Chapter 4: Observations: Cryosphere

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Executive Summary

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.1 Introduction

[The section will discuss the changing state of the cryosphere and our changing understanding of it. The sectopm will try to set themes, key ideas, underpinning viewpoint, that will sit with the reader and echo as they read the rest of the text, and add to overall understanding. Areas of rapid progress since AR4, and areas where progress is disappointing will be highlighted. There are likely to be relatively few references here, except to the more significant pieces of work. Discussion will avoid too many new results /breakthroughs in the introduction, since this would likely be repetitious of the Executive Summary and main text.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Statement of the components of the cryosphere and their primary impact on the Earth system, for example:

- Ice sheets and glaciers are the major control on global sea-level, and much of interest in cryospheric change is because of this, but there are many regional and local impacts that should not be ignored.

- Sea ice loss impacts ocean circulation, ocean productivity and climate and has direct impacts on shipping and mineral and oil explorations.

- Relatively minor alteration in permafrost can destroy substantial arctic infrastructure.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Statements on known fluctuations in cryospheric components in distant and recent past, and potential for rapid loss, compared to slow recovery particularly when related to sea-level rise.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Statements on the visibility (wide publicity etc) and perceived significance of the cryospheric as an indicator of global/local climate change.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Contrasting statement on the true significance of the changing cryosphere:

- Glaciers, ice caps and ice sheets are integrators of climate change that are in many ways longer-lived, more sensitive, less open to interpretation than instrumental records.

- But they should not be regarded as “simple thermometers”, rather they are complex “climate-meters”.

- Our understanding of the complexity of response is increasing, and is particularly improved by efforts to understand the apparently contradictory responses of individual glaciers etc.]

Glaciers, ice caps and ice sheets
[PLACEHOLDER FOR FIRST ORDER DRAFT: Short textbook explanation of glaciers and ice sheets as systems in dynamic equilibria; ice loss roughly balancing accumulation, with the possibility that a persistent change input or output will cause an imbalance that could either reach a new equilibria, or disappear entirely. Point to West Antarctic ice sheet as a system where a new equilibria may not be possible — introducing idea of “instability”.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Point out the rapid improvement in the understanding of ice sheet change, from AR3/4 to present.]

Sea ice
[PLACEHOLDER FOR FIRST ORDER DRAFT: Introductory statement on sea ice:

- Net impact of a retreating Arctic sea ice cover in the summer? Ice albedo feedback is expected to cause the ice to retreat faster but an extended period of large ice-free ocean in the autumn would cause the loss of the extra heat acquired by ice-free areas in the summer.

- What does the persistently low summer ice cover in the Arctic during the last 4 years mean? The summer ice shows a slow recovery but is it slow because of ice albedo feedback or other factors? Can it be viewed as a genuine recovery?

- What can be said about the net positive trend of sea ice in the Southern Hemisphere? There are substantial regional differences, some positive, some negative. The impact of ozone loss has gained ground as a sensible explanation but is SAM changing as well?]

Permafrost and snow-cover
[PLACEHOLDER FOR FIRST ORDER DRAFT:
- Discuss the current state of the permafrost and snow cover and evaluate the net impact of a thawing permafrost and a declining snow cover.

- Elucidate the possibility and evidence for release of methane from the thawing permafrost [likely needs cross-reference to other chapters].

- Discuss net impact of snow variability on ice albedo feedback effects in both hemispheres.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Introduction ends with any statements required to understand the main text that follows:

- Begin with global statements, and drill down to regional statements with specific examples and discussion of processes appearing where required.

- Where appropriate use units such as Gigatonnes (Gt=10^{12} tonnes). One Gt is roughly equal to 1 km^3 of water (1.1 km^3 of ice); and 361 Gt of ice added to the oceans is roughly equal to 1 mm of global sea level rise.]

[INSERT FIGURE 4.1 HERE]
Figure 4.1: Loss of end of summer (perennial) sea ice cover in the Arctic. The area in white and gray in the Central Arctic represents the extent of the perennial ice cover in 2007 when compared to the average value (that includes the area in gold) from 1979 to 2006. [Note on graphic: We need one strong graphic for the introduction – Figure 4.1 may be revised to suit. It strongly illustrates that dramatic and influential changes have been occuring in the Arctic. Note: AR4 did not report on the dramatics decline in the perennial ice cover in 2007.]

Table 4.1: Cryospheric components, sensitivity to climate and potential impacts.

	How are they quantified?	How are they measured?	Climate variables to which they are sensitive	Global assessment of health	Potential Impacts?
Snow on land	Areas and annual duration				Too close to WGII?
Sea ice	Area, duration and thickness				
River and lake ice					
Glaciers and ice caps	Area and thickness	Satellite inventory and field observations	Temperature, precip., global radiation	In most regions glaciers are now declining	Water supply, increasing flooding hazards
Ice shelves	Generally by area, occasionally by thickness	Satellite measurements			
Ice sheets					

Greenland
Antarctica
Seasonally frozen ground
Permafrost

4.2 Ice Sheets and Ice Shelves

Key Findings

- The Greenland ice sheet is losing mass, approximately half from enhanced surface melt and half from increased flow of its glaciers to the ocean.

- The total area of surface melt in Greenland has increased significantly over the past 3 decades of satellite observations, and the amount of surface runoff has doubled since the 1990s.

- The mass loss of the Greenland ice sheet has been increasing over the past 20 years and largely exceeds mass gained from increased interior snowfall.

- The Antarctic ice sheet is losing mass almost entirely from enhanced flow of its glaciers. Snowfall accumulation varies significantly on yearly to decadal time scales but exhibits no long-term trend over the past several decades.

- The mass loss of the Antarctic ice sheet is increasing with time, but the signal is less distinct than in Greenland due to the larger temporal variability in Antarctic snowfall.

- Both ice sheets contribute significantly to the observed sea level, their contribution is increasing with time and will dominate sea level rise in the 21st century if the trends continue.

- The cause and mechanism of surface melt in Greenland are well understood and are a direct consequence of climate warming.

- The cause and mechanisms of glacier acceleration in both Greenland and Antarctica are not well understood. Surface melt is not a strong participant. Melting of ice by the ocean is significant but the physical processes of ice-ocean interaction involved in ice flow at the ice sheet periphery, where the ice starts to float, are not well characterized or modelled at present.

- Floating ice shelves are thinning in Greenland and the more northern part of Antarctica and their surface area is declining. In the Antarctic Peninsula irreversible collapse of ice shelves has been followed by rapid increases in velocity of their tributary glaciers.

- Transmission of the climate signal to ice sheets is operating on shorter time scales than anticipated and occurs via glacier speedup, ice-shelf collapse, warming of the entire ice column by surface water, and vigorous ice-ocean interactions.

4.2.1 Background

[PLACEHOLDER FOR FIRST ORDER DRAFT: Update on total ice volume retained in ice sheets, equivalent sea-level contribution, annual turn over of mass and principal components (snowfall precipitation, surface melt, grounding line ice discharge).]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Explain what ice sheets constitute and what they do not. Ice sheets do not include surrounding glaciers and ice caps that are discussed in the mountain glacier chapter, but some techniques, e.g., time-variable gravity may include the signal from surrounding ice caps and glaciers in their estimates.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Explain how ice volumes evolve with time as a result of ice deformation and sliding, ice melting, formation of floating ice shelves, and iceberg calving, snow evaporation or re-deposition.]

AR5: We now have 3 decades of satellite/airborne observations of ice. Major advances have been made in our characterization of ice-sheet change since AR4. Ice sheet loss estimates and trends are converging, and major advances have been made in attributing the causes of glacier changes, but we remain far from a detailed understanding of the rapid evolution of ice sheets that might be sufficient to predict the future.

4.2.2 Mass Balance

Define the mass balance of an ice sheet as the difference between the mass that comes in as snowfall and the mass comes out as icebergs, melt water, blown snow, evaporation/sublimation and vapour deposition. In determining the mass balance, we are interested in the delicate balance between these two large numbers in the midst of large inter-annual variations.

Surface mass balance refers to the net accumulation of mass at the ice sheet surface, i.e., snowfall accumulation minus surface melt, surface evaporation, sublimation, vapour deposition and wind transport. This reports discusses surface mass balance estimates averaged over large areas (drainage basins of individual glaciers, or entire ice sheet), not measurements at single points.

Total mass balance refers to the difference between surface mass balance and any ice discharge at the grounding line, i.e., where ice detaches from the bed and becomes afloat in the ocean or in land-based lakes. Total mass balance is negative if losses (ablation) are greater than gains (accumulation). Total mass balance in this report is average over large areas or the entire ice sheet, not at single points. Total mass balance estimated discussed in this report also include floating ice shelves, but this is treated separately since floating ice shelves have negligible impact on sea level.

4.2.2.1 Techniques

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.2.2.1.1 Mass budget method

Historically the first approach to mass balance, calculates the difference between surface mass balance (input) and perimeter fluxes (output). Major achievements since AR4 include complete mapping of perimeter fluxes of both ice sheets (Rignot et al., 2008; [Rignot et al., 2011a]), new ice thickness data from radio echo sounding to reduce uncertainties [also Icebridge papers, and expected paper from Griggs and Bamber], ice velocity mapping of entire ice sheets in Greenland (Joughin et al., 2010b) and Antarctica [Rignot et al., 2011b], digital maps of surface mass balance for both ice sheets based on regional atmospheric climate models that are evaluated using in situ data instead of relying on in situ data to be derived (Ettema et al., 2009; Lenaerts et al., in press; van den Broeke et al., 2006b). [PLACEHOLDER FOR FIRST ORDER DRAFT: Summarize inputs and associated uncertainties (7% in Greenland, 5–30% in Antarctica) Summarize outputs and associated uncertainties (2–5% in Greenland and Antarctica on average). Comprehensive assessments and revisions not available for AR4, data gaps, regions of rapid change, residual uncertainties (snowfall, ice thickness). Note large annual to inter-annual variability in snowfall, especially Antarctica (Figure 4.2 and 4.3). Current mass loss is a small fraction of the annual turn over of mass in Antarctica (12%), but a large fraction in Greenland (70%).]

4.2.2.1.2 Repeated altimetry

Started in the 1980s but of definite accuracy only since the 1990s with better instruments and orbit determinations. ERS time series start in 1992 [Wingham et al, 2010]. Icesat time series in 2002–2010 (Pritchard et al., 2009; Zwally et al., 2011), Cryosat time series in 2011 [Reference needed: Wingham, xxxx], airborne laser altimetry since 1993 in Greenland and 2002 in Antarctica (Thomas et al., 2009) [possibly more Krabill papers]. Measures volume changes instead of mass changes. Issues with reflecting horizon of radars vs melt layers, reflecting horizon of laser vs power, correction for firn compaction, spatial sampling along coastal sectors of outlet glaciers, data gaps, and limited duration of missions. Radar altimetry confirms the increasing mass loss of Pine Island Bay (Wingham et al., 2009). Fluctuations are detected in coastal East Antarctica (Shepherd and Wingham, 2007); interior regions are dominated by the short-term variability in snowfall (Helsen et al., 2008), so that it is difficult to conclude on the sign of mass balance in interior East Antarctica. ICESAT time series suggest growth of interior Antarctica (Zwally et al., 2011), but there are issues with laser power correction [Schuman et al., 2010]. Results confirm the pattern of dynamic thinning along the coast, i.e., thinning concentrated along channels occupied by major glaciers (Pritchard et al., 2009; Thomas, 2004; [numerous other references before 2009]). Several estimates of mass balance have been produced for Greenland; fewer for Antarctica.

4.2.2.1.3 Temporal variations in Earth gravity

GRACE data since early 2002. Mission duration is two times longer than for AR4; major progress has been made using these data [suite of references on most recent studies, see below]. Correction for GIA remains a source of uncertainty for the absolute loss in Antarctica, but the correction does not matter for detecting changes in mass loss over time since the GIA is a constant signal. Papers reviewing the reasons behind spread of published values of mass loss using GRACE data in the literature ([Velicogna and Wahr, 2011]; Schrama and Wouters, 2011): results are affected by differences in processing schemes between data centers, calibration or lack of techniques, time period of observation, truncation of spherical harmonics to different orders, corrections or not for L1, inclusion of glaciers and ice caps surrounding the ice sheets, etc. New results are in better agreement, except for Wu et al. (2010) who combine GPS and GRACE data but this approach is limited by the existing GPS network.

Synergy: These three techniques help address complementary aspect of ice sheet change, including total mass loss, partitioning of mass losses between surface and ice dynamics , detection of areas of rapid changes and spatial details, and interior losses vs. losses along coastal regions.

4.2.2.2 Greenland

The Mass Budget Method (MBM) (Figure 4.2a-b) reveals that the partitioning of mass loss in Greenland is about 50% surface mass balance (SMB), 50% ice discharge (van den Broeke et al., 2009) (Figure 4.2d). The trend in SMB is dominated by a doubling in the amount of surface melt since the 1990s whereas snowfall accumulation appears to be steady. Short-term measurements of surface height suggest slight inland thickening (Figure 4.2f), but the impact on total mass balance is small. For ice discharge, major glacier speed up in central east, southeast and central west portions of Greenland in 2000–2005; some glaciers returned their speeds after 2005 in the southeast, but flow still faster than in 1996 in southeast and 2000 for central west and east Greenland (Howat et al., 2007); most recent changes are in the northwest, another high accumulation, high discharge sector of Greenland (Figure 4.2d-e).

[INSERT FIGURE 4.2 HERE]

Figure 4.2: a) cumulative mass loss in Greenland from the MB method for 1992–2010; b) time series of annually-resolved losses from MBM (black) and GRACE (red) 1992–2010; c) temporal pattern of mass loss from GRACE time-variable gravity; d) mass losses per sector detailing the partitioning between surface and dynamic losses combining RACMO2/GRE and MBM; e) velocity map from satellite interferometry 2009; f) ice thinning rates from ICESAT data 2003–2008.

GRACE results are in better agreement as a result of longer time series of GRACE data, progress in new data releases, and maturation of post processing techniques (Schrama and Wouters, 2011; Velicogna, 2009; [Luthcke et al., 2010; Rignot et al., 2011a]). The mass loss has decreased in the east and southeast but increased in central west and especially in the northwest ([Kahn et al., 2010]; Schrama and Wouters, 2011) (Figure 4.2 b-c).

Surface melt area climatology from passive microwave satellite shows increasing total melt area and change in equilibrium line altitude; this trend has continued since AR4 and accelerated in the past few years [Tedesco et al., 2011].

Altimetry missions report slightly lower losses (Zwally et al., 2011; [Csatho et al., in prep]) but sampling is sparse along the coast where loss is concentrated along narrow channels (Figure 4.2f), and the conversion from volume to mass is complex (Pritchard et al., 2009). Radar altimetry may be affected by a rise in reflecting horizon (Thomas et al., 2008), ICESAT altimetry may be biased by a dimming of the laser power during the mission [Schuman et al., 2011] and a short mission duration. There is, however, a strong agreement on the sign of the mass balance and on the increase in mass loss with time.

Table 4.2: Mass Balance of the Greenland Ice Sheet [Draft].

Method (reference)	Mass balance (Gt/yr)	Acceleration (Gt/yr2)	Time period
MBM + GRACE [Rignot et al., 2011]		–21.9±1	1992–2010
	–50±51		1992–2000
	–170±51		2000–2003
	–245±51		2003–2007
	–310±51		2007–2010
GRACE (Velicogna, 2009)		–30±11	2002–2009
	–137±33		2002–2003
	–286±33		2007–2009
GRACE (Schrama and Wouters, 2011)	–201±18	–8±6	2003–2010
GRACE (Cazenave et al., 2009)	–136±18		2003–2008
GRACE [Luthcke et al. 2011]	–187±73		2004–2009
ICESAT (Zwally et al., 2011)	–7±3		1992–2002
	–171±4		2003–2007
ICESAT (Sørensen et al., 2010)	–210±21		2003–2008
GRACE+GPS (Wu et al., 2010)	–104±23		2002–2008
MBM (Rignot et al., 2008)	–110±70		1960s
	–30±50		1970s–1980s
MBM (van den Broeke et al., 2009)	–166		2000–2008
	–238		2003–2008

Notes:
GIC removed.

4.2.2.3 Antarctica

AR4 estimates of mass balance of Antarctica were few. The MB method now includes 85% of the coastline (Rignot et al., 2008), and new radar echo sounding data (Rignot et al., 2011). A most outstanding advance has been the production of digital maps of snow accumulation that do not rely on in-situ data for derivation but only for evaluation (Lenaerts et al., in press; van den Broeke et al., 2006b) (Figure 4.3d). The new maps indicate similar levels of total accumulation but more snowfall along wet coastal sectors. Measurements of time-variable ice velocity indicate that a large fraction of the loss is due to the increase in ice discharge of several major glaciers (Figure 4.3e). The overall increase in mass loss is less than in Greenland but the absolute magnitude of the loss is comparable to that of Greenland (Figure 4.3a-b). Total snowfall exhibits large interannual variations but no long-term trend in the past few decades (Figure 4.3b, d).

GRACE results are more numerous than in AR4 and more mature, because of time series that is twice as long (Figure 4.3b-c). Discrepancies between the mass budget method and GRACE have been resolved, and indicate an accelerating loss of 13–14 Gt yr^{-2}. The examination of 18 years of data improves the reliability of the trends compared to that inferred from shorter missions like GRACE.

Laser altimetry indicates that most of the changes in volume are concentrated in narrow corridors occupied by outlet glaciers and ice streams, as illustrated by the striking correspondence between areas of thinning in ICESAT (Figure 4.3f) and areas of fast flow with InSAR (Figure 4.3e). It remains difficult to convert these volume changes into mass changes (Pritchard et al., 2009).

One area of uncertainty is the interior of East Antarctica. Earlier satellite radar altimetry results indicative of interior growth (Davis et al., 2005) were dominated by short-term variability in snowfall (Helsen et al., 2008). More recent updates with ICESAT altimetry for 2003–2009 suggest interior growth as well [Zwally et al., in prep]. ICESAT data may be affected by the short mission duration and other uncertainties [Schuman et al., 2011]. A recent comparison of global data reanalyses in Antarctica concludes that the ERA-Interim data is the most reliable reconstruction of Antarctic snowfall [Bromwich et al., in press]. Furthermore, the ERA-Interim reanalysis shows no long-term increase in Antarctic snowfall (Lenaerts et al., in press; Monaghan et al., 2006; van den Broeke et al., 2006a; [Bromwich et al., 2011]). GRACE data indicates no mass gain in East Antarctica (dependent on GIA), and no change in growth (independent of GIA). The discrepancy between GRACE/MBM and ICESAT is not resolved.

[INSERT FIGURE 4.3 HERE]

Figure 4.3: a) Cumulative mass loss from the MB method; b) annual mass loss from MB (black) and GRACE method (red); c) temporal evolution of mass loss from GRACE time-variable gravity; d) surface mass balance from RACMO2/ANT; e) ice sheet velocity from satellite radar interferometry 2007–2009; f) ice thinning rates from ICESAT 2003–2009.

Table 4.3: Mass Balance of the Antarctic Ice Sheet.

Method (reference)	Mass balance Gt yr^{-1}	Acceleration Gt yr^{-2}	Time period
MBM + GRACE (Rignot et al., 2011)		–14.5 ± 2	1992–2010
	–65 ± 91		1992–2000
	–165 ± 91		2000–2003
	–201 ± 91		2003–2007
	–258 ± 91		2007–2010
GRACE (Velicogna, 2009)*		–26 ± 14	2002–2009
	–104 ± 73		2002–2003
	–246 ± 73		2007–2009
GRACE (Chen et al., 2009)*	–190 ± 77		2002–2009
GRACE (Cazenave et al., 2009)*	–198 ± 22		2003–2008
GRACE (Wu et al., 2010)*	–83 ± 43		2002–2008
RALT (Wingham et al., 2006a)	+27 ± 29		1992–2003

Notes:
* Does not exclude glaciers and ice caps.

Table 4.4 lists freshwater fluxes for Greenland [Bamber et al., 2011] and Antarctica [Church et al., 2011]. This includes solid ice discharge at the grounding line and snow/ice melt along the periphery, mostly in Greenland.

Table 4.4: Net freshwater fluxes (solid ice discharge, ice sheet runoff and snowfield runoff).

Region	Freshwater flux (km^3 yr^{-1})	Time period
Antarctica [Church et al., 2011]	2390	1975–1992
	2550	2006–2010
Greenland [Bamber et al., 2011]	900	1958–1990
	1200	2003–2010

4.2.2.4 Floating Ice Shelves

Satellite radar altimetry suggest that most Antarctic ice shelves are thinning or near steady-state, except in Queen Maud Land and Amery Ice Shelf where ice shelves seem to be thickening [Shepherd et al., 2010]. Ground-based surface elevation measurements on the Amery extend the satellite altimetry record back to 1968 and show a near-zero thickness change over 4 decades (King et al., 2009). The large ice shelves in Ross and Filchner-Ronne appear to be slowly thickening but the changes are close to the level of detection, so the change is uncertain at present. ICESAT work from Pritchard et al. (2009) does not include floating ice shelves.

[INSERT FIGURE 4.4 HERE]
Figure 4.4: Average rate of Antarctic ice shelf thickness change, 1994–2008, determined from ERS and ENVISAT radar altimetry and a model of accumulation fluctuations [Helsen et al., 2008].

Ice shelves in the Peninsula and the Amundsen Sea are all thinning rapidly, which is consistent with the loss of buttressing of its glaciers and their widespread acceleration.

There is marked decline in ice shelf area in the Peninsula (Cook and Vaughan, 2010) initiated many decades ago which is ongoing, consistent with regional warming.

Larsen A and B have not recovered from the collapse and would take centuries to return. Glaciers upstream of these ice shelves are still flowing well above equilibrium in 2010. Larsen C does not appear to be on the verge of collapse [several new references on that topic coming up]. On the west coast, a large sector of Wilkins ice shelf broke up, as part of an ongoing retreat initiated decades ago [reference needed]. Update of calving of Mertz ice tongue and Thwaites ice tongue experienced major breakup however not directly related to climate change [reference needed].

Ice shelves in the north of Greenland broke off: Zachariae Isstrom, Ostenfeld Ice Shelf (Joughin et al., 2008) and ice shelves in Ellsmere Islands [Reference needed]. The recent break up of Petermann floating ice tongue is part of a long history of advance and calving events but could also be indicative of a warming trend [Johannessen et al., 2011; Falkner et al., 2011].

There is *in situ* evidence for increased ice shelf melting in Pine Island Glacier [Jenkins and Jacobs, 2010]. Elsewhere we have little information about changes in ice shelf melt rates.

Evidence for increased surface melting of ice shelves (Nghiem et al., 2007; Tedesco and Monaghan, 2009) showed a 30-year minimum in snow melt extent in 2009.

56% of mass ablation of Antarctic floating ice shelves is from basal melting vs 44% from iceberg calving [Rignot et al., 2011c]. Ice shelf melting exhibits a linear (Rignot and Jacobs, 2002; [Bindschadler et al., 2011]) to quadratic dependence on thermal forcing from the ocean (Holland et al., 2008a; Macayeal, 1984).

4.2.3 Causes of Changes in Ice Sheets

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.2.3.1 Changes in Snowfall and Surface Melt

Ice sheets experience a large inter-annual variability in snowfall, local trends may deviate significantly from the long-term trend in integrated snowfall, but — as in AR4 — we have little to no evidence for long term change in accumulation (Ettema et al., 2009; [van den Broeke et al., 2010]; Monaghan et al., 2006; [Bromwich et al., 2011]). There is however some divergence on that topic based on the results of laser altimetry (Zwally et al., 2011).

Surface melt is a major control on surface mass balance in Greenland [van den Broeke et al., 2009] but plays no role in Antarctica (Lenaerts et al., in press). In Greenland, *in situ* studies and regional atmospheric models note a large increase in runoff (Ettema et al., 2009), melt duration and melt extent [Tedesco et al., 2011], with 2010 being one of the strongest melt years on record in Greenland.

Studies of surface temperature conclude on widespread warming of Antarctica since 1957 (Barrett et al., 2009; Comiso, 2010; Steig et al., 2009).

4.2.3.2 Changes in Ocean Temperature Around Ice Sheets

Ocean temperature is on the rise in the southern ocean (Boning et al., 2008), confirming earlier studies (Gille, 2002), at a rate that exceeds the average rate of ocean warming, but the immediate vicinity of the Antarctic coastline is devoid of data, so it is not clear how much of this warming is able to reach glaciers.

In Greenland, a large fraction of the ongoing glacier changes has been attributed to the advection of warmer ocean waters along the coastline (Holland et al., 2008a; Howat et al., 2008; Rignot et al., 2010) [also articles in preparation], but significant uncertainties remain in the detailed understanding of the interaction between a warmer ocean and faster glacier flow.

[PLACEHOLDER FOR FIRST ORDER DRAFT: to b expanded; additional papers to be included, when available]

4.2.3.3 Partitioning of Mass Loss

50%/50% in Greenland from iceberg calving [Rignot, 2011]. In Antarctica, nearly 80% grounding-line ice discharge versus 20% SMB [Rignot, 2011].

4.2.3.4 Historical context

A subglacial ridge may have affected the retreat of Pine Island Glacier, WAIS (Jenkins et al., 2010) but does not explain the simultaneous thinning of all glaciers in this sector. Drainage of subglacial lakes in Antarctica induces short-lived speed-up that does not have much influence on ice sheet mass balance ([Hamilton et al., 2009]; Fricker et al., 2007; Gray et al., 2008; Wingham et al., 2006b).

Wake et al. (2009) show SMB for Greenland over the longer term. The Greenland ice sheet has experienced a similar forcing to that which it is currently experiencing in the 1920s and 1930s, but the effect was not as dramatic on JAKOBSHAVNS ISBRAE, so that other factors may be important.

Similar results are discussed by Mernild et al. (2010).

There is a growing number of studies on pre-satellite record ([Johannessen et al., 2011]; [Csatho et al., in prep]; Cook and Vaughan, 2010).

4.2.3.5 Dynamic Response to Recent Forcings

Physical processes capable of triggering glacier acceleration include changes in ice-ocean interactions, basal lubrication, ice-shelf buttressing, calving.

4.2.3.5.1 Basal lubrication

Basal lubrication occurs as witnessed by diurnal flow variations on land-terminating regions of the Greenland Ice Sheet (Das et al., 2008; Shepherd et al., 2009). Lake drainage — we know this happens to the base of the Greenland Ice Sheet, i.e., Das et al. (2008). In both cases speed-up is considerable, about 50–110% (Das et al., 2008; Joughin et al., 2008; Shepherd et al., 2009), but this is spatially and temporally (about 24 hours, Das et al., 2008) restricted. Speed-up over the annual cycle is about 10–15% (Joughin et al., 2008; Shepherd et al., 2009). Joughin et al. (2008) claim that melt-enhanced speed-up will have a "substantive but not catastrophic effect" on mass loss from Greenland *versus* [van de Wal et al.; 2008] no long-term impact on ice sheet motion over 17 years. Overall, these studies do not indicate that basal lubrication is an important factor in the recent speed up, nor will be important in the near term evolution of Greenland glaciers.

4.2.3.5.2 Ice shelf buttressing

Changes in ice-shelf buttressing are confirmed to have a major impact: Larsen B glaciers speed up by a factor 3 to 8 following the collapse of Larsen B, and the glaciers are still flowing 3 times faster than required to maintain equilibrium 8 years after the collapse, confirming that the collapse is irreversible, and the impact on the glaciers is long lived. Jakobshavns Isbrae doubled its flow speed as a result of the demise of its floating ice shelf.

4.2.3.5.3 Ice-ocean interaction

Ice-ocean interaction appears to be one of the most important forcing on ice sheets, but also the least well studied, characterized and understood at present. Ice-ocean interactions control the evolution of its floating ice shelves (melting at the bottom is 1-2 orders of magnitude larger than at the top) (Jacobs, 1992). Recent research suggests that the ocean is equally important for Greenland tidewater glaciers. The observed melt rates are large compared to ice shelves (Motyka et al., 2003; Rignot et al., 2010) and large compared to the surface mass balance. Model simulations suggest a linear to quadratic dependence on thermal forcing (Holland et al., 2008a)[Yun et al., 2011 in prep]. The intrusion of warm (tropical) waters in south-Greenland fjords is documented (Holland et al., 2008a; Straneo et al., 2010)[Also, Dowdeswell et al., 20xx, and papers in preparation], and is thought to have enhanced subaqueous melt and have an impact on grounding line dynamics (Payne et al., 2004; Schoof, 2007; Thomas, 2004)[Also, other papers]. This is the largest source of uncertainty for predicting the future of the ice sheets [Bindschadler , 2006].

In Greenland, there is evidence that the 1997 acceleration of Jakobshavn was related to ocean warming at depth (Holland et al., 2008a). Helheim, Kangerdlugssuaq and other SE Greenland glaciers accelerated (by up to 100%), thinned and retreated and then decelerated, thinning slowed or stopped and re-advanced synchronously (Howat et al., 2008; Murray et al., 2010), suggesting a regional control. These glaciers show possible relationship with sea surface temperature offshore (Howat et al., 2008), although there is an unexplained time-delay, so this explanation is not unequivocal. Waters have been shown to circulate rapidly to the fjord (Straneo et al., 2010) and models suggest that the response is consistent with change driven from front margin (Nick et al., 2009).

4.2.3.5.4 Calving

A calving law needs to be incorporated into predictive ice sheet models. Two empirical models have been used – but both have limitations, however, one, the flotation model, has been used to reproduce successfully the behavior of Heheim Glacier in SE Greenland (Nick et al., 2009). The Benn et al. (Benn et al., 2007)criterion (that calving occurs when crevasses reach to the water line) has begun to be included in models (Otero et al., 2010), but still needs both observational and model validation. Observations using stereo cameras have become relatively common, but are not yet directed at testing or deriving calving models e.g.,, (Ann and Box, 2010). Amundson et al. [Not identified, 2011] suggest that the melting of sikkusak has had a profound influence on calving dynamics in Greenland.

Percolation of melt water through moulins, crevasses and cracks has a major, rapid impact on the thermal regime of coastal ice, called "cryo-hydrologic warming" (Phillips et al., 2010). This phenomena implies a rapid transmission of surface warmth to the ice column, which also impacts ice rheology.

4.3 Glaciers and Ice Caps

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.3.1 Introduction

[PLACEHOLDER FOR FIRST ORDER DRAFT: The introduction will address the following issues:

The need to know the extent and the change of Glaciers and Ice Caps (GICs) as
(i) components of the regional hydrology,
(ii) contributors to sea level rise, and
(iii) indicators of climate conditions and changes.

(i) is addressed in WGII in various chapters, (ii) is addressed in WGI Chapter 13, and (iii) in WGI Chapter 10. In WGI Chapter 4 we assess the observed extents and changes of GIC in regional and global scales as a basis for the others and we address selected particular GIC behaviours. We aim to mention different approaches of estimating glacier changes which include field-based, remote-sensing, and – to certain extent – also reconstruction and modelling methods.

In each case it is the GIC mass and its change which is of concern. Yet, respective measurements and observations are very rare. Volumes and volume changes get us closest to the mass, area and length information is a much more indirect information but easier to obtain and thus better available. The latter allow us also for mass change reconstructions into the past and we aim for assessing mass changes back to the LIA for regions with available published material.

Directly measured changes area assessed but allow for model approaches that lead to spatial and temporal extrapolations as long as they are based on observed data.

[PLACEHOLDER FOR FIRST ORDER DRAFT: Products being considered include:
a) Regional and total areas and volumes/masses (as extrapolated from incomplete inventories (areas) and such as estimated from available area-volume relation algorithms);
b) Regionally grouped and globally integrated glacier length and GIC area changes;
c) Regionally grouped and globally integrated mass changes;
d) Regional peculiarities

[PLACEHOLDER FOR FIRST ORDER DRAFT: **Units:** if they address SLR totals and changes are in [Gt], [Gt/time], [m SLE], [m SLE/time]. Consideration to be given to expression in [kg /m^2] when we address hydrological or climate sensitivity relevant information (this is based on experiences from AR4). We propose to use double unit axes such as in Figs 4.14 and 4.15 in AR4 when appropriate.

[PLACEHOLDER FOR FIRST ORDER DRAFT: **Discussion on measurement and extrapolation methods**: methods from point measurements (which are not accessible to the community), full glacier mass balances, spatial and temporal extrapolations, to scaling methods including respective uncertainties shall be briefly described.

[PLACEHOLDER FOR FIRST ORDER DRAFT: **Glacier-climate relations:** A brief discussion on response times, mass turnover versus mass storage.

[PLACEHOLDER FOR FIRST ORDER DRAFT: The Introductory text to be cross checked with similar text portions prepared for Chapter 13.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: **AR4 state of the art on the change of glaciers and ice caps:**

From AR4 WGI Chapter 4: Executive Summary:

"Mass loss of glaciers and ice caps is estimated to be 0.50 ± 0.18 mm yr^{-1} in sea level equivalent (SLE) between 1961 and 2004, and 0.77 ± 0.22 mm yr^{-1} SLE between 1991 and 2004. The late 20th-century glacier wastage likely has been a response to post-1970 warming. Strongest mass losses per unit area have been observed in Patagonia, Alaska and northwest USA and southwest Canada. Because of the corresponding large areas, the biggest contributions to sea level rise came from Alaska, the Arctic and the Asian high mountains."

[PLACEHOLDER FOR FIRST ORDER DRAFT: This AR4 text will be reformulated and possibly extended in Section 4.3.1. A sentence about how glaciers and ice caps in GL and AA were (not) treated in AR4 to be included.]

4.3.2 Area/Volume Inventory

[PLACEHOLDER FOR FIRST ORDER DRAFT: From mountain ranges to global numbers (Table and/or Figure). If possible, we will maintain both tables in the Report; if not, the regional table to be included as Supplementary Material.

Table 4.5a: [PLACEHOLDER FOR FIRST ORDER DRAFT] Extents of Glaciers and Ice Caps [To be updated with new material coming up within the IPCC deadlines, from AR4 Table 4.3. (Lemke et al., 2007), also, others].

Reference	Area (10^3 km^2)	Volume (10^3 km^3)	SLE[f] (m)

Raper and Braithwaite, 2005[a,c]	522 ± 42	87 ± 10	0.24 ± 0.03
Ohmura, 2004[a,d]	512	51	0.15
Dyurgerov and Meier, 2005[a,e]	546 ± 30	133 ± 20	0.37 ± 0.06
Dyurgerov and Meier, 2005[b,e]	785 ± 100	260 ± 65	0.72 ± 0.2
IPCC, 2001[b]	680	180 ± 40	0.50 ± 0.1

Notes:
(a) Glaciers and ice caps surrounding Greenland and Antarctic Ice Sheets are excluded.
(b) Glaciers and ice caps surrounding Greenland and West Antarctic Ice Sheets are included.
(c) Volume derived from hypsometry and volume/area scaling within 1° × 1° grid cells.
(d) Volume derived from a statistical relationship between glacier volume and area, calibrated with 61 glacier volumes derived from radio-echo-sounding measurements.
(e) Volume derived from a statistical relationship between glacier volume and area, calibrated with 144 glacier volumes derived from radio-echo-sounding measurements.
(f) Calculated for the ocean surface area of 36 × 106 km2.

Table 4.5b: Total areas and volumes of mountain glaciers and ice caps. Excluding and including those in Antarctica and Greenland[a] (Radic and Hock, 2011).

Source	Area[b]		Volume[c]		Sea Level Equivalent (m)	
	Excluding	Including	Excluding	Including	Excluding	Including
Meier and Bahr (1996)	540	680	-	180 ± 4[d]	-	0.5 ± 0.1
Ohmura (2004)	521	-	51	-	~0.15	-
Dyurgerov and Meier (2005)	540 ± 30	785 ± 100	133 ± 20	260 ± 65	0.33 ± 0.05	0.65 ± 0.16
Raper and Braithwaite (2005)	522	-	87 ± 10d	-	0.241 ± 0.026	-
This study	518 ± 2	741 ± 68	166 ± 10	241 ± 29	0.41 ± 0.03	0.60 ± 0.07

Notes:
(a) Sea level equivalent is calculated assuming oceanic area of 3.62 × 108 km2 and a glacier density of 900 kg m^{-3}.
(b) Area values are × 103 km^2.
(c) Volume values are × 103 km^3.
(d) Volumes are given in water equivalent according to original reference, and not in ice equivalent as those of the other studies.

Table 4.6: Regional area and volume of glaciers [numbers as available from (Radic and Hock, 2011)].

Region	Glacier Area (km^2)		Glacier Volume (km^3)		Glacier Volume (mm SLE)	
Svalbard	36506	± 364	10260	± 823	26,00	± 2,00
Scandinavia	3057	± 18	224	± 11	0.56	± 0.03
Central Europe	3045	± 17	194	± 12	0.48	± 0.03
Franz Josef Land	13739	± 141	2248	± 176	5,60	± 0.40
Novaya Zemlya	23645	± 1132	9410	± 3388	23,00	± 8,00
Severnaya Zemlya	19397	± 566	6046	± 1231	15,00	± 3,00
Caucasus	1397	± 10	88	± 6	0.22	± 0.01
North and East Asia	2902	± 14	170	± 8	0.42	± 0.02
High Mountain Asia	114330	± 729	12483	± 462	31,00	± 1,00
Alaska	79260	± 1076	27436	± 3312	68,00	± 8,00

W. Canada and W. U.S.	21480	± 420	1892	± 361	4,70	± 0.90
Arctic Canada	146609	± 1068	80160	± 12151	199,00	± 30,00
Iceland	11005	± 821	4889	± 2244	12,00	± 6,00
South America I	7060	± 137	344	± 37	0.86	± 0.09
South America II	29640	± 663	8116	± 712	20,00	± 2,00
New Zealand	1156	± 13	83	± 11	0.21	± 0.03
Greenland	54400	± 4400	17865	± 2993	44,00	± 7,00
Sub-Antarctic islands	3740	± 129	363	± 44	0.9	± 0.10
Antarctica	169000	± 68000	59158	± 25829	147,00	± 64,00
TOTAL	741448	± 68186	241430	± 29229	600,00	± 73,00

inventory complete

inventory incomplete, numbers upscaled

Notes:
There are numbers such as in which are obtained in a consistent methodology, and there are other regional subclasses and methods available for erratic regions. Work is underway to provide uniform figures that lead to global numbers.

[PLACEHOLDER FOR FIRST ORDER DRAFT: **Outlook for AR5:** the GLIMS Inventory of glacier areas will likely cover appr. 95% within 2011. First, more coarse datasets (1 km grid) are expected to be ready within the next months. A major issue for the overall summary table will be to clearly define the boundary between the Greenland Ice Sheet and the local GIC, and where the GIC cover on the Antarctic Peninsula starts. Volume estimate methods different from the area/volume scaling (Bahr et al., 1997) are presently investigated by colleagues. All new results will be included here according to the availability within the IPCC deadline. The LAs are confident that an improvement from AR4 on the total GIC area and volume numbers will be obtained by then.

A Figure such as the placeholder Figure 4.5 may be produced instead of the tables: globe with e.g., regional circ plots including both area and volume with broken lines where we have little/no data + one summary plot.

[INSERT FIGURE 4.5 (OPTION A) HERE]
Figure 4.5 (a): [PLACEHOLDER FOR FIRST ORDER DRAFT: regions and numbers in this preliminary figure are from Radić and Hock (2010). Glacier outlines can possibly be added.

[Another option which is presently under examination is a "balloon" grouping of glacier regions.]

[INSERT FIGURE 4.5 (OPTION B) HERE]
Figure 4.5 (b): [PLACEHOLDER FOR FIRST ORDER DRAFT: [from G. Cogley, in progress]

4.3.3 Regional to Global Changes

[PLACEHOLDER FOR FIRST ORDER DRAFT: Global pictures/numbers as they accumulate from mountain ranges/latitudes/glacier types; incl. interpretation of forcing/processes [Note: Chapter will not be able to provide regional scales such as e.g., necessary for water supply questions. This will have to be done by WGII on examples as appropriate there.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: length variations]

[INSERT FIGURE 4.6 HERE]
Figure 4.6: [PLACEHOLDER FOR FIRST ORDER DRAFT: an update from this (Lemke et al., 2007) is expected.

[PLACEHOLDER FOR FIRST ORDER DRAFT: area changes
A compilation of a series of data sets from different sources and different time periods is underway and will be shown in a figure similar to the placeholder Figure 4.5.]

[PLACEHOLDER FOR FIRST ORDER DRAFT: volume/mass changes]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Sea level contribution (delivery to Chapter 13)

A small working group has started working on this in order to obtain information as indicated in the preliminary Figure 4.6. This figure shows mass change evolutions based on directly measured mass changes and geodetically obtained volume changes following Cogley (Cogley, 2009) subdivided into Radić and Hock regions (Radic and Hock, 2010). The plot shown here is one produced by G. Cogley for the Himalaya [Cogley, in prep]. Also results from (Radic and Hock, 2011) will be taken into account. We are confident to produce such plots for each R-H region.]

[INSERT FIGURE 4.7 (OPTION A) HERE]
Figure 4.7 (a): [PLACEHOLDER FOR FIRST ORDER DRAFT] regional evolution of mass changes in R-H regions. Example given here is the Himalaya one. Directly measured mass changes and geodetically obtained volume changes are merged. An extension back to LIA is planned for a selected number of regions. A symbol code will show how well data based are certain regions compared to others: e.g., good data coverage: solid/bold line, poor data coverage: broken/transparent line (or even no line e.g., for missing reconstructions of past mass changes) [Together with Chapter 13 the preparation of a world map including future GIC development scenarios is planned]

[INSERT FIGURE 4.7 (OPTION B) HERE]
Figure 4.7 (b): **[**PLACEHOLDER FOR FIRST ORDER DRAFT] regional evolution of mass changes of glaciers and ice caps in sectors (Zemp et al., 2007) [Also the potential for a simple gridded map like this may be a choice]

A discussion on different mass balance terms, the value and the uncertainties of different measuring/ reconstruction/ modelling/ extrapolation methods (e.g., Fischer, 2011) as well as on the total uncertainties (e.g., Cogley, 2004) will become part of this section.

[PLACEHOLDER FOR FIRST ORDER DRAFT] **Global – SL contribution**

[INSERT FIGURE 4.8 HERE]
Figure 4.8: the latest update (10 October, including some early measurement reports for 2008/2009) from Cogley (2009) of the global time series (glaciological only in red, glaciological plus geodetic balances in blue; simple arithmetic averages of measurements on the left, spatially interpolated estimates on the right) [from . G. Cogley, in progress]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Further updates from (Cogley, 2009) expected from Graham Cogley]

[PLACEHOLDER FOR FIRST ORDER DRAFT: In summary, an update from the following Figure as well as a table will be provided:

[INSERT FIGURE 4.9 HERE]
Figure 4.9: Compilation of SLE estimates (several methods) as available by June 2009 (Allison et al., in press). D: direct glaciological method, SMB. Surface mass balance, G: geodetic method. Compilation: G. Kaser, to be updated. Uncertainties to be partially added. (Hock et al., 2009, Lemke et al., 2007, Kaser et al., 2006, Oerlemans et al., 2007, Cogley, 2009, Meier et al., 2007). [Layout to be coordinated with the Ice Sheet chapter colleagues]

[PLACEHOLDER FOR FIRST ORDER DRAFT: It is planned to produce a table such as Table 4.4. in AR4 as a delivery to Chapter 13. In accordance with Ch 13, time periods will become 1970-2010 and 1993-2010.]

Table 4.7: [PLACEHOLDER FOR FIRST ORDER DRAFT] From AR4 (Table 4.4)] Global average mass balance of glaciers and ice caps for different periods, showing mean specific mass balance (kg m^{-2} yr^{-1}); total mass balance (Gt yr^{-1}); and SLE (mm yr^{-1}) derived from total mass balance and an ocean surface area of 362 × 106 km^2. Values for glaciers and ice caps excluding those around the ice sheets (total area 546 × 103 km^2) are derived from MB values in Figure 4.14. Values for glaciers and ice caps including those surrounding Greenland and West Antarctica (total area 785.0 × 103 km^2) are modified from Dyurgerov and Meier (2005) by applying pentadal DM05 to MB ratios. Uncertainties are for the 90% confidence level. Sources: Ohmura (2004), Cogley (2005) and Dyurgerov and Meier (2005), all updated to 2003/2004.

Period	Mean Specific Mass Balance[a] (kg m^{-2} yr^{-1})	Total Mass Balance[a] (Gt yr^{-1})	Sea Level Equivalent[a] (mm yr^{-1})	Mean Specific Mass Balance[b] (kg m^{-2} yr^{-1})	Total Mass Balance[b] (Gt yr^{-1})	Sea LevelEquivalent[b] (mm yr^{-1})
1960/1961–2003/2004	–283 ± 102	–155 ± 55	0.43 ± 0.15	–231 ± 82	–182 ± 64	0.50 ± 0.18
1960/1961–1989/1990	–219 ± 92	–120 ± 50	0.33 ± 0.14	–173 ± 73	–136 ± 57	0.37 ± 0.16
1990/1991–2003/2004	–420 ± 121	–230 ± 66	0.63 ± 0.18	–356 ± 101	–280 ± 79	0.77 ± 0.22

Notes:
(a) Excluding glaciers and ice caps around ice sheets
(b) Including glaciers and ice caps around ice sheets

4.3.4 Special Regional Features

[PLACEHOLDER FOR FIRST ORDER DRAFT: Karakoram (little changes, advances) (maybe highly complex behaviour of Himalaya glaciers). New results are expected from elevation and length change analysis; NZ and Norway. New results are expected from new mass balance analyses (NZ) and from new inventories (N).; Polythermal glaciers (Storglaciären and others). New results are under review from Storglaciären.; Very cold and dry glaciers (Kibo, Bolivia/Chile, Tibet, Dry Valleys); and Tropical glaciers (in AR4 this was an explicit question)]

[PLACEHOLDER FOR FIRST ORDER DRAFT: These examples had all been included in AR4 and will only be taken here if new developments and/or insights are made available.]

[Some material such as e.g., individual length variations, photo series may go to Supplementary Material]

4.4 Rapid Changes in Ice Sheets and Glacier Dynamics

This section reviews current knowledge on marine ice sheet instability and other processes capable of causing rapid changes in ice sheets.

While we have not seen signs of marine ice sheet collapse, we have evidence of the impact of ice shelf collapse on glacier flow and ice sheet mass balance from measurements in Greenland and Antarctica. These demonstrate that ice shelf buttressing is a major control on ice flow. There are also decades of observations that indicate that 1) ice sheet changes are more rapid than anticipated by AR4 models; and 2) some but not all marine-based sectors of Greenland and Antarctica are changing rapidly, ahead of anticipations by AR4 models. These sectors include J in Greenland and PIG in West Antarctica.

Knowledge gained since AR4 emphasizes the theoretical importance of processes of rapid changes (e.g., ocean warming, ice-shelf collapse), revealed new ones (e.g., cryo-hydrologic warming) but also de-emphasizes more traditional ones (basal lubrication from melt water).

4.4.1 Marine Ice Sheet Instability

[PLACEHOLDER FOR FIRST ORDER DRAFT: Define what it means, where it applies in Greenland (Sole et al., 2008) and Antarctica (Bamber et al., 2009).]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Evolution of marine-based sectors]

4.4.1.1 Greenland

The main channel of Jakobshavn Isbrae is marine-based and grounded 2 km below sea level (Figure 4.10). Following the loss of its floating ice tongue, the glacier doubled its speed and is continuing its inland retreat. Scan the literature for projections of J speed in the coming decades: Some numbers are quite large [Thomas, 2011] and suggest that J alone could be a significant source of mass loss in coming decades. Jakobshavn Isbrae drains 7% of the entire ice sheet.

[INSERT FIGURE 4.10 HERE]
Figure 4.10: Bed trough of Jakobshavn Isbrae, West Greenland mapped with radio echo sounding (Plummer et al., 2010).

Most glaciers that changed abruptly in northwest and east Greenland are grounded below sea level only over small regions. Other submarine sectors (Petermann, Humboldt, 79north, Zachariae) are showing slight signs of change but are not significant contributors to total ice sheet loss at present. Although Zachariae Isstrom lost half of its ice shelf, the glacier has only sped up 30% since 2004.

4.4.1.2 West Antarctica

Pine Island Glacier had been accelerating exponentially (Rignot, 2008) since 1975, but stopped accelerating in 2009 (Joughin et al., 2010a): its grounding line retreated 20 km since 1992. PIG is the closest analogue of marine instability in Antarctica. Opinions diverge on its future evolution, ranging from no acceleration in the coming two decades (Joughin et al., 2010a) to a tripling of glacier speed [Thomas, 2011]. Observations from the last twenty years however confirm a major evolution of this entire sector and a significant contribution to sea level. Large, rapid, future changes therefore cannot be excluded.

4.4.1.3 East Antarctica

Cook ice shelf, Totten Glacier, Denman Glacier, all marine-based, are thinning (Chen et al., 2009; Pritchard et al., 2009; Shepherd and Wingham, 2007) but are small contributors to mass loss at present. We do not understand the processes that drive their dynamic thinning. The glaciers are not accelerating [ref in prep]. There is little to no published information on ocean conditions. This sector is not included in Pfeffer et al. (Pfeffer et al., 2008) estimates of maximum contribution from Antarctica, but holds more ice than West Antarctica and therefore constitutes a large uncertainty for predictions. GPS-based studies suggest Totten glacier has been thinning for several decades [Allison et al., 2004].

4.4.2 Processes Capable of Triggering Rapid Changes

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.4.2.1 Ice Shelf Collapse

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.4.2.2 Ice-Ocean Interactions

An increase in subaqueous ice shelf melting is a most efficient way to collapse an ice sheet (Warner and Budd, 1998) (Warner and Budd, 1998; Huybrecht and de Wolde (Huybrechts and de Wolde, 1999)[Also, others], but prior models did not know how to constrain the melt rates as a function of thermal forcing of the ocean, and how to constrain future thermal forcing. Recent research suggests that the dependence of ice melt is of linear to quadratic nature, i.e., a fast response to ocean warming (Holland et al., 2008b)[Also others], but changes in ocean temperature around ice sheets are not well known (Boning et al., 2008). The impact of ice melting on grounding line dynamics is poorly known. This is a major source of uncertainty for predicting ice sheet evolution.

In Greenland, ice-ocean interactions affect tidewater glaciers, i.e., the vast majority of its glaciers. We also know the limits to this process: as glaciers reach higher ground and lose contact with the ocean, subaqueous melting will cease to be a driving process. In Greenland there are relatively few parts of the ice sheet that are sub-sea-level (Sole et al., 2008).

4.4.2.3 Basal Lubrication

Ice sheets are affected by basal processes and melt water in very much the same way as mountain glaciers [Niemow et al., in press]. Many recent studies have shown that the effect of enhanced melt water while significant is typically short lived (fraction of the summer), and its overall effect on ice sheet mass flow has been negligible so far. These recent studies call into question the possibility for this process of ice sheet de-stabilization to be able to trigger rapid glacier and ice sheet changes.

4.4.2.4 Cryo-Hydrologic Warming

Melt water penetrates ice to warm the ice and hence soften it [Thompsen et al. 1998] (Phillips et al., 2010). Below some elevation in west Greenland, the ice switches from cold, polar to nearly temperate (–2 to –3°C). Ice viscosity of coastal Greenland is lower than assumed in numerical models. A warmer climate at the surface has also a more direct, immediate impact on ice deformation than assumed previously. This process of rapid changes is new to AR5.

4.5 Sea Ice

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.5.1 Background

[PLACEHOLDER FOR FIRST ORDER DRAFT: Role sea ice cover in climate: Ice albedo feedback; Interface between ocean and atmosphere; Surface energy fluxes; Freshwater redistribution (Polyakov et al., 2008); Impact on meridional overturning circulation; Ecosystem]

[PLACEHOLDER FOR FIRST ORDER DRAFT: Important climate parameters considered here: Ice type and extent; Multiyear ice coverage; Ice thickness/age; Ice motion/export/deformation; Polynyas and Oddens; Dates of melt/freeze, duration of cover, length of melt season (Markus et al., 2009; Massom and Stammerjohn, 2010); Snow cover on sea ice (Maksym and Markus, 2008)and link to precipitation variability

4.5.2 Sea Ice Concentration and Extent

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.5.2.1 Historical Context

[PLACEHOLDER FOR FIRST ORDER DRAFT: Pre-satellite period (available data, inferred behavior of ice cover); a question of interest is when was the last time that the Arctic or the Antarctic was ice free in the summer.

[PLACEHOLDER FOR FIRST ORDER DRAFT: Satellite Era (data available for different ice parameters, length of dataset, The November 1978-February 2011 passive microwave data, error analysis)

4.5.2.2 Northern Hemisphere Record from Passive Microwave

[PLACEHOLDER FOR FIRST ORDER DRAFT: Seasonal, interannual, decadal trends; record setting events (year 2007);

The trends vary with season but this is in part because the perennial ice (and hence the summer ice) is declining rapidly. After 2007, the annual ice has recovered but the trend in the perennial ice continued to go down because the recovery was relatively minor. Arctic surface temperatures over sea ice covered and ice free regions have also shown significant positive trends . The trend of the pan-Arctic ice cover for the period 1978 to 2010 is about -4% per decade with the trend in winter much less than that in summer.

4.5.2.3 Southern Hemisphere Record from Passive Microwave

Seasonal, interannual, decadal trends. There are significant regional trends (increase in Ross Sea, decrease in Bellingshausen) and seasonal variability in the Antarctic (Comiso and Nishio, 2008; Massom and Stammerjohn, 2010; Parkinson and Cavalieri, 2008; Stammerjohn et al., 2008)Consider also links between sea ice extent variability and large scale circulation: ENSO, SAM, etc (Pezza et al., 2008; Kwok and Comiso, 2002; Yuan, 2004).

Record setting events. The record highest extent of the sea ice cover in the Antarctic occurred in the winter of 2008 during the satellite era. This was just a year after the record low perennial ice cover in the Arctic.

The environmental setting of the sea ice cover in the Antarctic is quite different from that of the Arctic and satellite data show that the trends in the sea ice cover go in opposite directions for the two hemispheres. Modeling studies have indicated that changes in the Antarctic sea ice extent will not be as large as in the Arctic. (Goosse et al., 2009; Holland and Raphael, 2006; Parkinson, 2004). The trend in the sea ice extent in the Antarctic is postive and about 2 % per decade. The reason for the positive trend is in part due to ozone hole which has caused a deepening of the lows in West Antarctica that in turn caused stronger winds and enhanced ice production in the Ross Sea (Goosse et al., 2009; Turner and Overland, 2009; Turner et al., 2009).

Fraser et al. [in press] use MODIS composite images for 2000-2008 to map landfast sea ice distribution and variability. In the Indian Ocean sector there is a strong positive trend (over this short period) of +4% per year.

[INSERT FIGURE 4.11 HERE]
Figure 4.11: Monthly ice extent anomalies in the (a) Northern Hemisphere and the (b) Southern Hemisphere. The anomalies were estimated by subtracting climatological monthly averages (as derived from 1978–2010 satellite SMMR and SSM/I data) from each monthly extent.

4.5.3 Multiyear/Seasonal Ice Coverage

Satellite-derived estimates of sea-ice age and thickness are combined to produce a proxy ice thickness record for the Arctic Ocean for 1982 to the present. These data show that in addition to the well-documented loss of perennial ice cover as a whole, the amount of oldest and thickest ice within the remaining multiyear ice pack has declined significantly.

4.5.3.1 Background and Data Sources

[PLACEHOLDER FOR FIRST ORDER DRAFT: Basin scale estimates from satellite passive and active microwave estimates; also from ice age estimates (Maslanik et al., 2007).

4.5.3.2 Variability and Trend

The thick component of the perennial ice, called multiyear ice, is declining at an even faster rate than the perennial ice (17.2%/decade vs 13.5%/decade) which is a clear indication that the average thickness of the perennial ice cover is also declining [Comiso, 2010] [Comiso, submitted 2011].

These data show that in addition to the well-documented loss of perennial ice cover as a whole, the amount of oldest and thickest ice within the remaining multiyear ice pack has declined significantly. The oldest ice types have essentially disappeared, and 58% of the multiyear ice now consists of relatively young 2- and 3-year-old ice compared to 35% in the mid-1980s (Maslanik et al., 2007).

[INSERT FIGURE 4.12 HERE]
Figure 4.12: Annual changes in (a) ice extent and (b) area of the perennial (blue) and multiyear (green line) ice cover as derived from passive microwave (SMMR and SSM/I) data.

4.5.4 Ice Thickness and Volume

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.5.4.1 Background and Data Sources

Until recently there have been no satellite remote sensing techniques capable of directly mapping sea ice thickness, and this parameter has primarily been determined by drilling, by under-ice sonar measurement of draft (the submerged portion of sea ice) from fixed moorings or submarines (Rothrock et al., 2008) , or by electromagnetic sounding of the ice cover (Haas, 2004).

An emerging new technique, using satellite radar or laser altimetry to estimate ice freeboard from the measured ranges to the ice and sea surface in open leads (and assuming an average floe density and snow depth), have provided a short record and offers promise for future monitoring of large-scale sea ice thickness (Kwok, 2004; Laxon et al., 2003).

4.5.4.2 Recent Changes in Sea Ice Thickness and Volume from Satellites

[PLACEHOLDER FOR FIRST ORDER DRAFT: State recent changes in ice thickness from radar and satellite altimetery (ERS, Envisat, and ICESat record) – (Giles et al., 2008; Kwok and Rothrock, 2009; Kwok et al., 2009); volume changes from altimetry (Kwok et al., 2009); what are the implications of these changes?

[INSERT FIGURE 4.13 HERE]
Figure 4.13: Decline in sea ice thickness (2004–2008) of the Arctic Ocean from ICESat. [more detail to be provided]

4.5.4.3 Combined Satellite and Submarine Record

[PLACEHOLDER FOR FIRST ORDER DRAFT: Multidecadal trend of thickness decline – changes not confined to the last decade (Kwok and Rothrock, 2009)

[INSERT FIGURE 4.14 HERE]
Figure 4.14: Decline in Arctic ice thickness from combined submarine and ICESat records [a more detailed description of the Figure to be provided].

4.5.4.4 Evidence of Change from Other Sources

All sources show strong decline over the last decade in parallel with the changes in ice coverage. Is ice volume recovering like ice extent? (no indication as yet due to lack of observations). However, in the Antarctic, there are no such records for estimation of changes in ice thickness and volume.

[PLACEHOLDER FOR FIRST ORDER DRAFT: Include also here model simulations (driven by reanalyses) of ice volume. For the Arctic AR4 used (Koberle and Gerdes, 2003; Rothrock et al., 2003) – there must be updates. For the Antarctic - (Goosse et al., 2009).

4.5.4.5 Changes in the Snow Cover over Sea Ice

[PLACEHOLDER FOR FIRST ORDER DRAFT: address long term changes in snow cover; references needed.]

4.5.5 Ice Motion

Pack ice motion influences ice mass locally, through deformation and creation of open water areas; regionally, through advection of ice from one area to another; and globally through export of ice from polar seas to lower latitudes where it melts. The drift of sea ice is primarily forced by the winds and ocean currents. On time scales of days to weeks, winds are responsible for most of the variance in sea ice motion. On longer time scales, the patterns of ice motion follow surface currents and the evolving patterns of wind forcing. Here we consider whether there are trends in the pattern of ice motion.

4.5.5.1 Arctic Sea Ice Export

Sea ice area is exported primarily through the Fram Strait; long-term volume flux is not available due to paucity of long time series of ice thickness estimates in the region. Recent estimates by (Spreen et al., 2009) – using ICESat - show no significant change in ice volume flux compared to(Kwok, 2009) estimates from the 1990s.

In recent years, there are estimates from other passages out of the Fram Strait; the export of thick ice at Nares Strait seems to be significant (Kwok et al., 2010; Samelson et al., 2006). Discuss in relation of fast-ice arch location in Kennedy Channels, and Lincoln Sea. What is the relative role of melt and ice export in the depletion of multiyear ice of the Arctic Ocean [e.g., Kwok et al., 2011] During the last decade, there is significant melt in the Beaufort Sea.

4.5.5.2 Arctic Drift Speed

Mean drift speed has increased over the last 29 years (+17% per decade for winter and +8.5% for summer). A strong seasonal dependence of the mean speed is also revealed, with a maximum in October and a minimum in April. (Rampal et al., 2009). They found at that this are unlikely to be consequences of a stronger atmospheric forcing suggestive that sea ice kinematics play a fundamental role in the albedo feedback loop and sea ice decline: increasing deformation means stronger fracturing, hence more lead opening, and changes in heat balance at the surface.

[INSERT FIGURE 4.15 HERE]
Figure 4.15: Export of sea ice area at the Fram Strait [volume estimates to be added].

4.5.5.3 Arctic Sea Ice Circulation Pattern

[PLACEHOLDER FOR FIRST ORDER DRAFT: What are the implications of shifts in pattern on sea ice mass balance? (Kwok, 2009).

4.5.5.4 Antarctic Sea Ice Circulation

[PLACEHOLDER FOR FIRST ORDER DRAFT: Significance of drift to determining Antarctic ice extent. Pattern of Antarctic drift [Comiso et al., in press]

4.5.6 Dates of Melt Onset, Freeze-up, and Melt Duration

Extended or more extensive sea ice melt in response to increasing atmospheric temperatures may be one of the primary drivers of reduced summer sea ice. The total amount of solar energy absorbed during the summer melt season was strongly related to the timing of when melt begins. Earlier melt onset allows for earlier development of open water areas that in turn enhance the ice-albedo feedback. Several approaches exist to determine melt and freeze onset of Arctic sea ice from satellite passive microwave data [e.g.,, Smith, 1998b; Drobot and Anderson, 2001; Belchansky et al., 2004].

4.5.6.1 Duration of Melt Season

[PLACEHOLDER FOR FIRST ORDER DRAFT: Markus et al. (2009); also, other studies indicate an increasing melt duration]

4.5.6.2 Trends and Variability

(Markus et al., 2009) analyzed trends in melt onset and freezeup for 10 different Arctic regions. In all regions except for the Sea of Okhotsk, which shows a very slight and statistically insignificant positive trend (0.4 d/ decade), trends in melt onset are negative, i.e., toward earlier melt. The trends range from 1.0 d/decade for the Bering Sea to 7.3 d/decade for the East Greenland Sea. Except for the Sea of Okhotsk all areas also show a trend toward later autumn freeze onset. The Chukchi/Beaufort seas and Laptev/East Siberian seas observe the strongest trends with 7 d/decade. For the entire Arctic, the melt season length has increased by about 20 days over the last 30 years. Largest trends of over 10 d/decade are seen for Hudson Bay, the East Greenland Sea, the Laptev/East Siberian seas, and the Chukchi/Beaufort seas. Those trends are statistically significant at the 99% level.

4.5.7 Polynyas and Oddens

Observed increase in the extent of coastal polynyas in the Ross Sea caused enhanced ice production that is primarily responsible for the positive trend in ice extent in the Antarctic [Comiso et. al., in press]. The frequency of occurrences of the Odden, a large sea ice feature that forms in the east Greenland Sea and which may protrude well eastward of the main sea ice pack, has declined substantially during the last two decades [Comiso, 2010]. This has the potential of decreasing deep ocean convection frequency in the Greenland Sea. Polynya in Nares Strait due to fast-ice arch in Kennedy Channel.

4.6 Seasonal Snow and Ice Cover: Variability and Trends

Key points:
Judiciously combining surface observations and satellite-derived snow cover extent (SCE) indicate declines in SCE in most months over the 1922–2010 period of record; declines (8%) are largest in spring and are strongly correlated with temperature. [tbc]

Other studies based on station observations indicate a range of trends, with trends likeliest to be negative at locations near freezing; later in the snow season (spring) than earlier; and for measures (e.g., date of last snow) sensitive to spring melting rather than to winter accumulation (e.g., maximum snow depth). Trends are likeliest to be positive at very cold locations (high mountains or high latitudes) where an increase in temperature is correlated with an increase in snowfall. [tbc]

Broadly, these changes in snow are consistent with a warming world.

[PLACEHOLDER FOR FIRST ORDER DRAFT: springtime snow-albedo. Feedback [added to first paragraph]; and increases in precipitation in very cold locations; information on lake and river ice to be added.

4.6.1 Introduction

Snow measurements include snow cover extent (SCE, typically measured by satellite), snow-covered area (SCA), the sum of daily snowfall, number of days with snow above a threshold depth, snow depth (SD), snow water equivalent (SWE), and other quantities. (Lemke et al., 2007) included a lengthier discussion of snow variability and trends using satellite and in situ data, than had previously appeared in IPCC report or appears here. A review paper (Brown and Mote, 2009) further synthesized a range of observational and modeling results, emphasizing patterns in where, when, and in which type of measurement of snow a response to changing climate could be expected. Their analysis helped explain why observational studies of changes in snow find increases in some locations and seasons, and decreases in others: decreases in snow are most likely to be observed in spring and at locations near the freezing point, where changes in temperature are most effective at either reducing snow accumulation or increasing snowmelt. Increases in snow are likeliest at very cold (winter temperatures below –15 or –20°C) locations, where precipitation generally is expected to increase in a warming world because the atmosphere holds more moisture. Disentangling the competing effects of rising temperatures and changing precipitation remains an important challenge in understanding and interpreting changes in snow.

Snow accumulates, very roughly, at either latitudes or altitudes where temperatures are sufficiently often below roughly 0°C - that is, poleward or at higher altitudes of that isotherm. Since in most mountainous areas, precipitation is enhanced, the snow accumulation can be substantially greater and persists longer into the spring. Satellite measurements have difficulty in mountainous terrain owing to extensive shielding by forests, strong variations in snow depth on very short spatial distances relative to the satellite footprint, and other factors. In the southern hemisphere, very little land area lies poleward of the 0°C isotherm, and most snow (as well as some glaciers) occurs in the mountain ranges - the Andes of South America, the southern Alps (and Mt Ruahepu on the north island) of New Zealand, and mountain ranges in southeast Australia.

Data sources include both satellite and *in situ* measurements. The weekly observations of SCA from National Oceanic and Atmospheric Administration (NOAA) visible data, dating to 1966 (Robinson et al., 1993), cover the northern hemisphere; for the southern hemisphere, mapping of SCE began only in 1978. Space-borne passive microwave sensors offer the potential for global monitoring since 1978 of not just snow cover, but also snow depth and SWE. Use of these sensors though requires resolving differences between Scanning Multichannel Microwave Radiometer (SMMR, up to 1987) and Special Sensor Microwave/Imager (1987 and beyond); for example, (Jezek et al., 1993) adjusted SMMR brightness temperatures based on the short period of overlap in 1987. The Gravity Recovery and Climate Experiment (GRACE) appears to be able to estimate SWE over Arctic basins that are snow-free in summer (to establish baseline; (Niu et al., 2007)), but GRACE was launched only in 2002.

Snow cover extent - for which the flatter areas of the continents are far more important than the much smaller area occupied by mountain ranges - can be by measured by satellite, by station observations of snow depth, or by blending the two. Whereas weather stations in snowy areas often report snow depth, standard weather stations with long complete records of snow depth are relatively rare in most mountainous areas and in the southern hemisphere. For predicting summer water supply, manual, and later automated, measurements of SWE have been made for decades at over a thousand sites in western North America, and at a much smaller number of sites in northern Europe. Very few locations in the SH have long records of snow data - six in the central Andes and four in southeast Australia. As noted in (Lemke et al., 2007), the only study for New Zealand explicitly looking at snow was published in 1995 and has not been updated, but airborne photography of end-of-summer snowline on 50 index glaciers since 1977 provides some additional knowledge.

4.6.2 Satellite Results for Snow Cover Extent

By blending *in situ* and satellite records, (Brown and Robinson, 2011) have updated the time series of northern hemispheric SCE, and as before reductions were largest in spring. Their analysis of the updated spring SCE series [Figure 4.16, will be revised for March-April] shows significant reductions over the past 90 years, with a higher rate of decrease during the past 40 years. Averaged March and April NH SCE declined around 8% (7 million km^2) over the 1970–2010 period relative to pre-1970 values. Declines in March are larger in Eurasia than in North America, but both continents exhibit significant reductions in April SCE.

That the trends in springtime SCE are linked to rising temperature is apparent in Figure 4.17, which shows correlation between spring temperature and SCE. The strength of the spring snow cover-albedo feedback contributes substantially to the hemispheric response to rising greenhouse gases and provides a useful test of GCMs (Fernandes et al., 2009)[Also, see also Chapter 9]. Declines in land snow cover and sea ice have contributed roughly equally to reductions in the cryospheric contribution to the surface energy balance, and the albedo feedback of the NH cryosphere is likely in the range 0.3–1.1W m^{-2} K^{-1}, substantially larger than estimates from 18 CMIP3 models (Flanner et al., 2011).

Satellite records for the southern hemisphere are still being developed.

4.6.3 In Situ Trends

A meta-analysis of variability and trends in measures of snow from surface observations (Brown and Mote, 2009) attempted to provide a taxonomic approach to the numerous, mostly country-based studies then available. Considering studies from Argentina, Australia, Austria, Bulgaria, Canada, Chile, China, Finland, Japan, Russia, Scotland, Slovakia, Switzerland, and the USA, they emphasized that declines were generally observed in measures of spring snow (e.g., spring snow water equivalent or the date of last snow), and that where temperature data were available, trends were near zero at cold locations (T~ –10°C) and become more negative as the temperature approaches, and passes, the freezing point.

Since the AR4, a few additional relevant studies have been published.

4.6.3.1 Europe

Marty (Marty, 2008) examined records of snow days at 34 long-term stations in Switzerland over the 1948–2007 period and found a significant step decrease by 20% to 60% in snow days at the end of the 1980's, corresponding to a step increase of the mean winter temperature. Skaugen et al. (Skaugen et al., Submitted) examined records of SWE at an unspecified number of stations in Norway 1931-2009 and found trends that were generally downward at lower elevation and flat or upward at higher elevation; they also noted a significant correlation with the North Atlantic Oscillation at many sites, though the strength and sign of the correlation depended on altitude and on the phase of the NAO. At a station in northern Sweden, Kohler (Kohler et al., 2006) found an increase of 2cm per decade (5% of the mean) in seasonal mean snow depth since 1913, but no change in the start, end, or length of snow season; these results too are consistent with a place that cold (DJF mean temp at nearby Kiruna is -13°C [worldclimate.com - need better source]). Hantel and Hirtl-Wielke (Hantel and Hirtl-Wielke, 2007) fitted a logistic curve to data on snow cover duration at 268 stations in Austria, France, Germany, Italy, Slovenia, and Switzerland, with a maximum response of 30 days reduction per 1°C warming, which occurred on the fitted curve at an average temperature of –1.6°C and an elevation of 700m. A reconstruction of spring snow using tree rings from Arctic willow in northeastern Greenland

4.6.3.2 North America

Dyer and Mote (Dyer and Mote, 2006) used a gridded dataset of snow depth derived from observations to examine trends over 1960-2000, finding minimal change in SD in early winter and regional decreases beginning in late January attributable to more rapid melt stemming from shallower snow cover. Pierce et al. (Pierce et al., 2008) performed detection and attribution [see Chapter 10] on measurements of SWE/P (April 1 SWE divided by winter precipitation) in the western US, and estimated that about half the average decline in western US SWE/P could be attributed to human-induced changes in radiative forcing.

4.6.3.3 Eurasia

Recent analysis of in situ data show significant increases in winter snow accumulation over large areas of Eurasia (Bulygina et al., 2009)[Also Shmakin, 2010] but with a shorter and more intense snowmelt season (Bulygina et al., 2009). Significant trends toward earlier snowmelt and a shortening of the snowmelt season have been identified over much of Eurasia [Takala et al., 2009] and the pan-Arctic region (Tedesco et al., 2009) since 1979 from analysis of passive microwave satellite data, with a trend toward earlier melt of about 0.5 days per year for the beginning of the melt season, and about 1 day per year for the end of the melt season. Variability associated with the Arctic Oscillation splains about 50% of the variability in melt onset over Eurasia, but only 10% of the variability over North America, similar to the variability of temperature patterns over the two continental areas.

4.6.3.4 Arctic

Brown (Brown et al., 2010) used a multi-dataset approach including satellite, reanalyses and in situ observations to document variability and trend in Arctic spring (May-June) SCE over the 1967–2008 period. The new estimates show a more linear reduction in spring SCE than previously characterized by the National Oceanic and Atmospheric Administration (NOAA) weekly chart dataset, with air temperature explaining 49% of the variability in Arctic SCE in May and 56% of the variability in June. May and June SCE were determined to have decreased 14% and 46% respectively over the pan-Arctic region over the 1967-2008 period in response to earlier snow melt. The observed reductions in June SCE over the 1979-2008 period were found to be of the same magnitude as reductions in June sea ice extent with both series significantly correlated to air temperature changes over the Arctic region and to each other. This result underscores the close relationship between the cryosphere and surface air temperatures over the Arctic region in June when albedo feedback potential is at a maximum.

4.6.3.5 South America

Foster (Foster et al., 2009) presented the first satellite study of variability and trends in any measure of snow, in this case SWE. They focused on the coldest months (May-September) and noted large year-to-year and lower frequency variability but no trends.

In the Andes mountains of South America, Masiokas (Masiokas et al., 2006) examined long (1966–2004 or longer) records of SWE from 6 high-elevation sites (2275–3500 m) and stream gauges; they found positive but generally not significant trends in both maximum SWE and in annual streamflow. They also found relationships with ENSO. In a more recent study, Masiokas (Masiokas et al., 2010) used a longer data set in the same region and observed regime shifts that appear to be related to the Pacific decadal oscillation (PDO). As Brown and Mote (Brown and Mote, 2009) noted, these sites are at high enough altitudes to have maximum SWE determined by precipitation, not by temperature. A complementary viewpoint comes from considering the altitude of the 0°C isotherm (ZIA). Carrasco et al. (Carrasco et al., 2008) obtained the ZIA from radiosonde data from aerological Chilean stations located at Antofagasta, Quintero/Santo Domingo, Puerto Montt and Punta Arenas, at latitudes from about 18°S to 53°S. Results for the 1958–2006 period (except for Punta Arenas which is for 1975-2006) revealed positive and statistically significant trends of 39 ± 2, 23 ± 1, 24 ± 1 and 7 ± 1 m per decade, respectively (statistically significant at 95% level). The climate shift of 1976/1977 was also clearly revealed by the time series with a negative (cooling) trend before 1976 and a positive (warming) trend after 1977. In fact, analysis of the mean ZIA indicates predominantly warmer conditions during 1977–2006 period, statistically significant (at the 95% level), compared to the 1958–1976 period.

4.6.3.6 Australia and New Zealand

Green and Pickering (Green and Pickering, 2009) noted that the extent of snow "patches" in Australia had declined over the period 1954–2007. Willsman et al. (Willsman et al., 2007) report on aerial surveys of end-of-summer snowline on 50 glaciers spanning the South Island of New Zealand, over the 1977–2007 period of record. Repeat photography - often hampered by un-flyable weather or early snowfall that obscured the snowline [offering an approximation of glacier mass balance]. Amid substantial variability they found little trend in the mean elevation of snowline elevation.

4.6.4 Changes in Snow Albedo

An important component of the response of snow cover to anthropogenic forcing is the change in albedo that stems from two related causes (Flanner et al., 2007): 1) darker snow grains as a result of increased combustion of both fossil fuels and northern forests, and 2) accelerated snow metamorphosis as a result of warming. Unfortunately there are extremely limited data on the changes of albedo over time, and we must rely instead on analyses from ice cores, direct recent observations, and modeling. Flanner et al. (Flanner et al., 2007), using a detailed snow radiative model coupled to a GCM and estimates of biomass burning in years with low (2001) and high (1998) amounts of Arctic wildfire, estimated that the human-induced radiative forcing by black carbon is roughly 0.05 W m^{-2}, of which 80% is from fossil fuels. Fernandes et al. (Fernandes et al., 2009) found, using Advanced Very High Resolution Radiometer data for 1982–1999, that surface albedo variation was about 1%/°C warming, of which roughly half was due to changes in SCE and half due to snow metamorphosis. However, spatially comprehensive surveys of impurities in Arctic snow in the late 2000s and mid-1980s indicate that if anything, impurities have decreased between those two periods (Doherty et al., 2010) and hence albedo changes have not been responsible for reductions in Arctic ice and snow.

4.6.5. River and Lake Ice

Most trends toward shorter freshwater ice durations over much of the circumpolar North closely correspond to increasing air temperature trends and, more specifically, timing of the 0 °C isotherm.

Broad spatial patterns in ice trends have also been linked to major atmospheric circulation patterns, different phases of which can cause contrasting ice conditions (e.g.,, shorter versus longer ice duration) across individual continents and between opposite sides of the circumpolar North.

Some important south-north contrasts, however, have also been identified in freshwater ice trends. Examples from Scandinavia show more pronounced change (later freeze-up and earlier break-up) occurring in southern than in northern lakes, perhaps indicating greater sensitivity to warming at the more temperate latitudes.

Contrasting results, albeit involving a more recent period, have been noted for south-north regions of Canada based on records obtained by remote sensing. The degree to which this reflects the effects of either more recent or higher-latitude warming or a combination of both is unclear.

[PLACEHOLDER FOR FIRST ORDER DRAFT: Possible other papers:
RL Armstrong, ed., Snow and Climate (ISBN-13: 9780521854542)
Brown, R., C. Derksen, and L. Wang (2007) Assessment of spring snow cover duration variability over northern Canada from satellite datasets. Remote sensing of environment, 11, 367-381. doi:10.1016/j.rse.2006.09.035. [compares several satellite datasets, noting strengths and weaknesses.]
Casola, J.H., L. Cuo, B. Livney, D.P. Lettenmaier, P.W. Mote, and J.M. Wallace, 2009: Assessing the impacts of global warming on Pacific Northwest snowpack. J. Climate, 22, doi: 10.1175/2008JCLI2612.1. [derives a sensitivity of ~15-20% decline in April 1 SWE per °C]
Deems et al. doi:10.1175/JHM487.1 fractal distribution
Déry and Brown 2007 [topic adequately covered by Fernandes et al.]
Edwards et al. snow and alpine soil temp
Niu, G.-Y., and Z.-L. Yang (2007), An observation-based formulation of snow cover fraction and its evaluation over large North American river basins, J. Geophys. Res., 112, D21101, doi:10.1029/2007JD008674.
Pierce et al. 2009 - attribution of western US snowpack to anthropogenic forcing [refer to Chapter 9].
Scherrer and Appenzeller: Swiss alpine snowpack variability [discuss if space].
Schmidt, N.M., C. Baittinger, and M.C. Forchhammer (2006). Reconstructing Century-long Snow Regimes Using Estimates of High Arctic Salix arctica Radial Growth. Arctic, Antarctic, and Alpine Research, Vol. 38, No. 2, 2006, pp. 257–262. [interesting paper, but (a) not clear whether paleo should be here; (b) paper builds a statistical relationship between tree ring data and "snow cover" but does not state what the source, method, or duration of snow cover data is; (c) shows a trend but does not say what it is.]
Flannery et al. 2009 carbonaceous
Qian et al. 2009
Wang et al. - GCM-satellite comparison of surface albedo (for Chapter 9) doi:10.1029/2005JD006728

[INSERT FIGURE 4.16 HERE]
Figure 4.16: Variability April NH SCE over the period of available data with13-term filtered values of the mean and 95% confidence interval. The width of the smoothed confidence interval is also [tbc] influenced by the interannual variability in SCE. From Brown and Robinson (2010).

[INSERT FIGURE 4.17 HERE]
Figure 4.17: Relationship between NH April SCE and corresponding land area air temperature anomalies over 40°N–60°N from the CRU dataset. Air temperature explains 48.7% of the variance. From Brown and Robinson (2010).

4.7 Frozen Ground

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.7.1 Background

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.7.2 Changes in Permafrost

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.7.2.1 Permafrost temperature

Measurements indicate that permafrost temperatures: (i) increase in recent decades, (ii) near-isothermal conditions close to or within a few tenth of a degree from the freezing point, and (iii) fluctuations with episodic cooling.

Observed permafrost warming is greatest in cold permafrost such as on North Slope of Alaska and northern Canada (Smith et al., 2010), Siberia (Romanovsky et al., 2010). Observed evidence shows that warm permafrost has less degree of temperature increase due primarily to the energy consumption of phase change from ice to water. Regional permafrost temperatures show a similar trend with different magnitudes such as Svalbard, Greenalnd, and the Nordic area (Christiansen et al., 2010), Qinghai-Tibetan Plateau (Cheng and Wu, 2007), Mongolia (Zhao et al., 2010), Tian Shan (Marchenko et al., 2007), Kamchatka and southern Siberia (Romanovsky et al., 2010), and North America (Smith et al., 2010).

Near-isothermal conditions of warm permafrost are observed mostly in mountain permafrost regions and southern margins of discontinuous permafrost regions at high latitudes such as the European Alps (Noetzli (Noetzli and Vonder Muehll, 2010), Scandinavia countries (Christiansen et al., 2010), the North American Western Cordillera (Smith et al., 2010), and the Qinghai-Tibetan Plateau (Wu and Zhang, 2010; Zhao et al., 2010).

Episodic cooling as part of temperature fluctuations is usually short-lived and mostly controlled by site-specific conditions.

The cause of permafrost warming is mainly due to increase in air temperature and changes in snow cover conditions. Over cold permafrost regions where permafrost temperature increase is greatest, changes in snow cover may play a more important role (Smith et al., 2010; Zhang, 2005).

4.7.2.2 Active Layer

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.7.2.2.1 Background and data sources

[PLACEHOLDER FOR FIRST ORDER DRAFT: CALM data [Brown et al., 2000; updates], InSAR data and technique (Liu et al., 2010; Lu et al., 2008; Wu and Zhang, 2008)[Also others].]

4.7.2.2.2 Changes in active layer thickness

Observations of active layer thickness (ALT) indicate a strong increasing trend in discontinuous and permafrost regions at high latitudes and mountain permafrost regions in middle latitudes. A progressive active layer thickening has been observed north European countries since the 1970s and has accelerated since 1995 (Akerman and Johansson, 2008; Callaghan et al., 2010). Observations conducted on Svalbard and Greenland show ALT increase since the late 1990s with substantial spatial and temporal variability (Christiansen et al., 2010). ALT has increased significantly over the Russian European North due to the recent climate warming (Mazhitova, 2008). ALT increase has also been observed during the past 15 years in East Siberia (Fyodorov-Davydov et al., 2008) and Chukotka (Zamolodchikov, 2008). A progressive increase in ALT has been observed in the Interior of Alaska during the past two decades (Viereck et al., 2008). Burn and Kokelj (Burn and Kokelj, 2009) reported that there was an increase of 8 cm in ALT between 1983 and 2008 in the northern portion of the Mackenzie valley. ALT has increased since the mid 1990s in the eastern portion of the Canadian Arctic with the largest increase occurring in the bedrock of discontinuous permafrost zone (Smith et al., 2010). The most pronounced ALT increase was observed in permafrost areas of Central Asia. Wu and Zhang (Wu and Zhang, 2010) reported ALT increase of about 7.5 cm/yr over a period from 1995 through 2007 on the Qinghai-Tibetan Plateau. Rates of ALT increase up to 40 cm/yr over the past decade were observed in Mongolian sites characterized by warm permafrost. A relative low rate of ALT increase occurred in shallow active-layer areas over ice-rich and colder permafrost (Zhao et al., 2010). A clear trend of increasing ALT is also visible in Tian Shan (Marchenko et al., 2007). Over the Alps, ALT changes respond to extreme years (Noetzli and Vonder Muehll, 2010).

Observations show that there is no pronounced trend in ALT change over the majority of continuous permafrost regions at high latitudes. Changes in ALT on the Alaskan North Slope have no visible trend over the 1995-2008 period (Streletskiy et al., 2008). Detailed analysis of long-term observations near Barrow, Alaska shows no significant trend in ALT change since the early 1990s (Shiklomanov et al., 2010). In fact, ALT measured in the early 1960s is generally higher than the values measured in the 1990s and is compatible with values measured in the 2000s. Smith et al. (Smith et al., 2010)found similar results in ALT changes over the past 15 years in the Mackenzie Valley. Vasiliev (Vasiliev et al., 2008) also reported no trend in ALT change in West Siberian continuous permafrost region.

No observed trend in ALT change in most continuous permafrost regions may be in part explained by observed surface subsidence. Thaw penetration into an ice-rich layer at the base of the active layer is often accompanied by loss of volume due to thaw consolidation manifested as a ground subsidence at the surface. Results from ground-based measurements at selected sites on the North Slope of Alaska indicate up to 11 cm in surface subsidence over 2001-2008 period (Streletskiy et al., 2008). Mazhitova and Kaverin (Mazhitova and Kaverin, 2007) reported up to 20 cm of surface ground subsidence due mainly to thawing of ice-rich permafrost underneath the active layer in the Russian European North. Using satellite remote sensing data, (Liu et al., 2010) reported a surface subsidence of 1 to 4 cm over 1992–2000 period on the North Slope of Alaska.

4.7.2.3 Talik Development

[PLACEHOLDER FOR FIRST ORDER DRAFT: Definition and description; Talik development in various regions; Causes: slightly more summer thaw and much less winter freeze.]

4.7.2.4 Permafrost Distribution

[PLACEHOLDER FOR FIRST ORDER DRAFT: Overview [Brown et al., 1997; Zhang et al., 2008]

Significant permafrost degradation has been reported in the Russian European North. Oberman (Oberman, 2007) reported that permafrost with thickness of 10 to 15 m was completely thawed out over 1975-2005 period in the Vorkuta area because of the recent climatic warming. As a result, the southern boundary of permafrost has moved northward by up to 80 km, while the boundary of continuous permafrost has moved northward by 15–20 km in the lowlands and by 30–50 km in the foothills (Oberman, 2008), 2008). Permafrost degradation is also evident in discontinuous permafrost of west Siberia (Romanovsky et al., 2010). Disappearance of permafrost in several mire landscapes in Nordic countries has also been reported (Akerman and Johansson, 2008; Callaghan et al., 2010).

Permafrost degradation has also been reported on the Qinghai-Tibetan Plateau (Zhao et al., 2010), and some Canada studies).

The Arctic Coast erosion is in a rapid during the past several decades and is accelerating in recent years [reference needed], resulting in large amount of terrestrial continuous permafrost into subsea permafrost. Once emerged under water, the cold continuous permafrost immediately thaws due to seawater effect. Similar case is also true for permafrost under thaw lakes. The number of thaw lakes is increasing and thaw lake is expanding [Smith et al., 200x]. Permafrost under these thaw lakes are also degrading.

4.7.3 Changes in Landforms in Permafrost Regions

Rock Glaciers: During recent years, fast, accelerating and destablized rock glaciers have received increased attention. Time series measured by terrestrial survey indicate a dramatic speed-up of roch glacier movement during recent decades as well as seasonal velocity changes related with ground temperatures (Bodin et al., 2009) [Schoeneich et al., 2010] (Delaloye et al., 2011) (Noetzli and Vonder Muehll, 2010). Diachroonic photo comparison and photogrammetry data indicate an increased activity and collapse-like features developing in rock glaciers (Roer et al.). The clear relationship between mean annual air temperature at rock glacier front and rock glacier velocity points to a temperature dependence and thus plausible climate impact on velocities (Kaab et al., 2007) and collapse. While these measurements are essentially from the Alps, this phenomenon very likely has global importance in delivering sediments pulses to steams and feeding debris-flow torrents. Strong surface lowering of rock glaciers has been reported in the Andes (Bodin et al., 2010), indicating melting of massive ground ice in rock glacier and permafrost degrading.

Many rock fall evens have originated from permafrost slopes during recent years (Ravanel and Deline, 2010 and Ravanel et al., 2010). Incrasing evidence based on exposed ice and on event statistics supports the hypothesis that this is in part due to thaw of permafrost in steep bedrock (Gruber and Haeberli, 2007).

[PLACEHOLDER FOR FIRST ORDER DRAFT: Thermokarst – satellite remote sensing, IPY (Osterkamp et al., 2009; Smith et al., 2009); thaw lakes – [Hinkel et al., 2008]; rock falls (Gruber et al., 2004); and solifluction and slope stability [Niu et al., 2006; French et al., 2008]

4.7.4 Changes in Seasonally Frozen Ground

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.7.4.1 Background and Data Sources

[PLACEHOLDER FOR FIRST ORDER DRAFT: Historical soil temperature data from Russia (Zhang, 2005); Mongolia [references will be provided]; China [references will be provided]; US NWS stations [HU et al., 2003 and updates]; Environmental Canada [references will be provided].]

4.7.4.2 Changes in Thickness

[PLACEHOLDER FOR FIRST ORDER DRAFT: This will be mainly from in-situ data. Thickness of seasonally frozen ground is decreasing rapidly [Frauenfeld et al., 2004 and updates](Zhao et al., 2010).

4.7.4.3 Changes in Area Extent

[PLACEHOLDER FOR FIRST ORDER DRAFT: Mainly from passive microwave remote sensing results [McDonald et al., 2007 and updates; Jin et al., 2011]; timing, duration, and number of days of soil freeze – satellite data [Zhang et al., 2009; Jin et al., 2011].]

4.7.5 Climate Controls for Changes in Frozen Ground

[PLACEHOLDER FOR FIRST ORDER DRAFT: Air temperature; precipitation; snow cover: timing and thickness; vegetation; modelling]

4.7.6 Uncertainty

[PLACEHOLDER FOR FIRST ORDER DRAFT]

4.7.7 Summary

[PLACEHOLDER FOR FIRST ORDER DRAFT]

[START FAQ 4.1 HERE]

FAQ 4.1: Are Mountain Glaciers Disappearing?

[PLACEHOLDER FOR FIRST ORDER DRAFT]

[END FAQ 4.1 HERE]

[START FAQ 4.2 HERE]

FAQ 4.2: How is Sea-Ice Changing in the Arctic and Around Antarctica?

[PLACEHOLDER FOR FIRST ORDER DRAFT]

[END FAQ4.2 HERE]

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Chapter 4: Observations: Cryosphere

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Figures

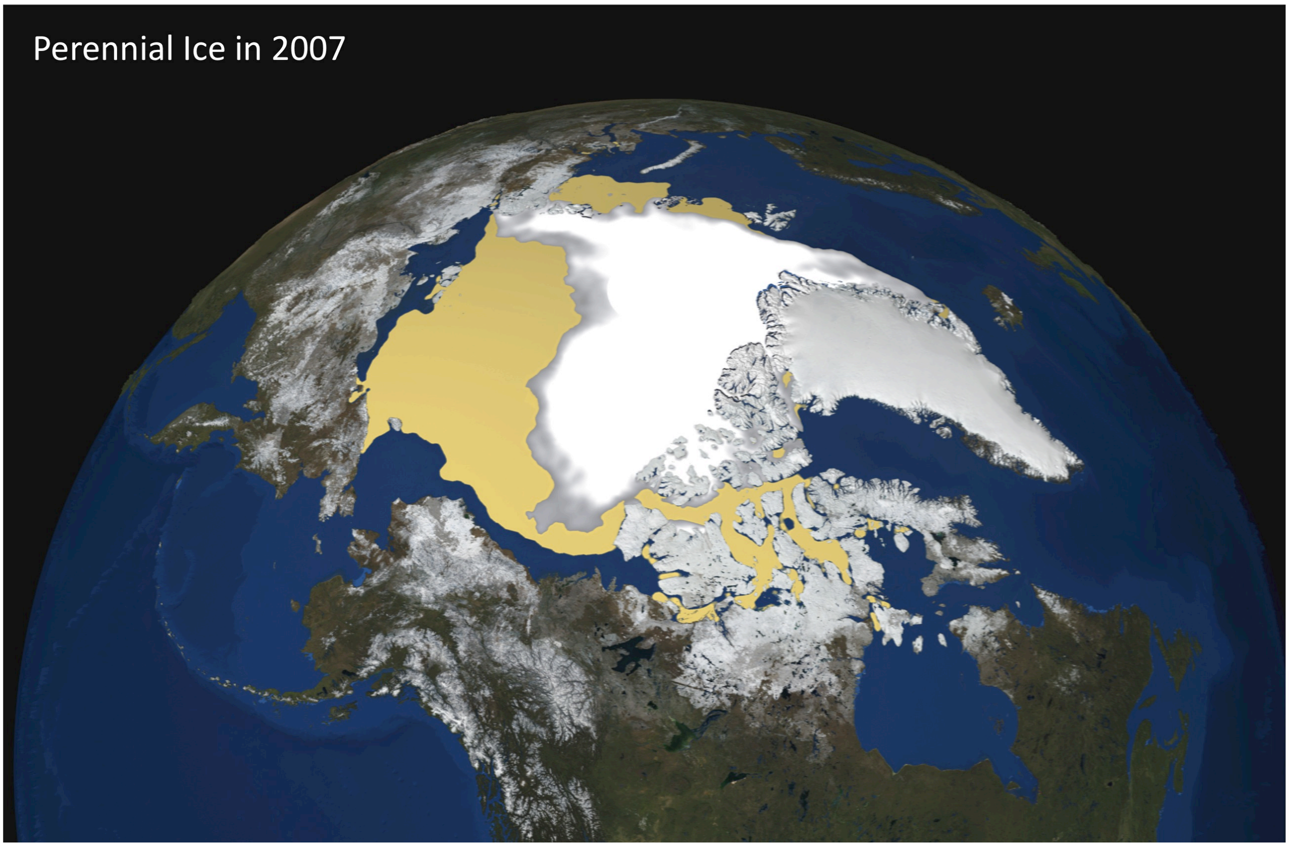

Figure 4.1: Loss of end of summer (perennial) sea ice cover in the Arctic. The area in white and gray in the Central Arctic represents the extent of the perennial ice cover in 2007 when compared to the average value (that includes the area in gold) from 1979 to 2006. [Note on graphic: We need one strong graphic for the introduction – Figure 4.1 may be revised to suit. It strongly illustrates that dramatic and influential changes have been occuring in the Arctic. Note: AR4 did not report on the dramatics decline in the perennial ice cover in 2007.]

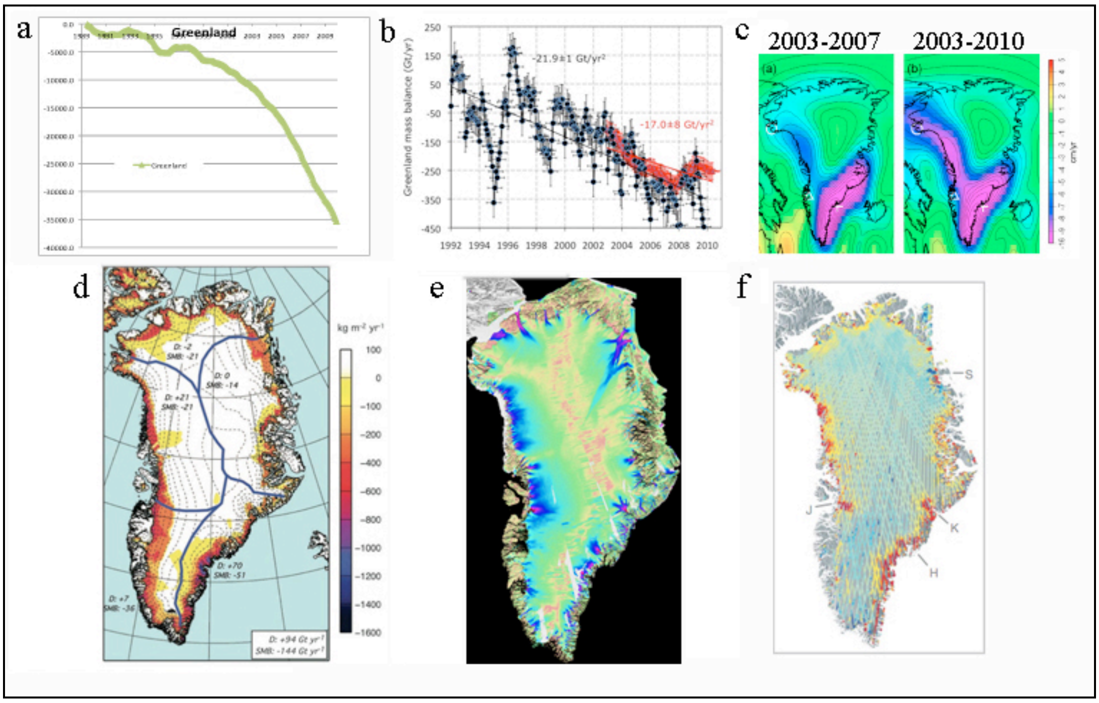

Figure 4.2: a) cumulative mass loss in Greenland from the MB method for 1992–2010; b) time series of annually-resolved losses from MBM (black) and GRACE (red) 1992–2010; c) temporal pattern of mass loss from GRACE time-variable gravity; d) mass losses per sector detailing the partitioning between surface and dynamic losses combining RACMO2/GRE and MBM; e) velocity map from satellite interferometry 2009; f) ice thinning rates from ICESAT data 2003–2008.

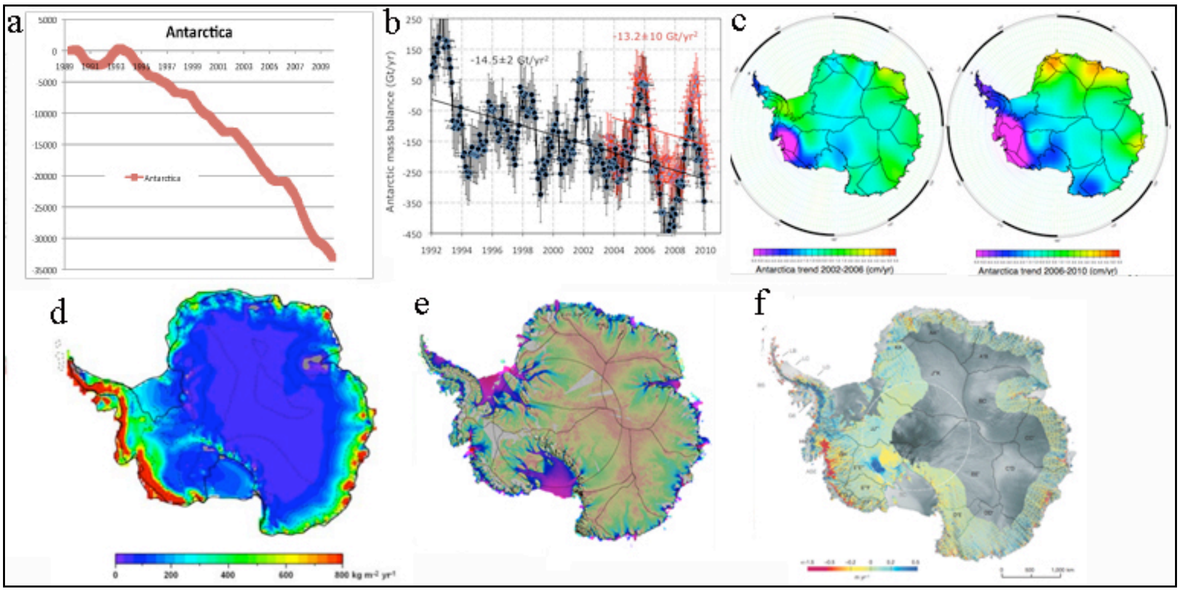

Figure 4.3: a) Cumulative mass loss from the MB method; b) annual mass loss from MB (black) and GRACE method (red); c) temporal evolution of mass loss from GRACE time-variable gravity; d) surface mass balance from RACMO2/ANT; e) ice sheet velocity from satellite radar interferometry 2007–2009; f) ice thinning rates from ICESAT 2003–2009.

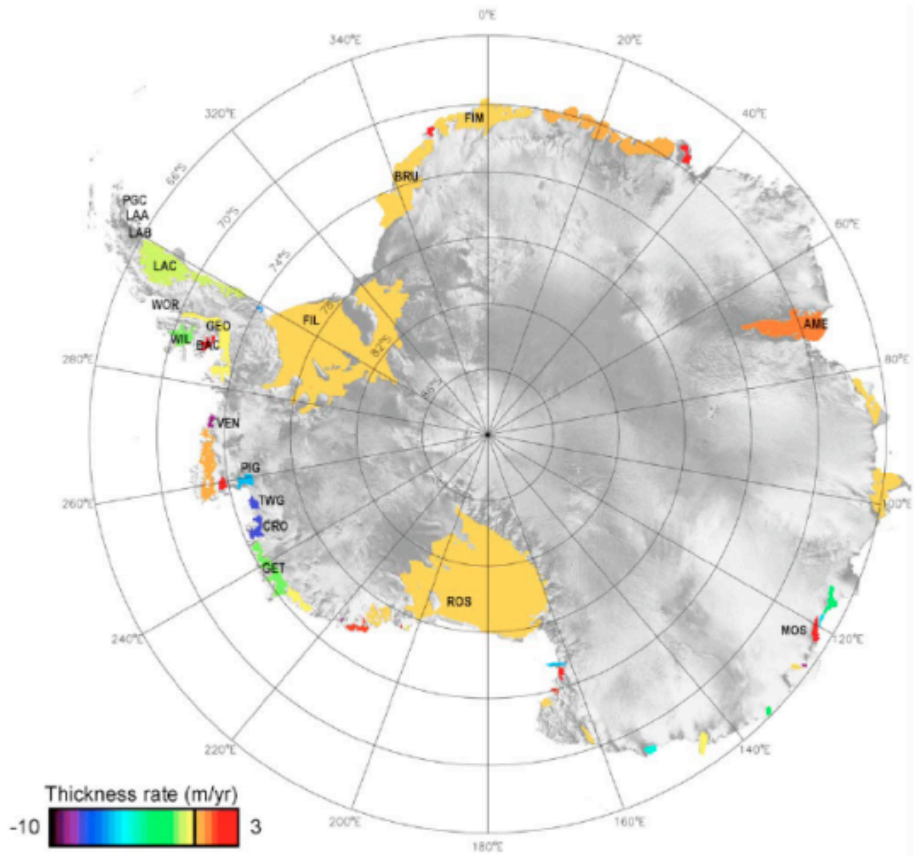

Figure 1. Average rate of Antarctic ice shelf thickness change, 1994 to 2008, determined from ERS and ENVISAT radar altimetry and a model of accumulation fluctuations [*Helsen et al.*, 2008].

Figure 4.4: Average rate of Antarctic ice shelf thickness change, 1994–2008, determined from ERS and ENVISAT radar altimetry and a model of accumulation fluctuations [Helsen et al., 2008].

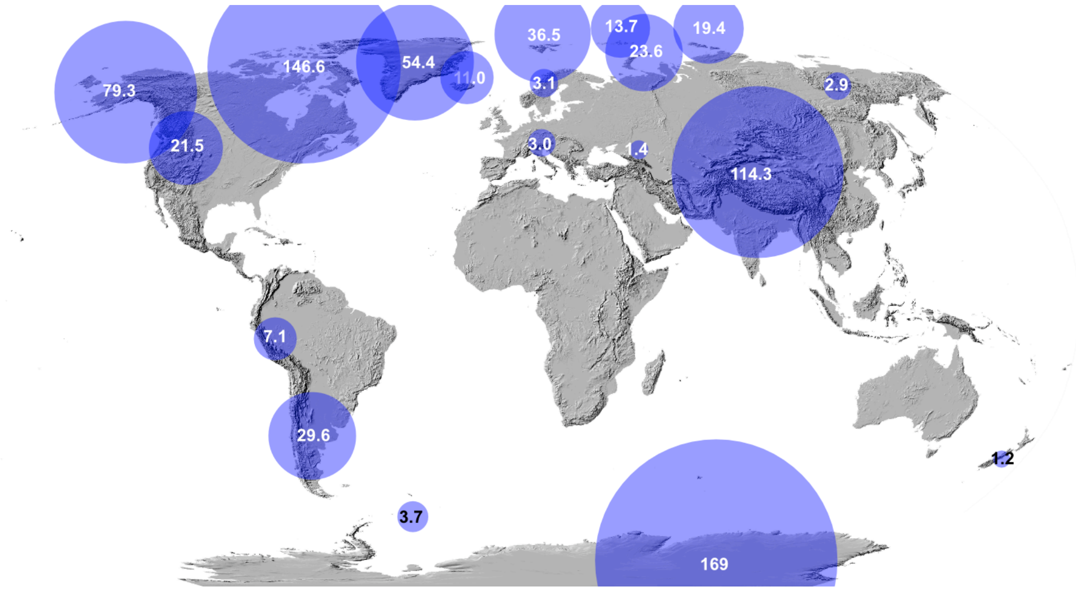

Figure 4.5 (a): [PLACEHOLDER FOR FIRST ORDER DRAFT: Option a [regions and numbers in this preliminary figure are from Radić and Hock (2010). Glacier outlines can possibly be added.]

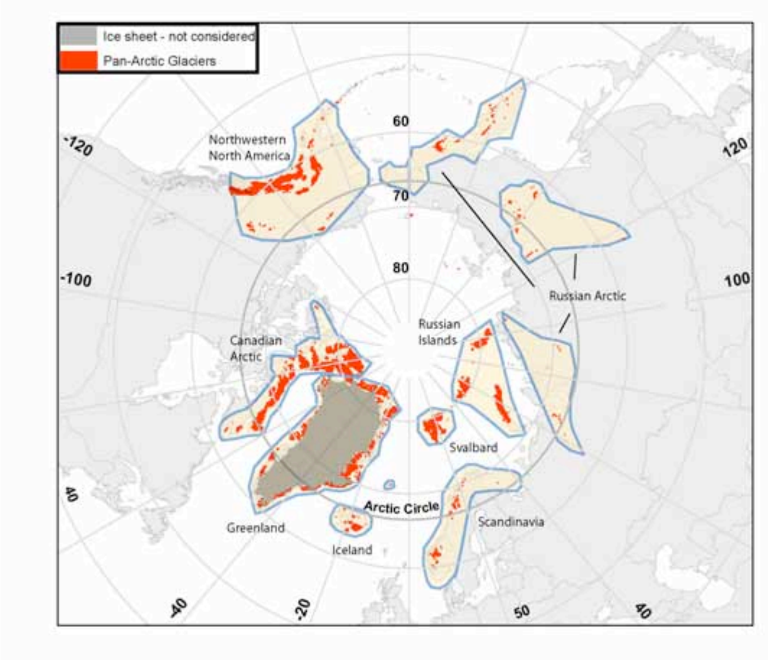

Figure 4.5 (b): [PLACEHOLDER FOR FIRST ORDER DRAFT: Option b [from G. Cogley, in progress]

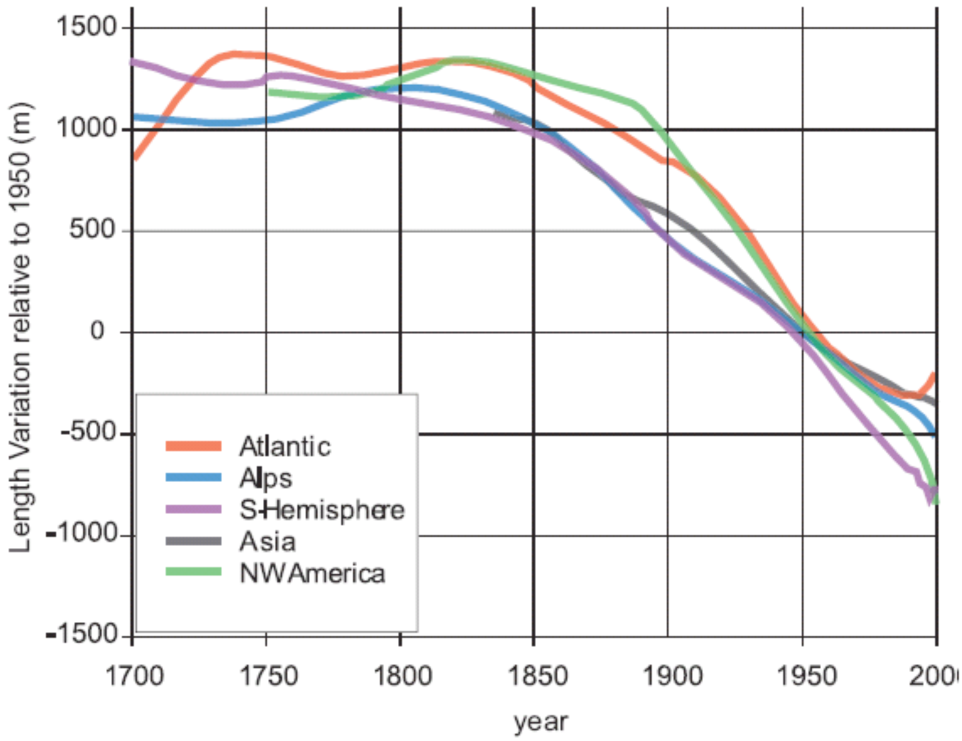

Figure 4.13. *Large-scale regional mean length variations of glacier tongues (Oerlemans, 2005). The raw data are all constrained to pass through zero in 1950. The curves shown are smoothed with the Stineman (1980) method and approximate this. Glaciers are grouped into the following regional classes: SH (tropics, New Zealand, Patagonia), northwest North America (mainly Canadian Rockies), Atlantic (South Greenland, Iceland, Jan Mayen, Svalbard, Scandinavia), European Alps and Asia (Caucasus and central Asia).*

Figure 4.6: [PLACEHOLDER FOR FIRST ORDER DRAFT: an update from this (Lemke et al., 2007) is expected.

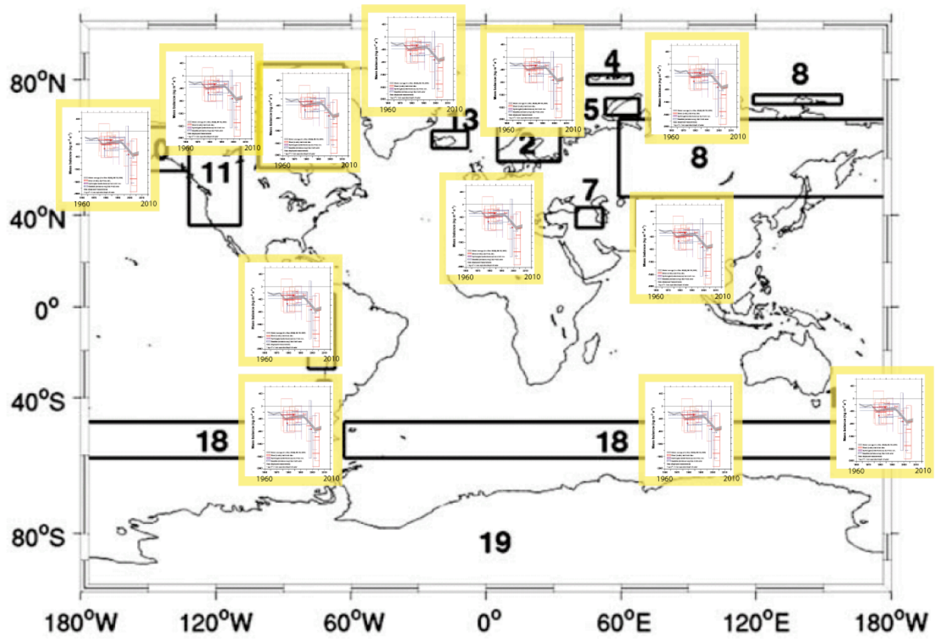

Figure 4.7 (a): [PLACEHOLDER FOR FIRST ORDER DRAFT] regional evolution of mass changes in R-H regions. Example given here is the Himalaya one. Directly measured mass changes and geodetically obtained volume changes are merged. An extension back to LIA is planned for a selected number of regions. A symbol code will show how well data based are certain regions compared to others: e.g., good data coverage: solid/bold line, poor data coverage: broken/transparent line (or even no line e.g., for missing reconstructions of past mass changes) [Together with Chapter 13 the preparation of a world map including future GIC development scenarios is planned]

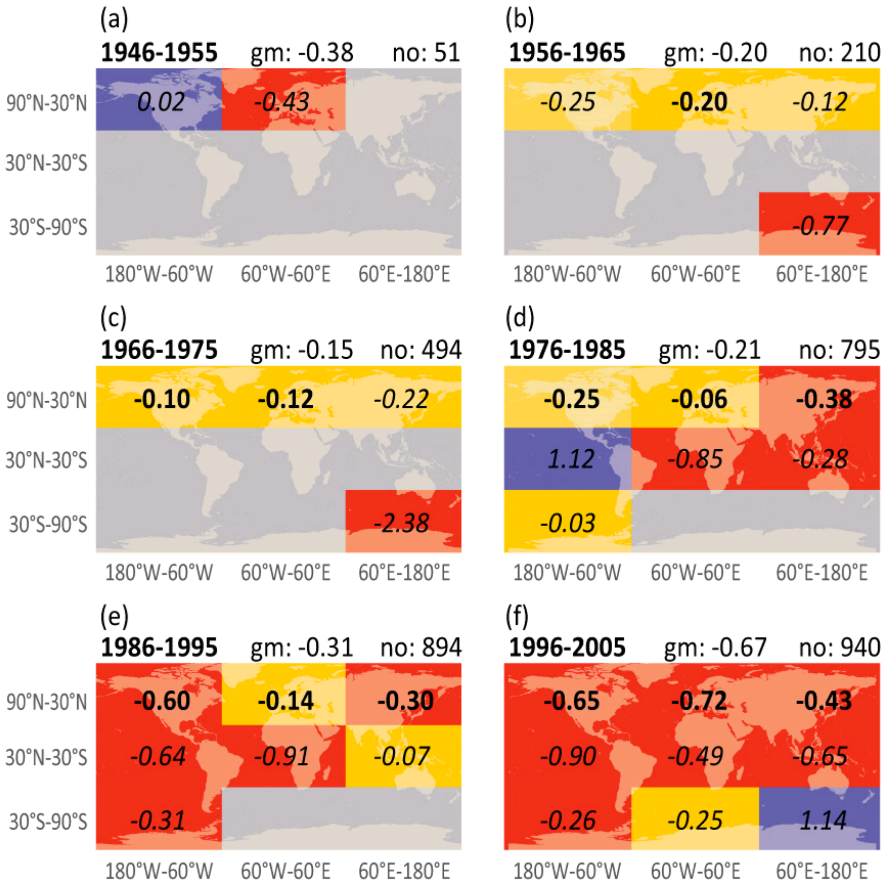

Figure 4.7 (b): [PLACEHOLDER FOR FIRST ORDER DRAFT] regional evolution of mass changes of glaciers and ice caps in sectors (Zemp et al., 2007) [Also the potential for a simple gridded map like this may be a choice]

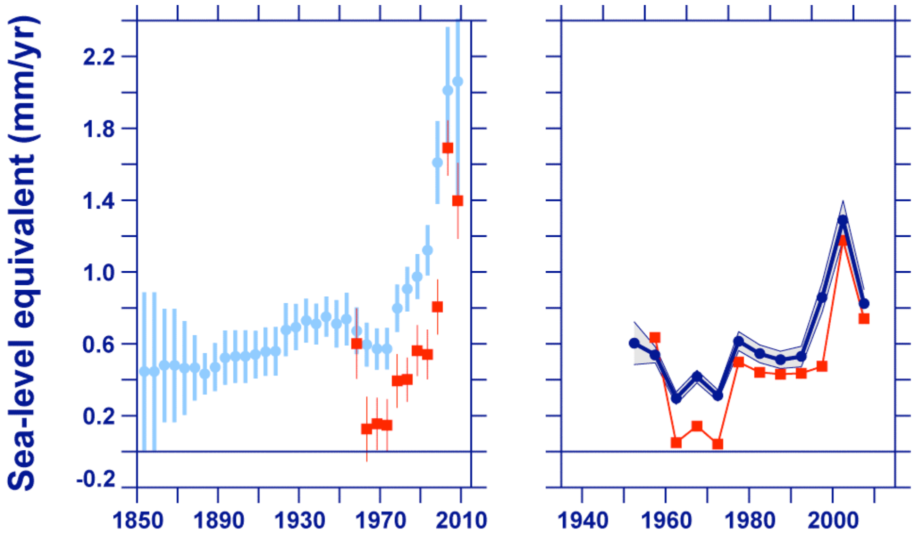

Figure 4.8: the latest update (10 October, including some early measurement reports for 2008/2009) from Cogley (2009) of the global time series (glaciological only in red, glaciological plus geodetic balances in blue; simple arithmetic averages of measurements on the left, spatially interpolated estimates on the right) [from . G. Cogley, in progress]

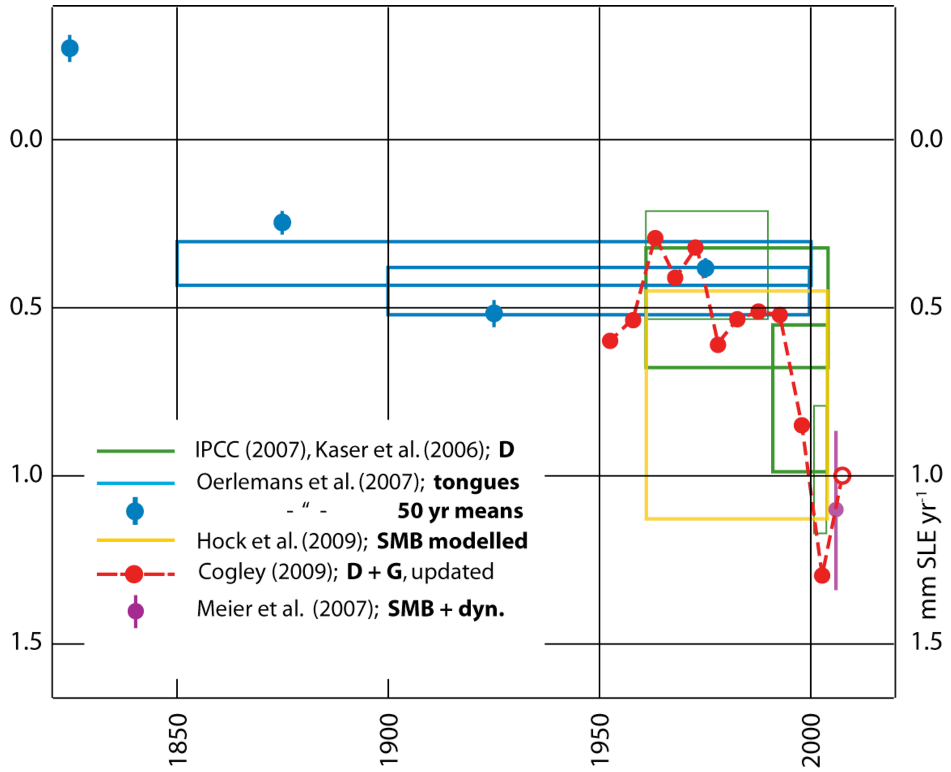

Figure 4.9: Compilation of SLE estimates (several methods) as available by June 2009 (Allison et al., in press). D: direct glaciological method, SMB. Surface mass balance, G: geodetic method. Compilation: G. Kaser, to be updated. Uncertainties to be partially added. (Hock et al., 2009, Lemke et al., 2007, Kaser et al., 2006, Oerlemans et al., 2007, Cogley, 2009, Meier et al., 2007). [Layout to be coordinated with the Ice Sheet chapter colleagues]

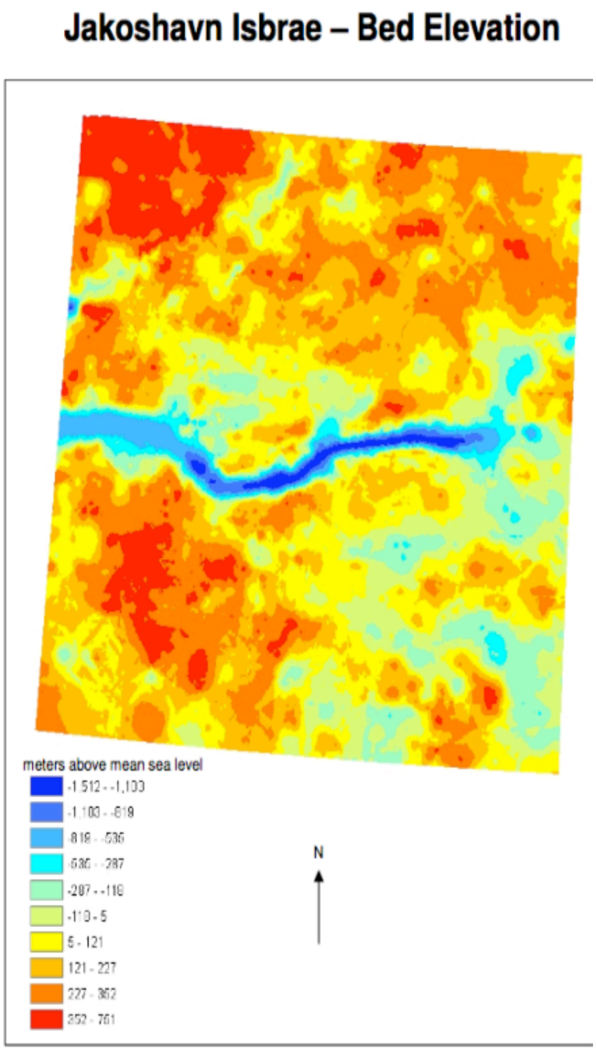

Figure 4.10: Bed trough of Jakobshavn Isbrae, West Greenland mapped with radio echo sounding (Plummer et al., 2010).

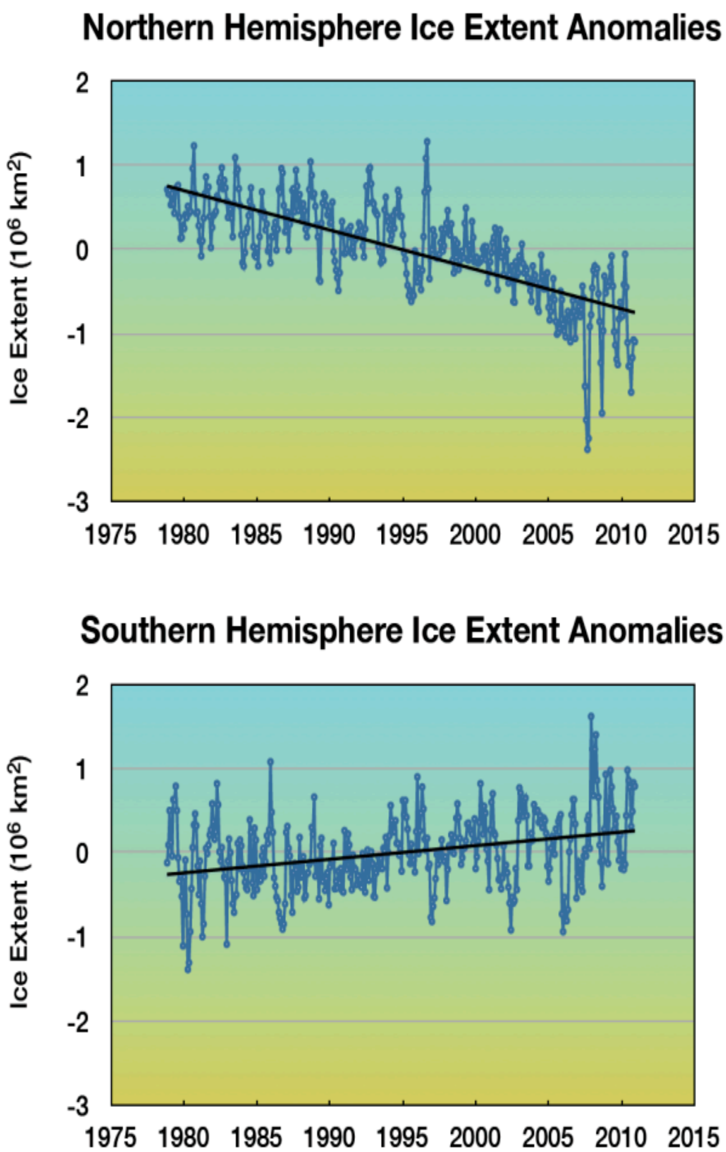

Figure 4.11: Monthly ice extent anomalies in the (a) Northern Hemisphere and the (b) Southern Hemisphere. The anomalies were estimated by subtracting climatological monthly averages (as derived from 1978–2010 satellite SMMR and SSM/I data) from each monthly extent.

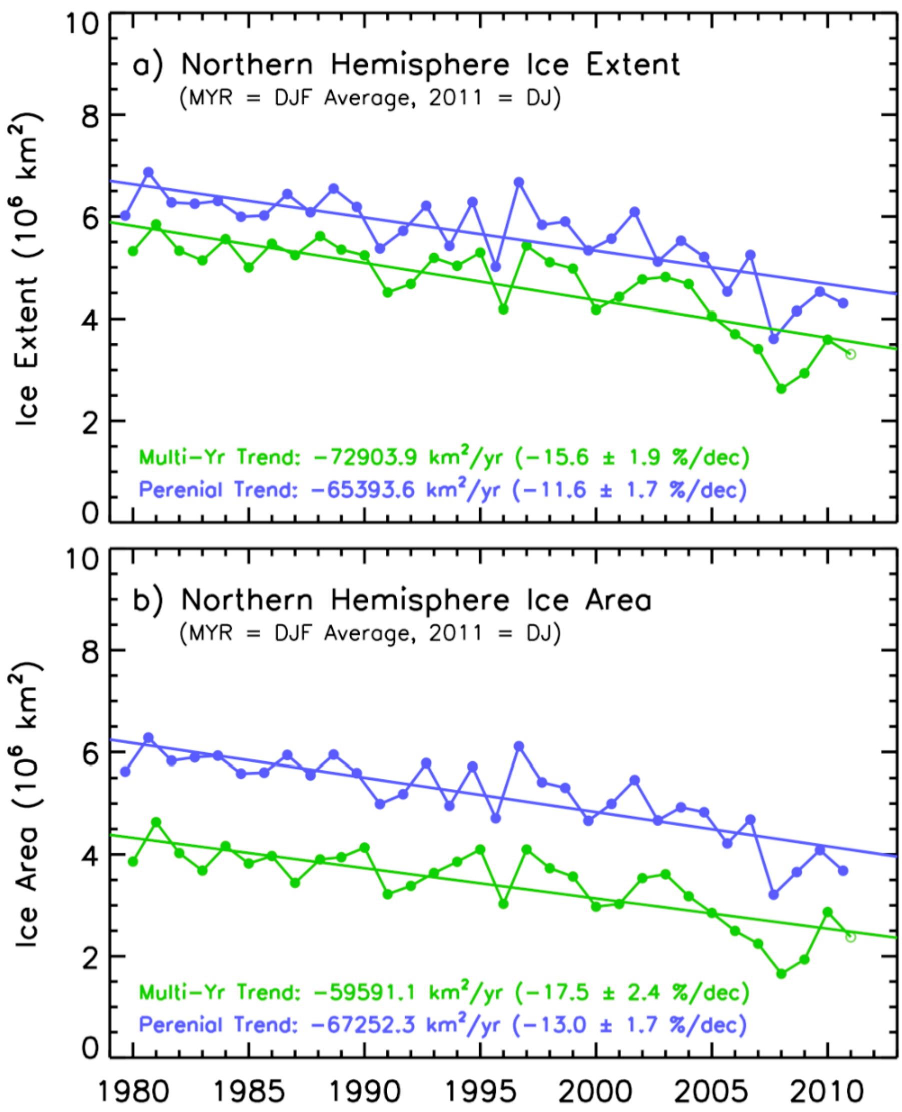

Figure 4.12: Annual changes in (a) ice extent and (b) area of the perennial (blue) and multiyear (green line) ice cover as derived from passive microwave (SMMR and SSM/I) data.

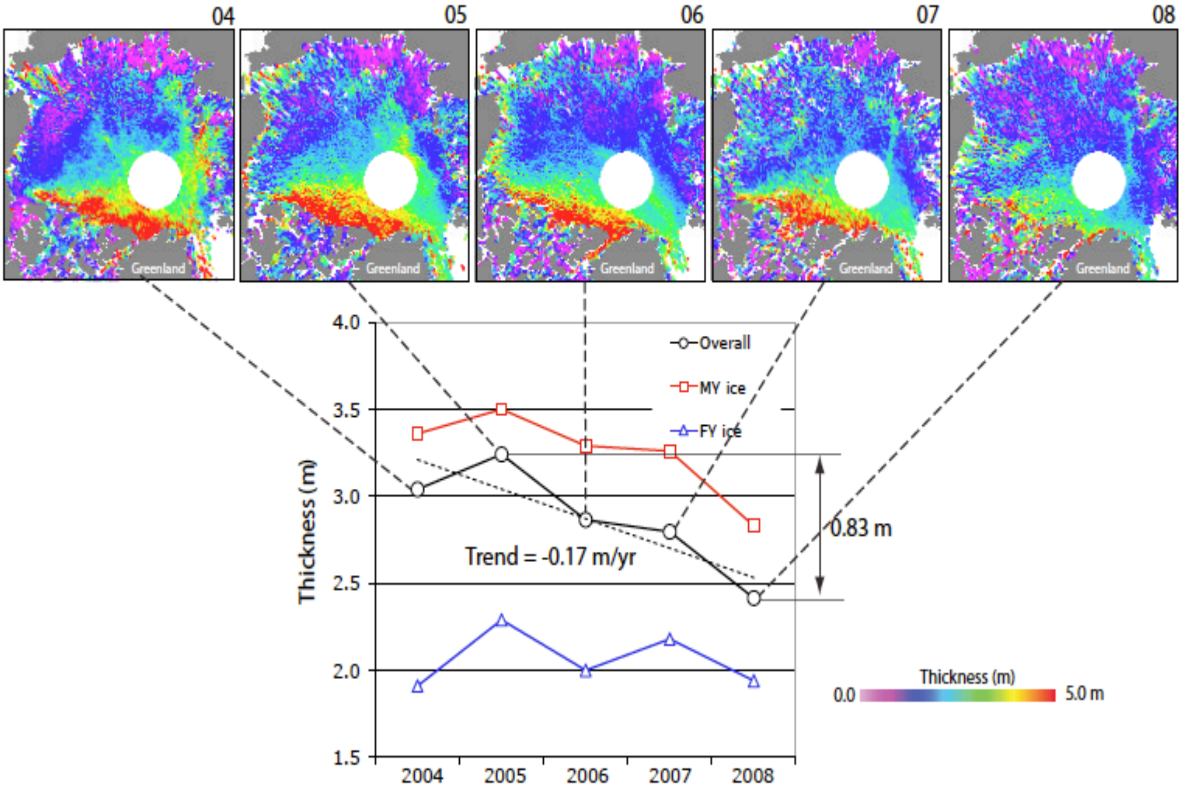

Figure 4.13: Decline in sea ice thickness (2004–2008) of the Arctic Ocean from ICESat. [more details to be provided]

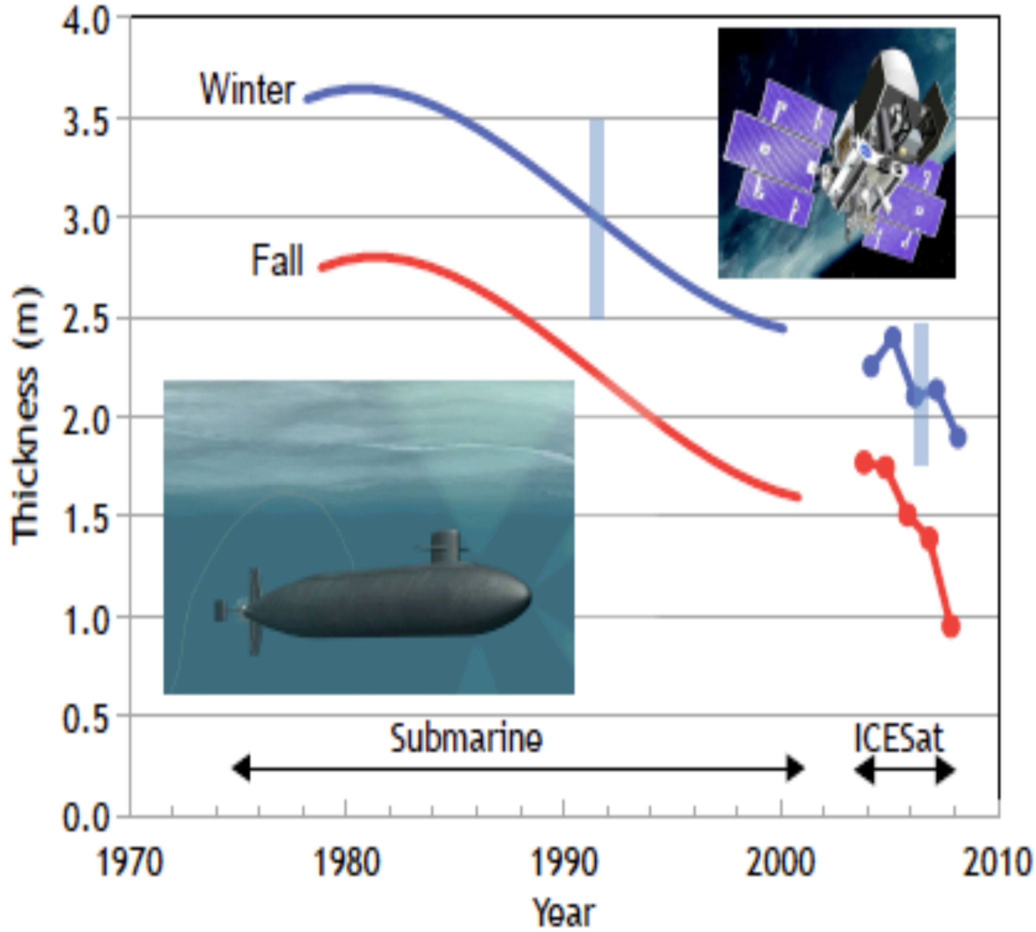

Figure 4.14: Decline in Arctic ice thickness from combined submarine and ICESat records. [more detailed description of the figure to be added]

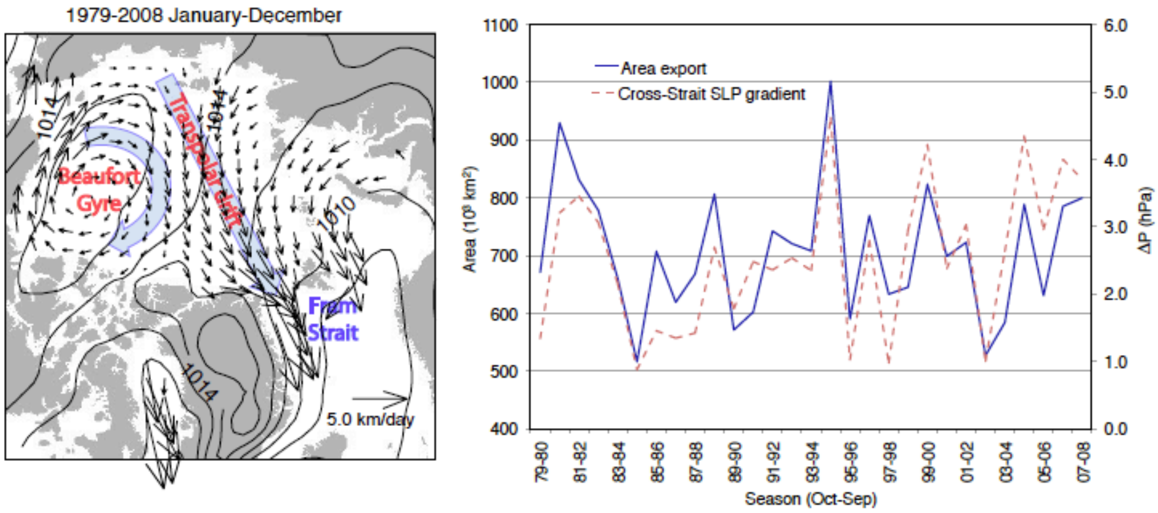

Figure 4.15: Export of sea ice area at the Fram Strait. [volume estimates to be added]

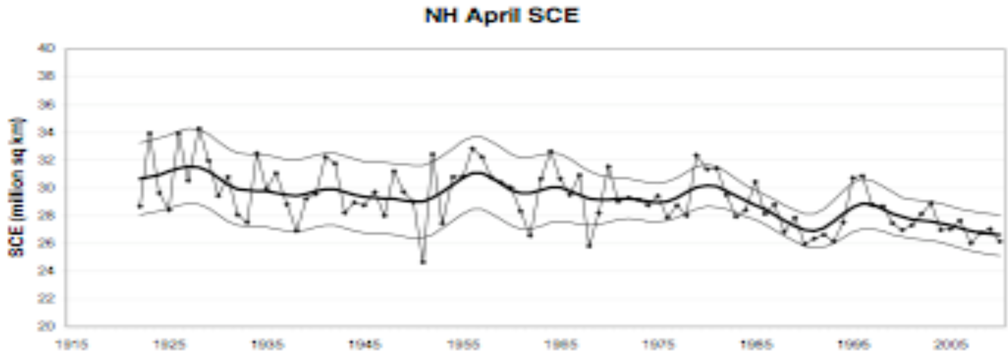

Figure 4.16: Variability April NH SCE over the period of available data with13-term filtered values of the mean and 95% confidence interval. The width of the smoothed confidence interval is also [tbc] influenced by the interannual variability in SCE. From Brown and Robinson (2010).

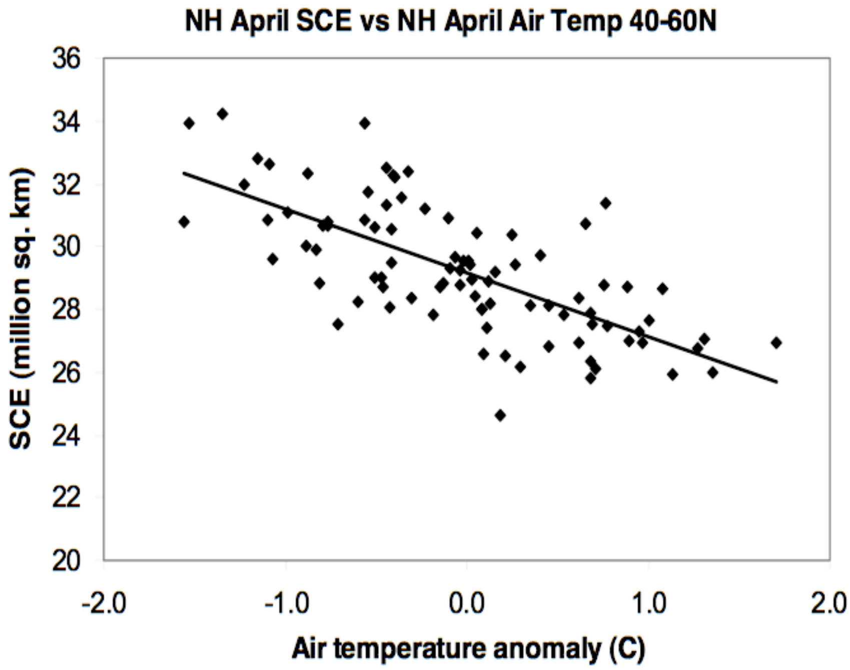

Figure 4.17: Relationship between NH April SCE and corresponding land area air temperature anomalies over 40°N–60°N from the CRU dataset. Air temperature explains 48.7% of the variance. From Brown and Robinson (2010).