Synchronous terminus change of East Antarctic outlet glaciers linked to climatic forcing

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In recent years there has been a growing trend of acceleration, thinning and retreat of ocean-terminating outlet glaciers in both the Greenland and West Antarctic Ice Sheets, resulting in an increasing contribution to sea level rise. Similar changes in glacier elevation have been observed in major marine basins of the much larger East Antarctic Ice Sheet, but there are few measurements of glacier terminus positions. In this study, the frontal position of 175 marine terminating glaciers along a 5,400 km stretch of the East Antarctic coastline is analysed. Overall, between 1974 and 2010 there was little change in glacier frontal position (median: 0.7 m a^{-1}). However, strong decadal trends have been observed, from 1974 to 1990, the majority of glaciers retreated (63%), whereas between 1990 and 2000, most glaciers advanced (72%), a pattern which dropped to fewer glaciers advancing (58%) in the most recent decade (2000 to 2010). The patterns in glacier frontal position change are consistent with a rapid and coherent response to atmospheric and oceanic/sea-ice forcing, which are ultimately driven by the dominant mode of atmospheric variability in the Southern hemisphere, the Southern Annular Mode. This indicates that the East Antarctic Ice Sheet may be more vulnerable to climate change than previously recognised. However, unlike in Greenland, there appears to be no clear link between recent changes in glacier elevation and frontal position, possibly due to the unconstrained (~90%) nature of the majority of glaciers in the study area.

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Introduction

1.1 Significance of Ice Sheets

It is important to monitor the state of the Earth's ice sheets, as they affect sea level and have the potential to actively influence climate (Clark *et al.*, 1999) through positive feedbacks associated with complex oceanic and atmospheric systems. To ascertain accurate predictions of future sea level contribution from ice sheets it is important to monitor the behaviour of outlet glaciers, because they account for a large proportion of ice discharge. (Rignot and Kanagaratnam, 2006; Howat *et al*., 2007; Moon *et al*., 2012). In recent years, there has been a growing trend for the acceleration of flow, thinning and recession of the world's major outlet glaciers and ice streams in both Greenland (Luckman *et al*., 2006; Howat *et al*., 2007 & Holland *et al*., 2008) and West Antarctica (Vaughan *et al*., 2010; Jenkins *et al*., 2010; Tinto & Bell, 2011). These changes are largely thought to be driven by thinning and/or retreat of the glacier termini (Nick et al., 2009) due to warming oceanic conditions (Holland et al., 2008; Jacobs et al, 2011). The subsequent reduction in buttressing initiates a series of positive feedbacks up glacier (Moon and Joughin, 2008), ultimately leading to enhanced mass loss and an increase in sea level rise contribution (van den Broeke *et al.*, 2009; Rignot *et al.*, 2011a; Moon *et al*., 2012). In addition, atmospheric warming has led to increased surface melt in Greenland (Mote, 2007), the reduction in area of ice shelves in the Antarctic Peninsula (Cook and Vaughan, 2010) and the retreat of glaciers there (Cook et al., 2005). Combined, this has led to the prediction that over the forthcoming decades, ice sheets will become the dominant contributor to sea level rise, potentially exceeding IPCC (2007) predictions (Meehl *et al*., 2007; Rignot *et al*., 2011;a).

1.2 Remote Sensing of Ice Sheets

Due to the remote and inhospitable nature of ice sheets and resources associated with fieldwork (Bolch, 2007), remote sensing is needed to acquire comprehensive, frequent and uniform global observations of glacier and ice sheet change at frequent intervals (Kargel *et al*., 2005). In the most recent decade, technological advancements have enabled the measurement of glacier terminus position (Frezzotti, 2002; Howat et al., 2006; Moon and Joughin, 2008), glacier velocity (Rignot et al., 2011a), elevation change (Pritchard et al., 2009; Flament and Remy, 2012) and mass loss (Chen et al., 2009). Despite the relatively small time period in which these data have been available, it has been invaluable in improving the understanding of recent ice sheet changes. However, the majority of this research focus has been concentrated in Greenland and West Antarctica, with relatively few studies on East Antarctica (Fig. 1.1).

1.3 East Antarctica

Despite East Antarctica representing the largest body of ice on the planet, equivalent to ~60 m sea level rise (SLR) (Lythe and Vaughan, 2001), it has been relatively sparsely researched in comparison to the much smaller ice sheets of Greenland, ~7 m SLR (Church & White, 2006) and West Antarctica, ~3-4 m SLR (Bamber *et al*., 2009), see Figure 1.1. Indeed, the large uncertainties in current mass balance estimates of the East Antarctic Ice Sheet (EAIS) (Rignot *et al*., 2008; 2011;a), highlight the need for further research, as any change in the EAIS mass balance, either positive or negative, is likely to have significant implications for future sea level change and subsequent mitigation strategies.

The lack of research focus on the EAIS may be partly attributed to its remoteness and the long standing belief that it is stable because of its relative isolation from climatic perturbations (Thomas, 1979; Alley & Whillans, 1984). However, recent studies using satellite altimetry, have demonstrated in some regions, there are dynamic changes occurring in major marine-based basins in East Antarctica (Pritchard *et al*., 2009; Flament and Remy, 2012). Studies in both Greenland and West Antarctica have shown that similar changes in glacier elevation can be linked to the retreat of glaciers, thinning of floating tongues and mass loss (Moon and Joughin, 2008). However, to date, there have only been a few small scale studies analysing the frontal position of East Antarctic outlet glaciers (e.g. Frezzotti et al., 1997; 1998; 2002). These show some evidence of cyclic behaviour and either no obvious trend or a reduction in glacier floating area since the 1950s. This study aims to bridge this gap by providing a valuable and comprehensive dataset of spatial changes in outlet glacier terminus position, over the past 50 years which is used to improve our understanding of the current state the EAIS.

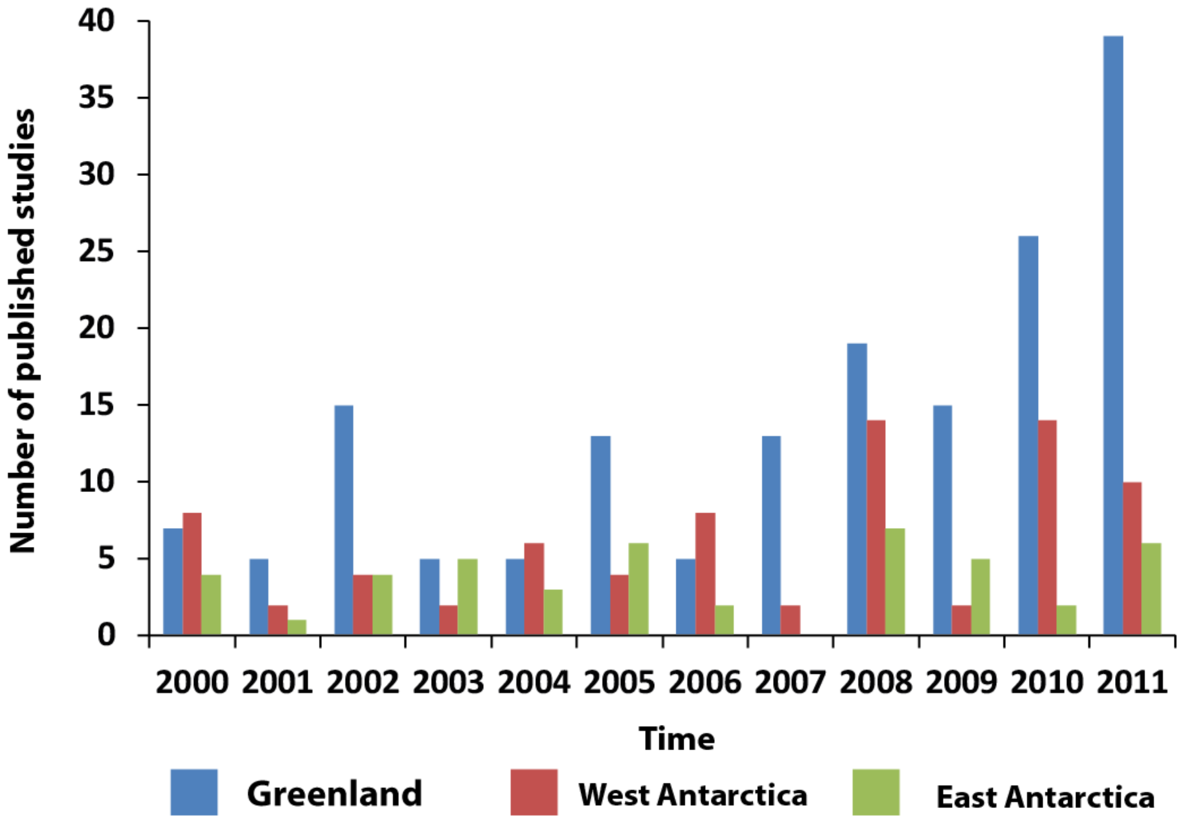


Figure 1.1: Number of published studies related to outlet glaciers in Greenland, West Antarctic and East Antarctic Ice Sheets. Data were taken from a series of searches on 'Web of Knowledge'; the search terms were: "Greenland" AND "outlet glacier" OR "outlet glaciers", "West Antarctica" OR "West Antarctic" OR "Antarctic Peninsula" AND "outlet glacier" OR "outlet glaciers" and "East Antarctic" OR "East Antarctica" AND "outlet glacier" OR "outlet glaciers".

1.4 Thesis aims and objectives

This project aims to examine and compare decadal scale fluctuations in outlet glaciers along a 5,400 km stretch of the EAIS margin and to determine their relationship to climatic perturbations and elevation change. To achieve this, the following specific objectives have been developed:

- To map changes in the terminus position of all glaciers within the study area at several approximately decadal time steps between 1963 and 2010.
- To assess the spatial and temporal trends in outlet glacier terminus position.
- To examine the relationship between glacier terminus position and available climatic/oceanic data.
- To explore the relationship between changes in glacier terminus position, velocity and size (width).

- To assess the relationship between glacier terminus position change and changes in glacier elevation, which will be extracted from Pritchard *et al*. (2009).

1.5 Study area

The study area consists of a 5,400 km stretch of the East Antarctic coastline stretching from 90° E (Shackelton Ice Shelf) to 175° E (Ross Ice Shelf) (Fig. 1.2). This section of coastline is thought to contain around 200 outlet glaciers, which is well suited for the purpose of this study. This region was selected because: (i) it encompasses two regions of pronounced mass loss and thinning (Wilkes and Victoria Land) (King *et al.*, 2012), (ii) large parts are grounded below sea level (Young *et al*., 2011), which may enhance its vulnerability to oceanic forcing, and (iii) the absence of large ice shelves makes individual glacier termini readily identifiable.

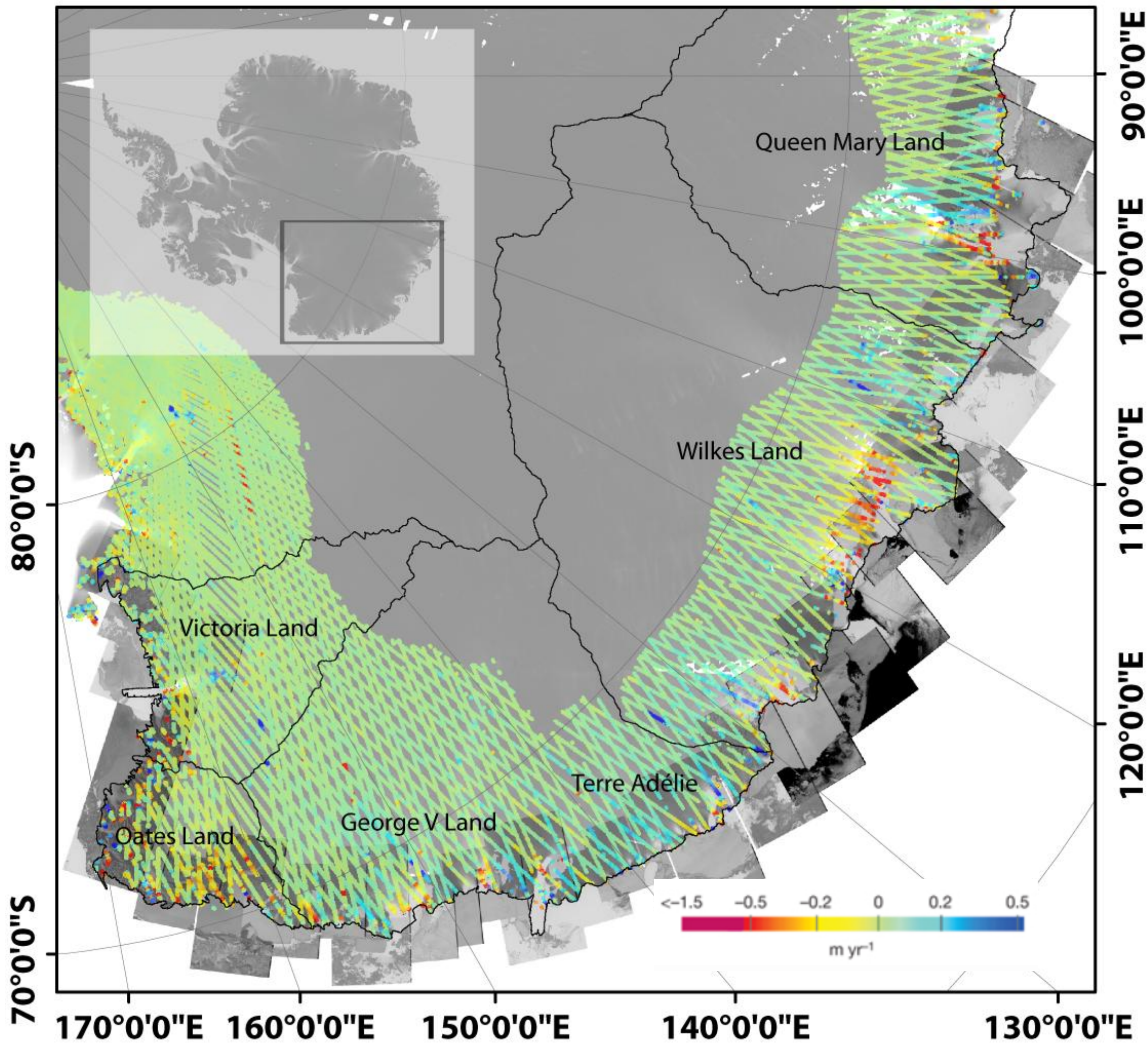


Figure 1.2: Map of the study area, with thinning rates overlain (red = thinning, blue = thickening) (Taken from Pritchard *et al*., 2009).

<u>A review into recent outlet glacier and ice sheet trends</u>

2.1 Introduction

Marine terminating outlet glaciers are more vulnerable to rapid change than their land terminating counterparts (Rignot *et al*., 2002). This can be attributed to their complex interaction with both the atmospheric and oceanic systems. This section will first review the global importance of dynamic glacier change. This is followed by a review of the processes which initiate changes in outlet glacier behaviour and the mechanical processes which may take place. Secondly, this chapter will analyse recent trends in ice sheet behaviour, focusing on the Greenland, West Antarctic and East Antarctic ice sheets.

2.2.1 Wider importance and drivers of outlet glacier evolution

Recently, the long standing view that variations in ice sheet mass balance are dominated by short term variations in surface balance has been raised (Howat *et al*, 2007) by the recent, dramatic increase in ice discharge from many of Greenland's outlet glaciers (e.g. Rignot and Kanagaratnam, 2006; Howat *et al*., 2007; Howat *et al*., 2008; Moon *et al*., 2012). Indeed, changes in the ice dynamics of outlet glaciers are now thought to be crucial for the recent negative trends in ice sheet mass balance (Rignot and Kanagaratnam, 2006), particularly in both the East and West Antarctic Ice sheets where it is largely deemed too cold for surface melt to significantly influence mass balance (Stearns *et al*., 2008).

In Greenland, it has been suggested that the mechanical process driving these changes in marine terminating outlet glaciers could be the downward propagation of lubricating meltwater to the glacier bed, resulting in an acceleration in glacier flow (Zwally *et al*., 2002; Parizek and Alley, 2004 Bartholomew *et al*, 2010). However, recent modelling work by Nick *et al*. (2009) has shown this to be unlikely, suggesting a more probable scenario of changing conditions at the calving terminus subsequently initiating of a series of positive feedbacks upstream (Howat *et al*., 2005, 2008; Joughin *et al.*, 2008). Specifically, it is hypothesized that complex changes in the ice/ocean interaction at the glacier terminus can lead to a reduction in ice buttressing (Nick *et al.*, 2009). Subsequently, this causes an increase in the near terminus velocity, which results in longitudinal stretching (thinning) of the near terminus. This increases the inland slope of the glacier, further accelerating glacier flow, driving the propagation of dynamic thinning up glacier and, ultimately, mass loss (Moon and Joughin, 2008; Vieli and Nick, 2011).

This highlights the importance of the interaction between the glacier calving front, bed geometry, lateral constraints, and the oceanic and atmospheric systems.

2.2.2 Conditions at the calving terminus and glacier de-buttressing

Changes in the magnitude of ice buttressing are ultimately driven by fluctuations in climate, which initially change conditions at the glacier terminus. The absolute magnitude of change in ice buttressing is thought to be driven by the individual glacier geometry (Vieli *et al*., 2001; Trabant *et al*., 2003). This is shown in the outlet glaciers of western Greenland, where a high degree of variability has been observed in the timing and magnitude of glacier change. Since these glaciers are undergoing a similar level of external forcing, it is presumed that the difference in glacier behaviour may be caused by variability in glacier geometry (McFadden *et al*., 2011). The flow of constrained marine terminating outlet glaciers is primarily resisted by lateral drag, so that thinning or the loss/removal of the floating ice tongue will reduce the contact area between glacier and fjord/ice walls, resulting in a reduction of resistive stresses and an increase in glacier flow (Thomas, 2004; McFadden *et al*., 2011; van der Veen *et al*., 2011); initiating the same set of positive feedbacks described in section 2.2.1. However, unconstrained glaciers are not buttressed by lateral forces and, consequently, their flow speed is not resisted by lateral drag (Rignot, 2008). Therefore, for unconstrained glaciers, thinning or changes in the extent of floating tongues is likely to have a minimal effect on glacier velocity (McFadden *et al*., 2011).

2.2.3 Calving

Calving, the mechanical loss of ice from marine terminating glaciers is an important component of the mass budget system, particularly in Antarctica (Jacobs *et al*., 2002). Calving takes place at the terminal face of a marine terminating glacier because the outward-directed cryostatic pressure is greater than the backward-directed hydrostatic pressure (Benn *et al*., 2007). Calving events occur at inherent lines of weaknesses (crevasses) which form mainly in response to longitudinal stresses. The calving rate is largely believed to be dictated by the flow speed of the glacier (van der Veen, 1996; 2002; Benn *et al*., 2007). In general, glacier velocity of flux is non-linearly proportional to glacier size (van der Veen and Whilians, 1996) and therefore larger glaciers tend to have a higher calving rate than smaller glaciers. Indeed, the largest calving events are formed from the release of tabular icebergs from the ice shelves and large floating tongues e.g.

Mertz, Denman, David glaciers, which fringe Antarctica (Lazzara *et al*., 1999; Joughin and MacAyeal, 2005). These events are thought to be highly episodic (Larour *et al*., 2004; Bassiss *et al*., 2005), with periodicities of around 50 years (Frezzotti *et al*., 2002). Whereas for smaller glaciers with no floating tongues calving is a more consistent process. However, calving rates can also change rapidly in response to fluctuations in climate (Rignot *et al*., 2003). The mechanisms in which changes in the climatic system can influence calving are summarised below (Vieli and Nick, 2011):

- **Basal melting** (thinning) of a floating glacier from warmer oceanic water can remove part of the lower layer of less brittle ice, making the glacier vulnerable to calving (MacAyeal and Thomas, 1986). Thinning near the glacier front can bring the glacier closer to floatation, leading to the break-up of the floating tongue and retreat (Howat *et al*. 2008; Pfeffer, 2007).
- **Meltwater driven crevasse propagation** has the potential to drive rapid changes in glacier frontal position (Rott *et al*., 1996; Scambos *et al*., 2000). This mechanism is driven by the influx of surface meltwater into surface crevasses. Due to the density difference between water and ice, crevasses propagate downward through the glacier or ice shelf, prompting fracturing and calving of elongated icebergs (van der Veen, 1998; Scambos *et al*., 2003, 2009). The subsequent overturning of these icebergs is likely to lead to a positive feedback initiating further calving (Guttenberg *et al*., 2011; Burton *et al*., 2012). Therefore, the calving rate is highly sensitive to variations in the amount of surface melt, which in turn, has the potential to drive rapid changes e.g. Larsen B (Scambos *et al.*, 2003)
- **Sea ice** at the glacier terminus exerts a back pressure, which resists the cyrostatic outward pressure of the calving face of a glacier. Indeed, analysis at Jakobshavn Isbrae, Greenland, has shown that sea ice growth and the stiffening of the proglacial ice mélange, during the winter prevents the calving of full thickness icebergs (Amundson *et al*., 2010). The stabilizing effect of such ice-mélange on calving has also shown to be relevant for Antarctic outlet glaciers and ice shelves (Larour *et al.*, 2004). Therefore, the relative concentration of sea ice at the glacier terminus is an important factor in controlling rate of calving.

2.3 Recent changes in ice sheets

2.3.1 Greenland

The most recent estimates of the current mass loss of the Greenland ice sheet (GIS) vary from 267 GT year $^{-1}$(Rignot *et al*., 2008) to 286 GT year $^{-1}$ (Velicogna, 2009). Around half of this mass loss has been attributed to surface melt as a result of warming temperatures (Box and Cohen, 2006; Mote, 2007; Hanna *et al*., 2008), whilst changes in glacier dynamics is thought to contribute the other half of the mass loss (Luckman *et al*., 2005, 2006; Howat *et al*., 2007). This has resulted in considerable attention being directed towards the GIS. In particular, there has been extensive research focus on Greenland's three largest outlet glaciers Jakobshavn Isbrae, Helheim and Kangerdlugssuaq, which between them drain 40% of the GIS (Bell, 2008). Jakobshavn Isbrae exhibited a sudden change from slow thickening to rapid thinning in 1997, associated with the velocity of its floating tongue doubling to 12.6 km yr^{-1} (Thomas *et al*., 2003; Joughin *et al*., 2004). Helheim and Kangerdlugssuaq glaciers showed a similar distinct change, albeit in 2000 (Holland *et al*., 2008; Howat *et al*., 2008). Additional analysis of Greenland's other major outlet glaciers reveals a similar trend of acceleration, thinning and retreat across the whole ice sheet margin (e.g. Moon & Joughin, 2008; Box, 2009; Thomas *et al*., 2009; Box & Decker, 2011; Howat & Eddy, 2011). The spatial extent of thinning between 2003 and 2007 can be seen in Figure 2.1 (Pritchard *et al*., 2009).

Some researchers argued that an increase in air temperature was the main factor driving this dramatic change (Zwally *et al*., 2002). It was suggested that warmer air temperatures may have led to more widespread surface melt, which when routed to the bed lowers basal friction, leading to an increase in ice velocity (Zwally *et al*., 2002). However, a recent study by Sole *et al*. (2008) suggests that the rapid retreat and thinning of many tidewater glaciers exceeds the expected changes induced from air temperatures alone. They analysed data from 25 land and tidewater terminating glaciers, arguing, that retreat and thinning rates should be the same if basal lubrication is the driving force behind the rapid retreat. However, tidewater glaciers thinned significantly faster between 1993 and 2006, with retreat increasing from 1998 onwards; no such change was seen in land terminating glaciers. This suggests that enhanced basal lubrication is not the main driving force behind the acceleration of Greenland's outlet glaciers. This is confirmed by recent modelling from Nick *et al*. (2009) which shows that dynamic glacier changes begin at the calving terminus and propagate up glacier, see section 2.2.1. Furthermore, Holland *et al* (2008) have shown that oceanic warming is likely to play a key part in changing conditions at the calving terminus. They show a clear correlation between the onset of warmer oceanic waters in the coastal areas of Greenland and the onset of rapid thinning and retreat in the late 1990s. In contrast, air temperatures in Greenland have been steadily increasing over the past two decades and have not undergone a rapid change (Holland *et al*., 2008).

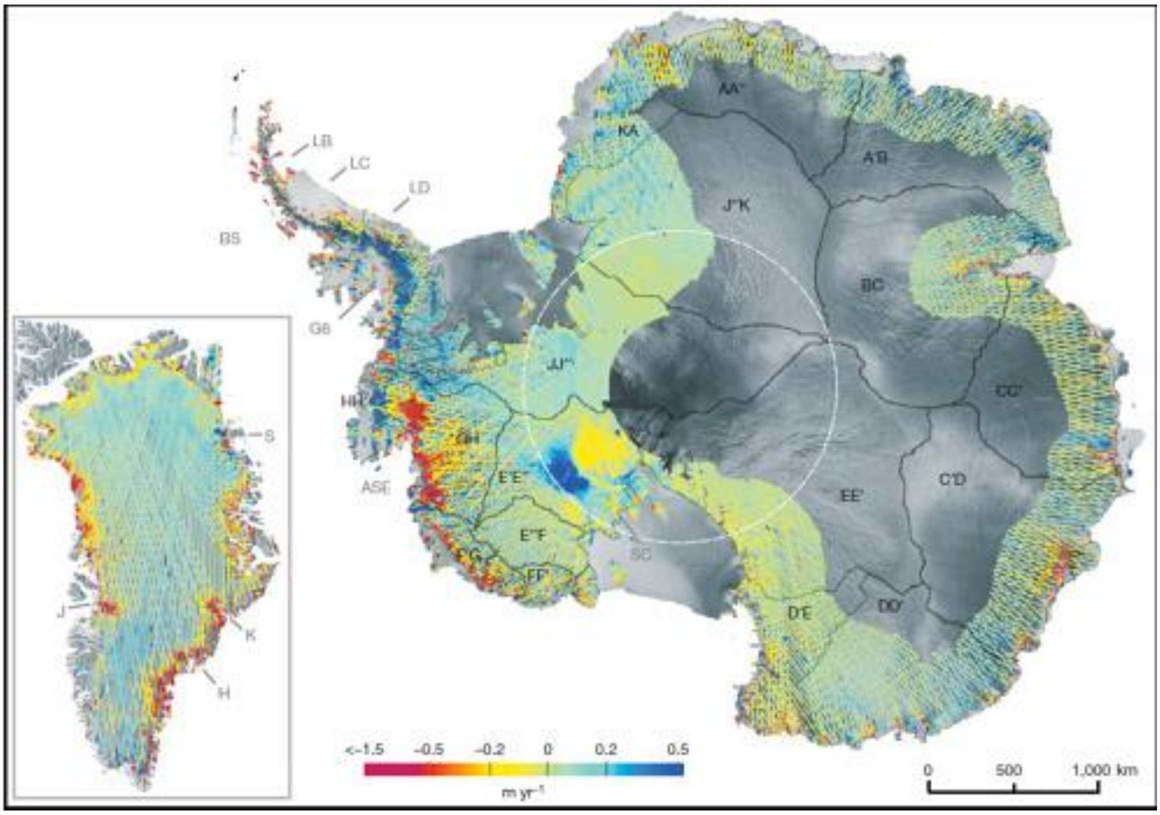


Figure 2.1: Thinning rates of the Greenland and Antarctic ice sheets (Taken from Pritchard *et al*., 2009).

2.3.2 Antarctica

Recent estimates for the mass loss in Antarctica range from -31 GT year^{-1} (Zwally and Giovinetto, 2011), -190 GT year^{-1} (Chen *et al*, 2009) and -246 GT year^{-1} (Veliconga *et al*., 2009), indicating considerable uncertainty. The large range mainly originates from uncertainties in the modelling of mass change due to glacial isostatic adjustment (King *et al*., 2012). However, a new glacial isostatic adjustment model from King *et al*. (2012), developed from geological constraints has improved the accuracy of these estimates. They estimate a lower mass balance of -69 +- 18 GT year^{-1}, between August 2002 and December 2010.

However, the mass change is not spatially consistent, with East Antarctica gaining mass (60 GT year^{-1}) and West Antarctica losing mass (117 GT year^{-1}). Indeed, even within each respective ice sheet there is considerable spatial variation, shown in Figure 2.2. In East Antarctica, most drainage basins are gaining mass between 0 and 90° E, whereas drainage basins between 90 and 180° E appear to be stable or losing mass. In West Antarctica, over 90% of total mass loss is concentrated on the Amundsen Sea coast (drainage basins 20-23), in areas drained by ice shelves with beds well below sea level.

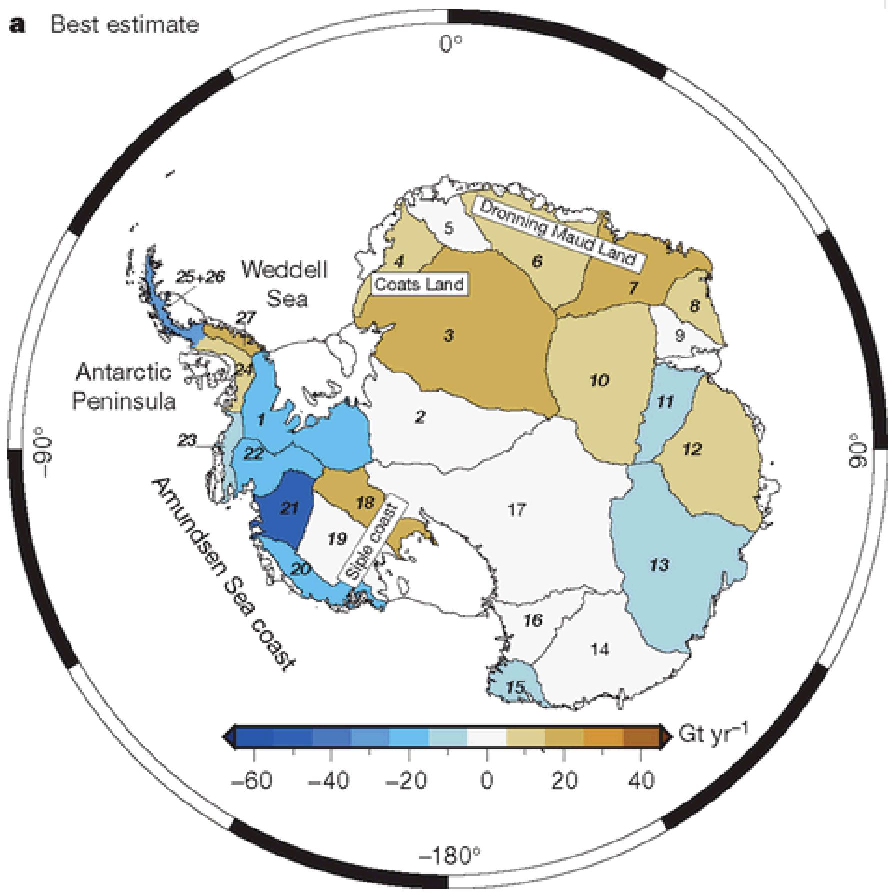


Figure 2.2: Best estimate of mass balance change in Antarctica between 2002 and 2010 (Taken from King *et al*., 2012).

2.3.2.1 West Antarctic Ice Sheet and Antarctic Peninsula

Recent changes in the West Antarctica have been driven by different factors including the increase in air temperature e.g. Antarctic Peninsula (Cook *et al*., 2005) and oceanic warming e.g. Pine Island bay (Jacobs *et al*., 2011). The Antarctic Peninsula (AP), is often cited as the most rapidly warming region on the planet, with the air temperature increase estimated at 3.7 +/- 1.6°C over the past century (Vaughan *et al*., 2003). Indeed, Cook *et al* (2005) linked this warming with the retreat of marine terminating glaciers and ice shelves in the AP (n = 244), with

87% of glaciers retreating over the study period (1940-2001). This link to air temperature is confirmed by the progressively southward migration of the boundary between mean advance and mean retreat as temperature increases. This suggests a rapid response of the frontal position of marine terminating glaciers to atmospheric warming. Indeed, this rapid response to climatic changes is shown by the collapse of the Larsen B ice shelf (Scambos *et al*., 2003), in which 3,200 km^2 of ice calved away in little over 5 weeks (Skvarca and De Angelis, 2003). This event is widely believed to be caused by the increased downward propagation of melt water (described section 2.2.1) driven by an anomalously intense and prolonged melt season which preceded its break up (van den Broeke, 2005). Overall, the warming of the AP has lead to a 28,000 km^2 reduction in ice shelf area over the past 50 years (Cook & Vaughan, 2010). This is important as the loss of the buttressing ice shelves in the AP has been associated with the acceleration and increasing ice discharge of the glaciers which feed them, which is important for sea level rise (Scambos *et al*., 2004; Rott *et al*., 2011).

In contrast to the AP, recent changes in the Amundsen Sea coast of the WAIS have been largely attributed to oceanic warming (Jacobs *et al*., 1996, 1997). Recently, there has been much attention in one of the WAIS three major drainage systems which flows into Pine Island Bay, following the discovery that the Pine Island Glacier's (PIG) buttressing ice shelf was melting rapidly in 1994 (Jacobs *et al*., 1996). This melting was attributed to anomalously warm deep water on the Amundsen Sea continental shelf (Jacobs *et al*., 1996, 1997). This resulted in the thinning of PIG's floating ice shelf and the retreat of its grounding line into deeper water (Shepherd *et al*., 2001, Pritchard *et al.,* 2012). It is believed that this has led to the observed acceleration of inland thinning of 3.1 m a^{-1} in 1995 to 10 m a^{-1} in 2006 (Wingham *et al*., 2009) and an increase in velocity of the PIG (Shepherd *et al*., 2001, Joughin *et al*., 2003; Joughin *et al*., 2010). Indeed, further research from Jacobs *et al*. (2011) has shown that the relative temperature and volume of Circumpolar Deep Water (CDW) in Pine Island Bay has increased, resulting in a 50% increase in melt water production and the formation of a cavity between the bottom of the ice shelf and the transverse ridge which it once rested on, exposing the grounding line to further melt (Jenkins *et al*., 2010; Jacobs *et al*., 2011). Mapping of the PIG bed has shown that it is likely that the positive feedback of thinning ice retreating into deeper waters will result in a further 200 km of retreat, until the next topographic pinning point (Jenkins *et al*., 2010); unless a series of complex glacial interactions lead to its stabilisation e.g. convergent ice flow or basal sediment forming a new pinning point (Benn *et al*., 2007).

The large scale changes observed in Pine Island Bay are considered by some, as evidence confirming the long standing belief that the WAIS is inherently unstable and vulnerable to catastrophic collapse through marine instability (Weertman, 1974; Mercer, 1978; Oppenheimer & Alley, 2004). This is because the majority of grounded ice in the WAIS lies on a bed that deepens inland and extends significantly below sea level. It is hypothesized that if the buttressing floating ice shelves of West Antarctica are lost (Thomas *et al*., 1979; Thomas *et al*., 2004; Dupont & Alley, 2006), the subsequent acceleration and thinning of ice in the ice sheet interior could lead to unstable retreat of the grounding line of WAIS glaciers due to its inward sloping bed topography (Bamber *et al*., 2009). This would lead to mass loss from the WAIS, with the potential for rapid sea level rise (Weertman, 1974; Rignot, 2008; Joughin & Alley, 2011).

2.3.2.2. East Antarctica

In comparison to the Greenland and West Antarctic Ice Sheets, East Antarctica has long been considered stable and less susceptible to melt beyond its marine based margins (Thomas, 1979; Sugden *et al*., 1993 and Pollard & DeConto, 2009). Nevertheless, uncertainty still remains into the effect of the landward dipping Aurora and Wilkes subglacial basins which extend significantly below sea level (Young *et al*., 2011), see Figure 2.3. This raises the possibility of marine instability in these areas of East Antarctica (Pollard & Deonto, 2009; Passchier, 2011). Recent research by Young *et al*. (2011) reconstructed the subglacial topography of the Aurora basin and revealed a previously unknown subglacial mountain range, which contained a series of preserved glaciated valleys which connect the coastal grounding zone with the continental interior. These valleys are important as they potentially aid the delivery of warmer CDW which can increase basal melting (Wahlin *et al*., 2010; Roberts *et al*., 2011).

Recent studies have suggested that some parts of East Antarctica are thinning and losing mass (e.g. Pritchard *et al*, 2009; King *et al*., 2012), see Fig. 2.1 and section 2.3.2. It is noted that the majority of this thinning and mass loss is concentrated between 90 and 165° E (Pritchard *et al*., 2009; Chen *et al*., 2009; King *et al*., 2012). Despite this, there have only been a couple of small scale studies analysing glacier frontal position change. On the Oates, George V and

Adélie coasts Frezzotti *et al*. (1998, 2002) note a reduction in the area of floating glaciers from the 1950s to the 1980s on the small number of glaciers measured (n = ~30). However, Frezzotti *et al*. (1998, 2002) also record a small increase in the area of floating ice from the late 1980s until the mid 1990s. Furthermore, Frezzotti (1997) shows a decrease in the area of floating ice from the 1950s until the 1970s and an increase between the 1970s and 1990's in Victoria Land. It is important to note that glacier change here appears to be a level of magnitude smaller than in the Oates, George V and Adélie coasts. The pattern of retreat was attributed to changes in the ice-ocean interaction, with Frezzotti *et al*. (2002) and De la Mare (1997) both noting a reduction in sea ice and fast ice between the 1950s and 1970s. Most of the glaciers measured are characterized by cyclic behaviour, with the growth of the glacier over approximately 15-50 years, followed by a calving event (Frezzotti *et al*., 2002). Although, there are hints that some of these glaciers still follow the overall trend (Frezzotti *et al*., 2002). Therefore, it appears that there has been a general reduction in the area of floating ice between the 1950s and 1970s, with Kim *et al*. (2001) also noting a reduction in ice shelf area in the Dronning Maud Land. For most areas, this pattern of retreat continued until the 1990s where there is a brief switch to advance, however, there have been no further studies using larger sample sizes.

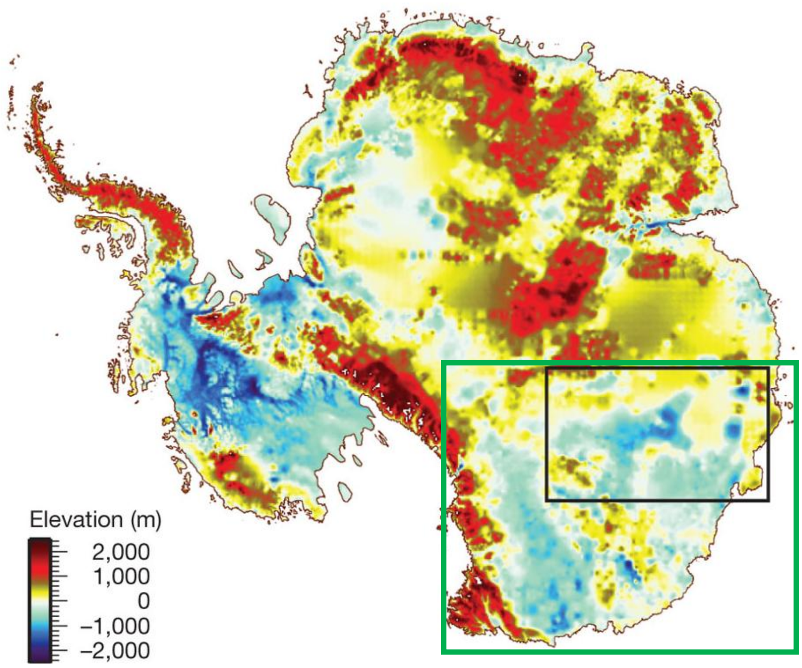


Figure 2.3: Bed topography of Antarctica, with the blue areas representing the subglacial basins which oceanic water can penetrate, image taken from Young *et al.*, 2011. The green box has been added highlighting the study area.

Methods

3.1 Introduction

In any large scale study focusing on ice sheet changes, satellite and aerial imagery are key as they represent the most realistic time and cost effective opportunity to obtain data. (Kargel *et al*., 2005; Bolch, 2007). Therefore, research focusing on glacier frontal changes is largely constrained by the availability of such imagery. The purpose of this chapter is to describe the types of imagery and processing methods used to map glacier frontal changes in East Antarctica.

3.2 Overview of Data Sources

Previous studies have used a range of different types of satellite imagery to study glacier frontal fluctuations, including Landsat (Moon and Joughin, 2008; Nye, 2010; Howat and Eddy, 2011), ASTER (Joughin *et al*, 2008), ARGON (Frezzotti *et al*., 2002), SAR (Csatho *et al*., 2002; Frezzotti *et al*, 1998) and Corona (Csatho *et al*., 2002; Joughin *et al*, 2008). However, each satellite is optimized to perform different tasks and their resolution and coverage throughout space and time is not uniform. In this study, virtually all mapping was done using Landsat imagery, as these images are freely available (http://www.glovis.usgs.gov) and provide an appropriate resolution (~30-80 m) for glacier mapping (Paul *et al*., 2002; Andreassen *et al*., 2008). An analysis of the available data revealed that the more recent Landsat satellites 4 & 5 Thematic Mapper (TM) and 7 Enhanced Thematic Mapper (ETM+) provide near complete coverage of the study area across the periods 1988-1991 and 2000-2010, respectively. The earlier Multispectral Scanner (MSS) satellite provide a more patchy coverage during the period 1972-1975 (Fig. 3.1), due to a loss of images through cloud cover and image corruption. It is important to note that outside the time periods detailed above there is no satellite imagery available on a large scale. Nevertheless, the imagery available enables decadal time steps to be derived, which are ideal for this study. Shorter time steps are more likely to detect inter-annual variability and stochastic calving, and are less likely to identify a meaningful trend. Therefore, decadal time steps combined with the large sample size of glaciers in this study will minimize the impact of the natural periodicity of glaciers and allow any potential climatic trend to be extracted.

In addition, a further time step was derived at 2006-2007 to look in more detail at recent glacier change. Finally, the time period was extended through the inclusion of an ARGON orthorectified mosaic of Antarctica from 1963 composed by Kim *et* *al* (2007). This enabled 6 time steps to be derived, simplified to: 1963, 1974, 1990, 2000, 2006 and 2010, see Table 3.1.

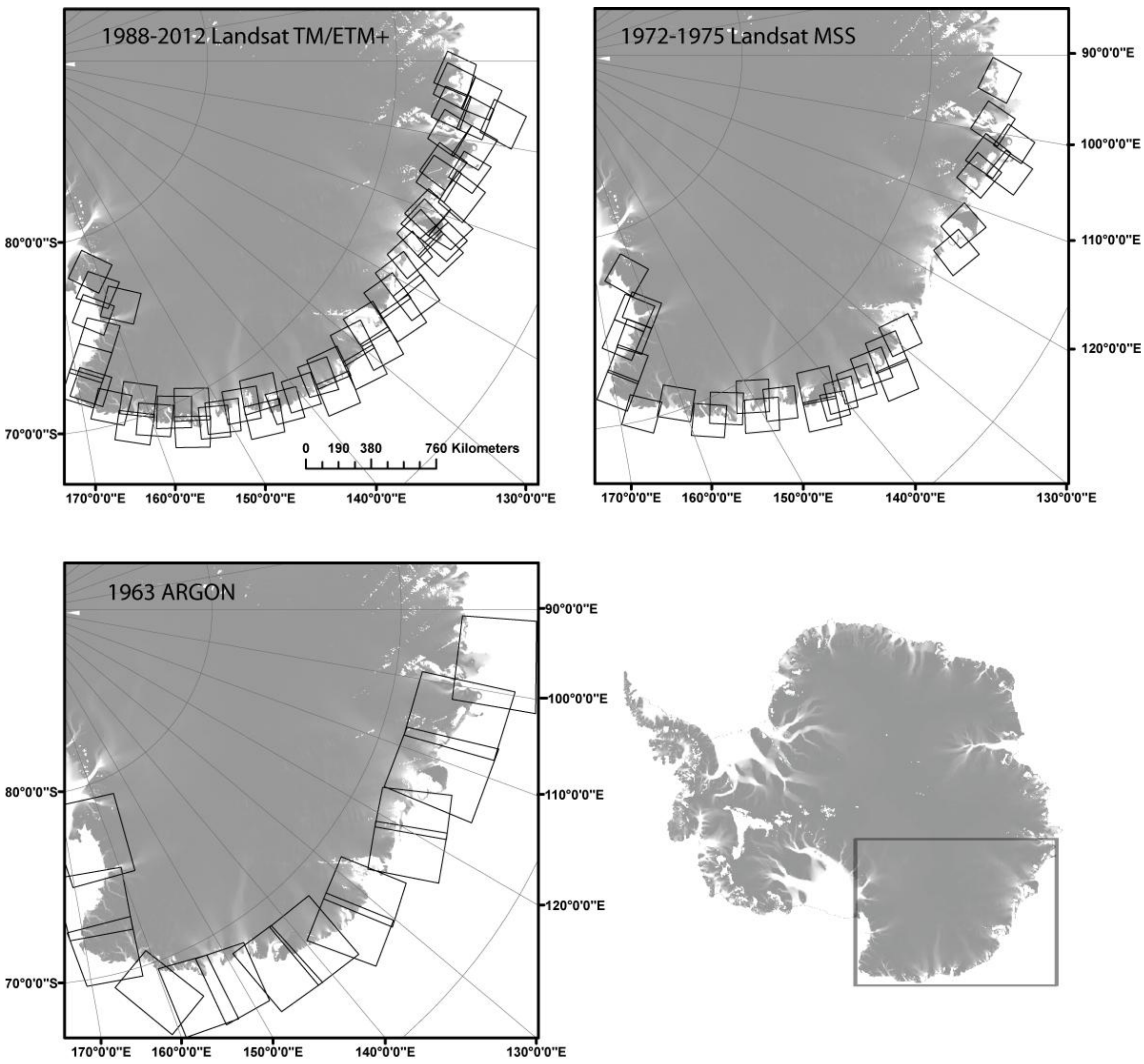


Figure 3.1: Overlay of the available images at each time period. **A)** Landsat TM/ETM+, **B)** Landsat MSS, **C)** ARGON and **D)** The study area It is important to note the gaps in image coverage in the 1972-75 time period; there is complete coverage in 1963, 1988-91, 1999-02, 2006-07 and 2009-12.

Table 3.1: Image resolution and scene size for each time step.

Time-Step	Imagery Dates	Imagery Type	Spatial Resolution (m)	Scene Size (km)
1963	1963	KH-5 Argon (missions 9058A & 9059A)	140	556 x 556
1974	1972-75	Landsat 1 (MSS)	60	170 x 185
1990	1988-91	Landsat 4/5 (TM)	30	170 x 185
2000	1999-02	Landsat 7 (ETM)	30	170 x 185
2010	2009-2012	Landsat 7 (ETM)	30	170 x 185

3.2.1 Landsat imagery

In total there have been seven Landsat satellite systems launched, with each carrying increasingly sophisticated sensors, producing better quality and higher resolution images. However, the imagery of the most recent satellite (Landsat-7 ETM+) was compromised in May 2003 when the Scan Line Corrector failed (SLC). This instrument compensates for the forward motion of the satellite, and without this, the ETM+ line of sight now traces a zig-zag pattern along the satellite track. This has resulted an estimated 22% of data loss, with the edges of each image principally effected (USGS, 2012), see Fig 3.2. Nevertheless, although this has been cited as a hindrance, many researchers mapping glacier frontal changes have been able to get around this through a combination of combining adjacent images and interpolation (e.g. Moon & Joughin, 2008; Bolch *et al*., 2010)

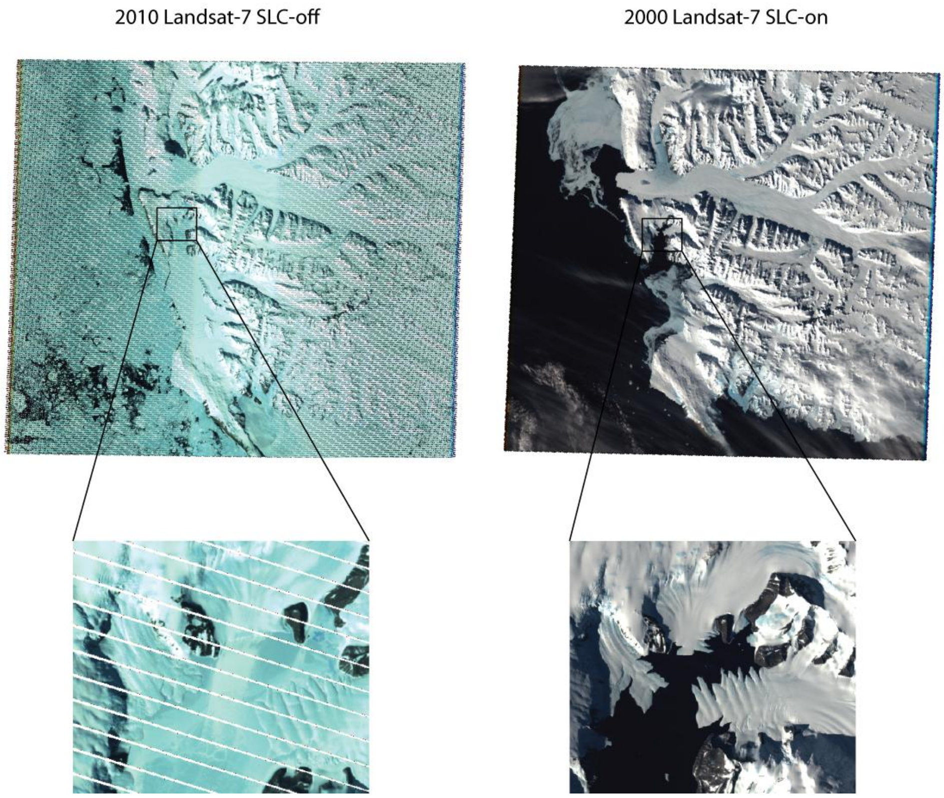


Figure 3.2: Contrast between Landsat ETM+ SLC on (right) and SLC-off (left) imagery.

3.2.2 ARGON imagery

ARGON was part of the first generation of US photo intelligence satellites along with CORONA and LANYARD, which collected 860,000 images of the Earth between 1960 and 1972. The primary use of the images was for reconnaissance for US intelligence agencies. However, since the satellites are no longer in use, the images were declassified in 1995 and made available for public use. The

ARGON satellite system carried out six successful missions between 1962 and 1964, with the 9058A and 9059A missions covering the whole of Antarctica. The satellites flew at an altitude of 322 km and in an orbit which allowed imaging to the pole, with a ground resolution of around 140 m.

3.3.1 Landsat co-registration

All Landsat images have already been rectified and transformed into the UTM WGS 84 projection by the USGS, and so only registration to a base scene was required. Here, images from the time period 2000 (ETM+) were selected as the base scene. This period was chosen as these images offer the most complete coverage of the study area, a high resolution (30 m), and are not affected by the SLC malfunction. In general, there is little or no shift between images taken with the same sensor. However, images taken with different sensors (e.g. TM to ETM+) can generate a much larger shift (Bolch *et al*., 2010).

To register images to a base scene, it is normal practice to place Ground Control Points (GCPs) on immovable or stable features visible in both images (Kaab *et al*, 2002; Silverio *et al*, 2005), with researchers recommending between 15 and 30 per image (Sidjak and Wheate, 1999; Paul and Andreassen. 2009). However, East Antarctica provides a unique challenge in that the vast ice cover leads to very few obvious GCPs in some images, which ideally would be nunataks or coastal rock outcrops (e.g. Giles *et al*, 2009). In order to get around this, individual images were merged into six mosaics covering the whole of the study area (Fig.3.3A) at each time period. In addition, to increase the amount of GCPs in each mosaic to a minimum of 20, clearly visible stable ice features such as bedrock bumps were also used as GCPs (Scambos *et al*, 1992 and Glasser *et al*, 2011). This allows images which originally did not have a suitable amount of GCPs to be co-registered and therefore included in the study, as these images can be 'absorbed' into the larger mosaic (Fig.3.3B). Generally, images were zoomed into pixel level to make sure features did not change in size or shape over time, allowing GCPs to be placed with precision.

Prior to the image transformation, those GCPs which appear to be the least accurate were deleted; generally these were the points with the highest root mean square error (RMSE). Finally, through the use of ERDAS IMAGINE software, image transformation was performed using a nearest neighbour sampling method with a polynomial transformation. After the transformation, registration accuracy was checked when the registered image was laid over the base image and gradually withdrawn across it using the ‘swipe’ tool on ARC GIS. Particular attention was paid to see if the images matched in all areas and that there was no image warping (see section 3.5 for error).

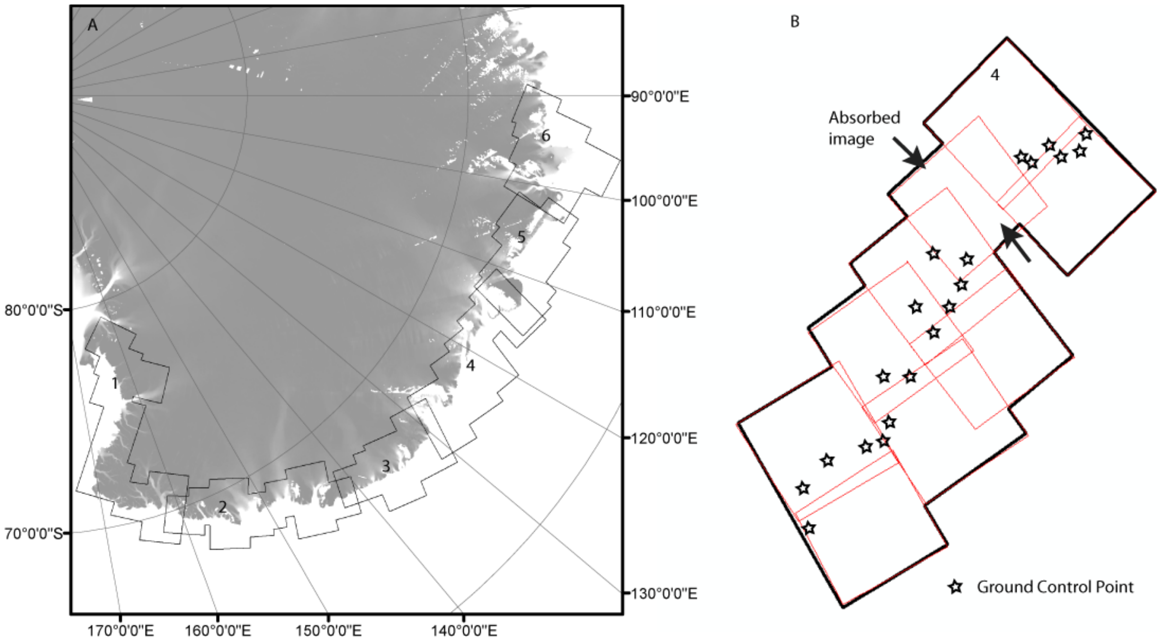


Figure 3.3: A) Overlay of the 6 mosaics created. B) Demonstrates the inclusion of an ‘absorbed image’ which does not have enough GCPs to be orthorectified alone, but can be included in a mosaic.

3.3.2 Argon co-registration

The Argon mosaic used in this study was taken from Kim *et al*. (2007). It was compiled using state of the art digital imaging methods. Kim *et al* (2007) were able to precisely assemble individual ARGON photographic images into a map quality mosaic. By means of an extended block adjustment technique they were able to determine positions of features in space with an accuracy which was comparable with the original resolution of the Argon imagery. The Argon images were orthorectified and projected in a polar stereographic format on a WGS84 ellipsoid. Further technical details on the compilation of the mosaic is provided in Kim *et al*. (2007). Before use in this study, the mosaic was opened in Arc Map and ‘swiped’ against the 2000 base scene, all features were shown to align within two pixels. Therefore, further orthorectification to the base scene was not required.

3.4 Data acquisition

3.4.1 Mapping

There is currently no definitive glacier inventory for all glaciers in this section of East Antarctica. Therefore, it was decided to map the entire 5,500 km stretch of coastline for each time period, locating individual glaciers thereafter. For the purpose of this study, the coastline can be defined as the boundary between ice attached to the main ice sheet and the ocean, regardless of whether the ice is floating or grounded. Mapping was achieved through means of manual delineation, which is the preferred method when mapping glaciers which are compromised by the presence of sea ice (Paul *et al*, 2002 and Bolch *et al*., 2007), as seen in many of the East Antarctic glaciers. All mapping was carried out using ARC MAP 10 software, generally mapped at a scale of 1:40,000 (Landsat TM/ETM+), 1:50,000 (Landsat MSS) or 1:80,000 (ARGON). However, in some cases, images were viewed at a higher or lower spatial resolution to aid mapping. All data has been stored in the form of shapefiles.

One of the key constraints in mapping the coastline accurately is distinguishing between the coastline and sea ice. By comparing the same image at different band combinations, it was shown the most appropriate band combination's were 6,4,2 for Landsat TM/ETM+ and 4,3,2 for Landsat MSS. It is important to note, that the Landsat TM/ETM+ satellites higher resolution panchromatic band (band 8) was not suitable as it proved too difficult to distinguish the glacier and sea ice. Indeed, even with a false colour composite band, it still remained impossible to distinguish between sea ice and the coastline in some areas. This led to some small sections of the coastline not being mapped. Further data loss was experienced in most time periods through excessive cloud cover, with extensive loss in the 1963 ARGON images. The combined loss of data caused by cloud cover and inability to distinguish the coastline is shown in Table 3.2.

Table 3.2: The amount of data lost at each time period

Time Period	Amount of data loss (km of coastline)	% of total (~5,500 km) coastline successfully mapped
1963	~3700	32
1974	~1900	65
1990	~500	90
2000	~440	92
2006	~440	92
2010	~500	90

3.4.2 Glacier Identification

In total, 175 glaciers were identified across the whole study area. Outlet Glaciers were identified through a combination of easily identifiable features (e.g. ice tongue or crevassing) and overlaying ice velocity data (described in 3.4.3) over the image. The SCAR Composite gazetteer (http://data.aad.gov.au/aadc/gaz/scar/) was used to name glaciers, although many of the glaciers identified were un-named. These unnamed glaciers were given their own identification numbers.

3.4.3 Calculating glacier change

Two principle methods for quantifying glacier frontal change are described in published literature. One method is to directly measure length change along the centreline of the glacier (Howat *et al*., 2005; Stokes *et al*., 2006; Kulkarni *et al*., 2007). The centreline is the recommended location for taking measurements, as it is simple to identify and tends to coincide with the location of the greatest advance/retreat (Stokes *et al*., 2006). However, if glacier change is not uniform across the glacier front, the decision as to where to take the initial measurement can severely influence results (Hall *et al*., 2003). Therefore, this method is generally recommended for land terminating glaciers where glacier frontal change tends to be more uniform. In contrast, frontal change across large marine terminating glaciers tends to be more heterogeneous as large icebergs are liable to calve at any point across the glacier front (MacGregor *et al*., 2012). Therefore, this method is not suitable for the present study.

Rather, this study uses a similar method described by Moon and Joughin (2008) known as the ‘box method’ (Fig. 3.4) and subsequently used by Howat *et al*. (2008) and MacGregor *et al*. (2012),. Here rectangles are digitized on identified glacier fronts to approximately delineate the sides of the glacier, making sure that the maximum and minimum extent of the glacier front is included in the rectangle. The shapefile containing the mapped coastline from earliest time period available is then overlain on the rectangles. Using the ‘feature to polygon’ tool in Arc Map, a new polygon is created between the mapped glacier front and the up ice edge back of the rectangle. The area of the polygon created is recorded for each glacier. This process is then repeated for each subsequent time step, with the same original rectangles used on each occasion. Glacier frontal change is calculated by dividing the difference in area of the polygons created by glacier width, which is taken as the width of the rectangle (Fig. 3.4). For example; to calculate the glacier frontal change between 1974 and 1990:

Frontal change = (1990 polygon area – 1974 polygon area) / glacier width

The main benefit of using this method is that it takes into account uneven changes across the glacier front (Moon and Joughin, 2008), which is particularly common in this study area.

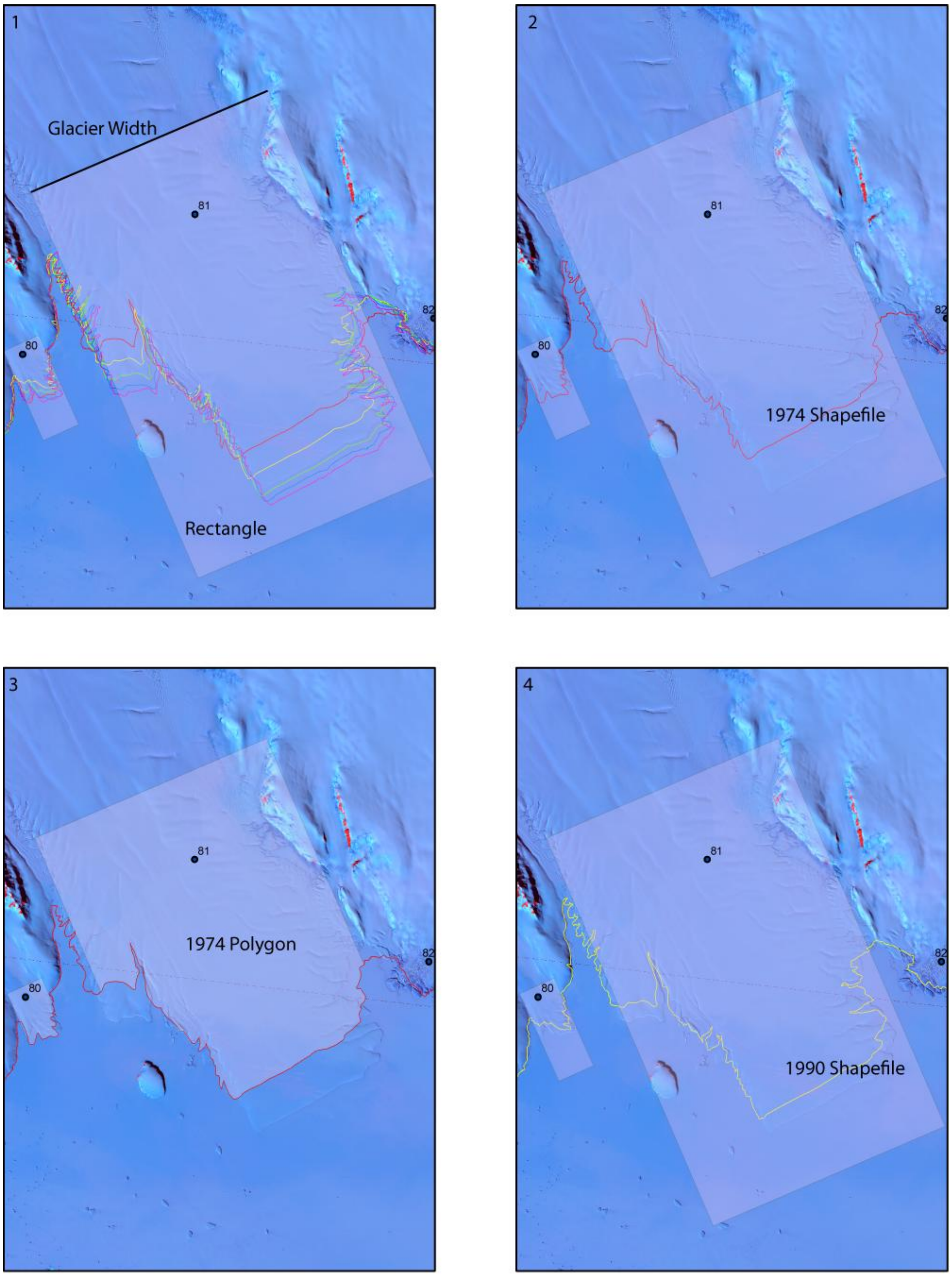


Figure 3.4: Example of the box method used for mapping glacier change: 1) Rectangle placed over the glacier, approximately delineating its sides. 2 & 3) Using Arc Map's 'spatial analyst' tool, the 1974 line is overlain over the rectangle and a new polygon is created. 4) The process is repeated using the next time period.

3.4.3 Grounding lines, ice velocity and elevation change

3.4.3.1 Grounding line

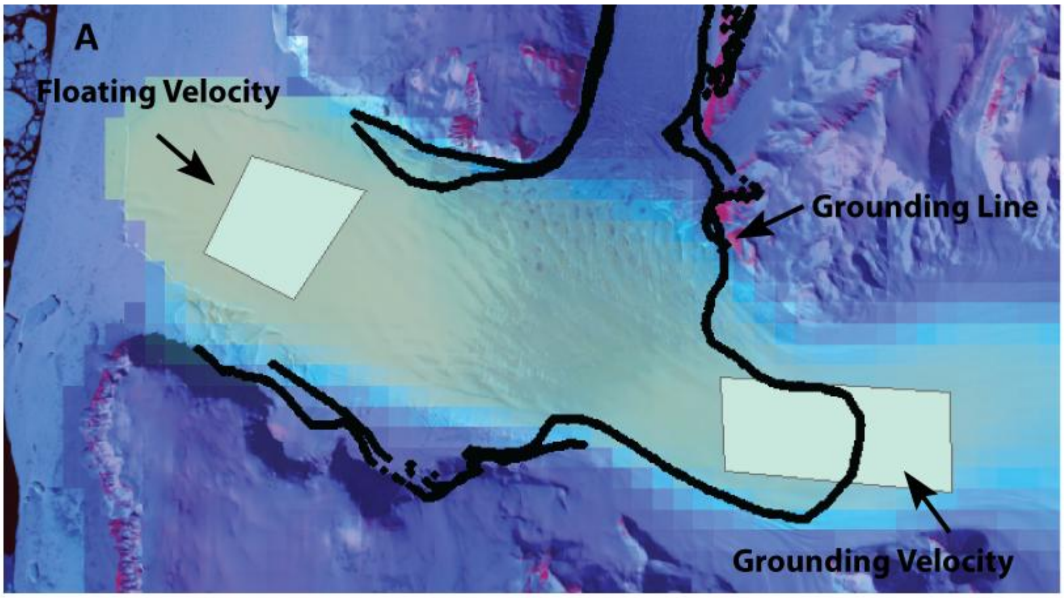
In order to aid data analysis, further secondary datasets were complied for the studies glaciers, namely grounding lines, ice velocity and elevation changes. A grounding line dataset was obtained from the National Snow and Ice Data Centre. This dataset, created by Rignot *et al*. (2011;b), was composed using differential satellite synthetic-aperture radar interferometry (DInSAR) data from the Earth remote Sensing Satellites 1-2 (ERS -1/2), RADARSAT -1 and 2, and the Advanced Land Observing System PALSAR for the years 1994 to 2009 (Rignot *et al*., 2011;a). Although the dataset's coverage is patchy in East Antarctica, it still provides valuable information**,** particularly when analysing ice velocity and elevation change, detailed below.

3.4.3.2 Ice Velocity

A high resolution, digital mosaic of ice motion in Antarctica was obtained from Rignot *et al*. (2011:b) (http://nsidc.org/data/nsidc-0484.html). It was assembled from multiple satellite interferometric synthetic-aperture radar data acquired between 2007 and 2009. The mosaic enabled the extraction of the ice velocity at each individual glacier mapped. This was achieved through digitizing a polygon on each glacier, and subsequently extracting the mean velocity value from the polygon using Arc Map's 'spatial analyst' tool. However, one of the key issues was where on the glacier to extract the velocity; on the floating section of the glacier or behind the grounding line. Preliminary extraction indicated that the trend between glacier velocity behind the grounding line and on the floating section of the glacier is virtually linear (r^2 = 0.85), see Fig. 3.5. Therefore, as the grounding line data coverage was incomplete, it was decided to take all measurements on the floating section of the glacier near the terminus.

3.4.3.3 Ice Elevation change

An ice elevation change dataset was obtained from Pritchard *et al*. (2009), who used the high resolution ICESat laser altimetry to map changes along the entire grounded margins of the Antarctic ice sheet between 2003 and 2007. The dataset enabled the extraction of elevation change data at individual glaciers to compare with changes in ice front position. Polygons were digitized just behind the grounding line of those glaciers where grounding data was available. As the resolution of the dataset was low, with measurements limited to satellite ground track crossover points, care was taken to assure that only glaciers with a suitable data crossover path were selected, see Fig. 3.6. This led to elevation change data being extracted for only 24% of glaciers. The spatial extent of the elevation coverage can be seen in Figure 3.7.



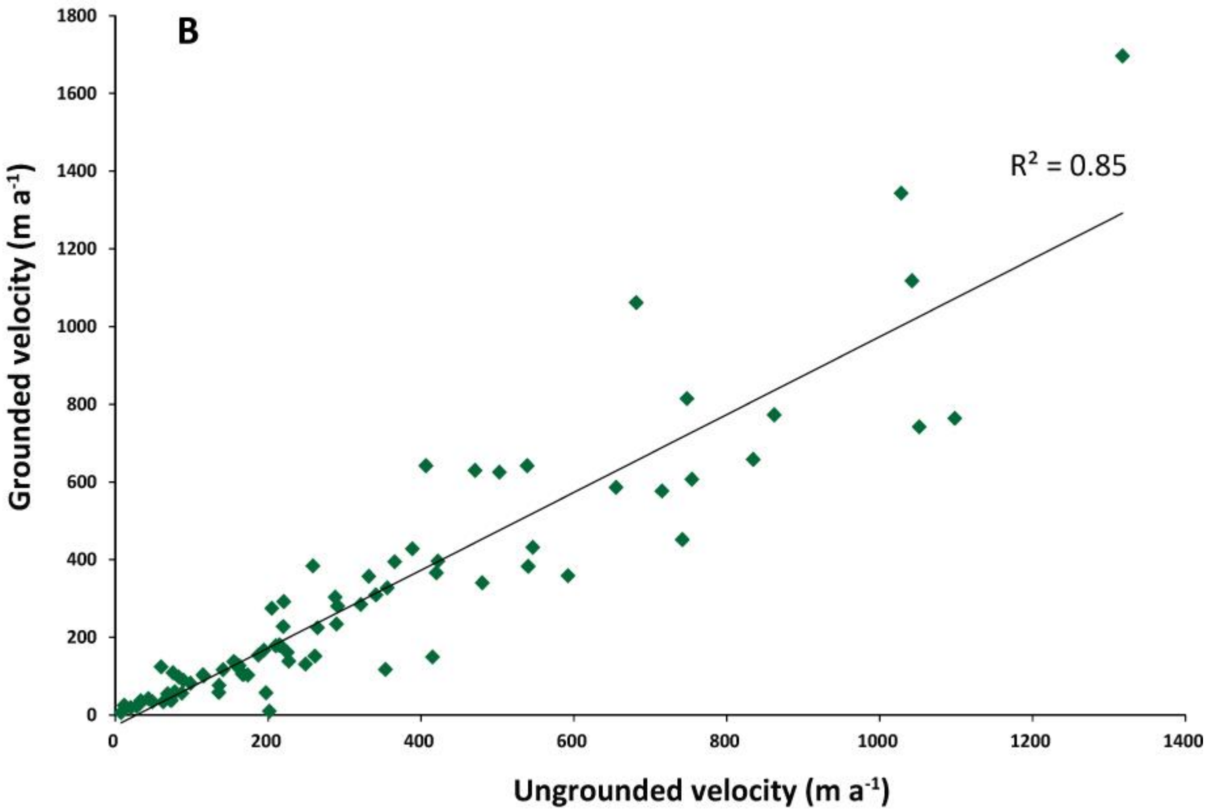


Figure 3.5: Extraction of flow speed data from Rignot (2011;b): a) Illustration showing where the grounding and floating velocity measurements were extracted. b) Scatter plot demonstrating the near linear relationship between flow speed extracted above and below the grounding line.

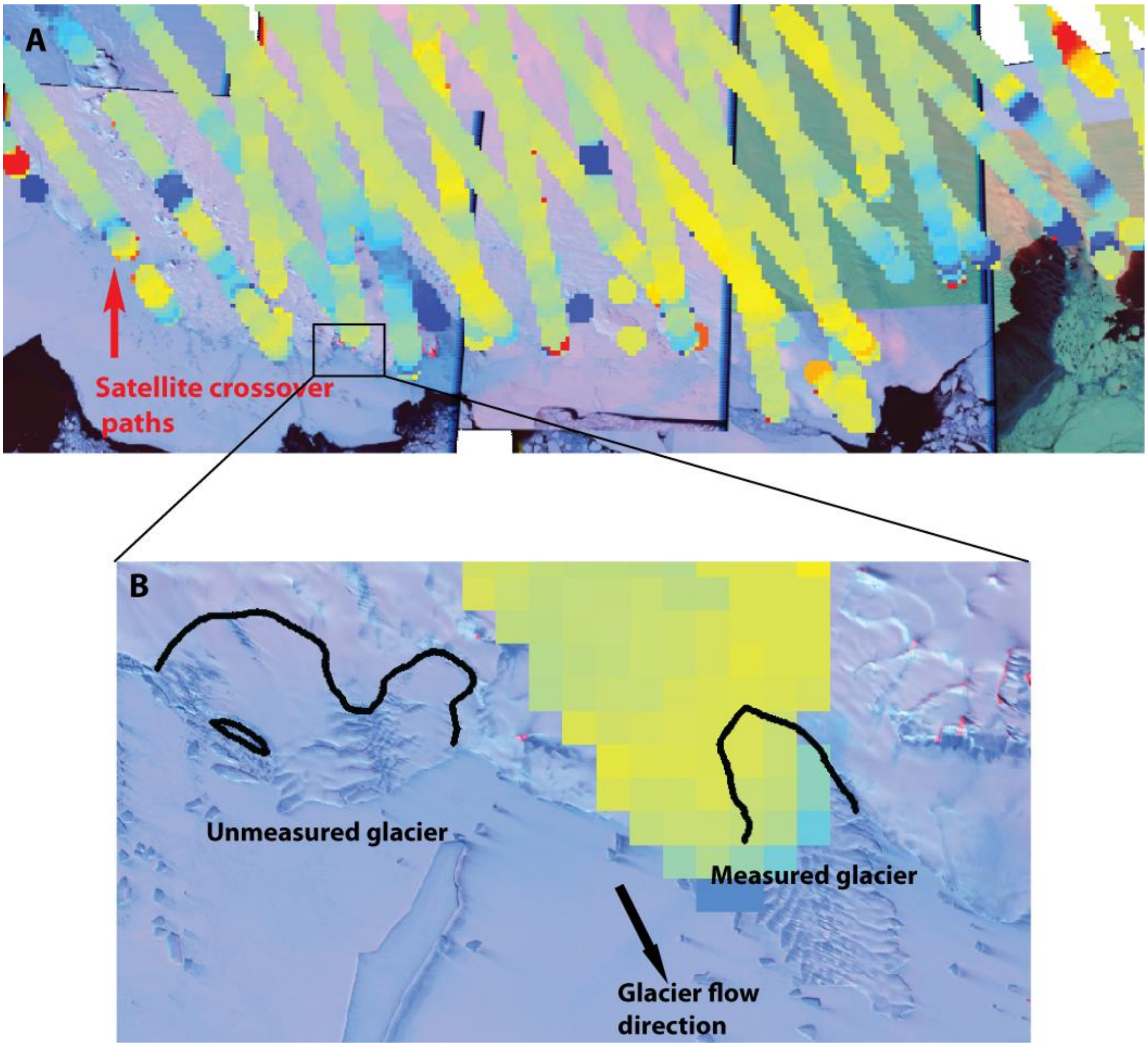


Figure 3.6: Images demonstrating the satellite crossover paths associated with the elevation change dataset.

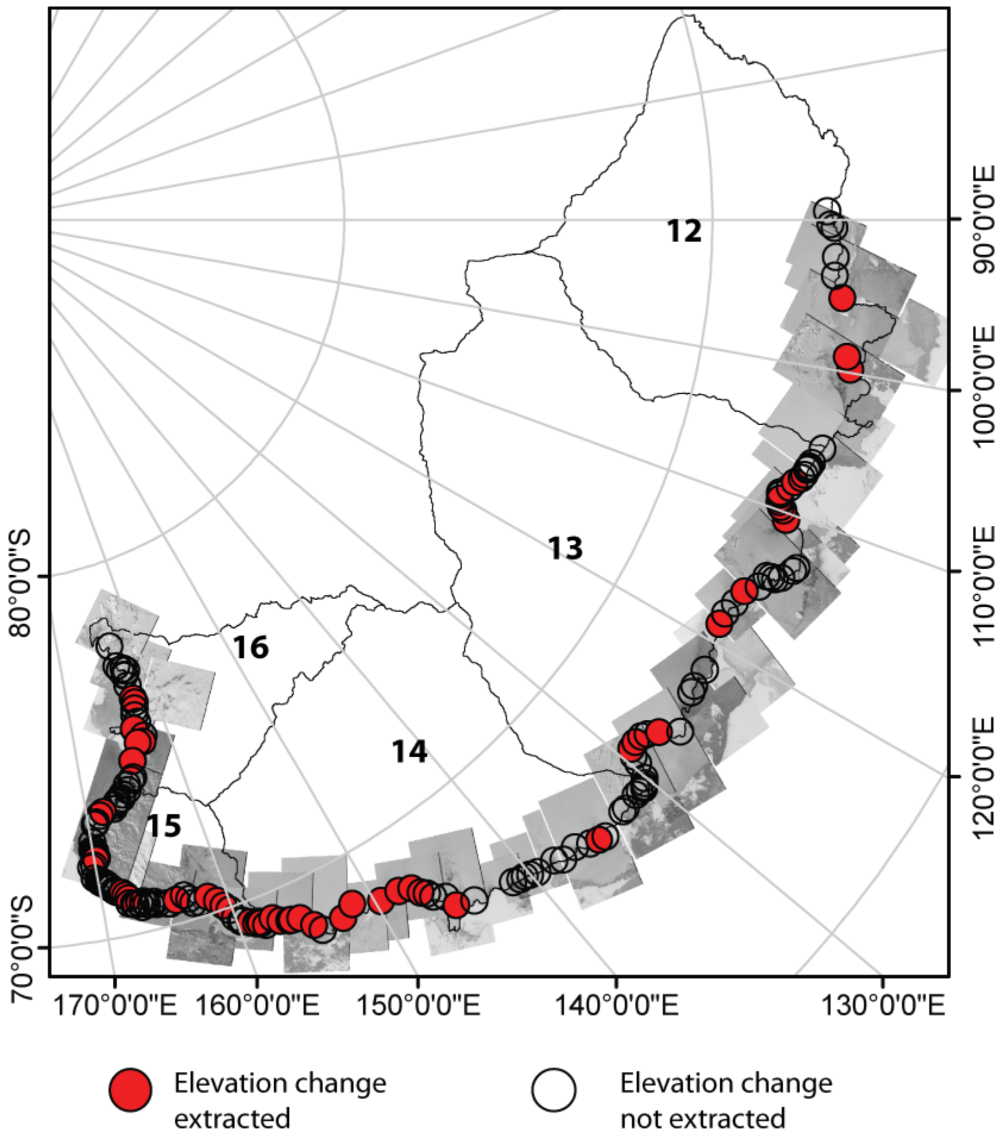


Figure 3.7: Spatial location of the elevation data collected.

3.6 Error Analysis

Two main sources of error affecting glacier frontal change data have been identified, and these are the accuracy of the terminus delineation and co-registration errors. Previous studies have shown that for relatively clean glacier termini, manual terminus delineation error is no greater than the pixel size of the image being mapped (Burgess and Sharp, 2004; De Beer and Sharp, 2007). However, many of the glaciers measured in this study have heavily crevassed and calved glacier fronts, see Fig. 3.8, making it difficult to determine the glacier front from recently calved ice. This could lead to an increase in error in comparison to clean glacier termini. In order to calculate the error associated with this, two sets of five glaciers were selected (with similar termini to that of Fig.3.8), with one set from the 2000 time period and one from the 1974 time period, to represent the change in image resolution at each time period. Each glacier terminus was then digitized

10 times independent of each other i.e. after each digitization a new shapefile was created and the previous shapefile was made invisible, with the maximum distance between any two digitized positions taken as the line placement error (e.g. Stokes *et al*, 2006; De Beer and Sharp, 2007). The results indicate that the maximum terminus digitization error in this study is in the order of a few 10s of meters, so roughly the same size as the pixel resolution of the image it is being mapped from. However, it is likely that this gives an over estimation of the actual error, as this assumes a maximum digitizing offset at all glaciers measured. Although in reality it is more likely that multiple random errors in the digitizing process roughly cancel each other out (De Beer and Sharp, 2007). Therefore, it was decided to estimate terminus digitization error in this study to be 50% of pixel resolution used during mapping (Table 3.3).

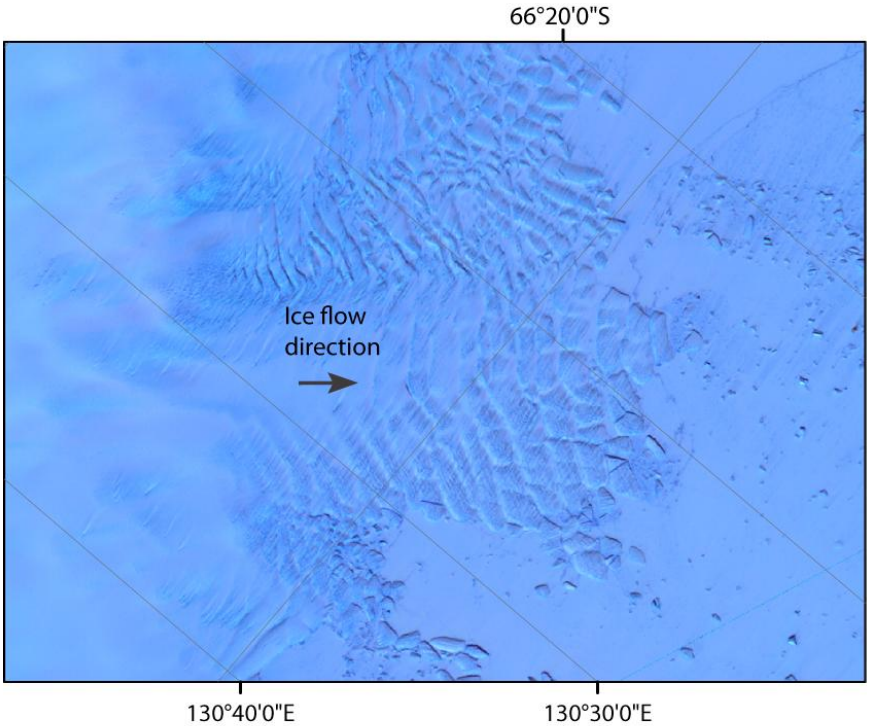


Figure 3.8: Example of a glacier with a heavily calved front, making it difficult to establish which blocks of ice are still attached to the glacier.

Table 3.3: Terminus digitization error.

Time Period	Resolution (m)	Digitization error (m)
1963	140	70
1974	80	40
1990	30	15
2000	30	15
2006	30	15
2010	30	15

Estimations of co-registration error were calculated for each mosaic at each time period, in relation to the Landsat 2000 base scene. This was calculated by placing a marker on three stable features in each individual base scene mosaic. Separate markers were then placed on the same features in each co-registered mosaic at every epoch. The average distance between the base scene and co-registered mosaic markers in each mosaic was taken as the co-registration error. The results indicated that co-registration error was dependent on both the spatial location of the mosaic and the sensor used to capture the image e.g. MSS, TM, ETM+. Mosaics centred near Victoria Land and Oates Land (mosaic's 1 and 2, Fig. 3.3a) tended to have a lower co-registration error in comparison to the others (Table 3.4). This can be attributed to the fact that these relatively mountainous regions contained more nunataks in comparison to other mosaics, enabling more accurate GCPs to be placed in these mosaics, hence leading to a lower co-registration error.

In general, the 1974 Landsat MSS images have the largest co-registration error (Table 3.4) because these images contain the largest pixel resolution (140 m). In contrast, the 2010 and 2006 Landsat ETM+ and 1990 Landsat TM images produced a comparatively smaller co-registration error as they have a lower pixel resolution (30 m). However, the 1990 TM images did produce a higher error compared to the ETM+ images. This is because they were measured with a different sensor (TM) compared to the base scene (ETM+). The exception to this is mosaic 6, which has a comparatively large error of 170 m in 1990. The co-registration error associated with the 1963 Argon mosaic was shown to be consistently around 350 m.

Table 3.4: Estimated co-registration error for each mosaic, at each time period.

	Co-registration error (m)				
Mosaic	1963	1974	1990	2006	2010
1	350	120	60	60	60
2	350	140	60	60	70
3	350	140	110	70	70
4	350	160	140	60	70
5	350	160	120	70	60
6	350	180	170	70	60

Overall, the total error is calculated by adding co-registration error to the digitization error (Table 3.5). This gives a maximum error ranging from 75 m to 210 m for Landsat based measurements, dependent on spatial location and satellite sensor. The Argon images have a much larger error at 420 m. In total, 85% of all measurements had an error of less than 180 m, which is broadly similar to Cook *et al.* (2005) who mapped glaciers in the Antarctic Peninsula.

Table 3.5: Estimated total error for each mosaic, at each time period.

	Total error (m)					
Mosaic	1963	1974	1990	2000	2006	2010
1	420	160	75	15	75	75
2	420	180	75	15	75	85
3	420	180	125	15	85	85
4	420	200	155	15	75	85
5	420	200	135	15	85	75
6	420	210	185	15	85	75

3.5 Climate Data

Secondary sources of monthly mean surface air temperature records from four research stations: Scott, Dumont d'Urville, Casey and Mirny, have been extracted from the SCAR Met reader project (http://www.antarctica.ac.uk/met/gjma/). The spatial location of each research station is detailed in Table 3.6. It is notable that Scott research station is located considerably further south than the others. All stations have complete monthly records between 1974 and 2010 allowing an annual mean and summer (December, January and February) temperature series to be extracted; although Scott base has data missing in early 1994 and, therefore, it was not possible to derive an annual average for this year.

Table 3.6: Spatial information of the four research stations used in this study (see Fig. 4.1).

Station	Region	Coordinates
Mirny	Wilkes Land	66°33'10"S, 93°00'34"E
Casey	Wilkes Land	66°16'55"S, 110°31'31"E
Dumont d'Urvile	Adélie Land	66°39'46"S, 140°00'5"E
Scott	Ross Island	77°50'58"S, 166°46'5"E

Results

4.1 Summary of glacier change measurements

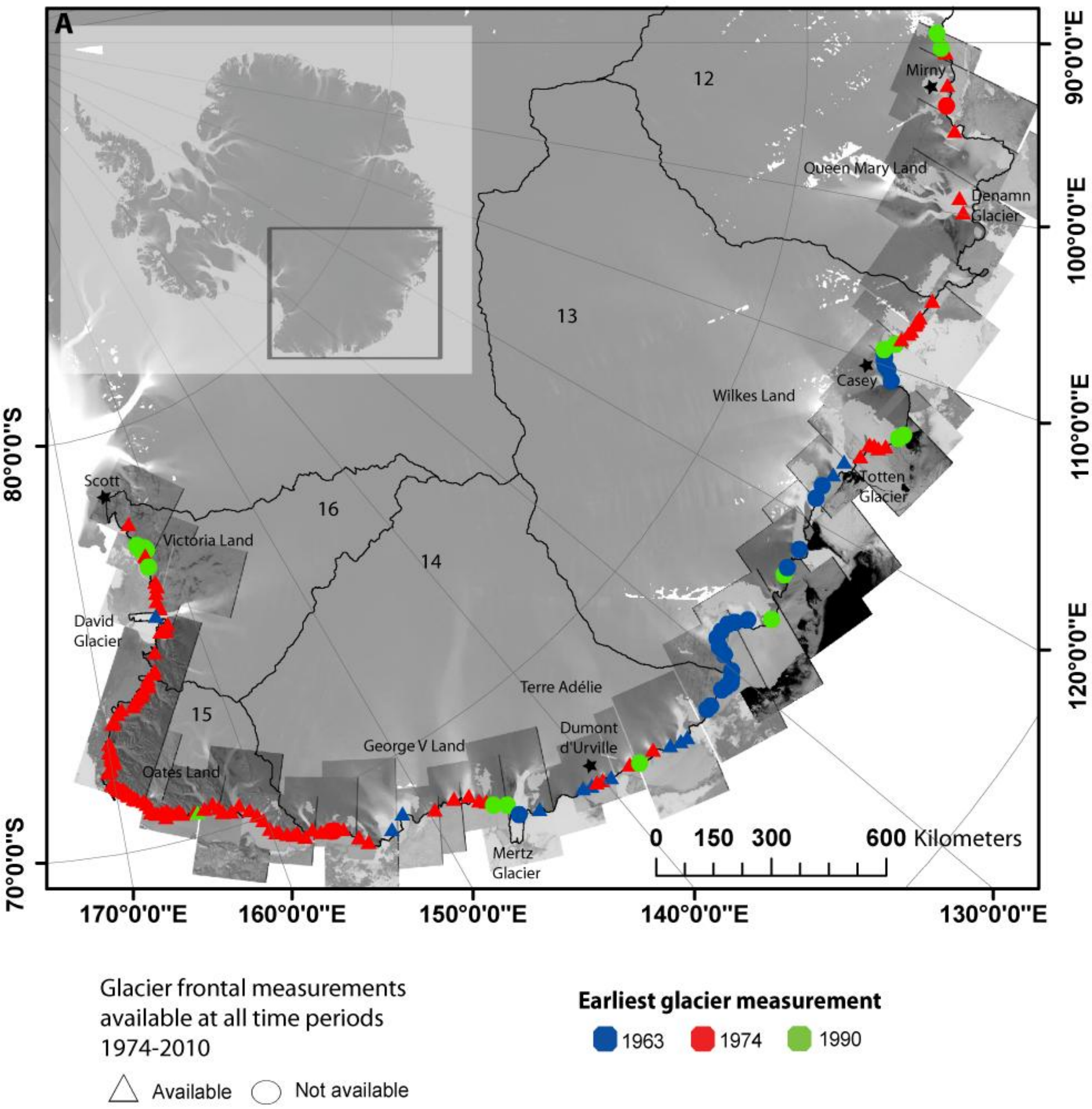
In total, 175 outlet glaciers were mapped from satellite imagery. It is important to note that the majority possess a floating tongue (~90%) and most are unconstrained (84%) by lateral boundaries. These glaciers were mapped at a maximum of six time steps: 1963, 1974, 1990, 2000, 2006 and 2010. However, due to image availability, it was not possible to map all glaciers at every time step. In fact, only 22% of glaciers were measured in 1963, with a majority (93%) having an oldest observation from 1974, see Table 4.1. The number of time steps at which individual glaciers were mapped varied from a minimum of three to a maximum of six. The temporal and spatial coverage of these variations in glacier mapping is shown in Figure 4.1.

Spatially, there is poor coverage in 1963. During this period glacier observations were concentrated between Wilkes Land and Terre Adélie (~110 – 140 ^{o}E), with virtually no measurements outside of this region. There is significantly better coverage in 1974, with 78% glaciers mapped. There is near complete coverage in Queen Mary Land (~90 – 105 ^{o}E) and between Terre Adélie and Victoria Land (~135 – 170 ^{o}E). The only area with poor coverage lies between Wilkes Land and Terre Adélie (~110 – 140 ^{o}E). Coincidently, this is the only section with good coverage in 1963. Due to the widespread availability of Landsat imagery, there is near complete coverage in the time periods 1990 (95%), 2000 (100%), 2006 (94%) and 2010 (98%).

Overall, this study will primarily focus on large sample sizes at four main decadal time steps: 1974, 1990, 2000 and 2010. In total, 73% of glaciers have observations at all of these time steps and 99% have observations at a minimum of three. The 1963 dataset will be used to discuss longer term patterns for a smaller population of glaciers, whilst the 2006 dataset will be used to analyse sub decadal glacier behaviour in the most recent decade.

Table 4.1: Number of glaciers successfully mapped at each time period.

Time Period	Number of Glaciers
1963	38
1974	136
1990	168
2000	175
2006	165
2010	171



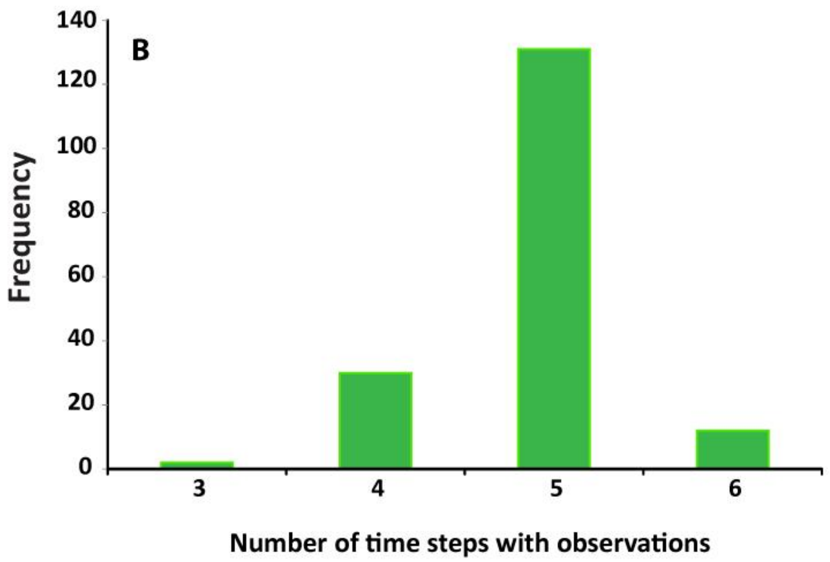


Figure 4.1 (A): Map showing the 175 glaciers mapped and which time steps they were mapped in. (**B)** Histogram showing how many time steps glaciers were mapped at each time step.

4.2 Glacier frontal position change

This section examines glacier frontal position change through time and space. To eliminate any bias associated with the differing lengths of the time periods, all results are expressed as glacier length change rate in meters per year. Furthermore, for the purpose of this section and thereafter, glaciers have been spatially defined into 5 drainage basins (DB), see Fig. 4.1. These drainage basins, derived by Zwally *et al.* (2012), are based on a 500 m DEM of Antarctica developed from ICESat/GLAS data. Individual drainage system have been delineated to identify regions broadly homogeneous regarding surface slope orientation relative to atmospheric advection, and additionally denoting the ice sheet areas feeding large ice shelves (Zwally *et al.*, 2012). The details of these drainage basins are shown in Table 4.2 and location in Figure 4.1.

Table 4.2: Drainage basin characteristics. * Note: Only a small proportion of this drainage section has been mapped in this study.

Drainage basin	Number of mapped glaciers	Total area (Km^2)	Grounded area (Km^2)
12	8	773,999*	722,224
13	39	1,126,542	1,109,771
14	37	726,359	710,953
15	71	133,755	125,183
16	20	271,666	265,243

4.2.1 An overview of glacier change between 1963/1974 and 2010

The small population of glaciers measured over the longest time period from 1963 to 2010 indicate an overall pattern of retreat, with 68% of glaciers retreating (mean frontal position change: -61.2 m a^{-1}; median -12.9 m a^{-1}; n=38). The larger sample of glaciers measured in 1974 and 2010 shows very little overall change in mean terminus position, with 47% of glaciers retreating (mean: -2.7 m a^{-1}; median: 0.7 m a^{-1}, n=136). Spatially, Table 4.3 demonstrates glacier frontal change between 1974 and 2010 broken down into drainage basins; DB 12, 13 and 14 indicate retreat, whilst DB 15 suggests advance. DB 16 shows a mean advance of 29.7 m a^{-1}; however, its median of -8.5 m a^{-1} suggests otherwise. It is important to note that the median provides a much more robust measurement in comparison to the mean. This is because the mean can be heavily influenced by large glaciers which are prone to large calving events and, therefore, mask the overall signal from smaller glaciers.

Table 4.3: Long term glacier terminus change of all mapped glaciers with red number referring to retreat rates and black to advance.

			1974-2010				
Dates	**1963-2010**	**1974-2010**	**DB12**	**DB13**	**DB14**	**DB15**	**DB16**
Number of glaciers (n)	38	136	6	15	26	70	15
Advanced (%)	29	54	33	13	31	74	40
Retreated (%)	71	46	67	87	69	26	60
Mean terminus change (m a^{-1})	-61.2	-2.7	-24.4	-35.2	-39.7	13.6	26.8
Median terminus change (m a^{-1})	-12.9	0.7	-4.9	-26.4	-12.9	9.0	-9.0

4.2.2 Decadal fluctuations

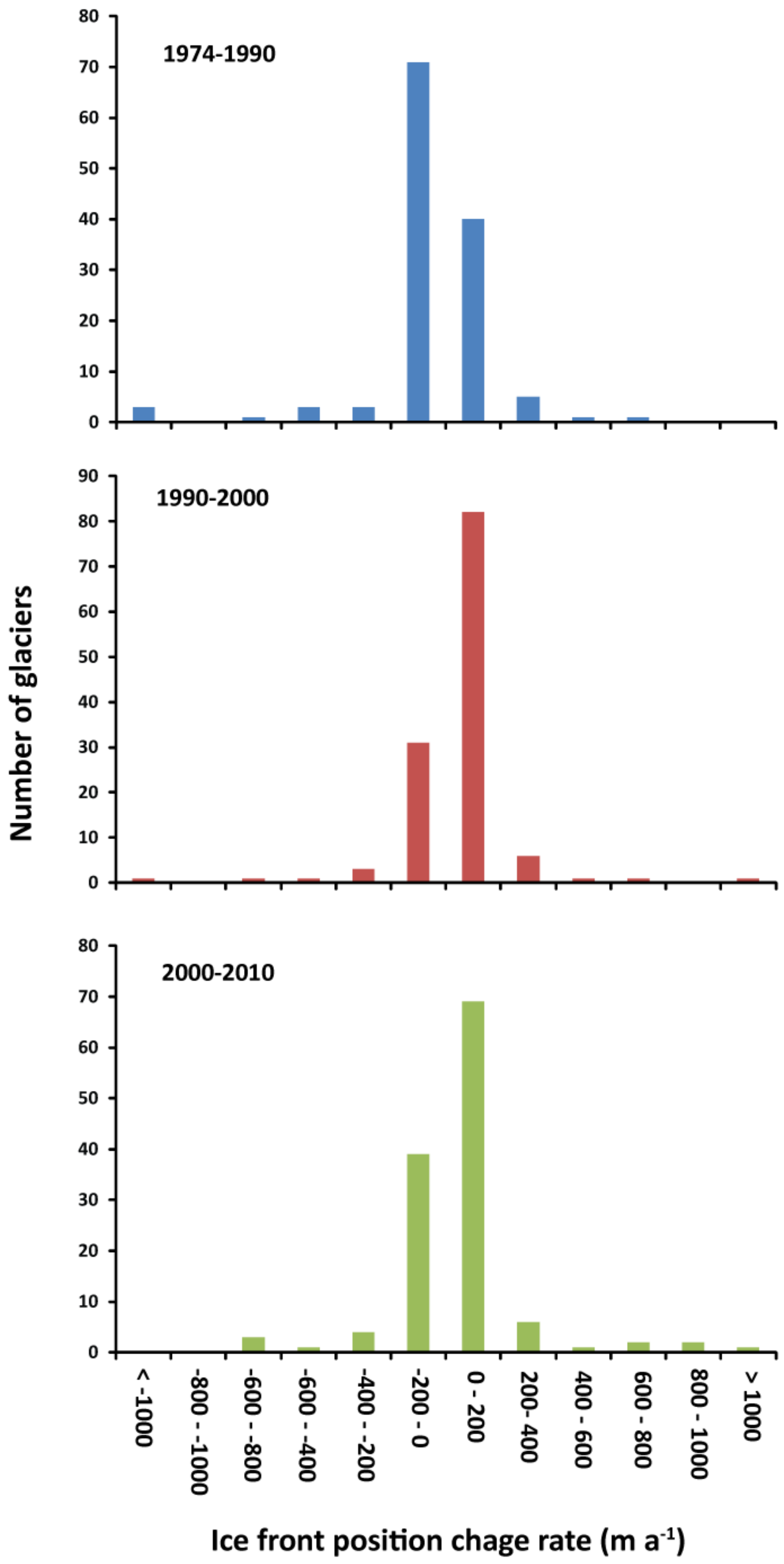
Strong decadal fluctuations were observed, with 63% of all observed glaciers retreating between 1974 and 1990 at a median rate of -12.5 m a^{-1} (mean: -43.4 m a^{-1}). From 1990-2000, however, this trend was reversed, when 72% of glaciers advanced at a median rate of 19.7 m a^{-1} (mean: 43.1 m a^{-1}). During the most recent period, 2000-2010, the number of glaciers which advanced or retreated was more balanced (58% advanced), resulting in a median of 8.4 m a^{-1} and mean of -17.9 m a^{-1}.

It is also important to consider glaciers which have a measurement in every time step. This trend is visualized in Figure 4.2, which shows histograms for each time period for the 128 glaciers which have observations in every time step from 1974 until 2010. The resulting histograms show a clear shift to the right in 1990-2000 and 2000-2010, compared to 1974-1990, indicating a tendency to advance. A comparison of the mean and median terminus position change between all glaciers and glaciers with a measurement in every time step can be seen in Table 4.4. In general, this shows that glacier trends are consistent in relation to the two sample sizes i.e. all glaciers or 128 with observations at each time period. The only exception to this is the mean rate change between 2000 and 2010. This is shown as a positive value (30.6 m a^{-1}) for the sample with observations at each time period whilst, for the sample with all glaciers this value is negative (-17.9 m a^{-1}). However, this can be attributed to the large calving event at Mertz glacier which is several orders of magnitude greater than the next largest calving event. This is not included in the sample with observations at every time period, hence resulting in a significantly larger mean rate.

Spatially, Figure 4.3 suggests trends are similar to the overall decadal pattern. For the period 1974-1990, most drainage basins show retreat dominating, and this was most pronounced in DB13 (87% retreat). The one exception is DB16, which shows a slight majority of glaciers advanced (61%). Between 1990 and 2000, the majority of glaciers advanced in all drainage basins, with the exception of DB16, where this advance was less prominent (53% advance). This predominance of glacier advance is also seen in most drainage basins for the period 2000-2010. Most drainage basins show similar levels of glacier advance in comparison to their 1990-2000 levels. The exception is DB13 which shows 77% of glaciers retreating during 2000-2010, representing a reverse in glacier behaviour compared to the 1990-2000 time period, where 77% of glaciers advanced. In fact, removing DB13 from the overall decadal pattern shows a 59% retreat and 69% advance of glaciers in 1974-1990 and 1990-2000 respectively, which is comparable to the original figures (61% retreat and 72% advance). For the period 2000-2010 and excluding DB13, 67% of the rest of the glaciers advanced during 2000-2010, considerably more than the 57% advance observed with the inclusion of DB13.

Table 4.4: Decadal changes in glacier frontal position. Negative values denote mean/median retreat (red) and positive denote advance (black). Data include all glacier measurements available for each time-step but, for comparability, values in brackets are for the same population of 128 glaciers, which reveal similar trends.

Dates	**1974-1990**	**1990-2000**	**2000-2010**
Number of glaciers (n)	131 (128)	168 (128)	171 (128)
Advanced (%)	37 (36)	72 (72)	58 (63)
Retreated (%)	63 (64)	28 (28)	42 (37)
Mean terminus change (m a^{-1})	-43.3 (-44.9)	43.1 (30.9)	-17.9 (30.6)
Median terminus change (m a^{-1})	-12.5 (-12.8)	19.7 (14.5)	8.4 (13.7)



Mean	-44.9
Median	-12.8
Standard Deviation	252.4
Range	2290.7
Minimum	-1611.7
Maximum	678.9
Sum	-5745.2

Mean	30.9
Median	14.5
Standard Deviation	247.7
Range	3196.8
Minimum	-1591.4
Maximum	1605.4
Sum	3950.5

Mean	30.6
Median	13.7
Standard Deviation	261.1
Range	2554.6
Minimum	-795.4
Maximum	1759.2
Sum	3918.9

Figure 4.2: Glacier frontal position change at each decadal time step, for 128 glaciers for observations: 1974, 1990, 2000 and 2010. Summary table values are in m a^{-1}.

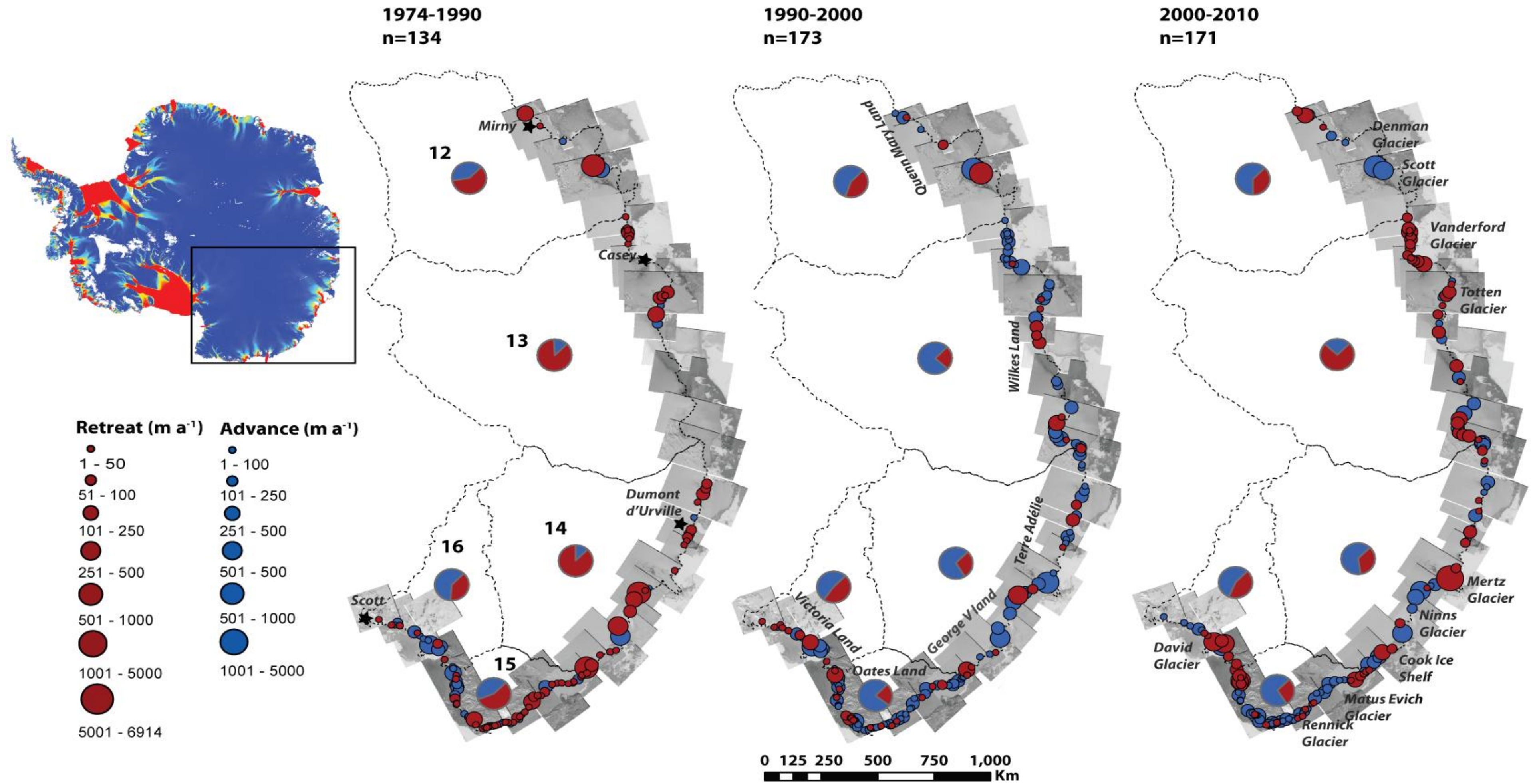


Figure 4.3: Spatial and temporal variations in glacier frontal position changes in East Antarctica. The pie charts show the percentage of advancing or retreating glaciers for each drainage basin and the circles refer to the glaciers terminus position change (red = retreat, blue = advance). Location map (top left) shows glacier velocity over Antarctica with fast flow zones in red (Rignot *et al.*, 2011;a).

It is imperative to express glacier frontal change results in terms of both percentage of glaciers that advance or retreat and in terms of absolute length change i.e. meters per year. This is because if results were just expressed in terms of percentages, it would treat all glacier change as the same magnitude, which in reality is unlikely. Equally, if all results were expressed in terms of absolute change, the greater magnitude of change associated with larger glaciers may dominate over the signal of the smaller glaciers (further detail in section 4.3).

Figure 4.4 expresses the rate of change of glacier frontal position (m a^{-1}) in the form of box plots for the different basins. All box plots show a similar pattern to the percentage advance or retreat results. DB 12, 13 and 14 all have negative medians during 1974-1990, indicating retreat, on the other hand DB 16 and 15, have a median close to 0. From 1990-2000, all drainage basins have a positive median, greatest in DB 13 (median = 54.9 m a^{-1}) and 14 (median = 30.3 m a^{-1}), whilst in DB 12, 15 and 16 the medians are slightly positive at 8.9, 12.6 and 1.7 m a^{-1}, respectively. In the period between 2000 and 2010, DB12, 15 and 16 showed a slight increase in their median's in comparison to 1990-2000 levels at 44.9, 20.6 and 10.9 m a^{-1}, respectively.

At sub-decadal time steps, DB15 and 16 show slight variations in median glacier terminus position change rates between the periods 2000-2006 and 2006-10, shown in Table 4.5. On the other hand, DB12 shows a marked change in median terminus position change rates from -24.4 m a^{-1} in 2000-2006 to 227.5 m a^{-1} for 2006-10. However, the magnitude of this change should be treated with caution due to the low sample size of glaciers in DB12 (n=8). DB14 shows a median terminus position change rate of 19.4 m a^{-1} for the period 2000-2010. Although positive and thus representing advance, it still represents a decrease in median compared to 1990-2000. However, it appears that this advance occurred primarily in 2000-2006 (median = 68.0 m a^{-1}), whilst in 2006-2010 there was a median of 5.1 m a^{-1}. In the period 2000-2010, neighbouring DB13 showed a median of –63.6 m a^{-1}, representing a strong signal of retreat. However, when the decade is split, there is a stronger retreat signal seen in 2000-2006 (median = -52.7 m a^{-1}), and in fact a weak advance signal in 2006-2010 (median = 6.7 m a^{-1}). Therefore, this indicates that the widespread retreat seen in DB13 between 2000 and 2010 was initiated by rapid retreat in 2000-2006, with glaciers then stabilizing in 2006-2010.

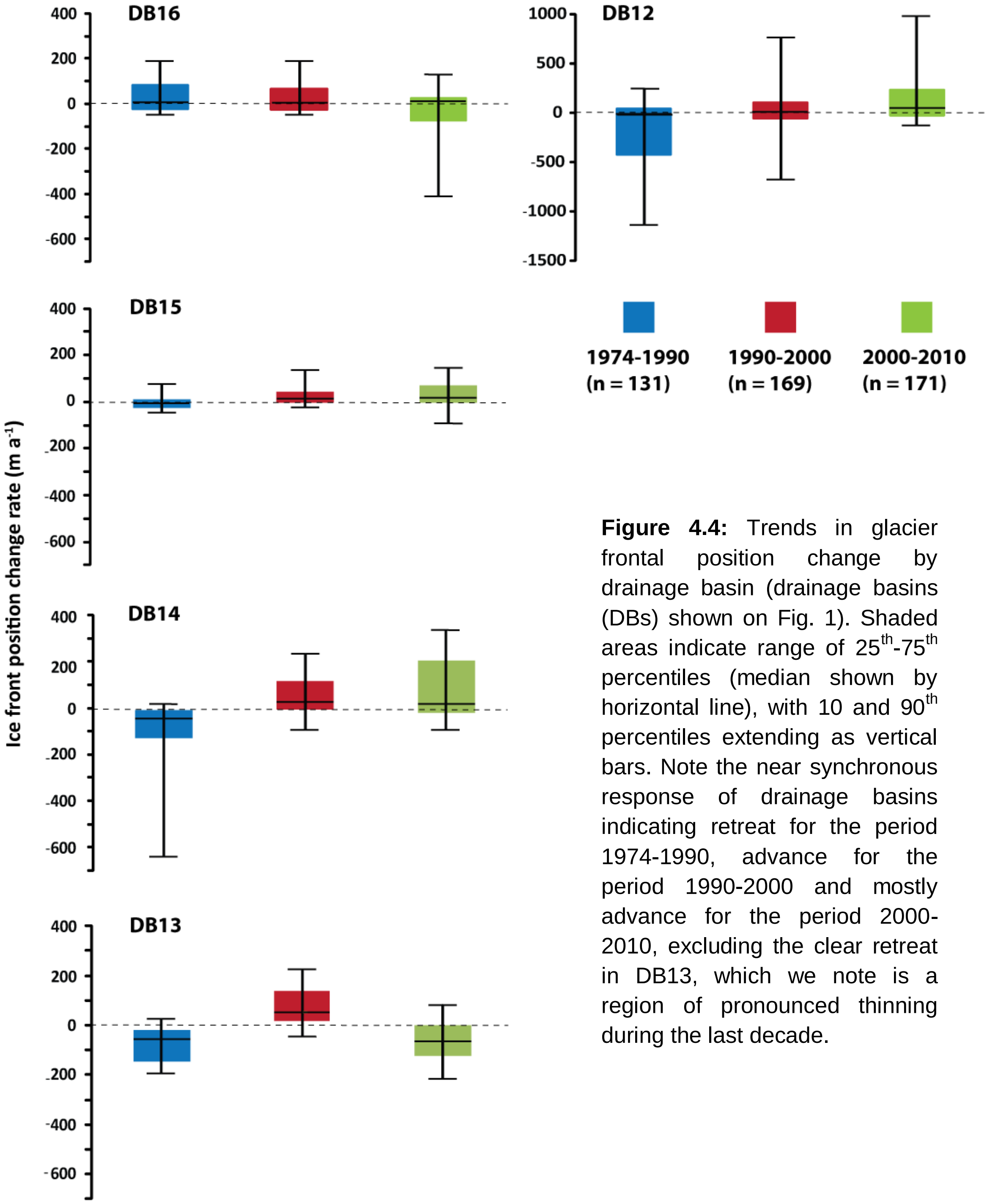


Figure 4.4: Trends in glacier frontal position change by drainage basin (drainage basins (DBs) shown on Fig. 1). Shaded areas indicate range of 25th-75th percentiles (median shown by horizontal line), with 10 and 90th percentiles extending as vertical bars. Note the near synchronous response of drainage basins indicating retreat for the period 1974-1990, advance for the period 1990-2000 and mostly advance for the period 2000-2010, excluding the clear retreat in DB13, which we note is a region of pronounced thinning during the last decade.

Table 4.5: Glacier frontal position change per drainage section (m a^{-1}), 2000-2006 and 2006-2010.

	2000 - 2006			2006 - 2010		
DB	**% advancing**	**mean**	**median**	**% advancing**	**mean**	**median**
12	38	97.4	-24.4	62	482.9	227.5
13	34	-36.4	-52.7	50	-67.4	6.7
14	70	146.2	68.0	51	-378.6	5.1
15	71	-10.2	35.1	71	48.9	21.1
16	65	49.4	16.4	69	-174.2	26.0

4.3 Glacier frontal position change in relation to width and velocity

The width of a glacier has been shown to be a good indicator for its overall size (Bahr, 2006). Previous studies have also found that larger glaciers often retreat or advance further than shorter glaciers in absolute terms (Stokes *et al.*, 2006; Joskoot *et al.*, 2009). This is also consistent in this study, with Figure 4.5 demonstrating a statistically significant relationship (r^2=0.32) between glacier width and range. In general, the flow of outlet glaciers is known to be non-linearly related to glacier width (van der Veen and Whillans, 1996). This is confirmed from this study's data which shows a statistically significant positive correlation between glacier width and velocity (r^2=0.29, p<0.001), therefore suggesting that larger (wider) glaciers tend to be faster and exhibit a greater magnitude of terminus fluctuation. Indeed, this is also shown in Figure 4.5 which shows that glaciers with the highest velocities undergo the largest ice front position changes. Thus, glacier velocity can be seen as proxy for glacier size.

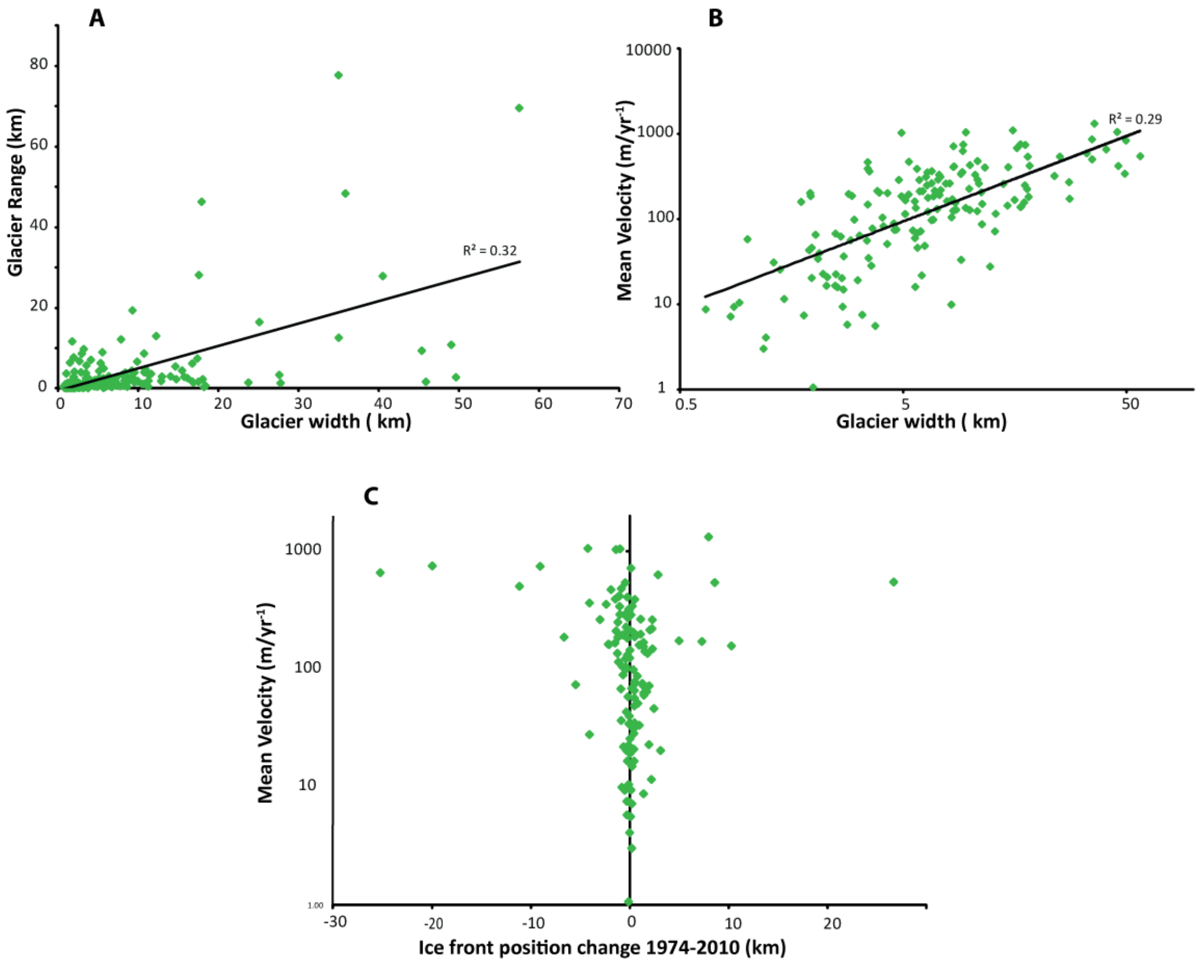


Figure 4.5 A & B: Statistically significant positive correlation between glacier width and **a)** glacier range (p<0.01) and **B)** mean velocity (p<0.01). **C)** Relationship between long term ice front position change and mean velocity.

It is of vital importance in the context of this study to establish if the decadal fluctuations described in section 4.2 are present in all glaciers regardless of their width class (size) (0-5, 5-10, 10-15 and >15 km). This is because large magnitude events from the wider glaciers may have to potential to skew the results. To investigate this, glaciers were broken down into four width classes. For each width class, the median frontal position change was calculated and plotted in Figure 4.6. There is a clear trend for the wider/faster glaciers to exert a larger magnitude of frontal position change. However, all glaciers regardless of width class still appear to follow the general decadal trends described in section 4.2. Indeed, this is shown by the fact that in every width class the lowest median rate (m a^{-1}) (either positive or negative) occurs between 1974 and 1990 and the highest median rates are recorded between 1990 and 2000. This confirms that in general, glaciers regardless of size (width/speed) appear to be behaving the same in respect to advance and retreat trends across a large section of East Antarctic coast through time. Thus, suggesting a common forcing driving observed decadal fluctuations.

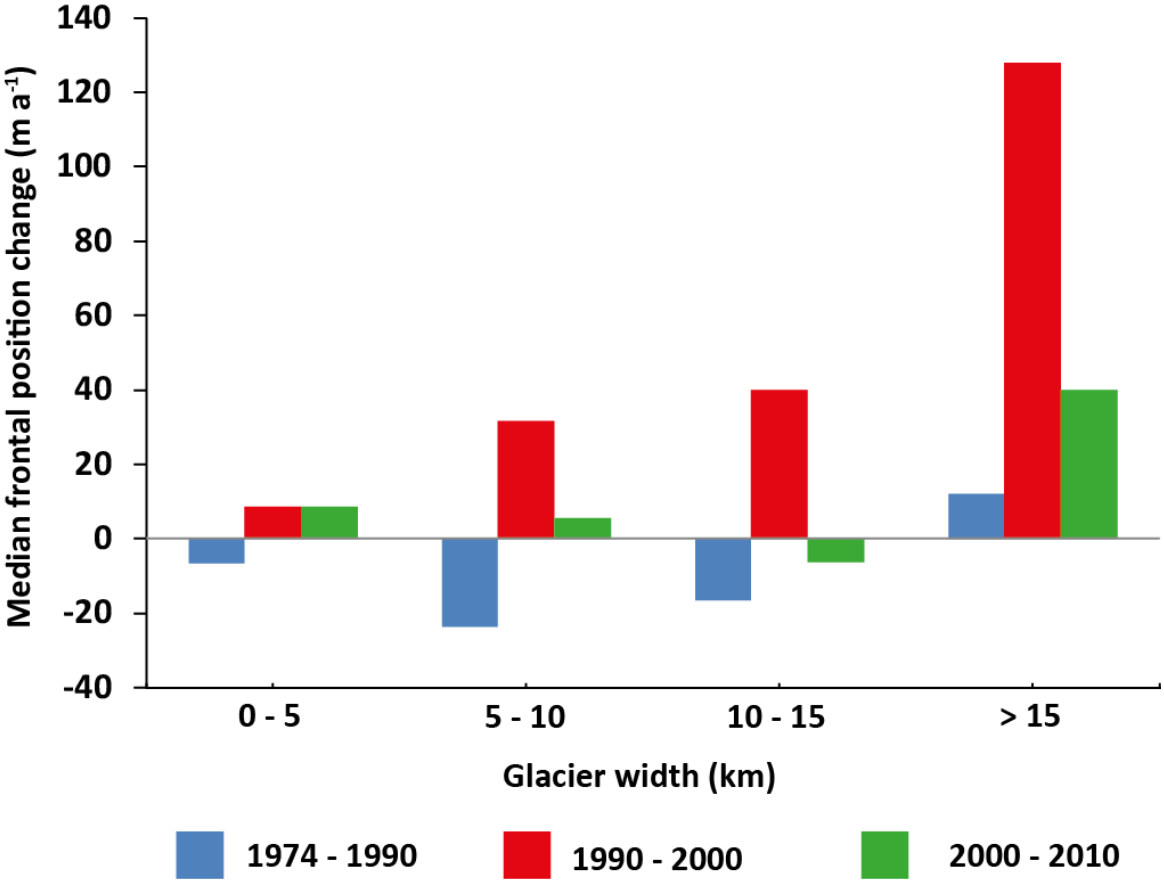


Figure 4.6: Median frontal position change for four glacier width classes, showing a general increase in magnitude in frontal position change as width increases and a consistent and robust trend showing the most negative frontal position change during the period 1974 to 1990 and the most positive between 1990 and 2000.

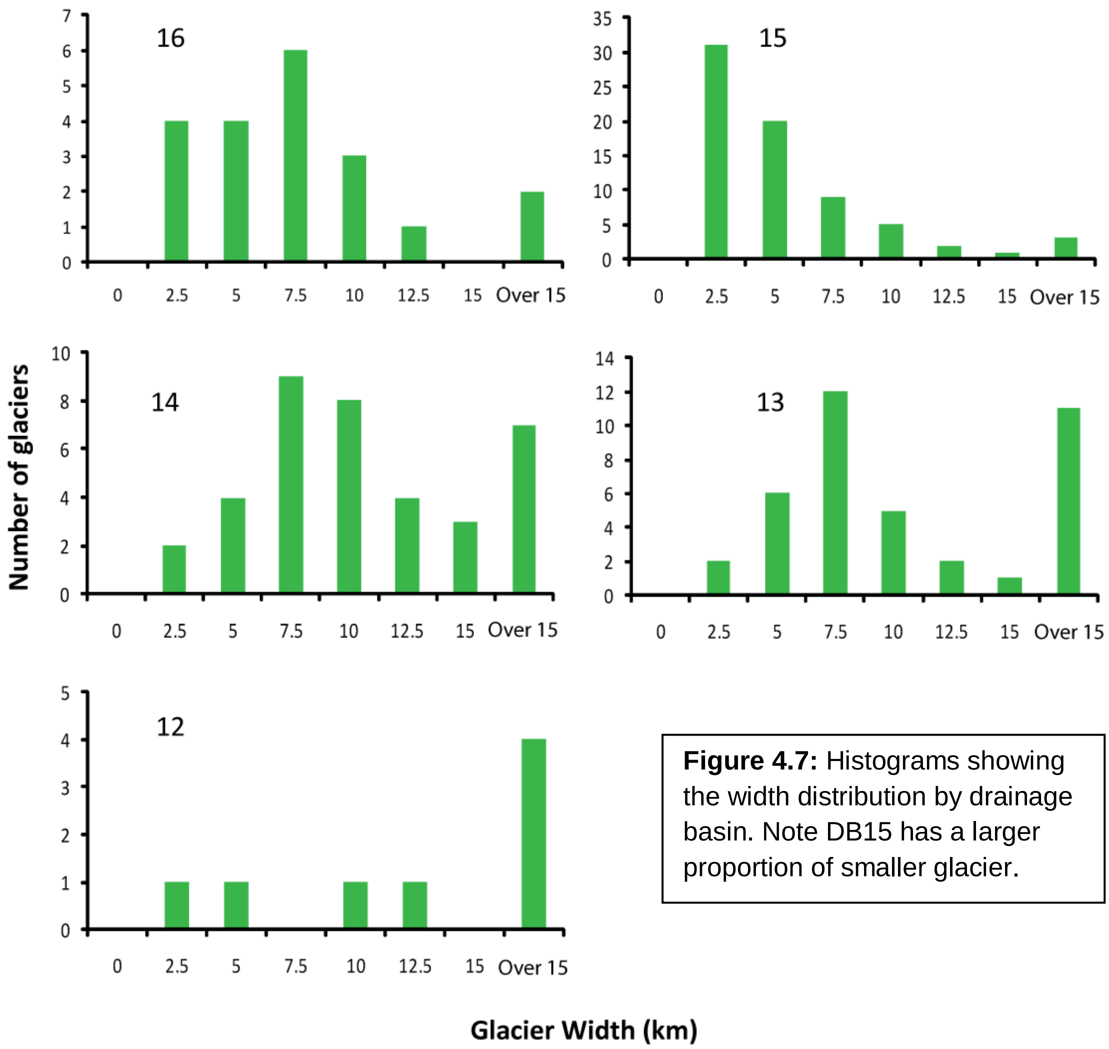


Figure 4.7: Histograms showing the width distribution by drainage basin. Note DB15 has a larger proportion of smaller glacier.

Spatially, glacier size (width/speed) is not distributed evenly across the study area, shown in Figure 4.7. DB15 has a much larger proportion of smaller glaciers in comparison to other drainage basins, with 44% of glaciers having a width under 2.5 km. The majority of glaciers in DB16, 14 and 13 have a width between 7.5 and 10 km. However, DB14 and 13 have a much larger fraction of glaciers with a width greater than 15 km at 19% and 28%, respectively, compared to DB16's 10%. DB12 has the largest proportion of glaciers with a width larger than 15 km at 50%.

As shown in Figure 4.5A, glaciers with a smaller width have been shown to have a lower range over the observation period. As DB15 has a comparatively large proportion of smaller glaciers, it would therefore be expected to have a lower magnitude of glacier change. This is confirmed in Figure 4.4 which shows the smallest averaged interquartile range (IQR) across each time step for DB15 (IQ = 89 m a^{-1}) , indicating a low magnitude of glacier change. Whereas DB12, which has the largest proportion of glaciers with a width greater than 15 km, shows the largest IQR (IQ = 301 m a^{-1}), demonstrating the relationship between glacier width and the magnitude of frontal change and that trends are robust to the effect of size.

4.3.1 Calving events

In cases where large glaciers have been mapped at up to 6 time steps, major calving events can shed several kilometres of ice (often 10s of km^2) and with periodicities of 10-40 years, see Figure 4.8 and 4.9. The most dramatic of such events occurring at glaciers with long floating tongues e.g. Mertz glacier, 2010 (Kusahara *et al*., 2011) and David Glacier (Parmiggiani and Fragiacomo, 2005) which is shown in Figure 4.8. These calving events are often followed by a subsequent re-advance and sometime further retreat e.g. Matus Evich glacier. In the case of smaller glaciers, it is likely the periodicities are under 10 years and therefore more than one cycle of advance and retreat could occur during the longer time steps (1974-1990 and 1990-2000). This is demonstrated in Figure 4.10 which shows that the majority of glaciers in the two smaller time steps 2000-2006 and 2006-2010 have a higher gradient compared to the earlier and longer periods.

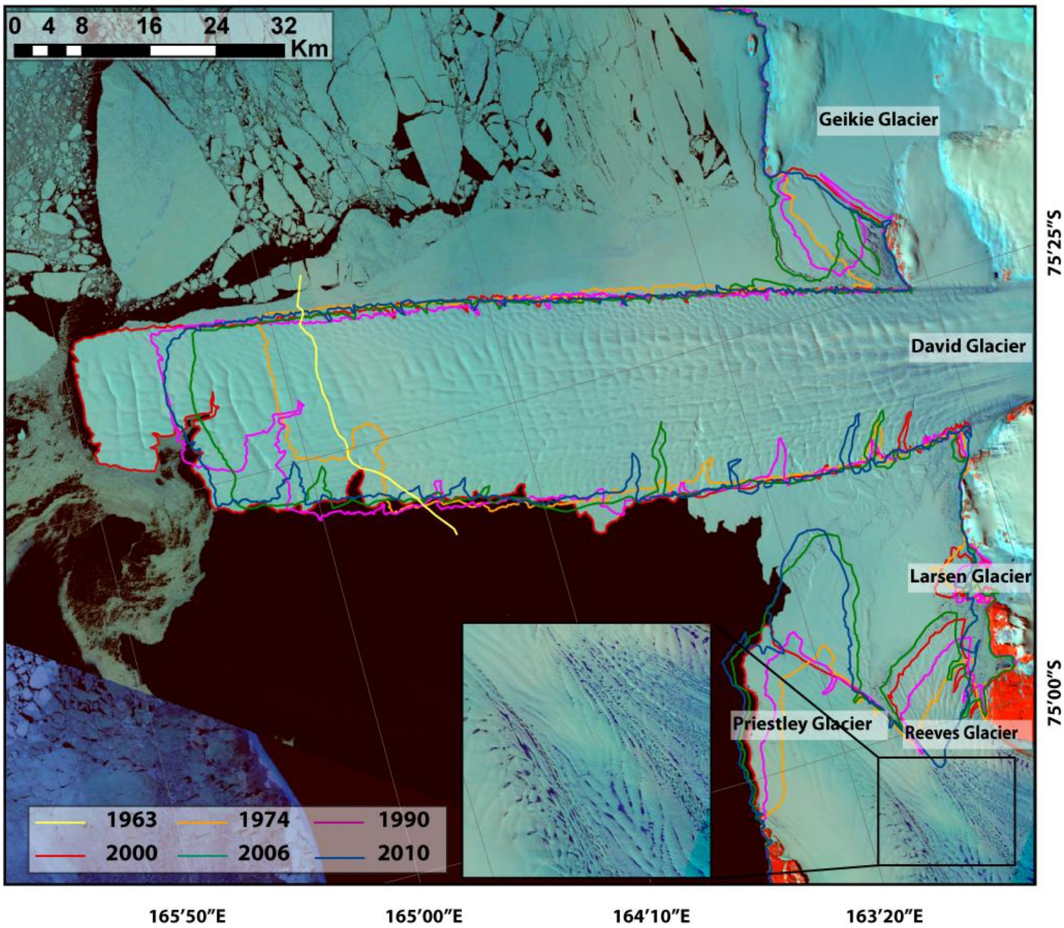


Figure 4.8: Fluctuations in ice front position on David Glacier, Victoria Land. Note advance from 1963 to 2000, followed by a large calving event and retreat to 2006, followed by subsequent re-advance. Also note meltwater ponding on the surface of Preistley Glacier (bottom right).

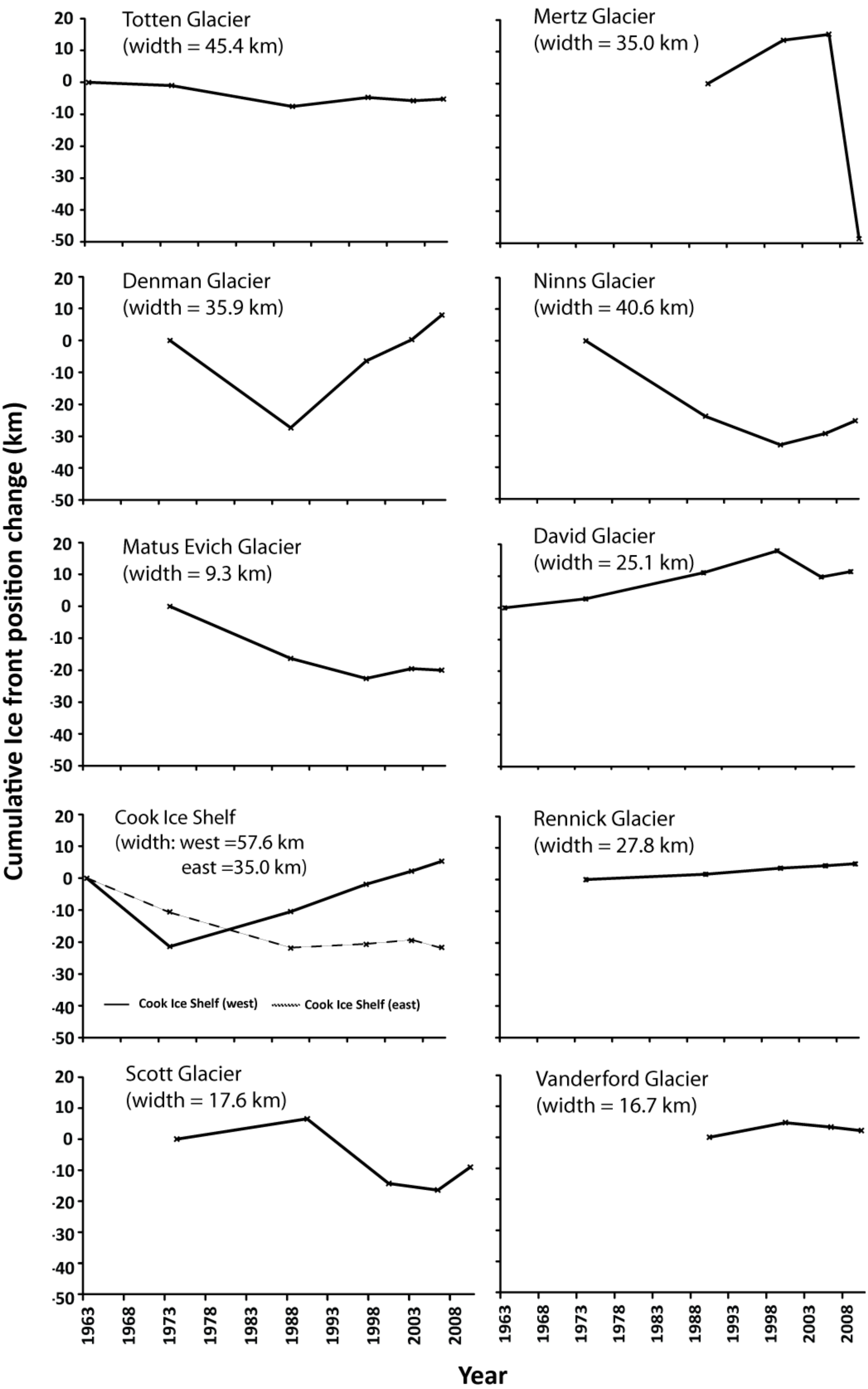


Figure 4.9: Evolution of mapped ice front positions with time on ten large, rapidly-flowing glaciers (see Fig.3 for glacier locations). Note the large magnitude of change (10s of kms) associated with major calving events and typically followed by re-advance. Cook ice shelf was divided into two parts due to differing ice front histories (Frezzotti *et. al.,* 1998).

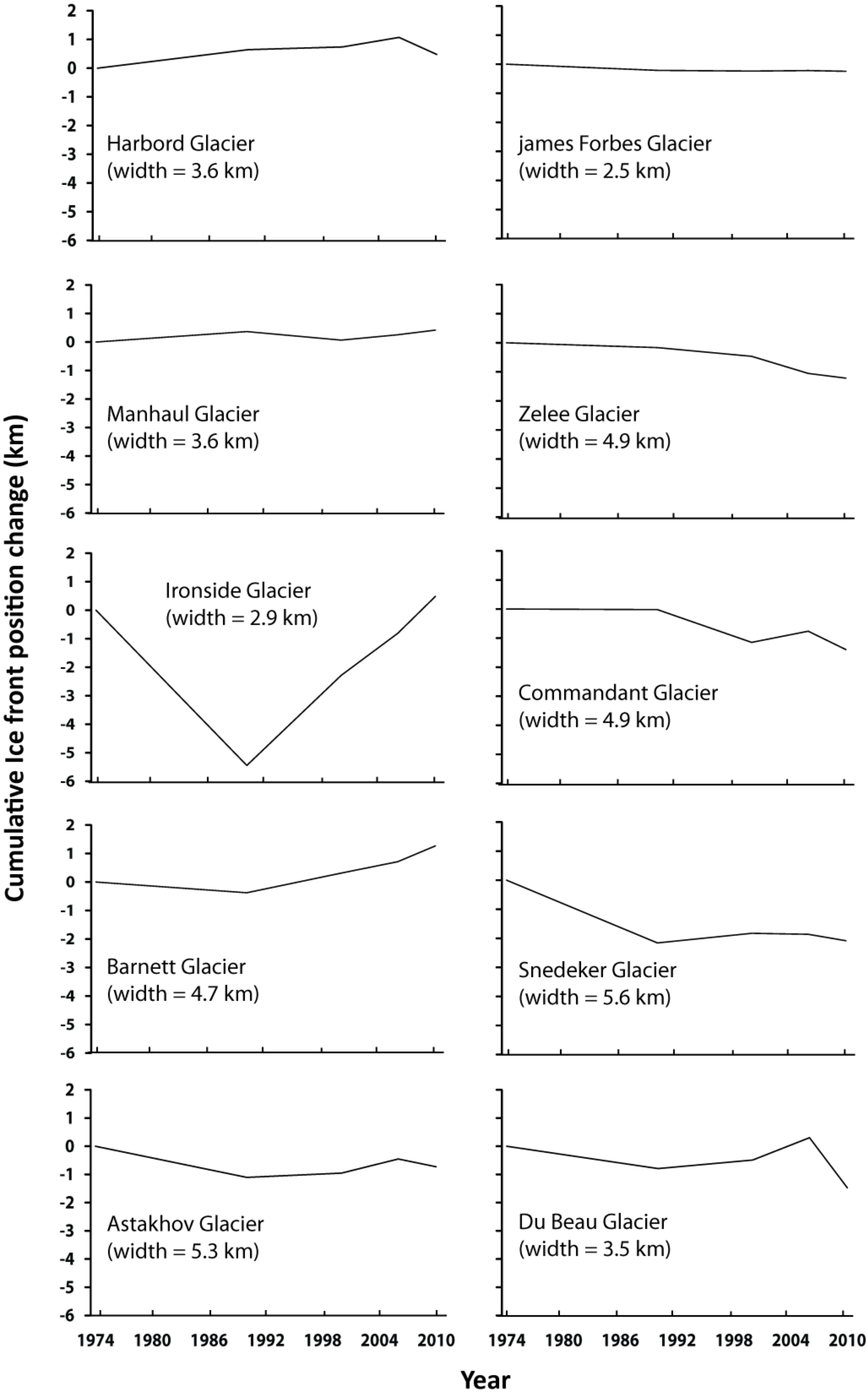


Figure 4.10: Cumulative changes in ice front position for smaller glaciers within the study area.

4.4 Glacier frontal position change and elevation

A comparison of glacier frontal changes on glaciers measured between 2000 and 2006 with elevation change reported between 2003 and 2007 (Pritchard *et al.*, 2009) provides no obvious trend, see Fig. 4.11. In general, those glaciers which are thinning tend to exert a larger magnitude of terminus change, with the majority advancing, whilst those which are thickening show little terminus change in either direction. In addition, the small population of glaciers which are constrained by lateral boundaries appear to be exerting little terminus change and can be associated with thinning or thickening. Those glaciers which are not constrained by lateral boundaries appear to follow the overall trend described above.

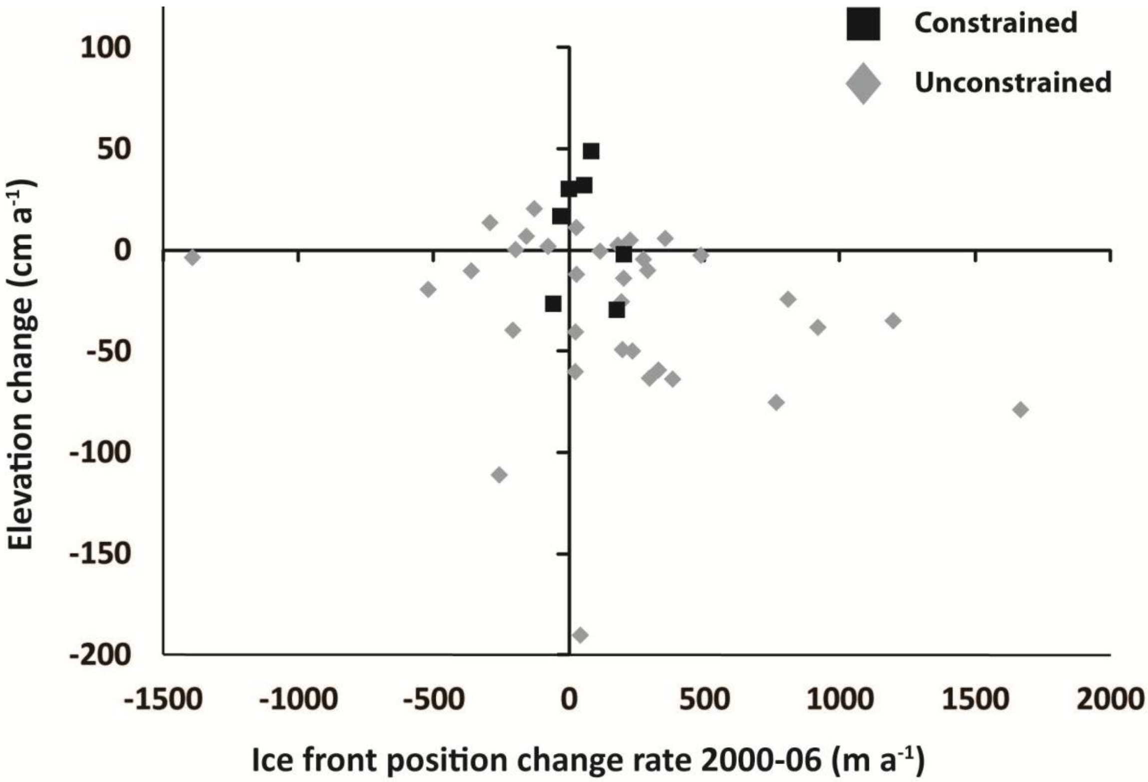


Figure 4.11: Glacier frontal position change (2000-2006) plotted against elevation change (2003-2007: from Pritchard *et al.*, 2009). Those glaciers which are constrained by lateral boundaries are marked by black squares, whilst those which are unconstrained are marked by grey diamonds.

4.5 Glacier change and climate

Annual and austral summer (mean December, January and February temperatures) air temperature data from four research stations: Scott, Dumont d'Urville, Casey and Mirny are shown in Figure 4.12. Despite being located several hundred kilometres apart trends are remarkably consistent. Summer temperatures reveal a consistent warming trend from the 1970s until the late 1980s. This is followed by an abrupt cooling period from the early 1990s until 2000, with summer temperatures around 0.75 ^{o}C cooler in comparison to 1970 temperatures, see Table 4.6. In the first half of the 2000s, there is shift to warming, particularly in Mirny, this is followed by a further cooling trend from the mid 2000s until 2010. However, the mean temperatures for each time period shown in Table 4.6 suggest that in Dumont d'Urville and Casey, the 2000-2010 temperature is still in the region of 0.5 ^{o}C cooler than 1974-1990 temperatures, and slightly warmer than 1990-2000 levels. In contrast, mean temperatures for Mirny indicate that 2000-2010 temperatures are similar to 1974-1990 levels. Scott base, located much further south, is considerably cooler, with average annual temperatures in the region of -20 ^{o}C, compared to annual temperatures of ~-10 ^{o}C for the more northerly stations (Table 4.7). Its summer air temperatures show little variation at each time step and do not show the same trends as the other research stations in the study area.

Air temperature trends are consistent with trends in glacier frontal position change, shown in Figure 4.12. The overall warming trend seen from the 1970s into the 1980s associates with glacier retreat in the 1974-1990 period. The subsequent cooling in the 1990s coincides with the trend of glacier advance during the same period. The fluctuation between warming and cooling between 2000 and 2010 can be associated with moderate overall advance of glaciers during the same period. Comparing glacier frontal position trends in the periods 2000-2006 and 2006-2010 shows a more positive median in 2006-2010 in DB12, 13 and 16, aligning with the cooling period from the mid 2000s until 2010. However, DB 14 and 15 do not follow this trend; they show a more positive median in 2000-2006 in comparison to 2006-2010. The closest station to these two drainage basins, Dumont d'Urville, shows an anonymously cold period between 2004 and 2006 which could be associated with glacier advance.

Table 4.6: Mean summer temperatures during the time periods of glacier change.

Mean summer temperature ^{o}C	**1974-1990**	**1990-2000**	**2000-2010**
Mirny	-3.00	-3.80	-3.09
Casey	-0.81	-1.56	-1.33
Dumont d'Urville	-1.84	-2.57	-2.19
Scott	-6.93	-6.95	-6.92

Table 4.7: Mean annual temperatures.

Mean annual temperature °C	1974-1990	1990-2000	2000-2010
Mirny	11.16	-11.58	-10.8
Casey	-8.86	-9.51	-8.9
Dumon d'Urville	-10.49	-10.89	-10.92
Scott	-19.73	-19.97	-19.82

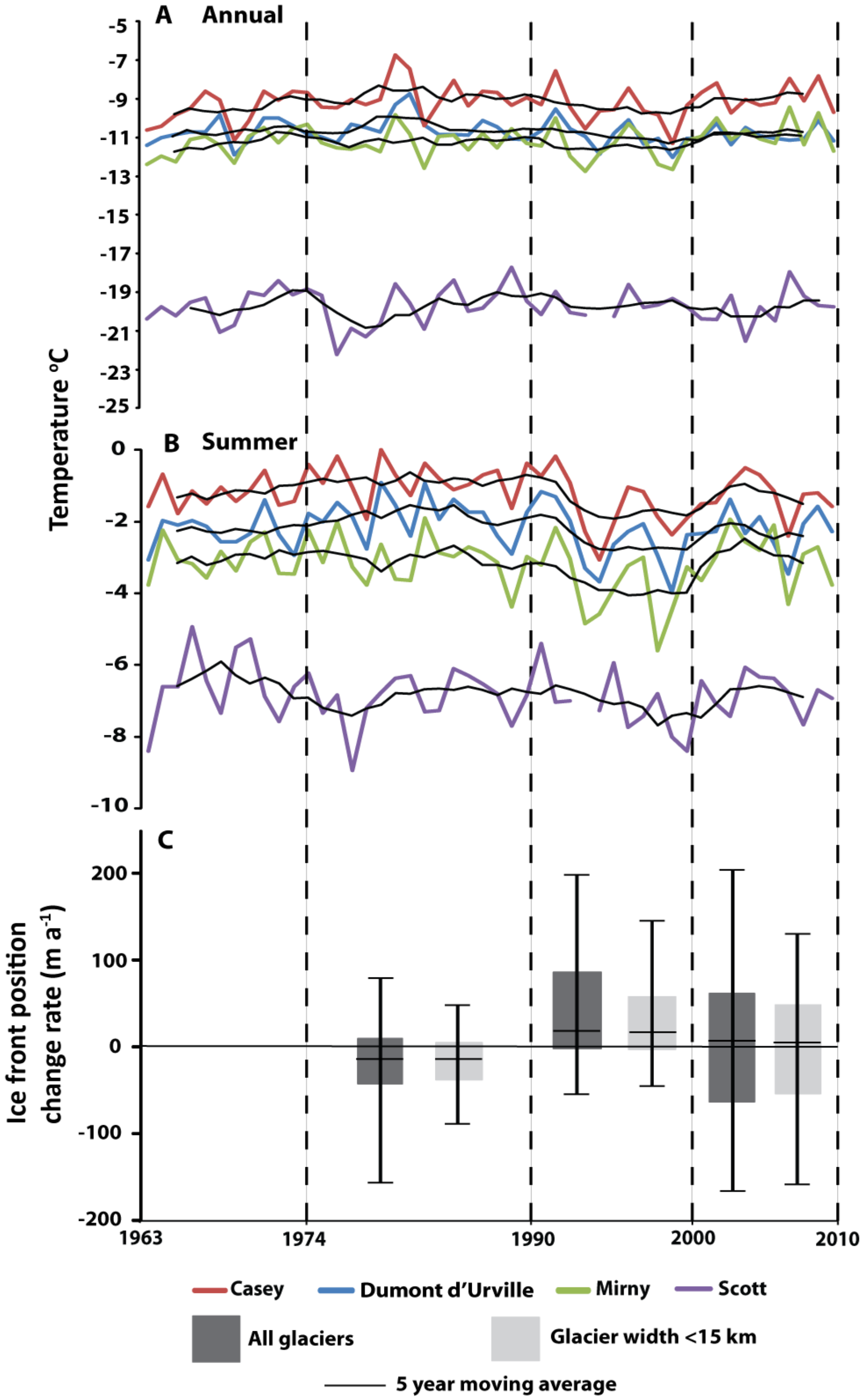


Figure 4.12: Mean annual and mean summer (December, January and February) temperature trends from four East Antarctic research stations, plotted with median ice front position change.

4.6 Statistical analysis

It is important to determine if the trends in both glacier frontal position and temperature from the different epochs are statistically significant. The key issue is whether the difference between the means from two different epochs (say the 1974-1990 period versus the 1990-200 period) might have been drawn from the same population by random sampling or are they drawn from two different populations. When data are normally distributed, this can be determined using the *t*-test of significance, which calculates the probability (*p*-value) that the two samples could be drawn from the same population. Convention dictates that a *p*-value <0.05 is sufficient to indicate that there is a 'significant' difference between the two populations (at 95% confidence) and with <0.01 'highly significant' (99% confidence). However, the data set used in this study is not normally distributed, and is positively skewed towards a few very high values. Therefore, it could be argued that these results carry less significance than using a non-parametric test. Although, this is improbable as the t-test is generally thought to be insensitive to violations of normality (Blair and Higgins, 1980), especially with large sample sizes, and is unlikely to lead to a type one error (i.e. find a significant difference which does not exist).

Given that this study is interested in the possibility of a significant difference in either direction (i.e. the 'second' epoch might be characterised by a mean frontal position change that is significantly less than or more than the 'first'epoch), all tests specified a two-tailed distribution. Two tests were then performed: (i) a paired t-test, where we only compare data from glaciers that were measured in both time-steps; and (ii) and unpaired t-test, where data were included even if the glacier was only measured at one of the time-steps. In these unpaired tests unequal variance (heteroscedastic) was assumed, which works equally well if equal variance holds but is more reliable if it does not.

The results show that there are highly significant trends when comparing the 1974-1990 period to the 1990-2000 period, analysing all glaciers, see Table 4.8. Spatially, there are highly significant trends in DB13 and DB14, and significant trends in DB15 (unpaired). However, there are no significant trends in DB12 and DB16. Although, in DB12 there is a considerably lower sample size (n=5), which may influence the significance value. Therefore, taking this into consideration, glacier change is consistent with temperature trends with Casey (DB13), Mirny (DB12) and Dumont d'Urville (DB14/DB15) stations, which all show significant trends when comparing mean summer temperatures between the 1974-1990 and 1990-2000 periods; whereas, in Scott station (DB16), there is not a significant trend, see Table 4.11. Indeed, the less significant (unpaired) and insignificant (paired) trends in DB15, may be explained by the fact DB15 has a large latitudinal range (see Fig. 4.1) and is effectively in a transition zone between colder and insignificant temperature trends (e.g. Scott station) and, the warmer and significant temperature trends (e.g. Dumont d'Urville). This is confirmed by performing the same *t*-test on the 20 most northern glaciers (Dumont d'Urville) and on the 20 most southerly glaciers (Scott). Results show a highly significant value for the most northern glaciers (p=0.0034), whilst an insignificant value for the 20 and colder glaciers (p=0.28), see Table 4.9.

In contrast, results comparing glacier frontal position between the 1990-2000 period and the 2000-2010 period are largely insignificant, see Table 4.10. This is consistent with the temperature trends, which show no significant trends between the 1990-2000 and 2000-2010 (Table 4.11). However, the significant trends in all glaciers <15 km and DB13 shows a divergence between the temperature record and glacier change, suggesting an influence from other climatic forcing's.

Table 4.8: Two-tailed t-tests for significance differences between glacier frontal position change between 1974-1990 and 1990-2000.

Sample	***t*-test***	**Epoch**	**n**	**Rate of frontal position change (m a^{-1})**				***p*-value#**
				Min.	**Median**	**Mean**	**Max.**	
All glaciers	Unpaired	1974-1990	131	-1611.7	-12.5	-43.3	678.9	**0.0026**
		1990-2000	168	-1591.4	19.7	43.1	1605.4	
	Paired	1974-1990	131	-1611.7	-12.5	-43.3	678.9	**0.0288**
		1990-2000	131	-1591.4	13.5	29.6	1605.4	
All glaciers < 15 km	Unpaired	1974-1990	115	-1091.5	-12.5	-25.3	341.4	**0.0004**
		1990-2000	143	-491.1	18.4	27.6	265.1	
	Paired	1974-1990	115	-1019.5	-12.5	-25.3	341.4	**0.0010**
		1990-2000	115	-491.1	11.9	22.0	265.1	
DB12	Unpaired	1974-1990	5	-1611.7	-13.2	-326.6	382.6	0.5012
		1990-2000	7	-1591.4	8.9	-326.6	1605.4	
	Paired	1974-1990	5	-1611.7	-13.2	-326.6	382.6	0.7314
		1990-2000	5	-1591.4	-49.6	-19.3	1605.4	
DB13	Unpaired	1974-1990	15	-406.8	-49.4	-83.5	96.7	**0.0004**
		1990-2000	37	-265.8	54.9	65.8	359.7	
	Paired	1974-1990	15	-406.8	-49.4	-83.5	96.7	**0.0173**
		1990-2000	15	-137.2	27.4	52.8	217.0	
DB14	Unpaired	1974-1990	24	-1388.2	-43.5	-161.7	678.9	**0.0266**
		1990-2000	35	-772.5	30.3	-161.7	1054.2	
	Paired	1974-1990	24	-1388.2	-43.5	-161.7	678.9	**0.0025**

		1990-2000	24	-772.5	24.0	19.6	713.1	
DB15	Unpaired	1974-1990	69	-302.6	-3.5	2.1	341.4	**0.0398**
		1990-2000	70	-289.5	12.6	30.5	265.1	
	Paired	1974-1990	69	-302.6	-3.5	2.1	341.4	0.0996
		1990-2000	69	-289.5	11.9	30.5	265.1	
DB16	Unpaired	1974-1990	18	-87.7	7.5	52.8	518	0.6912
		1990-2000	19	-380.9	1.8	32.3	528.8	
	Paired	1974-1990	18	-87.7	7.5	52.8	518	0.5049
		1990-2000	18	-380.9	4.5	34.2	528.8	

* All unpaired tests assume unequal variance. # p-values <0.05 are significant and highlighted in yellow

Table 4.9: Two-tailed t-tests for significance differences between glacier frontal position change between the 20 most northerly and 20 most southerly glaciers in DB15.

DB15 sample	***t*-test***	**Epoch**	**n**	**Rate of frontal position change (m a^{-1})**				***p*-value#**
				Min.	**Median**	**Mean**	**Max.**	
Northerly glaciers	Unpaired	1974-1990	18	-260.9	-22.3	-24.6	110.2	**0.0034**
		1990-2000	19	-73.0	28.8	51.7	232.9	
	Paired	1974-1990	18	-260.9	-22.3	-24.6	110.2	**0.0142**
		1990-2000	18	-73.0	25.1	52.7	232.9	
Southerly glaciers	Unpaired	1974-1990	20	-81.8	16.6	49.8	341.4	**0.2834**
		1990-2000	20	-289.5	8.3	13.7	193.0	
	Paired	1974-1990	20	-81.8	16.6	49.8	341.4	**0.3800**
		1990-2000	20	-289.5	8.3	13.7	193.0	

* All unpaired tests assume unequal variance.
# p-values <0.05 are significant and highlighted in yellow

Table 4.10: Two-tailed t-tests for significance differences between glacier frontal position change between 1990-2000 and 2000-2010.

Sample	***t*-test***	**Epoch**	**n**	**Rate of frontal position change (m a^{-1})**				***p*-value#**
				Min.	**Median**	**Mean**	**Max.**	
All glaciers	Unpaired	1990-2000	168	-1591.4	19.7	43.1	1605.4	0.2064
		2000-2010	171	-6913.7	8.4	-17.9	1759.2	
	Paired	1990-2000	165	-1591	20.6	44.3	1605.4	0.2391
		2000-2010	165	-6913.7	8.4	-19.6	1759.2	
All glaciers < 15 km	Unpaired	1990-2000	143	-491.1	18.4	27.6	265.1	**0.0240**
		2000-2010	144	-795.4	6.5	-6.9	350.9	
	Paired	1990-2000	140	-491.1	18.9	28.7	265.1	**0.0255**
		2000-2010	140	-795.4	6.5	-9.5	350.9	
DB12	Unpaired	1990-2000	7	-1591.4	8.9	17.5	1605.4	0.5421
		2000-2010	8	-255.1	44.9	281.5	1759.2	
	Paired	1990-2000	7	-1591.4	8.9	17.5	1605.4	0.4120
		2000-2010	7	-255.1	15.5	311.1	1759.2	
DB13	Unpaired	1990-2000	37	-265.8	54.9	65.8	359.7	**0.0002**
		2000-2010	39	-336.5	-63.6	-54.2	479.2	
	Paired	1990-2000	37	-265.8	54.9	65.8	359.7	**0.0017**
		2000-2010	37	-336.5	-63.6	-50.8	479.2	

DB14	Unpaired	1990-2000	35	-772.5	30.3	55.2	1054.2	0.4649
		2000-2010	38	-6913.7	19.4	-91.5	834.7	
	Paired	1990-2000	35	-772.5	30.3	55.2	1054.2	0.4983
		2000-2010	35	-6913.7	16.3	-106.8	834.7	
DB15	Unpaired	1990-2000	70	-289.5	12.6	30.5	265.1	0.3562
		2000-2010	71	-795.4	20.6	12.2	346.7	
	Paired	1990-2000	70	-289.5	12.6	30.5	265.1	0.3640
		2000-2010	70	-795.4	20.6	12.6	346.7	
DB16	Unpaired	1990-2000	19	-380.9	1.8	32.3	528.8	0.4232
		2000-2010	16	-778.2	10.9	-42.4	747.7	
	Paired	1990-2000	16	-380.9	4.5	42.6	528.8	0.4503
		2000-2010	16	-778.2	10.9	-42.4	747.7	

Table 4.11: Two-tailed *t*-test results for significant differences in mean austral summer (December, January, February) temperature between 1974-1990 and 1990-2000, and 2000-2010. All tests are unpaired and assume unequal variance.

Station	**Epoch**	**Mean summer temperature (° C)**	***p*-value**
Mirny	1974-1990	-3.0	**0.0496**
	1990-2000	-3.8	
	1990-2000	-3.8	0.0884
	2000-2010	-6.8	
Casey	1974-1990	-0.8	**0.0114**
	1990-2000	-1.7	
	1990-2000	-1.7	0.2411
	2000-2010	-1.3	
Dumont d'Urville	1974-1990	-1.8	**0.0124**
	1990-2000	-2.6	
	1990-2000	-2.6	0.1160
	2000-2010	-2.2	
Scott	1974-1990	-6.8	0.1569
	1990-2000	-7.3	
	1990-2000	-7.3	0.1797

4.7 Summary

The data presented in this section shows that glaciers measured between 1974 and 2010 show little change in terminus position (median: 0.7 m a^{-1}). However, within this period, strong decadal trends in glacier behaviour have been shown, with 63% of glaciers retreating between 1974 and 1990 (median: -12.5 m a^{-1}). Between 1990 and 2000, this trend reversed with 72% glaciers advancing at a median rate of 19.7 m a^{-1}. In the most recent period, between 2000 and 2010, 58% of glaciers advanced at a median rate of 8.4 m a^{-1}. Further analysis has revealed that wider and faster flowing glaciers undergo the largest changes in terminus position; however, it appears, such glaciers have little influence on decadal trends in glacier frontal position change. Therefore, this hints at a common external forcing(s) driving the clear trends in outlet glacier frontal position change. Furthermore, decadal trends in air temperature at four East Antarctic research stations appear to match decadal trends in glacier frontal position change.

East Antarctic climate and glacier change

5.1 Relationship between glacier frontal change and climate

The modern climate system in Antarctica is determined by the interplay of the ice sheet, ocean, sea ice and atmosphere system, and its response to both past and present climate forcing. This makes it difficult to identify any one part of the climate system likely to drive glacier change. This section will first describe each climatic factor, namely air temperature, sea ice, and the oceans, and how they can influence outlet glacier frontal position. Secondly, this section will describe how these factors are all interlinked through the dominant mode of atmospheric variability in the Southern Hemisphere (Gong and Wang, 1999; Thompson and Wallace, 2000) known as the Southern Annular Mode (SAM).

5.1.1 Air Temperature

Whilst simple relationships between glacier frontal changes and air temperatures have been invoked in similar studies focusing on the Antarctic Peninsula (Cook *et al.*, 2005) and in Greenland (Howat *et al.*, 2005; 2008; Luckman *et al.*, 2006; Van den Broeke *et al.*, 2009), temperatures in East Antarctica have been largely considered to be too cold to create enough surface melt to drive such changes (Stearns *et al.*, 2008). However, in the relatively warmer Antarctic Peninsula (AP), that has experienced rapid warming over recent years (Vaughan *et al.*, 2003); Cook *et al.* (2005) broadly demonstrated a link between air temperature records from Faraday research station (65’15° S, 64’16° W) and glacier frontal change between 64° S and 66° S (since the 1940s). Of interest, it is noted that January (austral summer) temperatures in Faraday between 1974 and 1990 are on average only ~0.7 °C warmer than in Casey, see Figure 5.1. Therefore, if air temperature is potentially forcing glacier change in the AP, it is also likely to influence frontal change in the warmer sections of East Antarctica e.g. Casey. In fact Torinesi *et al.* (2003) calculated an average summer temperature melt threshold of -4 °C for Wilkes Land. It is likely that a similar magnitude of melt threshold exists for Terre Adélie , George V Land and Oates Land, given the comparable summer temperatures between Dumont d’Urville and the Wilkes Land stations (Casey and Mirny). Thus, as summer temperatures rarely dip below this threshold (Fig. 4.12), some degree of melt does take place each summer. Indeed, in this study, examples of surface melt are evident on the imagery through water ponding on the surface in marginal areas of outlet glaciers, as seen in Fig. 4.8. Such surface melt can influence calving rates through enhancing crevasse opening through the process of hydro-fracturing (van der Veen *et al.*, 1998) e.g. Larsen B (Scambos *et al.*, 2003) and ultimately influence ice front position. This process is rapid and thus consistent with the rapid response of terminus position change to climate forcing.

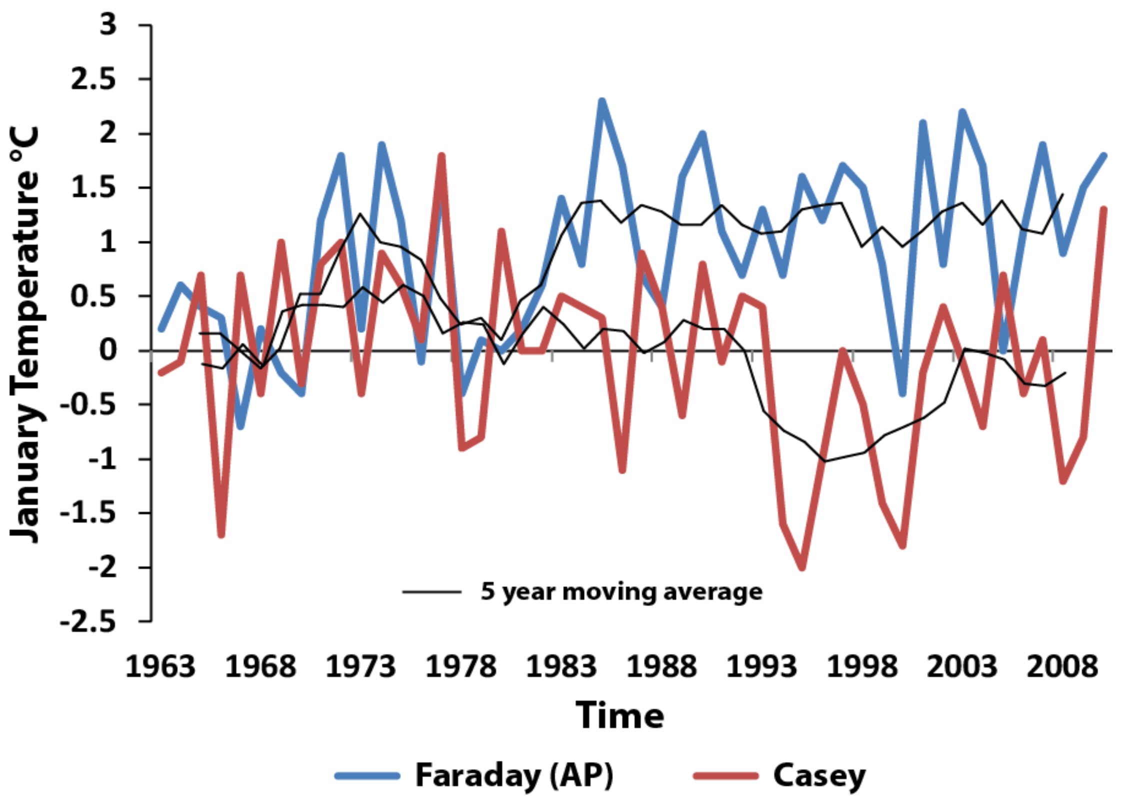


Figure 5.1: Casey and Faraday January temperatures from 1963 to 2010, data from MET reader (http://www.antarctica.ac.uk/met/READER/). Note the divergence in trends from 1980 onwards (Faraday = warming, Casey = cooling), driven by ozone depletion, see section 5.1.4.

5.1.2 Sea Ice

In general, measurements of sea ice around Antarctica are split into five sections, with the Western Pacific ocean sector (90-160° E) covering the majority of glaciers in the study area. This sector is chosen as the sole representation of sea ice in this study as it provides the best spatial representation. In the West Pacific ocean sector, sea ice extent (SIE) has been increasing at a rate of 1404 ± 1017 km yr^2 between 1978 – 2008 (Comiso *et al.*, 2011). However, this trend is not temporally consistent, with below average SIE in the 1980s, and above average SIE in the 1990s, followed by a return to average SIE in the early 2000s. This trend is broadly consistent with both glacier behaviour and air temperature trends (Fig. 5.2). Nevertheless, SIE may not exert a direct influence on glacier frontal position as most of these changes are occurring at the outer pack edge (Stammerjohn *et al.*, 2003) far away from the coastline and glacier termini. Instead, it is the concentration of ice at the coastline which can directly influence glacier calving. Indeed, the presence of sea ice and sea ice mélange (dense mixture of calved icebergs) at the glacier calving front has been shown to suppress calving (Joughin *et al.*, 2008; Amundson *et al.,* 2010) and several researchers have highlighted the importance of sea ice and mélange in both the short term and long term evolution and stability of glaciers e.g. Jakobshavn (Luckman and Murray, 2005; Joughin *et al.*, 2008; Howat *et al.*, 2010).

Another component of sea ice which is thought to influence glacier stability and calving is fast ice (a.k.a. Landfast). This is sea ice which becomes fixed or 'fast' to a shoreline, grounded ice bergs, islands, or ice fronts of a grounded ice sheet or floating ice shelves (World meteorological organization, 1970). Most fast ice forms annually in a narrow band around much of the Antarctic Ice Sheet (Massom *et al.*, 2010). However, some areas of thicker fast ice persist year around in certain locations e.g. sheltered embayments (Giles *et al.,* 2008), areas of favourable local bathymetry (Massom, 2001) and grounded iceberg distribution (Massom, 2003). Fast ice has been shown to ensure a greater stability of glacier floating tongues by protecting the glacier front from wave action and winds (Reeh *et al.*, 2001), suggesting the two are mechanically coupled (Massom *et al.*, 2010). Indeed, whilst observing the Nioghalvfjerdsfjorden glacier in Greenland, Reeh *et al.* (2003) observed no significant calving whilst fast ice was present on the glacier front. However, as the fast ice broke up, calving became extensive. Similar results have been found in other glaciers in Greenland (Seale et al., 2011; Joughin et al. 2012).

In the context of this study, it highlights the importance of the relative concentrations of mélange, sea ice and fast ice at the glacier terminus. Analysis of mean sea ice concentration (SIC) using data from the Nimbus-7 SMMR and DMSP SSM/I-SSMIS passive microwave data plotted using the KNMI climate explorer, at each of the time periods used in this study, hints at a lower SIC in 1974-1990, shown in Figure 5.3. This is confirmed when considering the absolute difference in SIC between the means of 1990-2010 and 1974-1990, shown in Figure 5.4.

In East Antarctica, coastal SIC seems to be largely driven by the intensity of off-shore katabatic winds, which create coastal polynyas (Gallee, 1997; Bromwich *et al.* 1982, 1984, 1994; Kern, 2009). Indeed, Kurtz and Bromwich (1983) note the decrease in the size of the Terra Nova polynya and subsequent rapid freezing of sea ice in response to the localised weakening of katabatic winds. Furthermore, Massom *et al.* (2009) highlight the sensitivity of fast ice formation off the coast Dumont d'Urville to wind direction. They infer below average or no fast ice formation when the annual wind direction is more off-shore (SE), whilst above average fast forms when the annual wind is directed more along-shore (E). Further analysis on the mean annual wind direction at Dumont d'Urville reveals a trend for less off-shore winds from the early 1990s to the mid 2000s, shown in Figure 5.5, suggesting greater fast ice extent during this period.

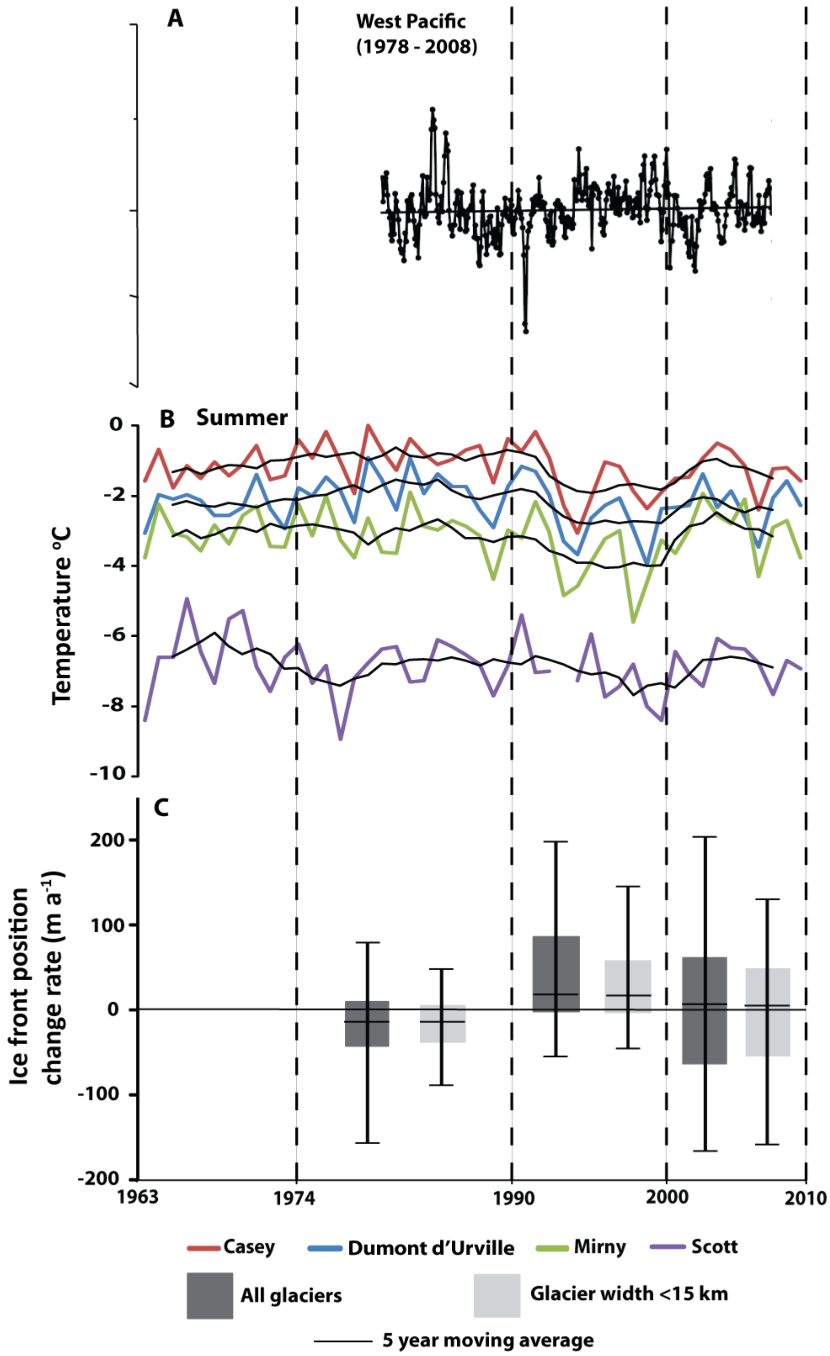


Figure 5.2: A) Sea ice area in the West Pacific ocean extent (from Comiso *et al.*, 2011) B) Summer temperature trends C) Median ice front position change.

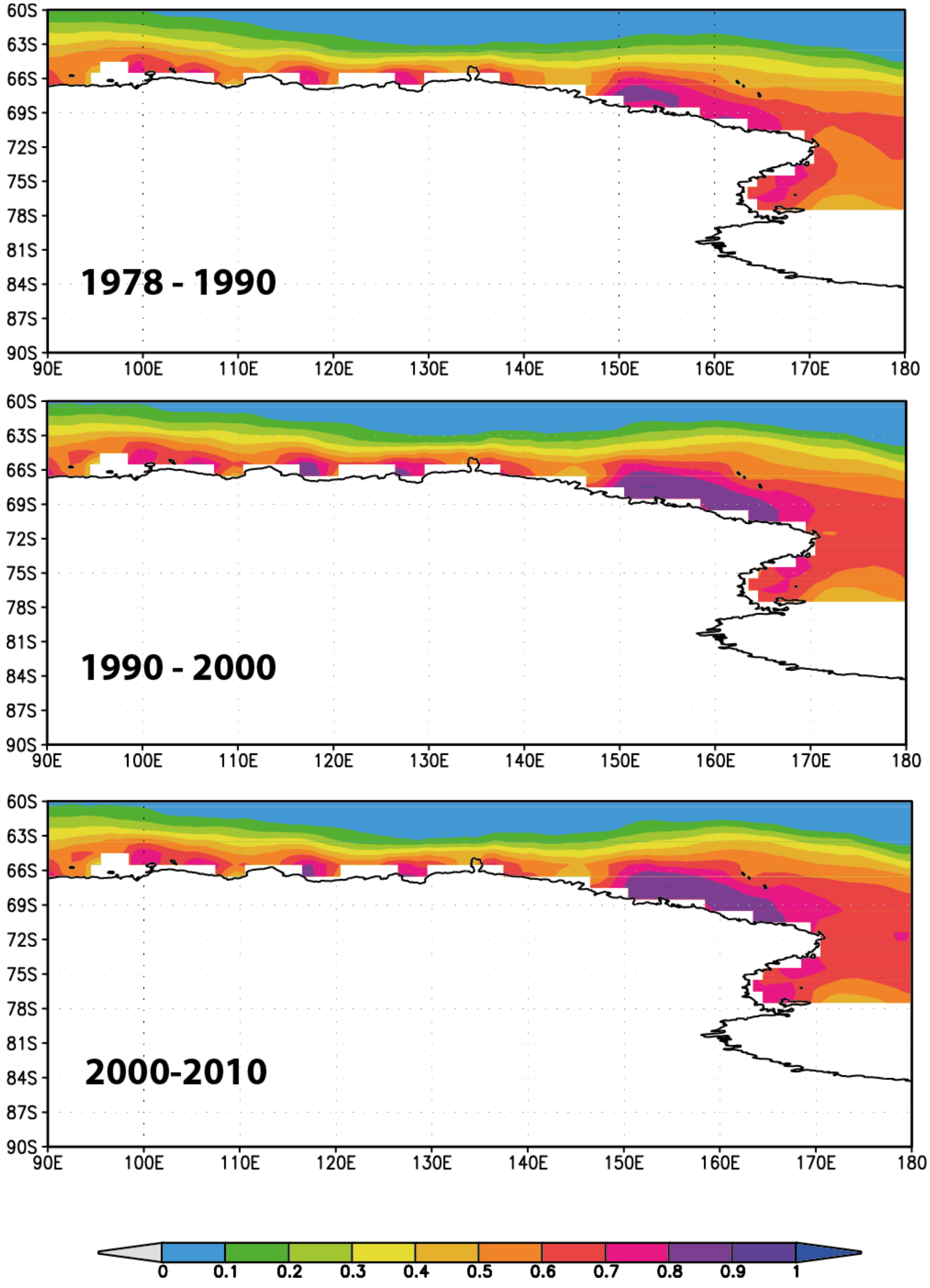


Figure 5.3: Mean coastal sea ice concentrations (1978-2010) plotted from Nimbus-7 SMMR passive microwave data, taken from the National Snow and Ice Data Centre and plotted using the KNMI climate explorer (http://climexp.knmi.nl/), note the earliest measurement available was from 1978.

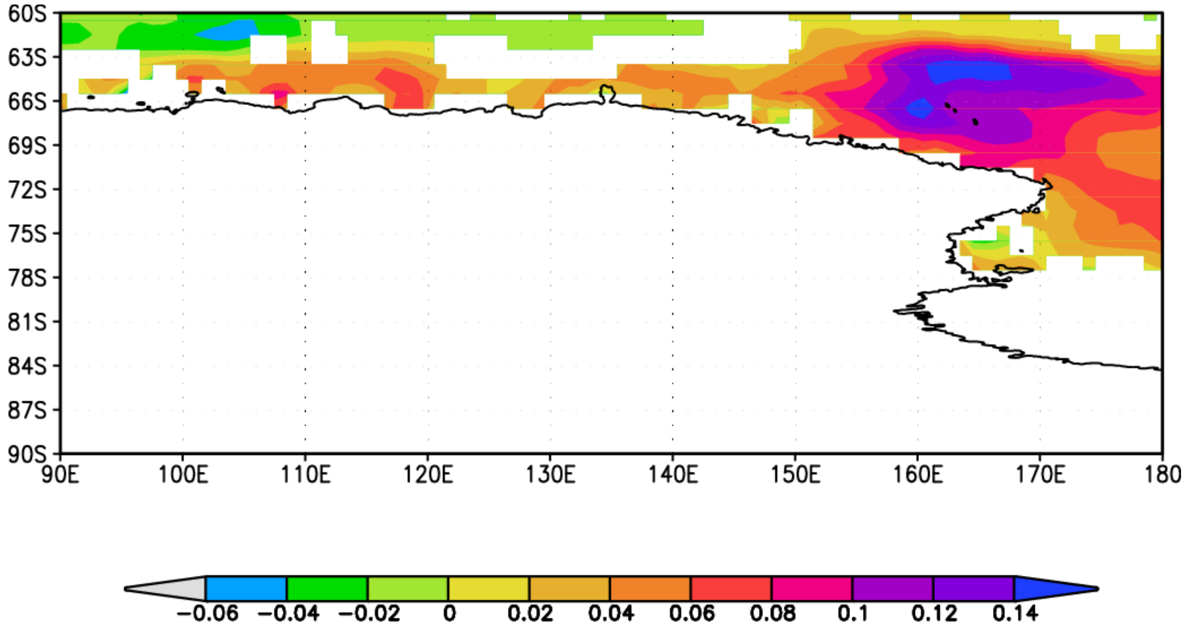


Figure 5.4: Mean difference in sea ice concentrations between 1990-2010 and 1974-1990 (shown in Fig.3), showing an increase in sea ice concentration in 1990-2000 compared to 1974-1990.

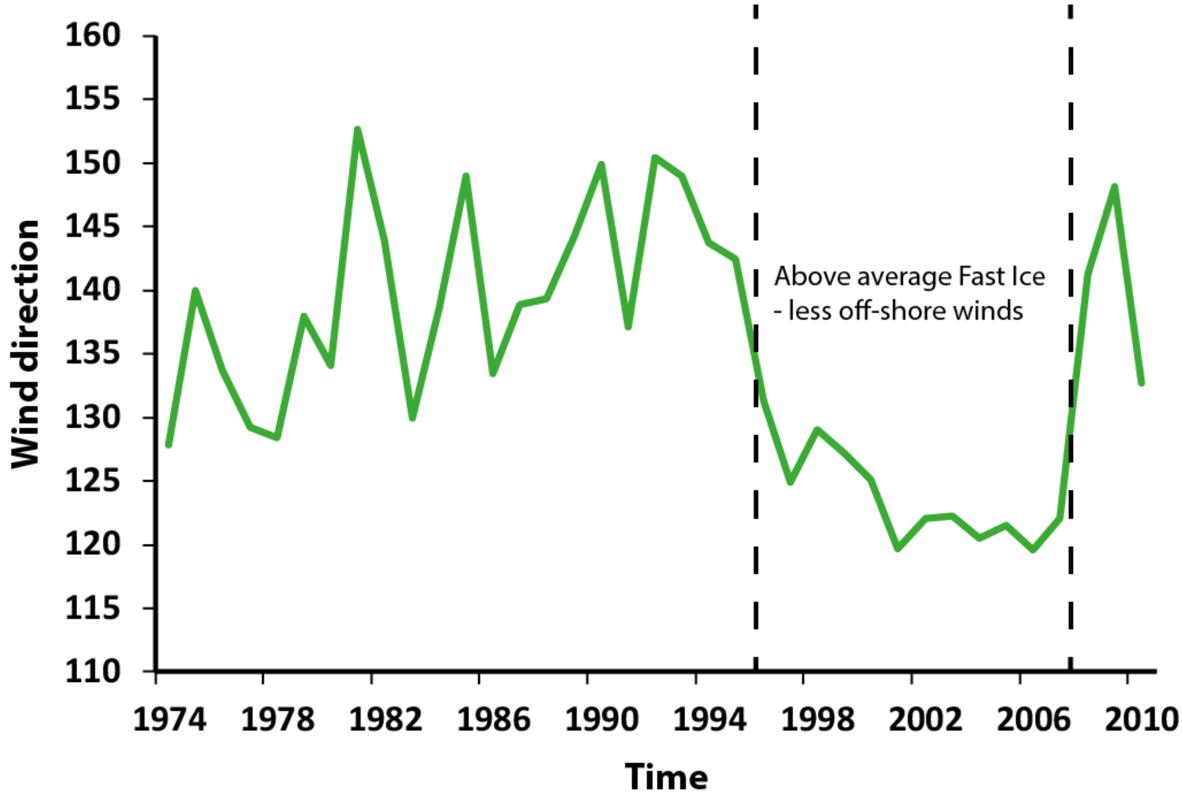


Figure 5.5: Mean Annual wind direction at Dumont d'Urville, data from MET reader (http://www.antarctica.ac.uk/met/READER/).

5.1.3 Ocean forcing

Despite limited oceanic data, the oceans are still thought to play a crucial role in marine terminating glacier evolution, see section 2. In Antarctica, ice shelf and glacier thinning by basal melt has been largely attributed to fluctuating intrusions of relatively warm Circumpolar Deep Water onto the continental shelf (Jacobs *et al.*, 1996, 2011; Holland *et al.*, 2010). The CDW is believed to be channelled at depth (~300 m) through bathymetric troughs in the sea floor below the ice shelf, exerting a potential for vigorous basal melt (Wahlin *et al.*, 2010). Indeed, Pritchard *et al.* (2012) note a correlation between higher thinning rates and ice shelf depth, inferring lower thinning rates for shallower ice shelves which do not reach the channelized CDW. The contrasting behaviour of neighbouring ice shelves in the Amundsen Sea sector, despite experiencing similar atmospheric forcing, suggests that the observed inland thinning in some sections of coastal Antarctica must be driven by increased basal melt of ice shelves, as oppose to glacier retreat or mass balance changes (Pritchard *et al.*, 2012). This is shown in Pine Island glacier sector, where basal melting of the PIG ice shelf (Jacobs *et al.*, 2011) has led to the unstable retreat of the glacier front (Shepherd *et al.*, 2001), an increase in glacier velocity (Joughin *et al.*, 2011) and a threefold increase in inland thinning (Wingham *et al.*, 2009).

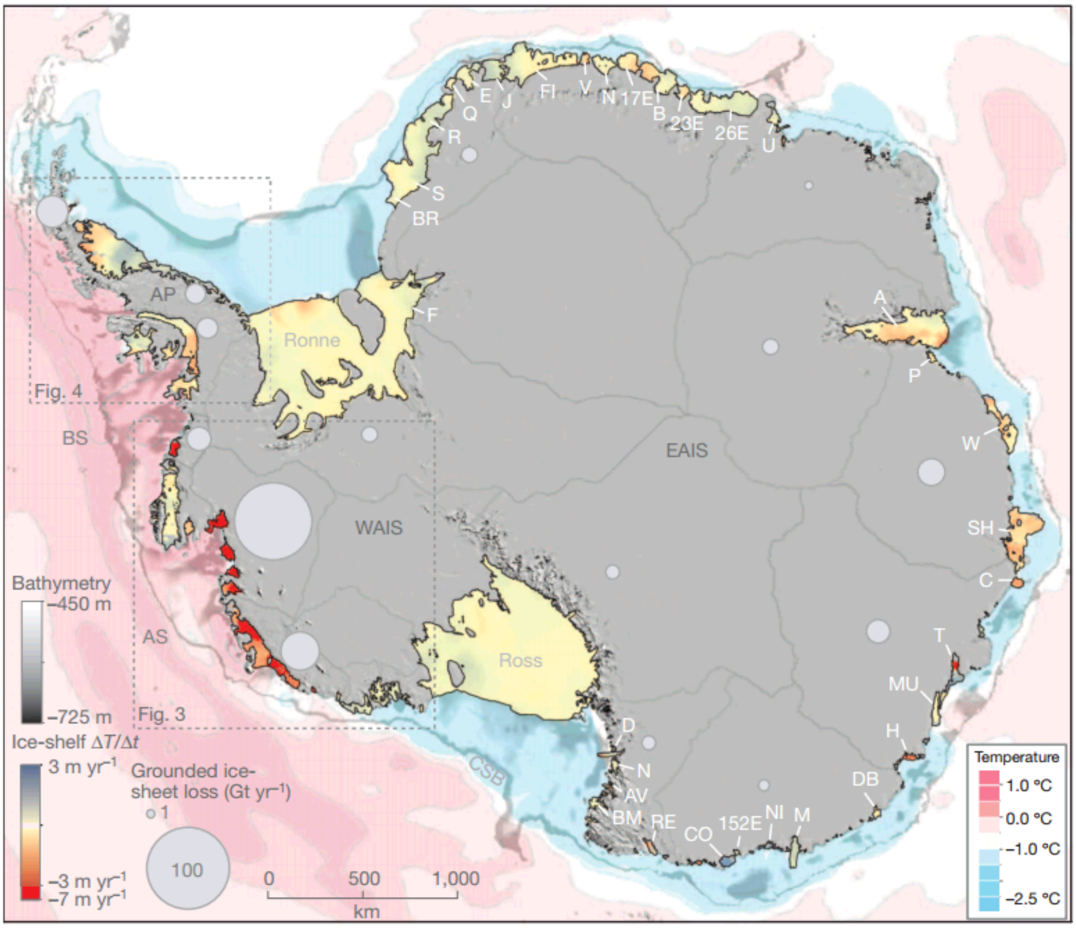


Figure 5.6: Antarctic ice shelf thickness change (Taken from Pritchard *et al.*, 2012). Ice shelf bathymetry (grey) is overlain on estimated sea floor potential temperatures. Ice shelf thickness change is shown in the labelled glaciers (blue = thickening; red = thinning).

In the region this study concentrates on, there have been no reports of CDW intrusions (Rignot et al., 2008). However, Pritchard *et al.* (2012) have noted the basal thinning of some of the thicker ice shelves in the study area, indicating some intrusions of CDW, see Fig. 5.6. Indeed, the area of most pronounced recent thinning, retreat and mass loss, DB13, overlaps the Aurora subglacial basin where preserved glaciated valleys have been recently discovered (Young *et al.*, 2011), see Figure 5.7.

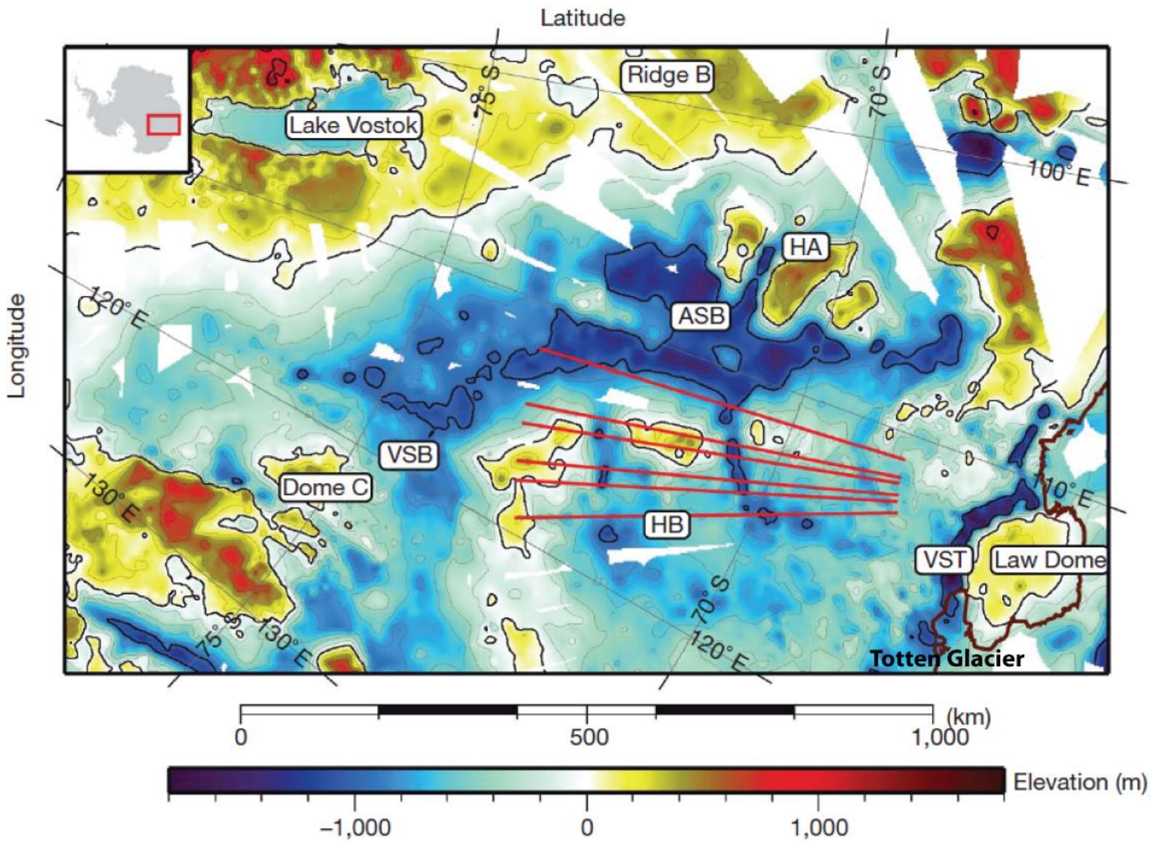


Figure 5.7: Bed topography of the Aurora subglacial basin region (Young *et al*., 2011), located in DB13, East Antarctica.

5.1.4 Southern Annular Mode

The Southern Annular Mode (SAM) is a large scale pattern of variability characterized by fluctuations in the strength of the circumpolar vortex (Marshall, 2002). The SAM dominates atmospheric variability in the Southern Hemisphere (Gong and Wang, 1999; Thompson and Wallace, 2000) and is known to influence wind speed and direction, sea ice extent and concentration, surface temperatures and sea-surface temperatures (Hall and Visbeck, 2002; Stammerjohn *et al.*, 2008; Massom *et al.*, 2009). Over recent decades, the SAM has shown significant positive trends, particularly during the austral autumn and summer (December – May) (Thompson and Solomon, 2002) (Fig. 5.8). Thompson and Solomon (2002) suggest that recent trends in SH tropospheric circulation can be interpreted as a bias towards a positive SAM, with several authors linking these trends to stratospheric ozone depletion (Thompson and Solomon, 2002; Gillet and Thompson, 2003) and increased greenhouse gases (Marshall *et al.*, 2004; Shindell and Schmidt, 2004). Indeed, Arblaster and Meehl (2006) find ozone to be the dominate forcer of recent trends in the SAM, although the increase of greenhouse gases is also a necessary factor to explain recent trends. They also suggest that solar forcing may be contributing to summer trends in the SAM.

It is hypothesized that the resulting intensification of circumpolar westerly airflow and delay in the annual breakdown of the polar vortex can be linked to the observed warming in the Antarctic Peninsula and cooling in East Antarctica (Fig. 5.1) (Kwok and Comiso, 2002; Thompson and Solomon, 2002; Van den Broeke and Van Lipzig, 2003). The stronger vortex also acts to suppress katabatic outflow throughout much of East Antarctica, through pushing coastal depressions further away from the coast, thus decreasing the pressure gradient and decreasing the intensity of katabatic winds (Van den Broeke and Van Lipzig, 2003), see Fig. 5.9. This favours an increased coastal SIC and fast ice during a positive SAM.

The signature of the positive trend in the SAM is also thought to manifest in the oceans. Hall and Visbeck (2002) show that the intensification of circumpolar westerly airflow can be associated with a stronger circumpolar current, causing strengthening northward Ekman flow which, in turn, results in a reduction of poleward heat transport. These heat transport anomalies are associated with decreased SST at ~55° S, although the drop in SST is relatively small (0.1 °C) due to the shallow SST gradients in this region. The increased northward surface ocean current flow also results in a greater sea ice extent, with surface currents acting as a conveyor belt driving ice further away from the continent (Hall and Visbeck, 2002). As a result of the increased Ekman flow, Hall and Visbeck (2002) also predict a 40% increase in coastal upwelling to satisfy mass continuity.

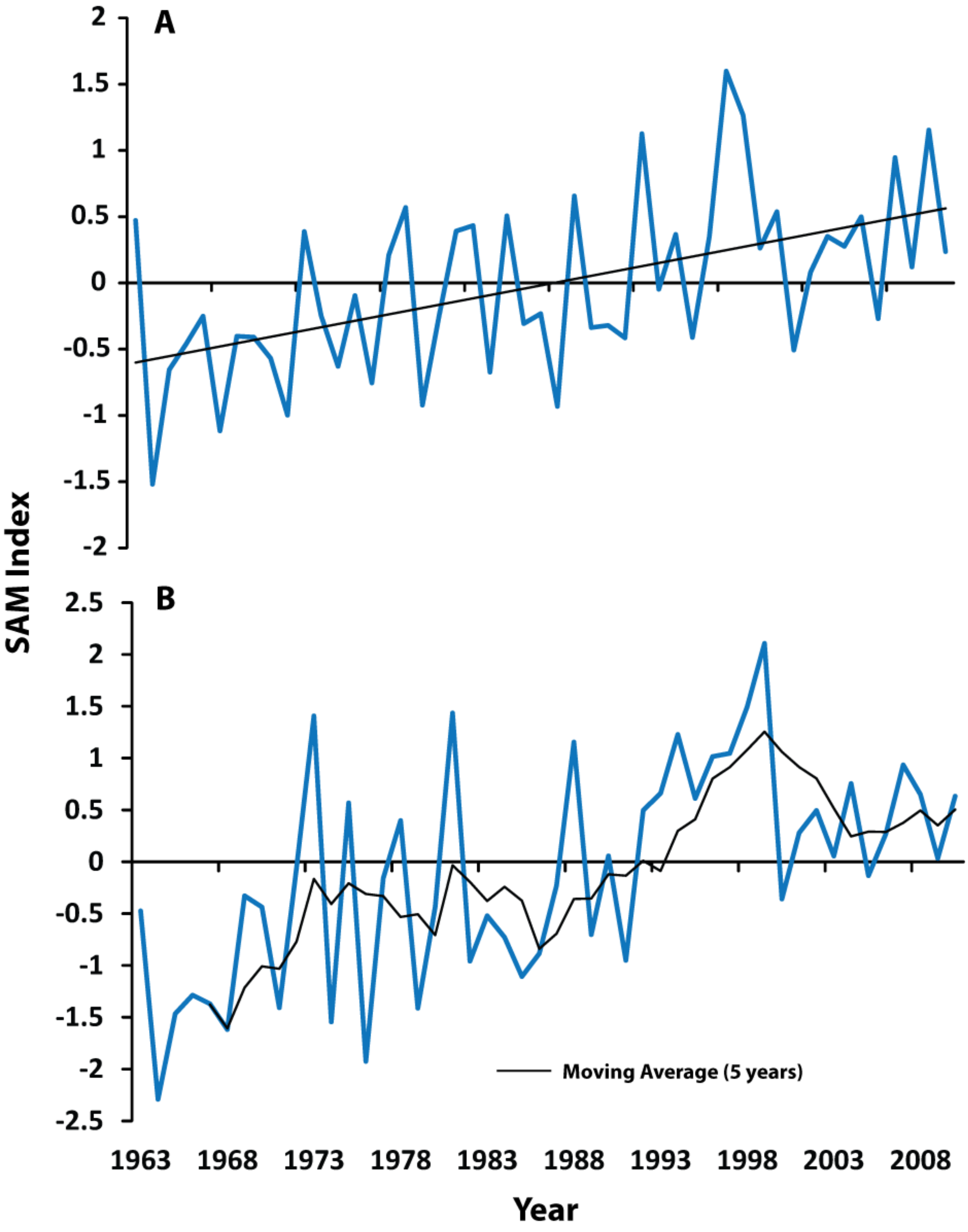


Figure 5.8: A) Annual averaged SAM index. B) December – May averaged SAM index. Data taken from http://www.nerc-bas.ac.uk/icd/gjma/sam.html (Marshall , 2003).

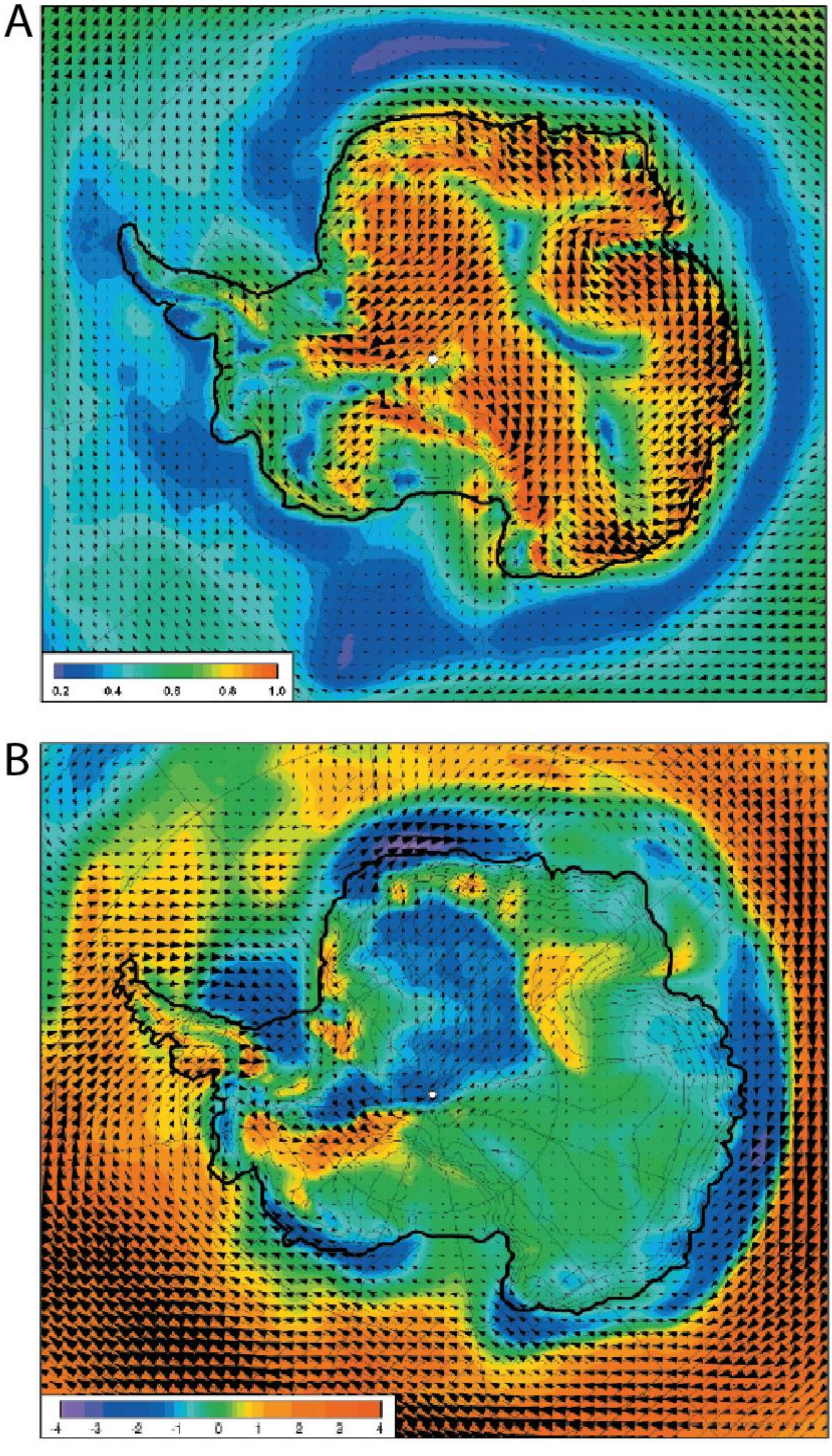
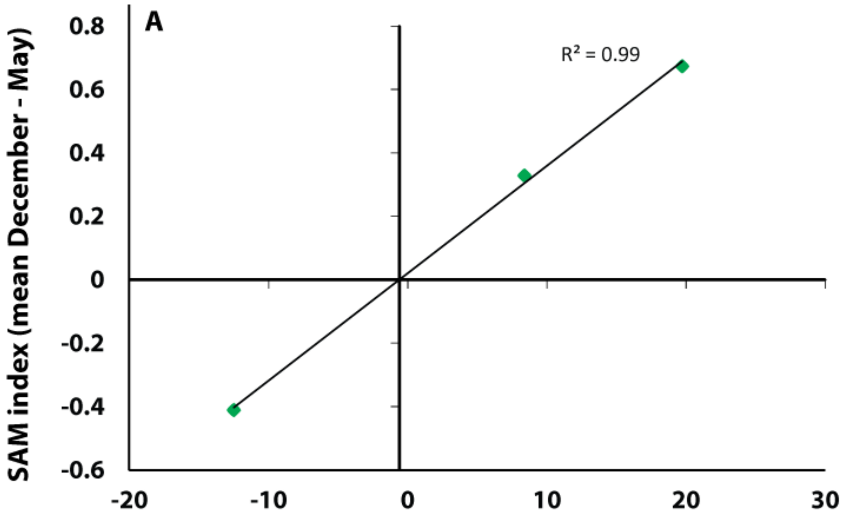


Figure 5.9: a) July surface wind vectors (arrows) and directional constancy (colours) for SAM negative polarity b) Difference in July surface wind vectors (arrows) and wind speed (colours) for SAM positive polarity minus SAM negative polarity. Note the decrease in wind speed around the coast of East Antarctica (suppression of katatbatics) and increase in wind speed around the continent (stronger circumpolar vortex). (Taken from Van den Broeke and Van Lipzig, 2003).

Discussion

6.1 Relationship to glacier frontal change

The general synchronicity in terminus change shown by glaciers of different width classes at each time period clearly suggests a common force driving the changes in glacier ice front position observed. The significant trend of retreat of glaciers between 1974 and 1990 to advance from the 1990s onwards is consistent with the shift from a negative to an increasingly positive SAM. From 1974 to 1990, the SAM was mainly negative; the relatively weaker polar vortex will have resulted in an intensification of katabatic outflow, driving sea ice away from the coast, lowering SIC at the glacier front. Furthermore, the relatively warmer summer temperatures linked to the weaker polar vortex will have increased the potential for surface melt. The combination of these climatic factors is likely to have driven the observed retreat. From the 1990s however, the general trend of glaciers to advance is more consistent with a mainly positive SAM. A cooling of summer temperatures will have reduced summer surface melt and a more intense polar vortex will have suppressed katabatic outflow. This, combined with the probable increase in available icebergs/mélange from the retreat phase (1974-1990) (Massom and Stammerjohn, 2010), will have aided an increase in SIC at glacier fronts, delaying calving and favouring advance. Figure 6.1 attempts to quantify the relationship between glacier frontal position change, the SAM (r=0.99) and temperature (r=0.93). These results provide a valuable insight into the relationship between the two climatic variables and glacier frontal position change. However, due to the low number of degrees of freedom, these results are statistically insignificant. Indeed, contrary to this relationship is the intensification of coastal upwelling of warm CDW also associated with an increasingly positive SAM. Intrusion of CDW onto the continental shelf, aided by deep bathymetric troughs, could result in increased basal melting and weakening of glaciers (Thoma *et al.*, 2008; Holland *et al.*, 2010; Jacobs *et al.*, 2011). Indeed, consistent with this are the recent observations of such basal thinning of some ice shelves within the area this study concentrates on (Pritchard *et al.*, 2012). In other regions basal thinning has been linked to glacier retreat e.g. Pine Island glacier (Jenkins *et al.*, 2010). However, it would appear that this process is yet to exert a major influence on glacier frontal position within this study area.



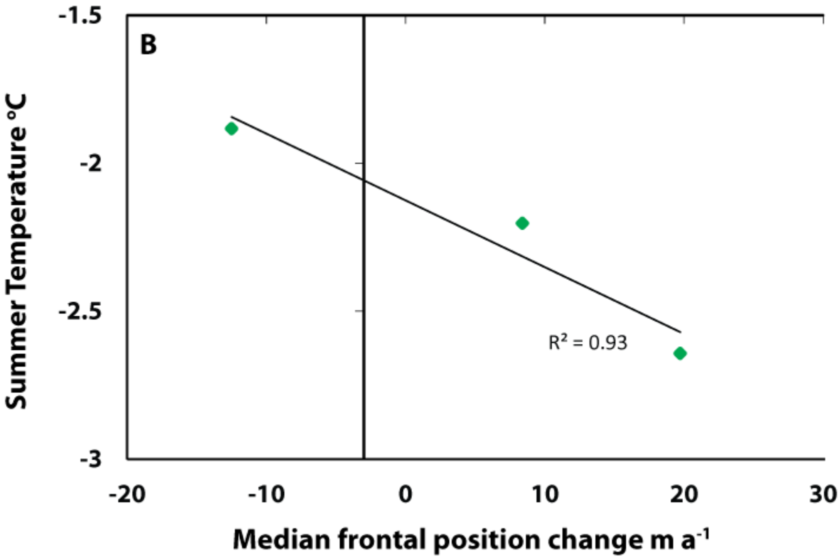


Figure 6.1: Scatterplots of median frontal position change versus A) December – May SAM index. B) Mean austral summer (December, January, February) temperature, data from the Casey, Dumont d'Urville and Mirny stations. It is important to note that temperature data from Scott station was not included. This is because it is considered too cold here for surface melt and, therefore, effect glacier frontal position.

Within the overall trend of glacier advance since the 1990s, there is considerable spatial variability, most notably the retreat of DB13 in 2000-2010. However, analysis at sub-decadal time steps indicates that this retreat took place between 2000 and 2006, before stabilization between 2006 and 2010, whereas in neighbouring DB14 the opposite occurs (advance 2000-2006; stabilization 2006-2010). These trends are also reflected in the relevant temperature records for DB13 (Casey) and DB14 (Dumont d'Urville). In general, the two temperate records are synchronous; however, there is an anomalous warming period for Casey between 2002 and 2005, diverging from the Dumont d'Urville trend, shown in Fig.

6.2. This is supported by the number of positive degree days and hours at Casey, which shows a notable spike at the 2004/5 melt season, in comparison to the rest of the decade, see Figure 6.3.

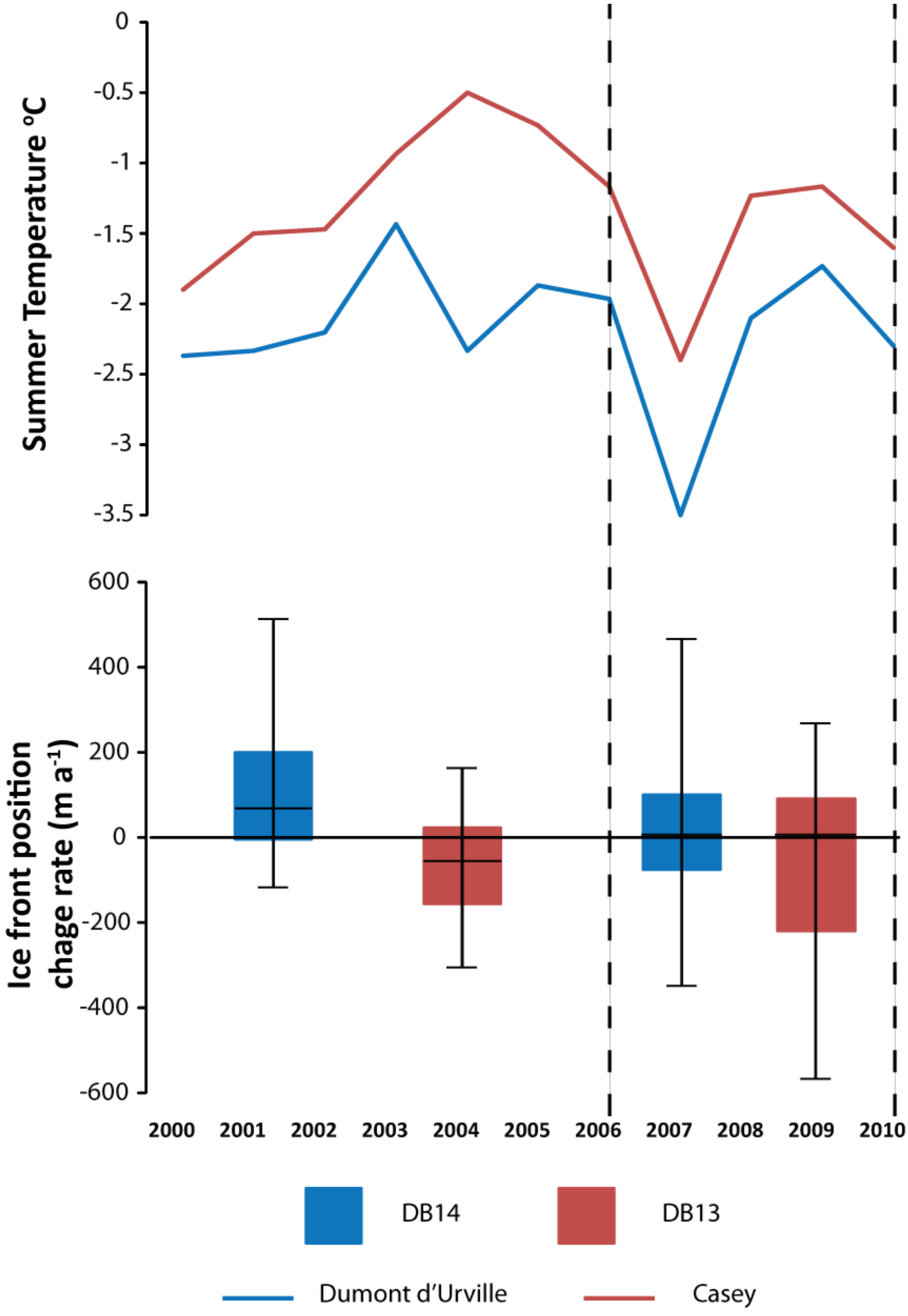


Figure 6.2: Summer temperature trends (top) at Dumont d'Urville (blue) and Casey (red), representing DB14 and DB13 respectively. These are compared to median ice front position change rates (bottom) for DB13 and DB14 at 2000-2006 and 2000-2010.

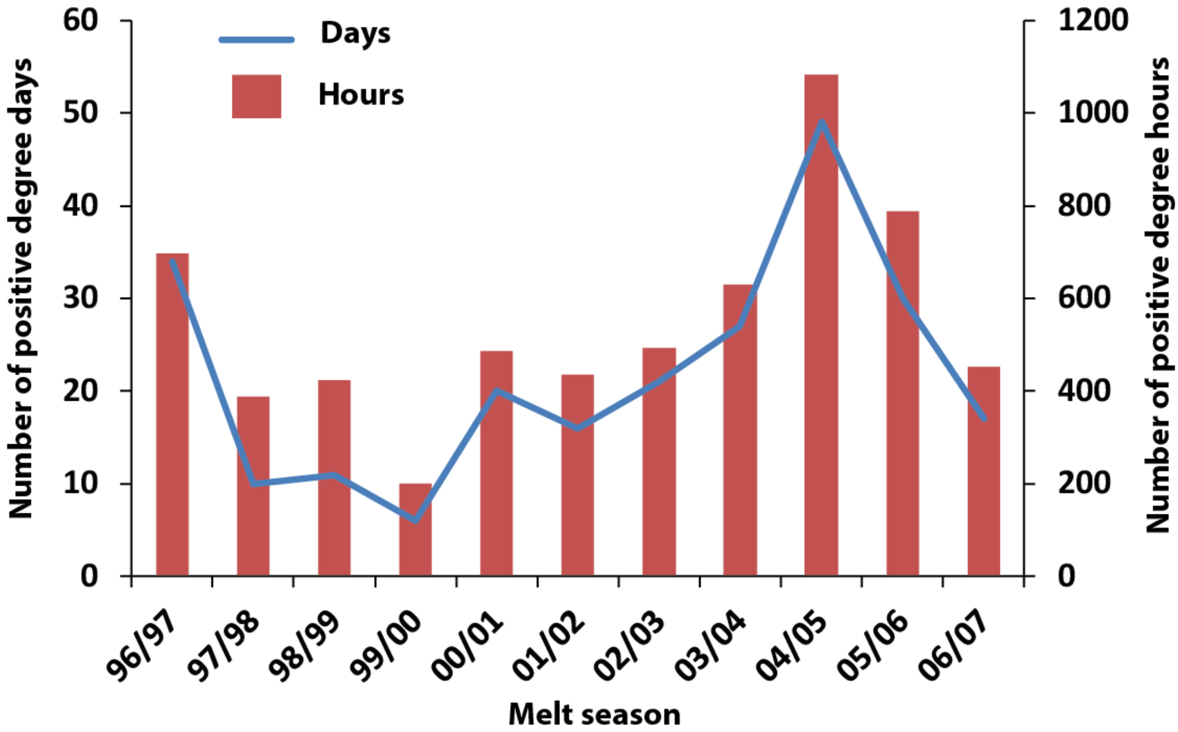


Figure 6.3: Positive degree days and hours trends from Casey AWS. Data obtained from the Australian Antarctic data centre (http://data.aad.gov.au/aadc/aws/)

In the Antarctic Peninsula, there is a precedent for comparatively intense melt seasons to drive rapid glacier and ice shelf change (Scambos *et al.*, 2000; 2003; Fahnestock *et al.*, 2002). Indeed, the disintegration of the Larsen B ice shelf in 2002 is largely believed to have been driven by an anomalously prolonged and intense summer melt season (van den Broeke, 2005). In the melt season (2001/2) which preceded the breakup of the Larsen B ice shelf, van den Broeke (2005) notes a 53% increase in melt days (96 days) and a 96% increase in the number of melt hours (802 hours), compared to the pre-2001 average (1995-2001). In comparison, Casey shows a 170% increase in melt days (49 days) and a 130% increase in melt hours (1083 hours) in the 2004/5 melt season, compared to the 1996-2004 average. This indicates that a similarly exceptional summer melt season may too have preceded the widespread retreat of outlet glaciers in DB13, suggesting that it may have been driven through increased surface melt and a similar mechanism to that of the recent ice shelf break up in the Antarctic Peninsula i.e. enhanced crevasse propagation (Scambos *et al.*, 2000). Indeed, recent modelling has highlighted the sensitivity of floating glacier tongues to deep penetrating crevasses (Bassis and Walker, 2011).

Despite the above considerations, the data from this study cannot demonstrate unequivocally that the retreat in DB13 initiated as a result of increased surface melting following the intense 2004/5 melt season. The exact timing of the retreat of glaciers in DB13 is not known. It is plausible that this could have happened at any point from 2000 to 2006. Indeed, it is also unknown if the temperature trends from Casey provide a good representation of the whole of DB13, as there are no other meteorological data available.

6.2 Glacier elevation change and frontal position change

Thinning has been observed in the last decade in some regions of East Antarctica, particularly Wilkes Land (DB12/13) and Victoria Land (DB14/15) (Pritchard *et al.*, 2009; Flament and Remy, 2012), consistent with the increased upwelling of warm CDW associated with a more positive SAM (Hall and Visbeck, 2002). Indeed, in DB13 this has been accompanied by mass loss (Chen *et al.*, 2009; King *et al.*, 2012). Elsewhere, similar rates of thinning have been coupled with retreat and a reduction in buttressing that is believed to drive flow acceleration, and subsequently alter the mass balance of ice sheets (Rignot and Kanagaratnam, 2006). However, no such accelerations have been reported in the study area (Rignot *et al.*, 2011a).

The region of most pronounced thinning (DB13) (Pritchard *et al.*, 2009) is the only drainage basin that shows a dominant pattern of retreat from 2000-2010. However, comparison between observed thinning between 2003 and 2007 (Pritchard *et al.*, 2009) and glacier frontal change 2000-2006, shows that whilst glaciers that are thickening exhibit very little terminus change in either direction; those that are thinning can be associated with both substantial advance and retreat, see Fig. 4.11. This can be explained by the fact that ~90% of glaciers measured in this study are unconstrained and lack any substantial buttressing. Therefore, glacier thinning in this section of East Antarctica is unlikely to result in a series of positive feedbacks resulting in glacier acceleration and retreat seen in Greenland (e.g. Moon and Joughin, 2008; Nick *et al.*, 2009). However, this may change if future warming removes the unconstrained ice tongues. In fact, the tendency for thinning glaciers to exhibit large changes in terminus position in both advance and retreat may be explained by the tendency for larger glaciers to thin and smaller glaciers to thicken, see Figure 6.4. Section 4.3 shows that larger glaciers exhibit a greater magnitude of frontal position change compared to smaller glaciers, potentially explaining the trend seen in Figure 4.11. Indeed, the cluster of thickening constrained glaciers seen in Fig. 4.11 may be explained by the tendency for constrained glaciers to be grounded and not have a floating tongue. Whereas, unconstrained glaciers tend to have a floating tongue and can be subjected to both basal melt and irregular calving, explaining the general trend of thinning and a large range in terminus behaviour.

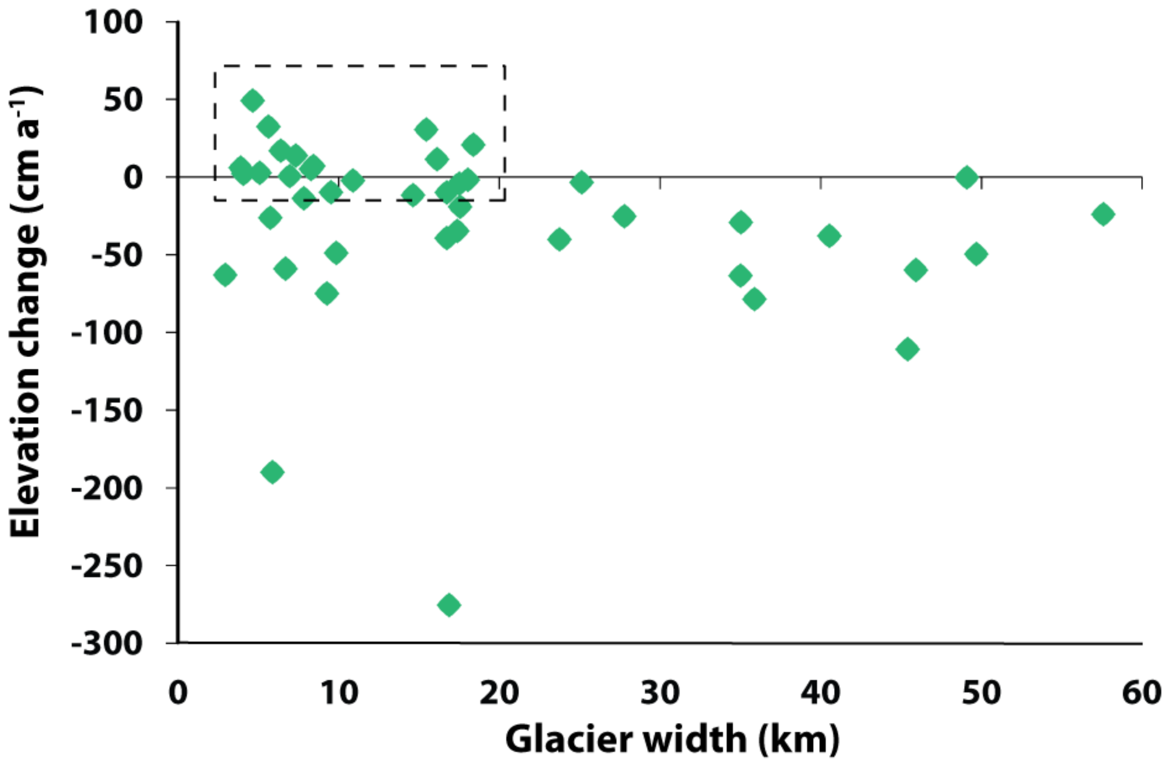


Figure 6.4: Elevation change plotted against glacier width, showing a tendency for narrower/slower glaciers to thicken (highlighted) and wider/faster glaciers to thin. Elevation data taken from Pritchard *et al.* (2009).

It has been demonstrated by Chen *et al.* (2009),that a rapid increase in the rate of mass loss in DB13 occurs after 2006, shown in Figure 6.5. This sudden increase in the rate of mass loss indicates an increase in ice discharge, driven by an increase in glacier velocity, as variations in surface melt have little direct influence on mass balance in East Antarctica (Torinesi *et al.*, 2003). The rapid increase in mass loss in DB13 appears to immediately follow the widespread retreat of glaciers in the region, particularly if these glaciers retreated following the 2004/5 melt season, as hypothesized in section 6.2. This suggests that despite the majority of glaciers in DB13 being unconstrained by lateral boundaries, the retreat of glaciers between 2000 and 2006 may have led to a reduction in buttressing and an increase in glacier velocity, subsequently increasing the rate of mass loss in the region. Indeed, it has been suggested that, in Greenland, sea ice mélange can generate lateral drag and act as a buttressing force slowing the glacier down (Lindsey and Dupont, 2012; Joughin *et al.*, 2012), effectively acting as an ice shelf (Dupont and Alley, 2005). In East Antarctica, both seasonal fast ice and multi year fast ice (MYFI) have been positively correlated to the size of the glacier tongue (Massom *et al.*, 2010). Fast ice can become fixed to glacier tongues (Massom *et al.*, 2010), particularly in the autumn and winter months (Massom *et al.*, 2009), most likely exerting some buttressing on the ice tongue. Therefore, the retreat of glaciers in DB13 will have reduced the contact area between the ice tongues and fast ice, decreasing the buttressing force, which, in turn may have increased glacier velocity and subsequently driven the observed mass loss.

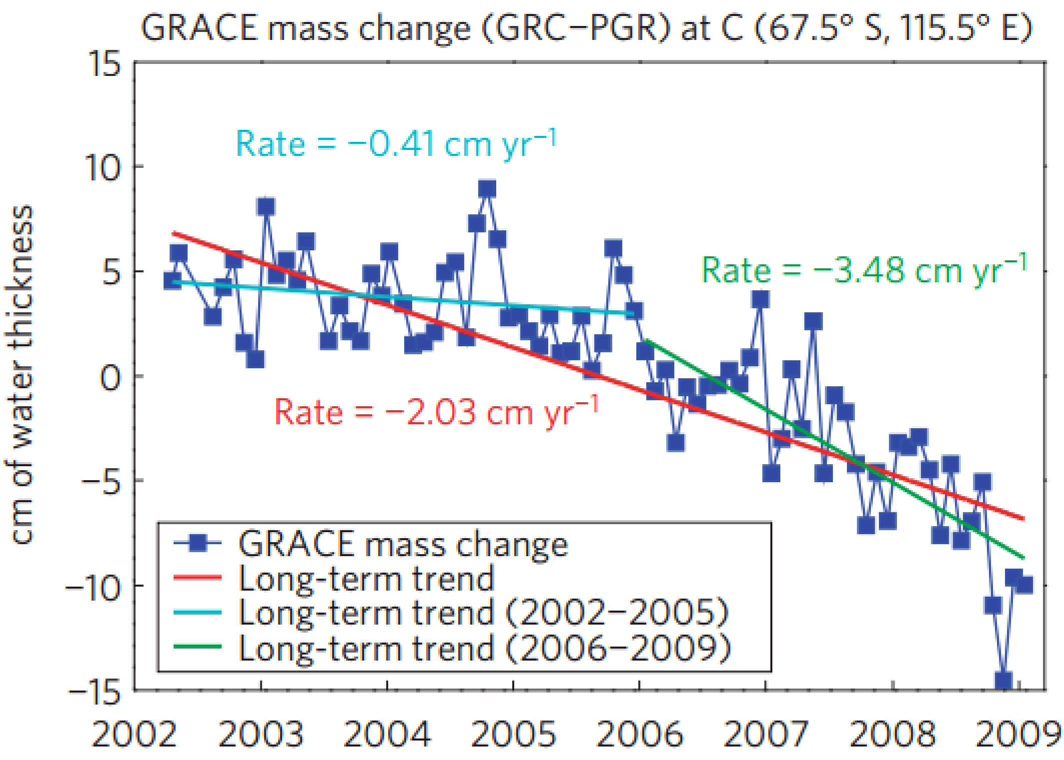


Figure 6.5: GRACE surface mass time series for Wilkes Land (Taken from Chen *et al.*, 2009).

Conclusions

In this study, the terminus position of 175 East Antarctic outlet glaciers has been mapped at a number of time steps between 1963 and 2010, along with the extraction of glacier velocity, glacier width and elevation change from available datasets in the 2000s. The analysis of these data in conjunction with meteorological and oceanic data allows the following key conclusions to be drawn:

1) **There was little overall change in glacier terminus position between 1974 and 2010.**

   The 132 glaciers with measurements between 1974 and 2010 indicate little overall change in glacier terminus position, with a median rate of 0.7 m a^{-1} and with a relatively even number of glaciers advancing (53%) and retreating (47%). However, a smaller population of glaciers measured between 1963 and 2010 (n = 38) indicated a long term overall pattern of retreat, with a median rate of -12.9 m a^{-1} and 68% of glaciers retreating.

2) **Strong decadal patterns in glacier behaviour exist within the study period, with the majority of glaciers retreating from 1974 to 1990 and the majority advancing from 1990 onwards.**

   Between 1974 and 1990, 63% of glaciers retreated at a median rate of -12.5 m a^{-1}.From 1990 to 2000, however, this trend was reversed, when 72% of glaciers advanced at a median rate of 19.7 m a^{-1}. During the most recent period, 2000 to 2010, the number of glaciers that advanced was reduced to 58% and the median fell to 8.4 m a^{-1}. These trends appear to be consistent across all drainage basins with the exception of DB16 (Ross Sea) where there is little discernible trend in glacier behaviour throughout the study period. Furthermore, between 2000 and 2010 in DB13, 74% of glaciers retreated (median rate: -63.6 m a^{-1}). This, compared to the overall decadal trend of 58% advance of glaciers (median rate: 8.4 m a^{-1}) during the same period.

3) **Wider and faster flowing glaciers undergo the largest changes in terminus position, but appear to have little influence on decadal trends in glacier frontal position change, hinting at a common external forcing.**

The observed relationship between glacier width and velocity confirms the theory that as glacier width increases, glacier velocity increases non-linearly. Indeed, it is demonstrated that the absolute magnitude of the range of glacier terminus position change over the study period can be linked to glacier width and consequently velocity, with the wider and faster flowing outlet glaciers in East Antarctica undergoing the largest changes in terminus position in both advance and retreat. However, despite wider glaciers exerting terminus position changes that are several orders of magnitude greater than smaller glaciers, it has been shown that these larger glaciers (width > 15 km) exert little influence on overall trends. Therefore, the broad decadal patterns across drainage basins and glaciers of all sizes are robust, hinting at a common external forcing(s).

4) **Decadal trends in glacier frontal position change can be linked with patterns in air temperature, suggesting a heightened sensitivity of East Antarctic outlet glaciers to perturbations in climate than previously thought.**

Air temperature records from four climate stations distributed throughout the study area reveal a consistent trend of above average temperatures during 1974 to 1990, coinciding with glacier retreat. From 1990 to 2000, however, this trend is reversed and a period of cooling is associated with glacier advance. Warming in the first half of the 2000s may explain the increase in the number of glaciers retreating during 2000 to 2010, compared to the period 1990 to 2000. Initially, this discovery was considered surprising due to the extent of the cold temperatures. However the presence of melt ponding is some sections of the study area confirmed the presence surface melt. Indeed, this is consistent with the coldest region in the study area, DB16, showing the weakest relationship to air temperature patterns.

5) **The rapid retreat of DB13 between 2000 and 2006 can be linked to an anomalously prolonged and intense melt season in 2004/05.**

An anomalously prolonged and intense melt season recorded at Casey in 2004/05, comparable in intensity to that of the ‘exceptional’ melt season which preceded the breakup of the Larsen B ice shelf (van den Broeke, 2005), may be linked to the widespread retreat of glaciers in DB13 measured between 2000 and 2006. This anomalous warming event is only recorded at Casey, thus explaining why only DB13 shows of pattern of retreat between 2000 and 2006.

6) The shift from the majority of glaciers retreating, to a majority of advancing in 1990 is consistent with the change in polarity of the Southern Annular Mode, which in turn, is linked to ozone depletion and an increase in greenhouse gases.

In recent decades the SAM has switched from a negative polarity to an increasingly positive polarity, hypothesized to be driven by ozone depletion and an increase in greenhouse gases (Thompson and Solomon, 2002). This shift is characterized by an increase in strength of the polar vortex, which reduces poleward heat transport, resulting in the cooling of summer temperatures and the suppression of katabatic outflow, resulting in an increase in coastal sea ice concentration. Therefore, a more positive SAM favours a larger proportion of glaciers to advance, explaining the general pattern of advance from 1990 onwards.

7) Glaciers which are thickening show little change in terminus position, whilst those glaciers which are thinning can be associated with both retreat and substantial advance.

In other regions similar rates of thinning have been associated with a reduction in buttressing, retreat and acceleration of flow e.g. Greenland (Rignot *et al.*, 2011;a). Here, a comparison between glacier elevation change and terminus position suggests that while those glaciers which are thickening show little changes in terminus position, those which are thinning shows signs of both advance and retreat, differing from other regions. This can be explained because the calving front of the vast majority of glaciers in this study is laterally unconstrained and therefore lacks any substantial buttressing.

This study provides the first large scale analysis of multi-decadal trends in East Antarctic outlet glacier frontal position. Taken together, results from this study are the first to suggest a rapid and synchronous response of East Antarctic outlet glaciers to external forgings, driven by large scale atmospheric variability. This effectively questions the long standing view that the EAIS may be immune to the dynamic changes that have been observed in both Greenland and West Antarctica. Indeed, the recent retreat and subsequent mass loss of DB13, suggests that this, in part, may already be happening.

The correlation between a positive SAM and glacier advance may provide a rare example of anthropogenic induced changes driving glacier advance, as oppose to retreat. This is because the recent trend for a positive SAM has been primarily attributed ozone depletion. Therefore, it may be that future glacier change in East Antarctica will be largely influenced by the balance between the predicted ozone recovery, and further increase in the concentration of greenhouse gasses, which are thought to play a secondary role in recent SAM trends. This highlights the importance in incorporating predicted SAM trends in modelling future changes in East Antarctic outlet glacier frontal position.

It is of crucial importance that future research establishes the relationship between glacier frontal position changes, elevation changes and glacier velocity in East Antarctica. Results from this study indicate a more complex relationship between these factors in comparison to Greenland, attributed to the laterally unconstrained nature of the majority of glaciers in the study area. In particular, future research needs to be directed towards increasing our understanding of the complexities of sea ice, which may be of heighted importance in determining the evolution of unconstrained glaciers. This combined, will allow an increase our understanding of the driving forces behind changes in glacier velocity, which is important in determining future changes in the EAIS mass balance and contribution to sea level change.

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