
Ice shelf changes on the Antarctic Peninsula

A. J. Cook and D. G. Vaughan

Title Page
Abstract
Introduction
Conclusions
References
Tables
Figures
◀
▶
Back
Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

Bach Ice Shelf





Radarsat Antarctic Mapping Project (1997) image of Bach Ice Shelf front. Image courtesy of National Snow and Ice Data Center (Liu et al., 2001).

Ice shelf Area

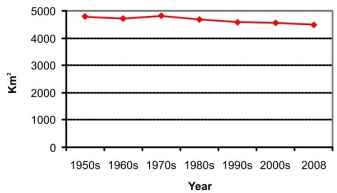


Fig. 4f. Bach Ice Shelf

TCD
3, 579–630, 2009

Ice shelf changes on the Antarctic Peninsula

A. J. Cook and D. G. Vaughan

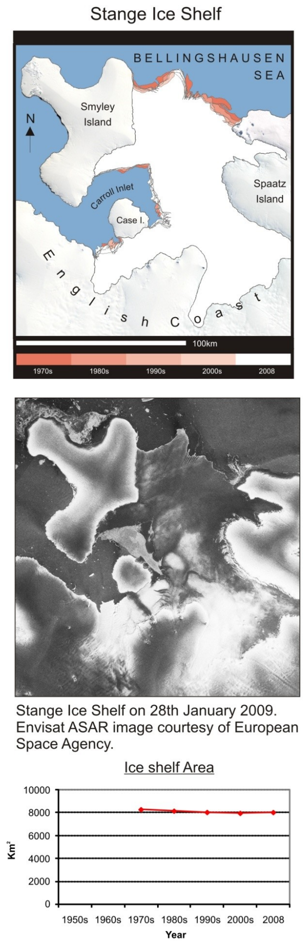


Fig. 4g. Stange Ice Shelf

TCD

3, 579–630, 2009

Ice shelf changes on the Antarctic Peninsula

A. J. Cook and
D. G. Vaughan

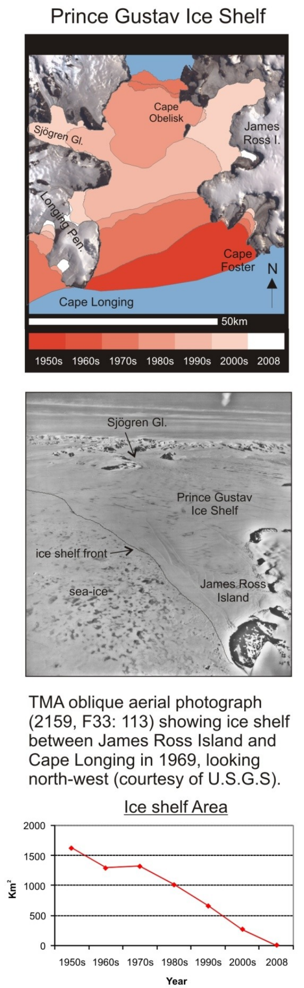


Fig. 4h. Prince Gustav Ice Shelf

TCD

3, 579–630, 2009

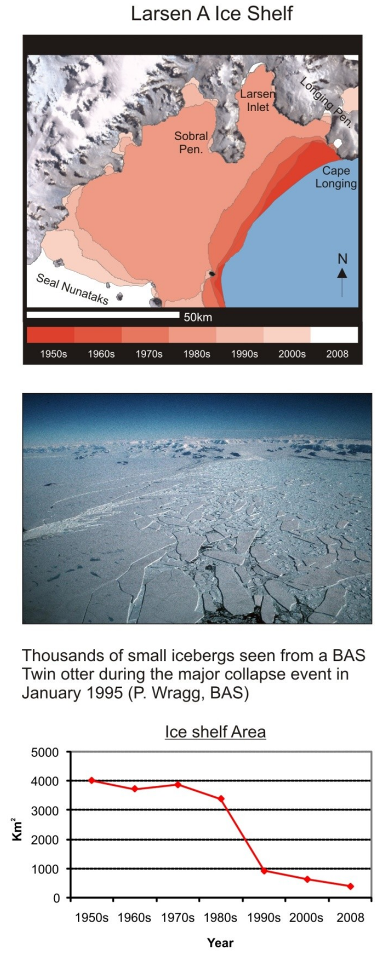


Fig. 4i. Larsen A and Larsen Inlet Ice Shelves

TCD
3, 579–630, 2009

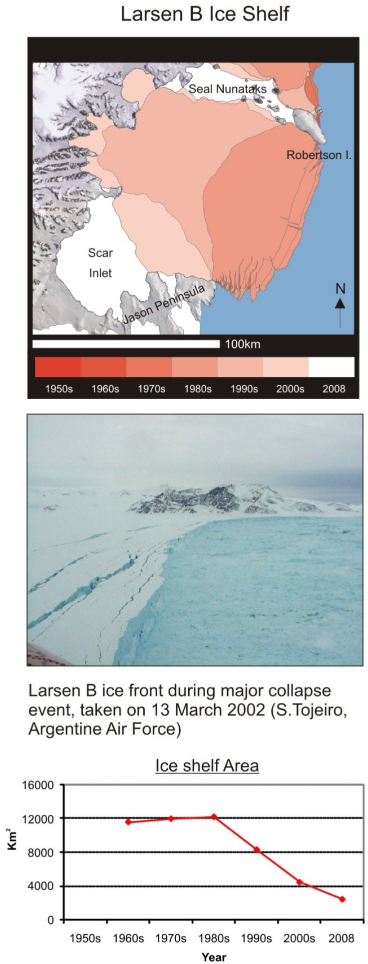


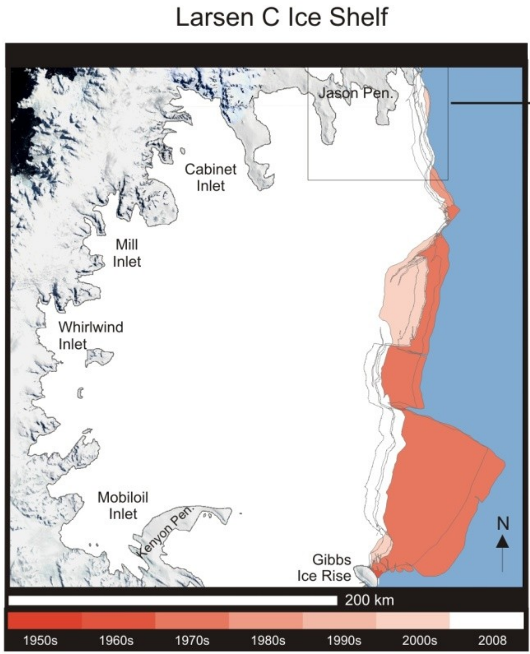
Fig. 4j. Larsen B Ice Shelf

TCD
3, 579–630, 2009

Ice shelf changes on the Antarctic Peninsula

A. J. Cook and
D. G. Vaughan

TCD
3, 579–630, 2009
Ice shelf changes on the Antarctic Peninsula
A. J. Cook and D. G. Vaughan



Radarsat Antarctic Mapping Project (1997) image showing surface features surrounding Jason Peninsula. Image courtesy of National Snow and Ice Data Center (Liu, H. et al., 2001).

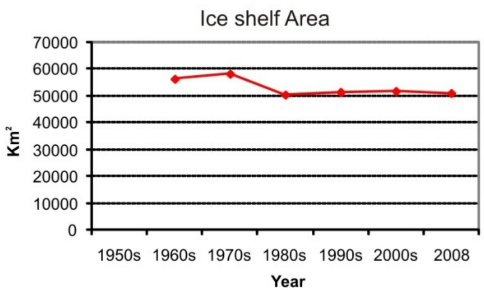


Fig. 4k. Larsen C Ice Shelf

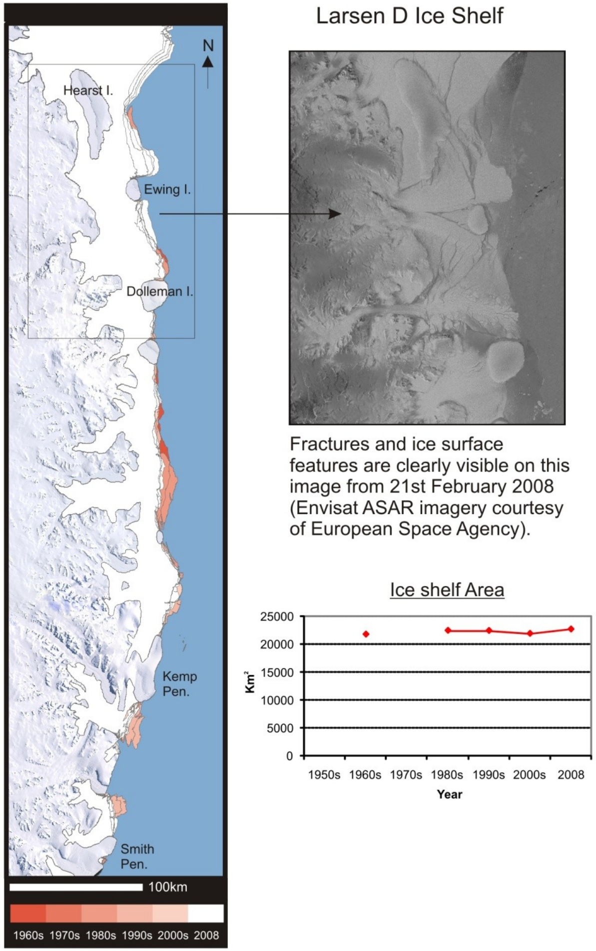


Fig. 4l. Larsen D Ice Shelf

TCD
3, 579–630, 2009

Ice shelf changes on the Antarctic Peninsula

A. J. Cook and D. G. Vaughan