

Chapter 4: Observations: Cryosphere

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Executive Summary

The cryosphere, comprising snow, river and lake ice, sea ice, glaciers, ice shelves and ice sheets, and frozen ground, plays a major role in the Earth's climate system through its impact on the surface energy budget, sea level change, water cycle, primary productivity and surface gas exchange. It is thus a fundamental control on physical, biological and social environment over substantial areas of the Earth's surface. Given the inherent temperature-sensitivity of all components of the cryosphere over a wide range of time scales, the cryosphere is a natural integrator of climate variability that provides some of the most visible signatures of climate change.

Since the AR4, observational technology has improved, and key time-series of measurements have been lengthened, such that our measurement of changes and trends in all components of the cryosphere has been substantially improved and our understanding of the specific processes governing their responses has been refined. These new observations show that ice is being lost from many of the components of the cryosphere, although there are significant regional differences in the rate of loss. The major changes occurring to the cryosphere are as follows.

- The strong and significant decrease in Arctic sea ice extent and area reported in AR4 has continued, and has been accompanied by many other changes in the characteristics of the Arctic sea ice cover (robust evidence in high agreement). The average decadal extent of Arctic sea ice has decreased in every season and in every successive decade since satellite observations commenced. The overall trend in sea ice extent over the period 1979–2011 has been –3.9% per decade with larger changes occurring in summer and autumn. The largest changes of all are the decline in the coverage of perennial ice (the summer minimum extent; –12.2 % per decade) and multiyear ice (more than 2 years old; –15.6 % per decade) Sea ice concentration has also decreased and the rate of decrease in ice area has been greater than that in extent. Robust evidence shows decline in perennial and multiyear sea ice coverage and decreases in ice thickness, and in ice volume. The overall mean winter thickness decreased by 48% to only 1.89 m between 1980 and 2009. With decrease in both concentration and thickness, sea ice has less resistance to wind forcing, and the rate of drift has also increased. Other significant changes to the Arctic Ocean sea ice include lengthening in the duration of the period of surface melt on perennial sea ice of 6 days per decade over the period 1979–2010, and a nearly 2-month lengthening of the ice-free season in the region from the East Siberian Sea to the western Beaufort Sea over the period 1979–2011.
- In Antarctica, there was a small but significant increase in total sea ice extent of 1.4% per decade between 1979 and 2011, and a greater increase in sea ice area, due to an increase in concentration. Robust evidence shows strong regional differences within this total, with some regions increasing in extent/area and some decreasing. There are also contrasting regions around the Antarctic where, over the period of satellite observations, the ice-free season has lengthened, and others where it has shortened. Decadal trends in ice drift show acceleration of the Ross Sea Gyre and deceleration of the Weddell Gyre. Available data are inadequate to assess the status of change of many other characteristics of Antarctic sea ice (e.g., thickness and volume).
- The total land-surface area covered by glaciers was not precisely known in AR4, resulting in large uncertainties for all related (area-dependent) calculations. For AR5, a new globally complete vector dataset of glacier outlines has been compiled indicating a total area of 739,820 km², with the peripheral glaciers around the ice sheets accounting for about 30% of the total. The related volume estimates are, however, still highly uncertain and range from about 165,000 to 212,103 km³ (equivalent to 0.46–0.59 m of global sea level).
- The time series of measured changes in glacier length and area, as well as estimates of volume and mass change give robust evidence in high agreement that globally, glaciers continue to shrink and lose mass. The number of regional-scale estimates of glacier change have grown with newly available data sets (e.g., from satellite remote sensing) and methods (e.g., gravimetry) becoming available, but results still show some spread due to methodological limitations and uncertainties. There are, however, notable regional exceptions to these trends in all three characteristics (length, area, mass changes) resulting from: regionally-specific climatic conditions (e.g., increased

precipitation), or from glacier-specific characteristics unrelated to climate (e.g., surge-type, calving, or debris-covered).

- In the period 2005–2009 global glacier ice loss amounted to 371 ± 50 Gt/yr, or 262 ± 67 Gt yr⁻¹ if glaciers around the periphery of the ice sheets are excluded. A recent multi-method estimate over the period 2003–2009 shows only 251 ± 65 Gt yr⁻¹ and 210 ± 65 Gt yr⁻¹, respectively. There is thus robust evidence and high agreement that, globally, mass loss from glaciers is ongoing, but there is less agreement about the actual rate. Glacier area changes averaged over entire mountain ranges varied between 0 and –1% per year from the 1960s to 2000, and are regionally even higher (–1 to –2% per year) for the past two decades. Several hundred glaciers globally have completely disappeared in the past 30 years (robust evidence, in high agreement). Estimates from different methods that yield long time series, indicate steady increases in ice loss from glaciers since about 1985 with a slight decline in the most recent years. Two estimates indicate that in the 1920s to 1940s, ice loss from glaciers in the Arctic, mainly from the Greenland peripheral glaciers was higher than today (medium evidence medium agreement).
- There is robust evidence in high agreement that the Greenland Ice Sheet has lost mass since the early 1990s, and that the rate of loss has increased. The average ice-loss from Greenland was 123 ± 22 Gt yr⁻¹ over the period 1993–2010, and 228 ± 54 Gt yr⁻¹ in the period 2005–2010. This loss is equivalent to a sea level rise of 0.34 mm yr⁻¹ for 1993–2010, and 0.63 mm yr⁻¹ for 2005–2010. This loss resulted from increased surface melt and runoff, and increased glacier discharge: melt has increased with warming temperature and discharge has increased with increased glacier speed, particularly in southeast, central west and northwest Greenland.
- Overall, the Antarctic Ice Sheet is also very likely currently losing mass (Robust evidence high agreement). The average ice-loss from Antarctica was 65 ± 33 Gt yr⁻¹ over the period 1993–2010, and 112 ± 58 Gt yr⁻¹ over the period 2005–2010. This loss was equivalent to a sea level increase of 0.18 mm yr⁻¹ for 1993–2010, and 0.31 mm/yr for 2005–2010. The largest ice losses from Antarctica have occurred on the northern Antarctic Peninsula and from the Amundsen Sea sector of West Antarctica. In the last two decades, East Antarctica is likely to have experienced a small gain in mass. As in AR4, reconstructions of snowfall spanning, now covering the period 1979–2011, do not suggest any change in total snowfall in Antarctica.
- In both ice sheets, substantial changes in the speed of some outlet glaciers have continued trends reported in AR4. Some glaciers slowed down following acceleration, others stabilized at high flow speeds, but none have returned to their conditions prior to the 1990s. Recent observations (Robust evidence, high agreement) suggest that observed changes in ice flow are likely due to oceanic changes in glacial fjords (Greenland) and beneath ice shelves (West Antarctica), but the observational record of ice-ocean interactions around both ice sheets remains poor.
- Ice shelves in the Antarctic Peninsula are very likely continuing a long-term trend of retreat and partial collapse that began decades ago, already reported in AR4, and very likely due to climate warming. Similarly, a progressive weakening of the floating parts of glaciers is likely taking place in Greenland. In the Antarctic, many floating ice shelves are virtually certain to be thinning in areas of rapid glacier changes due to enhanced oceanic thermal forcing. Many ice shelves in East Antarctica, especially the large ice shelves in the Ross and Weddell seas, however, are likely stable at present.
- In total, the sea level rise from the combined contribution of the Antarctic and Greenland ice sheets and global glaciers was 1.2 ± 0.2 mm yr⁻¹ for 1993–2010 and 1.7 ± 0.5 mm yr⁻¹ for 2005–2010. For the latter period, the sea level rise from melting ice was greater than that from thermal expansion of a warming ocean (see Chapter 3).
- Both satellite and in-situ observations show significant reductions in the Northern Hemisphere snow cover extent (SCE) over the past 90 years, with most reduction occurring in the 1980s. Snow cover decreased most in spring when the average extent decreased by around 8% (7 million km²) over the period 1970–2010 compared with the period 1922–1970. Because of earlier spring

1 snowmelt, the duration of the NH snow season has declined by 5.3 days per decade since the
2 1972/1973 winter. Since AR4, the rate of reductions in June SCE – both absolute and relative - has
3 surpassed the rate of reduction of March-April SCE. These trends in spring SCE are very likely
4 linked to rising temperature. In addition to reductions in snow cover extent, there is some evidence
5 that the reflectivity (albedo) of the snow itself may also be changing in response to human
6 activities. In the southern hemisphere, very few long records exist; satellite records of snow water
7 equivalent date from 1979, but show no trends.

- 8
- 9 • The limited evidence available for freshwater lake and river ice indicate that winter ice duration is
10 contracting, with delays in autumn freeze-up proceeding more slowly than advances in spring
11 break-up, and there is some evidence of recent acceleration in both (to 0.16 days yr⁻¹ later freezeup
12 and 0.19 days yr⁻¹ earlier breakup over 1975–2004 for more than 30 sites across the NH). The
13 seasonal contrast in rates is not as strong as for NH snow cover, which shows no trends in autumn
14 and large trends in spring and early summer.
 - 15 • Over wide areas, there is robust evidence with high agreement that permafrost temperatures have
16 increased by up to 3°C during the past three decades in response to increased air temperature and
17 changing snow cover. However, in some regions, permafrost temperatures show little change, or
18 even slight decrease. Generally, the temperature increase for colder permafrost has been greater
19 than for warmer permafrost. Significant permafrost degradation has occurred in the Russian
20 European North (robust evidence high agreement), where: where taliks have developed in the
21 discontinuous permafrost zone; permafrost with thickness of 10 to 15 m completely thawed in some
22 regions over the period 1975–2005; the southern limit of the permafrost boundary moved north by
23 about 80 km; and the boundary of the continuous permafrost moved north by 15–50 km.
 - 24 • The thickness of the active layer (that portion of the soil above permafrost that thaws in summer
25 and re-freezes in winter) has increased by up to 90 cm since the 1980s, although this change varies
26 from a few centimetres to tens of centimetres. In some areas, especially in northern North America,
27 the active layer thickness (ALT) shows large inter-annual variations and no significant trend. The
28 in-situ measurements and satellite data show significant surface subsidence over permafrost in the
29 past two to three decades. Globally, the thickness and duration of seasonally-frozen ground has
30 responded to increases in air temperature. For example, in Russia over the period 1930–2000, it
31 decreased by about 32 cm, and in western China it decreased by 20 to 40 cm since the early-1960s.
32
33
34

4.1 Introduction

The cryosphere is the collective term for the components of the Earth system that comprise a substantial fraction of water in the frozen state (Table 4.1). The cryosphere comprises several components: snow, river and lake ice; sea ice; glaciers in mountainous regions, ice caps and ice sheets; and finally, frozen ground which exists, both on land and beneath the oceans (Figure 4.1). The lifespan of each component is very different. River and lake ice, for example, are transient features that generally do not survive from winter to summer; sea ice advances and retreats with the seasons but especially in the Arctic can survive to become multi-year ice, lasting several years: the East Antarctic ice sheet, on the other hand, is believed to have survived for more than 30 Million years (DeConto and Pollard, 2003). Nevertheless, all components of the cryosphere are inherently sensitive to changes in surface temperature and precipitation, and hence to a changing climate (see Chapter 2).

[INSERT FIGURE 4.1 HERE]

Figure 4.1: The cryosphere in the Northern and Southern Hemispheres in polar projection. The map of the Northern Hemisphere shows the sea ice cover during minimum extent (9th September 2011). The yellow line is the average location of the ice edge (15% ice concentration) for the yearly minima for 1979 to 2011. Areas of continuous permafrost are shown in darker pink, discontinuous permafrost in lighter pink. The shaded area over land and permafrost shows snow cover as derived from MODIS data (July 2009 to March 2010) with the greatest extent during that period represented by the black line. The Greenland ice sheet (blue/grey) and locations of glaciers (yellow) are also shown, but the glaciers within the ice sheet are shown as part of the ice sheet. The map of the Southern Hemisphere shows approximately the maximum sea ice cover during an austral winter (9th September 2011). The yellow line shows the average ice edge (15% ice concentration) during maximum extents of the sea ice cover for 1979 to 2011. Some of the elements (e.g., some glaciers and snow) located at low latitudes are not visible in this projection (see Figure 4.8). The source of the data for sea ice, permafrost, snow and ice sheet are datasets held at the National Snow and Ice Data Center (NSIDC), University of Colorado, on behalf of the North American Atlas, Instituto Nacional de Estadística, Geografía e Informática (Mexico), Natural Resources Canada, U.S. Geological Survey, Government of Canada, Canada Centre for Remote Sensing and The Atlas of Canada. Glacier outlines were derived from multiple datasets (Weidick et al., 1992; Zheltyhina, 2005). Figure courtesy of the NASA Visualization Group.

Changes in the long-lived components of the cryosphere (e.g., glaciers) are the result of an integrated response to climate, and they are often referred to as ‘natural thermometers’, but as our understanding of the complexity of this response has grown, it is becoming increasingly clear that elements of the cryosphere should rather be considered as ‘natural climate-meters’, responsive not only to temperature, but also to other climate variables (e.g., precipitation). However, it remains the case that the conspicuous and widespread nature of changes in the cryosphere (in particular; sea ice, glaciers and ice sheets) means these changes are frequently used emblems of the impact of changing climate, and for this reason, it is imperative that we understand the context of current change within the framework of past changes and natural variability.

The cryosphere is, however, not simply a passive indicator of climate change; changes in each component of the cryosphere have a significant and lasting impact on physical, biological and social systems. Ice sheets and glaciers exert a major control on global sea level (see WGII Chapter 5), and the loss of glaciers may have direct impacts on water resources used by many populations (see WGII Chapter 24). Similarly, reduced sea ice extent has altered, and in future may continue to alter, ocean circulation, ocean productivity and regional climate, and will have direct impacts on shipping and mineral and oil explorations (see WGII Chapter 24). Furthermore, declines in snow cover and sea ice will tend to amplify regional warming through the ice albedo feedback effect (see Chapter 9). Finally, changes in frozen ground (in particular, permafrost) will damage some vulnerable Arctic infrastructure (see WGII, Chapter 28), and could substantially alter the carbon budget through the release of methane (see Chapter 6).

Since the AR4, substantial progress has been made throughout cryospheric research. Satellite technologies now permit estimates of regional and temporal changes in the volume and mass of the ice sheets. The longer time-series now available enable more accurate assessments of trends and anomalies in sea ice cover and rapid identification of unusual events such as the dramatic decline of perennial sea ice in 2007. Similarly, sea ice thickness can now be measured using satellite altimetry, allowing pan-Arctic measurements of changes in volume and mass. A nearly complete glacier inventory covers now almost all glaciers worldwide (42% in AR4) and allows for better estimates of the total ice volume. Remote sensing measurements of regional glacier volume change are also now available and modelling of glacier mass change has improved

considerably. Finally, fluctuations in the cryosphere in distant and recent past have been mapped with increasing certainty, demonstrating the potential for rapid loss, compared to slow recovery, particularly when related to sea level rise.

This chapter describes the current state and the observed variability and change of the cryosphere with a focus on recent improvements in understanding, addressing each of the important components of the cryosphere in turn. Although observed trends are presented, projections of future cryospheric changes are discussed elsewhere (e.g., Chapter 13). Earlier IPCC reports used cryospheric terms that have specific scientific meanings (see Cogley et al., 2011), but have rather different meanings in everyday language. To avoid confusion, this chapter uses the term “glaciers” for what was previously termed, “glaciers and ice caps” (e.g., Lemke et al., 2007). For the largest glaciers, those covering Greenland and Antarctica, we use “ice sheets”. For simplicity, we use units such as Gigatonnes (Gt, 10^9 tonnes, or 10^{12} kg). One Gt is approximately equal to one cubic kilometre of water (1.1 km^3 of ice), and 362 Gt of ice removed from the land and immersed in the oceans will cause roughly 1 mm of global sea level rise (Lemke et al., 2007).

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

Ice on Land	Percent of Global Land Surface ^j	Sea Level Equivalent ^k (m)
Antarctic ice sheet ^a	9.4%	58
Greenland ice sheet ^b	1.2%	7.4
Glaciers ^c	0.5 %	0.6
Terrestrial Permafrost ^d	9–14%	Not applicable
Seasonally frozen ground ^e	33%	Not applicable
Seasonal snow cover (seasonally variable) ^f	1.3% to 30.6%	0.001–0.01
Total	30.6% to 57.9%	64.6 m
Ice in the Ocean	Percent of Global Ocean Area ^j	Volume ^l (10^3 km^3)
Antarctic ice shelves ^g	0.21%	~761
Antarctic sea ice (seasonally variable)	0.8% to 5.2%	4.5–19.0
Arctic sea ice (seasonally variable)	1.7% to 3.9%	18.0–35.0
Sub-sea permafrost ^h	~0.6%	Not applicable
Total	5.1% to 7.3%	37.7 to 40.2

Notes:

(a) (Fretwell et al., Submitted). Area is 13.924 Mkm^2

(b) (Griggs and Bamber, 2011b)

(c) (Arendt et al., 2012), includes glaciers around Greenland and Antarctica

(d) (Gruber, 2012) this excludes permafrost under the Antarctic ice sheet

(e) (Zhang et al., 2003); excludes Southern Hemisphere

(f) (Lemke et al., 2007)

(g) Values derived from published data (Griggs and Bamber, 2011a)

(h) Few estimates of the area of sub-sea permafrost exist in the literature. The estimate shown, to which significant uncertainty is attached, arises from a map given by (Osterkamp, 2001) and various citations, including personal communications, cited therein.

(j) Assuming a global land area of 147.6 Mkm^2 , and ocean area of 362.5 Mkm^2

(k) Assuming an ice density of 917 kg m^{-3} , a seawater density of $1,028 \text{ kg m}^{-3}$, with seawater replacing ice currently below sea level

(l) Calculated assuming average Antarctic austral summer (winter) thicknesses of 1.0 (1.5) m, and average Arctic boreal summer (winter) thickness of 2.5 (3.0) m (Kwok et al., 2009)

4.2 Sea Ice

4.2.1 Background

1 Sea ice is an important component of the climate system. A sea ice cover on the ocean changes the surface
2 albedo, insulates the ocean from heat loss, and provides a barrier to the exchange of momentum and gases
3 such as water vapour and CO₂ between the ocean and atmosphere. Salt ejected by growing sea ice alters the
4 density structure and modifies the circulation of the ocean. Regional climate changes affect the sea ice
5 characteristics and those changes can feed back on the climate system, both regionally and globally. Sea ice
6 is also a major component of polar ecosystems; plants and animals at all trophic levels find a habitat in, or
7 associated with, sea ice.

8
9 Most sea ice exists as pack ice, and wind and ocean currents drive the drift of individual pieces of ice (called
10 floes). Divergence and shear in sea ice motion create areas of open water where, during colder months, new
11 ice can quickly form and grow. On the other hand, convergent ice motion causes the ice cover to thicken by
12 deformation. Two relatively thin floes colliding with each other can “raft”, stacking one on top of the other
13 and thickening the ice. When thicker floes collide, thick ridges may be built from broken pieces, with a
14 height above the surface (ridge sail) of 2 m or more, and a much greater thickness (~10 m) and width below
15 the surface (ridge keel).

16
17 Sea ice thickness also increases by basal freezing during winter months. But the thicker the ice becomes the
18 more it insulates heat-loss from the ocean to the atmosphere and the slower the basal growth is: there is an
19 equilibrium thickness for basal ice growth that is dependent on the air temperature, and heat from the deep
20 ocean below. Snow cover lying on the surface of sea ice provides additional insulation, and also alters the
21 surface albedo and aerodynamic roughness. But also, and particularly in the Antarctic, a heavy snow load on
22 thin sea ice can depress the ice surface and allow seawater to flood the snow. This saturated snow layer
23 freezes quickly to form “snow ice” (see FAQ 4.2).

24
25 Because sea ice is formed from seawater it contains sea salt, mostly in small pockets of concentrated brine.
26 The total salt content in newly formed sea ice is only 25–50% of that in the parent seawater, and the residual
27 salt ejected as the sea ice forms alters ocean water density and stability. The salinity of the ice decreases as it
28 ages, particularly for multiyear ice where melt ponds can form on the surface in summer and subsequently
29 drain through and flush the ice. The salinity of sea ice affects its mechanical strength, its thermal properties
30 and its electrical properties – the latter being very important for remote sensing.

31
32 Geographical constraints play a dominant (but not exclusive) role in determining the quite different
33 characteristics of sea ice in the Arctic and the Antarctic (see FAQ 4.2). This is one of the reasons why the sea
34 ice in is changing differently in the north and south. We also have much more information on Arctic sea ice
35 thickness than we do on Antarctic sea ice thickness, and so discuss Arctic and Antarctic separately in this
36 assessment.

37 38 **4.2.2 Arctic Sea Ice**

39
40 Regional sea ice observations, which span more than a century, have revealed significant inter-annual
41 changes in sea ice coverage (Walsh and Chapman, 2001). Since the advent of satellite passive microwave
42 imaging systems, which now provide more than 32 years of continuous coverage, it has been possible to
43 monitor the entire extent of sea ice with a temporal resolution of just a few days. A number of procedures
44 have been used to convert the observed microwave brightness temperature into sea ice concentration – the
45 fractional area of the ocean covered by ice (Comiso and Nishio, 2008; Markus and Cavalieri, 2000) – and
46 thence to derive sea ice extent and area. Sea ice extent being defined as the integral sum of ice covered areas
47 with concentrations of at least 15%, while ice area is the product of the ice concentration and area of each
48 data element within the ice extent. Ice concentrations and trends in extent and area derived from different
49 procedures are generally consistent (Parkinson and Comiso, 2008), but the results presented in this report are
50 based on a single technique (Comiso and Nishio, 2008).

51
52 Arctic sea ice cover is seasonal, with the average ice extent varying between about 6×10^6 km² in the
53 summer and 15×10^6 km² in the winter (Comiso and Nishio, 2008; Gloersen et al., 1992). The summer ice
54 cover is confined to mainly the Arctic basin, while winter sea ice reaches as far south as 44°N, into the
55 peripheral seas. At the end of summer, Arctic ice cover consists primarily of thick, old and ridged ice types.
56 Interannual variability is largely determined by the extent of the ice cover in the peripheral seas in winter,
57 and by the ice cover that survives the summer melt in the Arctic basin.

4.2.2.1 Total Arctic Sea Ice Extent and Concentration

Figure 4.2 (derived from passive microwave data) shows both the seasonality of the Arctic sea ice cover, and the large decadal changes that have occurred over the last 32 years. Typically, Arctic sea ice reaches its maximum seasonal extent in February or March while the minimum occurs in September at the end of summer-melt. Decadal changes in Arctic ice extent are more pronounced in summer than in winter. The change in winter extent between 1979–1988 and 1989–1998 was negligible. Between 1989–1998 and 1999–2008, there was a decrease in winter extent of around $0.6 \times 10^6 \text{ km}^2$. This can be contrasted to a decrease in ice extent at the end of the summer (September) of $0.5 \times 10^6 \text{ km}^2$ between 1979–1988 and 1989–1998, followed by a further decrease of $1.2 \times 10^6 \text{ km}^2$ between 1989–1998 and 1999–2008. Figure 4.2 also shows that the change in extent from 1979–1988 to 1989–1998 was significant mainly in spring and summer while the change from 1989–1998 to 1999–2008 was significant during all seasons. The largest interannual changes occur during the summer minima when only the thick components (called perennial ice) survives the summer melt (Comiso, 2011a; Comiso et al., 2008).

Changes have been large in the last three years: the average extent for 2009–2011 was less than in earlier periods in all seasons, especially summer. The summer minimum extent was at a record low in 2007 (Comiso et al., 2008; Stroeve et al., 2007). The ice extents for each day of the growth season from December 2010 to March 2011 were significantly lower than those in previous years, and the values for most days were at record lows for the satellite era. The 2011 extent in the spring and summer was comparable to the 2007 record low.

Although relatively short as a climate record, the 32-year satellite record is long enough to allow determination of significant and consistent trends between monthly anomalies of ice extent, area and concentration (i.e., difference between the monthly and the averages over the 32-year record). The trends in ice concentration for the winter, spring, summer and autumn for the period November 1978 to December 2011 are shown in Figure 4.2. The seasonal trends for different regions, except the Bering Sea, are predominantly negative. Ice cover changes are relatively large in the eastern Arctic basin in winter and spring, while in the western basin they are more pronounced in summer and spring. Changes also occur in the peripheral seas and near the sea ice edge.

From the monthly anomaly data, the trend in sea ice extent in the Northern Hemisphere for the period from November 1978 to May 2012 is $-3.9 \pm 0.2\%$ per decade (see FAQ 4.2). The trends for different regions vary greatly, ranging from $+4\%$ per decade in the Bering Sea to -8% per decade in the Greenland Sea. This large spatial variability is associated with the complexity of the atmospheric circulation system as influenced by the Arctic Oscillation (Thompson and Wallace, 1998). The trends also differ with season (Comiso, 2011a): for the entire Northern Hemisphere, the trends in ice extent are -2.3 ± 0.4 , -2.3 ± 0.4 , -5.9 ± 0.7 , and $-6.5 \pm 0.9\%$ per decade in winter, spring, summer and autumn, respectively. The corresponding trends in ice area are -2.8 ± 0.4 , -2.7 ± 0.4 , -7.2 ± 0.8 , and $-7.3 \pm 0.9\%$ per decade. Similar results were obtained by (Cavalieri and Parkinson, 2012) using data from 1979 to 2010. The trends for ice extent and ice area are comparable except in the summer and autumn when the trend in ice area is significantly more than that in ice extent. This is due in part to increasing open water areas within the pack that may be caused by more frequent storms and more divergence in the summer (Simmonds et al., 2008). The trends are larger in the summer and autumn mainly because of the rapid decline of the multiyear ice cover as discussed below.

[INSERT FIGURE 4.2 HERE]

Figure 4.2: (a) Plots of decadal averages of daily sea ice extent in the Arctic (1979 to 1988 in red, 1989 to 1998 in blue, 1999 to 2008 in gold) and a five-year average daily ice extent from 2009 to 2011; ice concentration trends (1979–2011) in (b) winter, (c) spring, (d) summer and (e) autumn (Comiso and Nishio, 2008).

[INSERT FIGURE 4.3 HERE]

Figure 4.3: (a) Yearly and (b) seasonal ice extent in the Arctic using averages of mid-month values derived from in situ and satellite data from 1870 to 2011 (updated from, Walsh and Chapman, 2001). The yearly and seasonal averages for the period from 1979 to 2011 are from satellite passive microwave data (triangle fonts) which have also been updated.

4.2.2.2 Longer Records of Arctic Ice Extent

1 Since 1979, satellite data have provided consistent near-daily mapping of the spatial distribution of the
2 global sea ice cover, however, the variability of the sea ice cover prior to the satellite era is also of great
3 interest. There have been several studies, some based on regional in situ observations taken from ships or
4 aerial reconnaissance (e.g., Polyakov et al., 2003; Walsh and Chapman, 2001) while others on terrestrial
5 proxies (e.g., Fauria et al., 2010; Kinnard et al., 2011). Polyakov et al. (2003) focused their studies on the
6 marginal seas near the Russian coastline using ice extent data from 1900 to 2000 and found a low frequency
7 multidecadal oscillation near the Kara Sea that shifted to a dominant decadal oscillation in the Chukchi Sea.
8 A more comprehensive database compiled by Walsh and Chapman (2001) covered the entire Arctic showed
9 very little interannual variability until the last three to four decades. For the period 1901 to 1998, their results
10 show a summer mode that includes an anomaly of the same sign over nearly the entire Arctic that captures
11 the sea-ice trend determined from recent satellite data.

12
13 Figure 4.3 shows an updated data set with longer time coverage (i.e., 1870 to 2011) that is more robust since
14 it includes additional historical data (e.g., from Danish meteorological stations). The updated data also
15 includes an update of the satellite passive microwave (PM) data that were used for the more recent part of
16 the data set. A comparison of this updated data set with that originally reported by Walsh and Chapman
17 (2001) shows similar interannual variability that is dominated by a nearly constant extent of the annual ice
18 cover from 1870 to the 1950s. An important factor that may have contributed to the lack of significant
19 interannual variability during that period is the heavy use of climatology to fill gaps, which would have
20 masked the natural variability. Since the 1950s, these data sets show a consistent decline in the sea ice cover
21 that is relatively moderate during the winter but more dramatic during the summer months. The drastic
22 reduction of the sea ice extent from 1978 to 1979 is in part due to the change from one data set to another
23 (i.e., ground observations to satellite data). Kinnard et al (2011) and Fauria et al (2010) made use of terrestrial
24 proxies to reconstruct sea ice extent data and observed that the decline of sea ice in the last few decades is
25 unprecedented over the past 1450 years. Again, the uncertainties of such studies are likely very high.

26 27 **[INSERT FIGURE 4.4 HERE]**

28 **Figure 4.4:** Yearly perennial (blue) and multiyear (green) ice extent (a) and ice area (b) in the Central Arctic for each
29 year from 1979 to 2011 as derived from satellite passive microwave data. Perennial ice values are derived from summer
30 ice minimum values, while the multiyear ice values are averages of December, January and February data. The gray
31 lines (after 2002) are derived from AMSR-E data (Comiso, 2011b).

32 33 *4.2.2.3 Multiyear/Seasonal Ice Coverage*

34
35 The winter extent and area of the perennial and multiyear ice cover in the Central Arctic (i.e., excluding
36 Greenland Sea multiyear ice) for the period 1979–2010 are shown in Figure 4.4. Perennial ice is that which
37 survives the summer, and the extent at summer minimum has been used as a measure of its coverage
38 (Comiso, 2002). Multiyear ice (as defined by World Meteorological Organization) is ice that has survived at
39 least two summers. Generally, that ice is less salty and has a distinct microwave signature that differs from
40 the seasonal ice, and thus can be monitored with satellite radiometers (Comiso, 2011b; Johannessen et al.,
41 1999; Zwally and Gloersen, 2008).

42
43 Figure 4.4 shows large but similar interannual variability for perennial and multiyear ice for the period 1979
44 to 2011 (2012 for multiyear ice). The extent of the perennial ice cover, which was about $8 \times 10^6 \text{ km}^2$ in the
45 early-1980s, decreased to about $5 \times 10^6 \text{ km}^2$ in the latter part of the 2000s. Similarly, the multiyear ice extent
46 decreased from about $6.2 \times 10^6 \text{ km}^2$ in the 1980s to about $3.5 \times 10^6 \text{ km}^2$ in the late-2000s. The trends in
47 perennial ice extent and ice area were strongly negative at -12.7 ± 1.5 and $-14.3 \pm 1.5\%$ per decade
48 respectively. These values indicate an increased decline from the -9% per decade reported by Comiso
49 (Comiso, 2002) for the 1979 to 2000 period. The trends in multiyear ice extent and area are even more
50 negative, at -17.1 ± 2.0 and $-19.1 \pm 2.4\%$ per decade, respectively, for the period from 1979 to 2012
51 (Comiso, 2011a). The higher negative trend in ice area compared to that in ice extent indicates that the
52 average ice concentration of multiyear ice in the Central Arctic has also been declining. The rate of decline
53 in the extent and area of multiyear ice cover is consistent with the observed decline of old ice types from the
54 analysis of ice drift and ice age by Maslanik et al. (2007), confirming that older and thicker ice types in the
55 Arctic have been declining significantly. The higher negative trend for the thicker multiyear ice area than
56 that for the perennial ice area implies that the average thickness of the ice, and hence the ice volume, has also
57 been declining.

1
2 Drastic changes in the multiyear ice coverage from QuikScat (scatterometer) data, validated using high
3 resolution SAR data (Kwok, 2004; Nghiem et al., 2007), have also been reported. Some of these changes
4 have been attributed to the near zero replenishment of the Arctic multiyear ice cover during the summer
5 (Kwok, 2007).

6 7 *4.2.2.4 Ice Thickness and Volume*

8
9 For the Arctic there are several different techniques available for estimating the ice thickness distribution.
10 These show strong and broadly consistent decreases in Arctic sea ice thickness over recent years.

11 12 *4.2.2.4.1 Submarine ice draft*

13 Data collected by upward-looking sonar on submarines operating beneath the Arctic pack ice provided the
14 first evidence of 'basin-wide' decreases in ice thickness (Wadhams, 1990). Sonar measurements are of draft
15 (the submerged portion of sea ice), which is converted to thickness by assuming an average density for the
16 measured floe, including its snow cover. Rothrock et al. (1999) found that ice draft in the mid-1990s was less
17 than that measured between 1958 and 1977 in each of six locations within the basin. The change was least (–
18 0.9 m) in the Beaufort and Chukchi Seas and greatest (–1.7 m) in the Eurasian Basin (with an estimated
19 overall error of less than 0.3 m). The decline averaged about 42% of the average 1958 to 1977 thickness.
20 This average decline matched the decline measured in the Eurasian Basin between 1976 and 1996 using UK
21 submarine data (Wadhams and Davis, 2000), which was 43%.

22
23 A subsequent analysis (Rothrock et al., 2008) used a much richer data set from 34 submarine cruises within a
24 data release area that covered almost 38% of the area of the Arctic Ocean, rather than just select locations.
25 These cruises were equally distributed in spring and autumn over a 25-year period from 1975 to 2000.
26 Multiple regression was employed to separate the interannual change (Figure 4.7), the annual cycle, and the
27 spatial distribution of draft. They show that the annual mean ice draft declined from a peak of 3.1 m in 1980
28 to a minimum of 2.0 m in 2000, a decrease of 1.1 m (1.2 m in thickness). Over the period, the steepest rate of
29 decrease is -0.08 m yr^{-1} in 1990. The most recent submarine data, reported by (Wadhams et al., 2011), found
30 that tracks north of Greenland repeated between the winters of 2004 and 2007 showed a continuing change
31 towards less multiyear ice.

32 33 *4.2.2.4.2 Satellite freeboard and thickness*

34 It has been demonstrated that satellite altimetry techniques are now capable of mapping sea ice freeboard to
35 provide a spatially comprehensive estimates of the distribution of Arctic sea ice thickness. Like ice draft,
36 satellite measured freeboard (the floating portion of sea ice) is converted to thickness, assuming an average
37 density of the ice and snow cover. The principal challenges to accurate thickness estimation are in the
38 discrimination of ice and open water, and in estimating snow cover thickness.

39
40 Radar altimeters on the ESA, ERS and Envisat satellites have provided circum-Arctic observations south of
41 81.5°N . Even though the ERS-1 estimates of ice thickness (Laxon et al., 2003) show a downward trend in ice
42 thickness, the high variability and short time-series (1993–2001) indicate that the trend in a region of mixed
43 seasonal and multiyear ice cannot be considered as significant. Their data also reveal a realistic geographic
44 variation in thickness (increasing from about 2 m near Siberia to 4.5 m off the coasts of the Canadian
45 Archipelago and Greenland). Envisat observations between 2002 and 2008 showed a large decrease (0.25 m)
46 following the September 2007 ice extent minimum (Giles et al., 2008b). However, this decline in thickness
47 was regionally confined to the Beaufort and Chukchi Seas, with no significant changes found in the eastern
48 Arctic. Results from the NASA ICESat laser altimeter over the same region also show thinning between
49 2007 and 2008, although the ICESat retrievals suggest stronger interannual variability Figure 4.7 (Kwok,
50 2009). A large decrease in thickness due to the 2007 record minimum in summer ice is clearly seen in both
51 the radar and lidar thickness estimates.

52
53 The lidar altimeter on ICESat (which ceased operation in 2009) provided a broader picture of the Arctic
54 Basin as the coverage extends to 86°N . ICESat thickness estimates also provided seasonal contrast between
55 the spring and autumn. Data from five winter ICESat campaigns between 2003 and 2008 show thinning and
56 volume loss of the Arctic Ocean ice cover (Kwok, 2009). Regions covered by multiyear ice thinned by ~ 0.6
57 m over four years (Figure 4.5), while the average thickness of the first-year ice in mid-winter (~ 2 m),

1 exhibited a negligible trend. Over the period, the ICESat estimates show that average winter sea ice volume
2 inside the Arctic Basin was $\sim 14,000 \text{ km}^3$. In the four years since 2005, the total multiyear ice volume in the
3 winter experienced a net loss of 6300 km^3 ($>40\%$). The large volume loss of $-1237 \text{ km}^3 \text{ yr}^{-1}$ (October), while
4 highlighting the rapid changes during the short ICESat record (2004–2008), can be put within the context of
5 a more moderate multidecadal annual loss of $-280 \pm 100 \text{ km}^3 \text{ yr}^{-1}$ (1979–2010) from a sea ice reanalysis
6 study (Schweiger et al., 2011). With a 42% decline in the area of multiyear sea ice over this period, first-year
7 ice became the dominant ice type of the Arctic Ocean, increasing in both area and volume.

8 9 **[INSERT FIGURE 4.5 HERE]**

10 **Figure 4.5:** The distribution of winter sea ice thickness in the Arctic and the trends in average, first-year (FY) ice, and
11 multiyear (MY) ice thickness derived from ICESat records, 2004–2008 (Kwok, 2009).

12 13 4.2.2.4.2 *Airborne electro-magnetic (EM) sounding*

14 EM sounding measures the distance between an EM instrument and the ice/water interface, and provides
15 another approach to measure ice thickness. The technique is based on the amplitude and phase of a
16 secondary EM field induced in the seawater relative to the primary field generated by the instrument
17 deployed on an airborne platform. Repeat EM surveys, even though limited in time and space, have provided
18 a regional view of the changing ice cover. From repeat ground-based and helicopter-borne EM surveys, Haas
19 et al. (2008) found significant thinning in the region of the Transpolar Drift. Between 1991 and 2004, modal
20 thicknesses decreased from 2.5 m to 2.2 m, and then dropped to 0.9 m in 2007. Mean thicknesses also
21 decreased strongly. This thinning was associated with reductions of the age of the ice, and replacement of
22 second-year ice by first-year ice in 2007 as seen in satellite observations. Airborne EM measurements
23 performed in the Lincoln Sea since 2004 (Haas et al., 2010) within a latitudinal band between 83°N and
24 84°N showed some of the thickest ice in the Arctic, with mean and modal thicknesses of more than 4.5 m
25 and 4 m, respectively. Since 2008, mean and modal thicknesses have decreased to less than 4.6 m and 3.5 m,
26 which is most likely related to the narrowing of the remaining band of old ice along the northern coast of
27 Canada.

28 29 4.2.2.5 *Arctic Sea Ice Drift*

30
31 Pack ice motion influences ice mass: locally, through deformation and creation of open water areas;
32 regionally, through advection of ice from one area to another; and globally, through export of ice from polar
33 seas to lower latitudes where it melts. The drift and deformation of sea ice is primarily forced by winds and
34 ocean currents, but depends also on ice strength, top/bottom surface roughness, and regional ice
35 concentration. On time scales of days to weeks, winds are responsible for most of the variance in sea ice
36 motion. On longer time scales, the patterns of ice motion follow surface currents and the evolving patterns of
37 wind forcing. Changes in ice drift affect impacts surface heat and mass balance of sea ice.

38 39 4.2.2.5.1 *Drift speed*

40 Drifting buoys have been used to measure Arctic Ocean ice displacements over the last 29 years. From these
41 data Rampal et al. (2009) found an increase in average drift speed between 1978 and 2007 of $+17 \pm 4.5\%$ per
42 decade in winter and $+8.5 \pm 2.0\%$ per decade in summer. Spreen et al. (2011) used daily satellite ice motion
43 fields, which provide a better regional depiction, to show that between 1992 and 2008, the spatially averaged
44 trend in winter ice drift speed was $+10.6 \pm 0.9\%$ per decade, but varied regionally between -4 and $+16\%$ per
45 decade. Increases in drift speed are seen over much of the Arctic except in areas with thicker ice (Figure 4.7,
46 e.g., north of Greenland and the Canadian Archipelago). The largest increases occurred during the second
47 half of the period, coinciding with the years of rapid ice thinning. From examination of wind fields in
48 atmospheric reanalyses, both of these investigations concluded that the observed spatial trends suggest a
49 weaker and thinner ice cover, especially during the period after 2005, rather than stronger winds.

50 51 4.2.2.5.2 *Ice export*

52 Sea ice export through the Fram Strait, together with growth and melt, is a major component of the Arctic
53 Ocean ice mass balance, annually exporting approximately 10% of the area of Arctic Ocean ice. Over a 31-
54 year satellite record (1979–2010), the mean annual area outflow through Fram Strait was $\sim 699 \pm 112 \times 10^3$
55 km^2 with a peak in 1994/1995 (Kwok, 2009) But there has been no significant decadal trend in Fram Strait
56 area flux in the satellite records. Decadal trend in ice volume export – a more definitive measure of changes
57 in mass balance – is far less certain due to the lack of an extended record of the thickness of ice exported

1 through Fram Strait. Comparison of recent estimates of volume outflow (Spreen et al., 2009) with earlier
2 estimates by Kwok and Rothrock (1999) and Vinje (2001) ($\sim 2,200\text{--}2,900\text{ km}^3$) shows no discernible change.
3

4 4.2.2.5.3 *Export versus in-situ melt*

5 Between 2005 and 2008, more than a third of the thicker and older sea ice lost occurred by transport of
6 multiyear ice (typically found west of the Canadian Archipelago) into the southern Beaufort Sea where the
7 ice was lost to melt during the summer (Kwok and Cunningham, 2010). Uncertainties in the relative
8 contributions of in-basin melt and export to the observed Arctic ice loss remain. It has also been shown that
9 thicker/older MY ice has been exported through Nares Strait (Kwok, 2005), which could account for a small
10 fraction of the volume loss.

11 4.2.2.6 *Time of Arctic Sea Ice Advance, Retreat and Ice Season Duration; Length of Arctic Melt Season*

12 Seasonality collectively describes the annual timing of sea ice advance, retreat and resultant duration. It is
13 derived from the daily satellite ice-concentration records, which have been used to determine, for each year
14 and each satellite pixel location, the day sea ice advanced to that location and the day it retreated from that
15 location. Maps of the timing of sea ice advance, retreat and ice season duration (the time between day of
16 advance and retreat), as well as maps of trends over time are derived from these data (for detail methods see,
17 Parkinson, 2002; Stammerjohn et al., 2008)
18

19 Most regions in the Arctic show trends towards shorter ice season duration. However, the strongest seasonal
20 trends were found in an area extending from the East Siberian Sea to the western Beaufort Sea. Here, over
21 1979/1980–2006/2007, sea ice advance has become 26 ± 7 days later (at $+0.93 \pm 0.25\text{ days yr}^{-1}$), sea ice
22 retreat 35 ± 8 days earlier (at $-1.25 \pm 0.29\text{ days yr}^{-1}$), and ice season duration 59 ± 11 days shorter (at -2.09
23 $\pm 0.40\text{ days yr}^{-1}$). There has been a nearly 2-month lengthening of the summer ice-free season (Stammerjohn
24 et al., 2012).
25

26 The timing of melt onset during spring, and freeze-up in autumn, can be derived from satellite passive
27 microwave brightness temperature (Belchansky et al., 2004; Drobot and Anderson, 2001; Smith, 1998) as the
28 emissivity of the surface changes significantly with snow melt. The length of a melt season is the number of
29 days between the onset of surface melt in spring and the onset of surface freeze in autumn. The amount of
30 solar energy absorbed by the ice cover increases with the length of the melt season. Longer melt seasons
31 with lower albedo surfaces (wet snow, melt ponds, and open water) increase absorption of incoming
32 shortwave and melt, creating a positive feedback. Hudson (2011) estimates that the observed reduction in
33 Arctic sea ice has contributed approximately 0.1 W m^{-2} of additional global radiative forcing, and that an
34 ice-free summer Arctic Ocean will result in a forcing of about 0.3 W m^{-2} .
35

36 The satellite record (Markus et al., 2009) shows trends toward earlier melt and later freeze-up nearly
37 everywhere in the Arctic. Over the last 30 years, the mean melt season over the Arctic ice cover has
38 increased by about 20 days. The largest and most significant trends (at the 99% level) of >10 days per decade
39 are seen in the coastal margins and peripheral seas: Hudson Bay, the East Greenland Sea, the Laptev/East
40 Siberian seas, and the Chukchi/Beaufort seas.
41

42 4.2.2.7 *Arctic Odden and Polynyas*

43 The Odden sea ice feature is a tongue of sea ice, related to cold surface waters, that extends more than 1000
44 km eastward from the normal east Greenland ice edge at about 73°N . Time series (1951–2005) of the Odden
45 ice extent have been analyzed in the context of sea level pressure, surface wind, air temperature, cloud, and
46 energy flux variations using NCEP-NCAR reanalyses (Rogers and Hung, 2008). The Odden was a recurring
47 winter feature in 1966–1972, during the Great Salinity Anomaly (GSA), appeared occasionally in the 1980s
48 and 1990s, but has occurred rarely since 2000.
49

50 High ice production in coastal polynyas (anomalous regions of open water or low ice concentration) over the
51 continental shelves of the Arctic Ocean is responsible for the formation of cold saline water, which
52 contributes to the maintenance of the Arctic Ocean halocline. A new passive microwave algorithm has been
53 used to estimate thin sea ice thicknesses ($<0.15\text{ m}$) in the Arctic Ocean (Tamura and Ohshima, 2011),
54 providing the first circumpolar mapping of sea ice production in coastal polynyas. High sea ice production is
55
56
57

1 confined to most persistent Arctic coastal polynyas, with the highest ice production rate being in the North
2 Water Polynya. Sea ice production in the 10 major Arctic polynyas decreased by 462 km³ between 1992 and
3 2007.

4 4.2.2.8 Arctic Land-Fast Ice

6
7 Fast ice along the Arctic coast is typically grounded in shallow water, with the seaward edge around the 20–
8 30 m isobath (Mahoney et al., 2007). In fjords and confined bays, fast ice extends into deeper water. Since
9 the interannual variability of fast ice coverage and thickness is often high, long-term observations over at
10 least several decades are necessary to detect trends in Arctic fast ice. Depending on the region and
11 observation periods, both significant and non-significant trends have been observed over the past decades.
12 Long-term monitoring near Hopen, Svalbard, revealed thinning of fast ice in the Barents Sea region by 11
13 cm per decade between 1966 and 2007 (Gerland et al., 2008). Between 1936 and 2000, the thickness trends
14 (in May) from five different individual Siberian sites (Kara Sea, Laptev Sea, East Siberian Sea, Chucki Sea)
15 are insignificant (Polyakov et al., 2003). More recently, a composite time series of fast ice thickness from 15
16 locations (recorded between mid 1960s and early 2000s) along the Siberian coast revealed an average rate of
17 (relative) thickness loss of 0.33 cm yr⁻¹, defined by linear trend (Polyakov et al., 2010). At four sites in the
18 Canadian Arctic, Brown and Cote (1992) found no significant trend in ice thickness from observations
19 between the 1950s and the late 1980s. Even though the trend in the fast ice extent near Barrow, Alaska is not
20 significant (Mahoney et al., 2007), relatively recent observations by Mahoney et al (2007) and
21 Druckenmiller et al. (2009) found longer ice-free seasons and thinner fast ice in northern Alaska compared to
22 earlier records (Barry et al., 1979; Weeks and Gow, 1978). As freeze-on happens later, the growth season
23 shortens and the thinner ice breaks up and melts earlier.

24 4.2.3 Antarctic Sea Ice

26
27 The Antarctic sea ice cover is more seasonal than that in the Arctic, with average extent varying from a
28 minimum of about 3×10^6 km² in February to a maximum of about 18×10^6 km² in September (Comiso et
29 al., 2011; Zwally et al., 2002). The relatively small fraction of Antarctic sea ice that survives the summer is
30 found mostly in the Weddell Sea, but with some perennial ice also surviving on the western side of the
31 Antarctic Peninsula and in small patches around the coast. As well as being mostly first-year ice, Antarctic
32 sea ice is also on average thinner, warmer, more saline and more mobile than Arctic ice. These
33 characteristics, which reduce the capabilities of some remote sensing techniques, together with its more
34 distant location from inhabited continents, result in far less being known about the properties of Antarctic sea
35 ice than of that in the north.

36 4.2.3.1 Total Antarctic Sea Ice Extent and Concentration

37
38
39 Figure 4.6 shows the seasonal variability of Antarctic sea ice extent from 32 years of satellite passive
40 microwave data. In contrast to the Arctic, decadal monthly averages almost overlap with each other, and the
41 seasonality of the total Antarctic sea ice cover has not changed much over the period. In winter, the values
42 for the 1999–2008 decade were slightly higher than those of the other decades while in autumn, the values
43 for 1989–1998 and 1999–2008 decades are higher than those of 1979–1988. The plot for 2009–2011 shows
44 more seasonal variability than the decadal plots with relatively high values in late autumn, winter and spring.

45
46 Trend maps for winter, spring, summer and autumn extent are presented in Figure 4.5 b, c, d, and e
47 respectively. The seasonal trends are significant mainly near the ice edge, with the values alternating
48 between positive and negative around Antarctica. Such an alternating pattern has been described previously
49 and associated with the influence of the Antarctic Circumpolar Wave (White and Peterson, 1996). In the
50 Austral winter, negative trends are evident at the tip of the Antarctic Peninsula and also at the opposite side
51 of the continent, while positive trends are prevalent in the Weddell and Ross seas. The patterns in Austral
52 spring are very similar to those of winter while in summer and autumn, negative trends are mainly confined
53 to the Bellingshausen/Amundsen Seas, while positive trends are dominant in the Ross Sea.

54
55 The regression trend in the Antarctic sea ice extent monthly anomalies from November 1978 to May 2012 is
56 slightly positive, at $1.4 \pm 0.2\%$ per decade (see FAQ 4.2). For the seasonal trends in ice extent in winter,
57 spring, summer and autumn Comiso et al. (2011) report -0.8 ± 0.4 , 0.7 ± 0.4 , 1.0 ± 1.4 , and $3.1 \pm 1.3\%$ per

1 decade, respectively. The corresponding trends in ice area are 1.4 ± 0.4 , 1.2 ± 0.4 , 2.1 ± 1.6 , and $4.7 \pm 1.5\%$
2 per decade. The values are all positive except for a negative but insignificant trend in winter extent. The
3 trends are consistently higher for ice area than ice extent, indicating less open water (possibly due to less
4 divergence and storms) within the pack in later years. Similar trend results were reported by Parkinson and
5 Cavalieri (Submitted) using data from 1978 to 2010. The overall inter-annual trends for the various sectors
6 are given in FAQ 4.2, but such trends have to be interpreted carefully because of an atmospheric circulation
7 pattern that is influenced strongly by the Southern Annular Mode and the Antarctic Circumpolar Wave.

8 9 **[INSERT FIGURE 4.6 HERE]**

10 **Figure 4.6:** (a) Plots of decadal averages of daily sea ice extent in the Antarctic (1979 to 1988 in red, 1989 to 1998 in
11 blue, 1999 to 2008 in gold) and average values of daily ice extents in 2009 to 2011; ice concentration trends (1979–
12 2011) in (b) winter, (c) spring, (d) summer and (e) autumn.

13 14 *4.2.3.2 Antarctic Sea Ice Thickness and Volume*

15
16 Since AR4, advances have been made in determining the thickness of Antarctic sea ice, particularly in the
17 use of ship-based observations and satellite altimetry. However there remains an ongoing lack of data
18 providing routine global coverage for monitoring and no information on large-scale Antarctic ice thickness
19 change. Worby et al. (2008) compiled 25 years of ship-based data from 83 Antarctic voyages on which
20 routine observations of ice and snow properties were made. Their compilation included a gridded data set
21 that reflects the regional differences in sea ice thickness. A subset of these ship observations, and ice charts,
22 was used by DeLiberty et al. (2011) to estimate the annual cycle of sea ice thickness and volume in the Ross
23 Sea, and to investigate the relationship between ice thickness and extent. They found that maximum sea ice
24 volume was reached later than maximum extent. While ice advects to the northern edge and melts, the
25 interior is supplied with ice from higher latitudes and continues to thicken by thermodynamic growth and
26 deformation. Satellite retrievals of freeboard and thickness in the Antarctic (Mahoney et al., 2007; Xie et al.,
27 2011; Zwally et al., 2008) are under development but progress is limited by our present knowledge of snow
28 thickness and the paucity of suitable validation data sets.

29 30 *4.2.3.3 Antarctic Sea Ice Drift*

31
32 Using a 19-year dataset (1992–2010) of satellite-tracked sea ice motion, Holland and Kwok (Submitted)
33 found large and statistically significant decadal trends in Antarctic ice drift that in most sectors are caused by
34 changes in local winds. These trends suggest acceleration of the Ross Gyre and deceleration of the Weddell
35 Gyre. The changes in meridional ice export affect freshwater budgets, implying changes in brine release in
36 near the Antarctic coast. This is consistent with the increase of $30,000 \text{ km}^2 \text{ yr}^{-1}$ in the net export of sea ice
37 from the Ross shelf/coastal polynya region between 1992 to 2008 (Comiso, 2011b). Assuming an average
38 thickness of 0.6 m, this is equivalent to a volume transport of $20 \text{ km}^3 \text{ yr}^{-1}$ which is similar to the rate of
39 production in the Ross Sea coastal polynya region for the same period discussed below.

40 41 *4.2.3.4 Time of Antarctic Sea Ice Advance, Retreat and Ice Season Duration*

42
43 There have been contrasting regional patterns of change in the Antarctic ice duration. In the northeast and
44 west Antarctic Peninsula and southern Bellingshausen Sea region, later ice advance (48 ± 11 days later) and
45 earlier retreat (35 ± 9 days earlier) has shortened the ice season by 83 ± 23 days over the period 1979/1980–
46 2006/2007 (a trend of $-2.97 \pm 0.81 \text{ days yr}^{-1}$). The opposite is true in the adjacent western Ross Sea, where
47 the ice season has lengthened by 57 ± 13 days (at $2.02 \pm 0.46 \text{ days yr}^{-1}$) due to earlier advance (29 ± 7 days
48 earlier) and later retreat (28 ± 6 days later) (Stammerjohn et al., 2012). The pattern of change across the
49 extensive East Antarctic sector is more complex, with mixed signals on regional-to-local scales and areas of
50 strongly positive ($+2$ to $+3 \text{ days yr}^{-1}$) and negative trends (-2 to -3 days yr^{-1}) in ice season duration
51 occurring in near juxtaposition in certain regions such as Prydz Bay (Massom et al., Submitted).

52
53 The magnitude of the seasonal/regional sea ice changes in both the Arctic and Antarctic point to strong
54 positive feedbacks. However, the asymmetric seasonal response appears to indicate different sensitivities and
55 underlying mechanisms. For example, an earlier sea ice retreat is consistent with an ice albedo feedback
56 (e.g., Perovich et al., 2008), while a delayed sea ice advance is consistent with an enhanced ocean heat
57 feedback (e.g., Steele et al., 2008). Additionally, changes in winds can also affect the ice-edge advance and

1 retreat (Massom et al., 2008). Thus, different atmosphere-ocean forcings and sea ice sensitivities are likely
2 present in different regions and seasons, but we currently lack the data to fully resolve these differences.

3 4 4.2.3.5 *Antarctic Polynyas*

5
6 Polynyas are commonly found along the coasts of Antarctica. An increase in the extent of coastal polynyas
7 in the Ross Sea caused increased ice production that is primarily responsible for the positive trend in ice
8 extent in the Antarctic (Comiso, 2011b). Variability in the ice cover in this region is linked to changes in the
9 Southern Annular Mode (SAM) and secondarily to the Antarctic Circumpolar Wave. Ocean convection that
10 injected relatively warm deep water into the surface layer was observed between 1974–1976 creating the
11 large Weddell Sea Polynya, but since the late-1970s the SAM has been negative or slightly positive,
12 resulting in warmer and wetter condition forestalling the Weddell Polynya (Gordon et al., 2007). A recent
13 study shows that the net ice export equals the polynya ice production of approximately 400 km³ in 1992 for
14 the Ross Sea in Antarctica, with an annual increase in ice production of 21 km³ per year to 2008, whereas the
15 ice production, which is three times less in the Weddell Sea, has a statistically insignificant trend over the
16 same period (Drucker et al., 2011).

17 18 4.2.3.6 *Antarctic Land-Fast Ice*

19
20 Landfast ice (or “fast ice”) occurs in both the Arctic and Antarctic, but is more extensive in the Antarctic.
21 Around East Antarctica, where it typically occurs in narrow coastal bands of varying width up to 150 km
22 from the coast and in water depths of up to 400–500 m, it is generally between 5% (winter) and 30%
23 (summer) of the overall sea ice area (Fraser et al., in press), and a greater fraction of ice volume (Giles et al.,
24 2008a).

25
26 Variability in the distribution and extent of fast ice is sensitive to processes of ice formation and to climate-
27 related processes (such as ocean swell and waves, and strong wind events) that cause the fast ice to break-up.
28 Historical records of Antarctic fast ice extent, such as that of Kozlovsky et al. (1977) covering 0° to 160°E,
29 were limited by sparse and sporadic sampling. But using cloud-free MODIS composite images, Fraser et al.
30 (in press) derive a high resolution time series of landfast sea ice extent along the East Antarctic coast,
31 showing a statistically-significant increase ($1.43 \pm 0.30\% \text{ yr}^{-1}$) between March 2000 and December 2008.
32 There is a strong increase in the Indian Ocean sector (20°E to 90°E, $4.07 \pm 0.42\% \text{ yr}^{-1}$), and a non-significant
33 decrease in the Western Pacific Ocean sector (90°E to 160°E, $-0.40 \pm 0.37\% \text{ yr}^{-1}$). An apparent shift from a
34 negative to a positive trend was noted in the Indian Ocean sector from 2004, which coincided with greater
35 interannual variability. Although significant changes are observed, this record is only 9 years in length.

36 37 4.2.4 *Synthesis of Sea Ice Changes*

38
39 The strong and significant decrease in Arctic sea ice extent and area reported in AR4 has continued, and is
40 accompanied by many other changes in the characteristics of the Arctic sea ice cover. These changes are
41 robust and in high agreement.

42
43 The average decadal extent of the ice has decreased in every season and in every successive decade since
44 satellite observations commenced. The overall trend in extent over the period 1979–2010 has been –4% per
45 decade (-0.47 million square kilometres per decade; Figure 4.7), with larger changes occurring in summer
46 and autumn, and the largest changes of all to the perennial ice (the summer minimum extent; -12.2% per
47 decade) and multiyear ice (more than 2 years old; -15.6% per decade; Figure 4.7). The rate of decrease in ice
48 area has been greater than that in extent because the ice concentration has also decreased (Figure 4.7a).

49
50 The decrease in perennial and multiyear ice coverage has resulted in a strong decrease in ice thickness, and
51 hence in ice volume. Declassified submarine sonar measurements, covering ~38% of the Arctic Ocean,
52 indicate an overall mean winter thickness of 3.64 m in 1980, which decreased by 1.75 m to only 1.89 m
53 during 2009. Between 1975 and 2000, the steepest rate of decrease was -0.08 m yr^{-1} in 1990 compared to a
54 slightly higher winter/summer rate of $-0.10/-0.20 \text{ m yr}^{-1}$ in the five-year ICESat record (2003–2008) (Kwok
55 and Rothrock, 2009). This combined analysis (Figure 4.7b) shows a long-term trend of sea ice thinning that
56 spans five decades. The decrease of thickness means that Arctic ice is becoming increasingly seasonal and
57 because of the time it takes to form multiyear ice, would take at least several years for any recovery.

1
2 The decrease in both concentration and thickness means that the ice has less resistance to wind forcing, and
3 the rate of ice drift has also increased (Figure 4.7e) (Rampal et al., 2009; Spreen et al., 2011). Other
4 significant changes to the Arctic Ocean sea ice include lengthening in the duration of the surface melt on
5 perennial ice of 6 days per decade (Figure 4.7d) and a nearly 2-month lengthening of the ice-free season in
6 the region from the East Siberian Sea to the western Beaufort Sea.

7
8 For Antarctica, the status of change in many of these sea ice characteristics is not known. There has been a
9 small but significant increase in total ice extent of 1.1% per decade between 1979 and 2010, and a greater
10 increase in ice area, indicating an increase in concentration. But there are strong regional differences within
11 this total, with some regions increasing in extent/area and some decreasing. There are also contrasting
12 regions around the Antarctic where the ice-free season has lengthened, and others where it has decreased
13 over the satellite period. There are still inadequate data to make any assessment of changes to Antarctic sea
14 ice thickness and volume.

15 [INSERT FIGURE 4.7 HERE]

16 **Figure 4.7:** Summary of linear decadal trends (red) in: (a) Arctic ice extent and concentration from satellite passive
17 microwave observations (Comiso and Nishio, 2008) ; (b) sea ice thickness from submarine (blue), satellites (black), and
18 in-situ/EM surveys (circles); trend in submarine ice thickness is from multiple regression of available observations to
19 separate the interannual changes, the annual cycle, and the spatial field of thickness within the data release area (Haas et
20 al., 2008; Kwok and Rothrock, 2009); (c) multiyear sea ice coverage from analysis of QuikSCAT (Kwok, 2009); (d)
21 length of melt season (Markus et al., 2009); and (e) satellite-derived drift speed (Spreen et al., 2011).
22
23

24 4.3 Glaciers

25
26 This section considers all perennial surface land ice masses outside of, and disconnected from, the ice sheets
27 of Antarctica and Greenland. Glaciers occur where climate conditions and topographic relief allow snow to
28 accumulate over several years and gradually transform to firn (old snow lasting more than one year) and
29 finally to ice. Under the force of gravity, this ice flows downslope to regions with higher temperatures where
30 various processes of ablation (loss of snow and ice) occur. The relief modifies atmospheric conditions and
31 hence also influences accumulation (gain of snow and ice) and ablation processes. Accumulation is in most
32 regions mainly due to solid precipitation, but also results from refreezing of liquid water (important in polar
33 regions), and re-distribution of snow by wind and avalanches. Ablation is in most regions mainly due to
34 melting with subsequent runoff, but loss of ice by calving into lake or ocean water or sublimation (important
35 in low latitude regions) and loss of wind-blown snow also contribute. The sum of all accumulation and
36 ablation processes determines the mass budget of a glacier. The related energy and mass fluxes are modified
37 by the relief and are directly linked to the atmospheric conditions. Glaciers are sensitive climate indicators as
38 they react to changes in climatic conditions (e.g., temperature and precipitation) by adjusting their size
39 (Figure 4.1, FAQ 4.1). Glaciers are also important seasonal to long-term hydrologic reservoirs [WGII] on a
40 regional scale and a main contributor to sea level rise on a global scale (Section 4.10). The determination of
41 their overall state (extent) and their changes through time is reported in the following.
42

43 4.3.1 Current Glacier Extent and Volume

44
45 The total area covered by glaciers was not precisely known in AR4, resulting in large uncertainties for all
46 related calculations. Since AR4, the formerly used world glacier inventory (WGMS, 1989) was spatially
47 extended by (Cogley, 2009a) and (Radic and Hock, 2010). For AR5, a new globally complete vector dataset
48 of glacier outlines was compiled from a wide range of data sources (Arendt et al., 2012) (Figure 4.8 and
49 Table 4.2). Though glacier outlines in this inventory refer to the past 50 years, the new dataset (version 2.0),
50 referred to as the Randolph Glacier Inventory (RGI), is a key improvement over previous inventories and has
51 been used as a base in several studies assessed here. Glacier-covered areas were determined for 19 regions
52 (modified from , Radic and Hock, 2010) reflecting different physio-geographic zones and supplemented with
53 the percentage of the area covered by tidewater glaciers (which are in direct contact with the ocean) in each
54 region according to Gardner et al. (Submitted). While a new detailed inventory is available for the local (or
55 peripheral) glaciers on Greenland (Rastner et al., 2012), for Antarctica the assessment of the area covered by
56 glaciers is still incomplete. Calculations presented in this chapter refer to ice masses that are spatially
57 completely disconnected from the ice sheet (e.g., on surrounding islands) (Bliss et al., Submitted) as
58 indicated in Figure 4.1.

1
2 Measurements of glacier thickness and corresponding volumes estimates are only available for relatively few
3 glaciers worldwide. The required up-scaling of these measurements to unmeasured glaciers results in
4 considerable uncertainty (Table 4.2). A frequently applied method determines glacier volume from area
5 using a power-law relation and is based on geometric and ice-dynamic considerations (Bahr et al., 1997).
6 New approaches have been developed recently that model the spatial ice thickness distribution based on
7 glacier geometry, digital terrain information and basic glaciological principles (Clarke et al., submitted;
8 Farinotti et al., 2009; Linsbauer et al., 2012). One of these new approaches was applied in a simplified form
9 by Huss and Farinotti (Submitted) (S1 in Table 4.2) to model the volume of all 181,500 glaciers from the
10 Arendt et al. (Arendt et al., 2012) compilation. For a total glaciated area of about 740,000 km², a volume of
11 171,600 km³ (0.43 m sea level equivalent, SLE) was calculated (Table 4.2). Other studies (e.g., Giesen and
12 Oerlemans, 2012; Grinsted, submitted; Radic et al., submitted) used different ways of calculating the
13 volume-area scaling, as well as different glacier areas and entity allocations, and obtained slightly different
14 volumes. Glacier volumes derived by Radić et al. (submitted) (S2a and S2b in Table 4.2) indicate the range
15 of possible values reaching from 734,572 to 786,882 km² for global glacier area and from 140,778 to
16 212,838 km³ (0.35 to 0.53 m SLE) for global volume. The values used in AR4 (area: 795,000 km², volume
17 260,000 km³) were higher than the new numbers, and the values from Radić and Hock (2010) are in good
18 agreement for glaciated area (741,000 km²), but are higher for volume (241,000 km³). About 42% of the
19 glacier covered area is found in the three regions *Alaska* [region 1], *Greenland* [5] and *Antarctic and*
20 *Subantarctic* [19] with another 31% added from *Arctic Canada North and South* [3 and 4], *Svalbard* [7] and
21 the *Russian Arctic* [9] (Table 4.2 and Figure 4.8). The number of glaciers in each region (Table 4.2) can only
22 be seen as a rough indicator, as the count depends on the rules used to divide an ice mass into individual
23 glaciers as well as other factors.
24
25

26 **Table 4.2:** Global glacier number, area, sea level equivalent (SLE) and volume for the 19 RGI regions according to
27 study S1 by Huss and Farinotti (Submitted) and glacier volumes from study S2 by Radic et al. (submitted) using
28 methods (S2a and S2b) and slightly different areas for each region. The percentage area of tidewater glaciers are from
29 Gardner et al. (Submitted).

Nr.	Region Name	Number of Glaciers	Area [km ²]	Percent of total area [%]	Tidewater fraction [%]	SLE S1 [mm]	Volume S1 [km ³]	Volume S2a [km ³]	Volume S2b [km ³]
1	Alaska	24250	90437	12.2	13.0	50.6	20290	30056	19890
2	Western Canada and US	15379	14661	2.0	0	2.5	1021	1275	1275
3	Arctic Canada North	4150	105930	14.3	45.7	86	34526	55545	34716
4	Arctic Canada South	6981	40723	5.5	7.5	24.2	9723	9356	6994
5	Greenland	13906	87099	11.8	35.5	47.1	18896	17130	13588
6	Iceland	291	10981	1.5	0	11	4400	2655	2655
7	Svalbard	2514	33917	4.6	6.5	24.7	9917	9089	6799
8	Scandinavia	1796	2828	0.4	0	0.6	254	201	202
9	Russian Arctic	382	51211	6.9	0	43.2	17340	17778	12240
10	North Asia	3473	2803	0.4	0	0.3	139	274	274
11	Central Europe	3904	2055	0.3	0	0.3	116	139	139
12	Caucasus	1336	1119	0.2	0	0.2	61	68	68
13	Central Asia	30236	64408	8.7	0	12.5	5020	6480	6072
14	South Asia (West)	23260	34256	4.6	0	8.1	3254	4475	3792
15	South Asia (East)	13993	22074	3.0	0	3.3	1320	1852	1803
16	Low Latitudes	6053	4097	0.6	0	0.4	145	231	231
17	Southern Andes	22818	32420	4.4	19.0	16.6	6643	6842	5118
18	New Zealand	3012	1158	0.2	0	0.2	70	79	79
19	Antarctic & Subantarctic	3770	137633	18.6	99.0	95.9	38479	49314	49314
	Total	181504	739820			427.7 ± 57.8	171625 ± 20860	212838	165250

30

[INSERT FIGURE 4.8 HERE]

Figure 4.8: Global distribution of glaciers (yellow) and area covered (size of the circle), sub-divided into the 19 RGI regions (white number) referenced in Table 4.3. The area percentage covered by tidewater glaciers in each region is shown in green. Data from Arendt et al. (2012) and Gardner et al. (Submitted).

4.3.2 Background on Glacier Changes

Glacier changes illustrate climatic changes in a highly visible and also physically understandable way (e.g., Molnia, 2007). Yet the changes occur across a wide range of spatial and temporal scales and it is non-trivial to monitor all of them adequately. Hence, a wide range of observation techniques have been developed to measure changes in glacier length, area, mass and volume regularly. They have individual benefits for specific space and time ranges and are summarized in Table 4.3. If not stated otherwise, we refer in the following to ‘normal’ glaciers, i.e., without peculiarities such as surging (an internally triggered cyclic flow instability), calving, disintegration, or being heavily debris covered.

4.3.2.1 Length Change Measurements

Length changes are usually obtained from annually repeated measurements of the glacier terminus position in the field or from aerial photography and satellite imagery. Past terminus positions are reconstructed from maps, photographs, satellite imagery, paintings and dated moraines (e.g., Davies and Glasser, Submitted; Lopez et al., 2010; Masiokas et al., 2009; Nussbaumer et al., 2011; Rabatel, 2012). Length changes (i.e., front variations) are a smoothed and delayed reaction to climate change, as glacier dynamics modify the response (e.g., Leysinger Vieli and Gudmundsson, 2004). But length changes that occur gradually can also be seen as a strong amplification of a climate forcing (e.g., a temperature increase of 0.1°C over 10 years can result in a retreat of 100 m or more), making them easier to measure than the related temperature change. While for some individual glaciers terminus fluctuations have been reconstructed more than 3000 years back in time (Holzhauser et al., 2005), a much larger number of records go back into the 16th or 17th Century (Zemp et al., 2011, and references therein). These records were used, for example, to determine the contribution of glaciers to global sea level rise (Leclercq et al., 2011) (Section 4.3.4), and for an independent temperature reconstructions at a hemispheric scale (Leclercq and Oerlemans, 2012) over this period. However, some glacier types show front variations that occur independently from the climatic forcing (e.g., calving or surging glaciers) and should be excluded in climate-related analysis (Yde and Pasche, 2010). Length changes are discussed here for the past ca. 150 years over which they are well constrained by direct measurements (Figure 4.9).

4.3.2.2 Area Change Measurements

Glacier area changes are increasingly reported from the comparison of repeat satellite imagery in all parts of the world (WGMS, 2008). Some studies also reconstruct glacier extent at the so-called Little Ice Age (LIA) from mapping of moraines and trimlines which are visible in the field or on aerial photography and satellite images (e.g., Citterio et al., 2009; Davies and Glasser, Submitted) and link them to the records of direct measurements (Leclercq et al., Submitted). Due to the highly variable ice-thickness distribution of glaciers, a direct correlation of area changes with climate changes is difficult to establish. However, the observed geometric changes (e.g., separation of tributaries, emerging rock outcrops) provide indirect evidence for climate-related impacts such as surface lowering (Diolaiuti et al., 2012; Paul et al., 2007; Pelto, 2010). Annual area-loss rates have been reported for many mountain ranges globally and can thus be compared (Figure 4.10). However, as changes in relative area depend (in most regions) on glacier size, it is only possible to compare studies that have analysed entire mountain ranges rather than individual glaciers and thus refer to a regional characteristic. Furthermore, the start and end of the time periods of analyses vary with the data available and results cannot be compared for the same intervals.

4.3.2.3 Volume and Mass Change Measurements

Glacier mass changes can be measured in four ways: Traditionally, seasonal mass gains and losses and, when summed up, the net annual surface mass balance are derived from repeated snow density and snow/ice stake readings on individual glaciers. This laborious method is generally restricted to a limited number of

1 accessible glaciers, and uninterrupted time series spanning more than 40 years are only available from 37
 2 glaciers worldwide (Zemp et al., 2011). Mass loss by calving or sub-glacial ablation is not included in this
 3 method. Uncertainty is introduced firstly by the extrapolation of the point measurements to the entire glacier
 4 and secondly by applying spatial extrapolations to unmeasured glaciers to obtain the overall regional and
 5 global mass changes. At present, it is not possible to objectively quantify all sources of uncertainty in overall
 6 mass budgets extrapolated from single-glacier measurements (Cogley, 2009b). In some cases, for example,
 7 without any sampled glaciers in a particular region there is no better estimate of the regional mass budget
 8 than the corresponding global one. Individual repeat measurements of surface-elevation changes on
 9 individual glaciers but also on regional scales, which are growing in number, provide a second way to
 10 determine overall volume changes (Section 4.3.3 and Figure 4.11). The information is derived from (i)
 11 subtracting digital terrain models (used preferably for smaller glaciers in complex topography), or (ii) repeat
 12 airborne or satellite altimetry (suitable for larger and flatter ice surfaces) of two epochs. The conversion from
 13 volume to mass can cause a major uncertainty over short periods as density information is required but is
 14 generally lacking. A third way to estimate overall mass change is by repeat measurements of the changing
 15 gravity field from satellites (GRACE mission). The coarse spatial resolution (about 300 km) and the
 16 difficulties of separating different mass change signals such as hydrological storage and glacial isostatic
 17 adjustment, (Larsen et al., 2005; Sato et al., 2011) limit this method to regions with large ice extent (Gardner
 18 et al., Submitted). A fourth method is developing with application of models that calculate the mass balance
 19 of individual glaciers or a glacier region. These models either transfer other glacier variables such as length
 20 changes (e.g., Leclercq et al., 2011) or equilibrium line altitudes (e.g., Luethi et al., 2010) into mass changes
 21 or scale from temperature and other meteorological variables using distributed mass balance models (e.g.,
 22 Machguth et al., 2009) or work as glacier by glacier statistically refined approaches (Hirabayashi et al.,
 23 submitted; Hock et al., 2009; Marzeion et al., submitted). They are improving fidelity and physical
 24 completeness and add value to the scarce direct measurements by providing calibrated modelling estimates
 25 for an individual glacier to global scale. Results from the method that determines glacier mass changes as
 26 residuals of a water balance are not considered here since they are made for hydrological basins rather than
 27 for glacier regions.

28

29

30

31

Table 4.3: Overview of methods used to determine changes in glacier length, area and volume along with some typical characteristics. The last three columns provide only indicative numbers. LIA is ‘Little Ice Age’.

Parameter	Method	Techniques	Advantages	Disadvantages	Number of Glaciers	Repeat Interval	Earliest Data
Length	Various	Reconstruction	long temporal extension	dating is difficult	dozens	decadal - centuries	LIA (Holocene)
	Field	in-situ measurement	precise	requires tongue access	hundreds	annual	19th century
	Remote Sensing	Photogrammetric survey	precise	clouds, shadow, expensive	hundreds	annual	20th century
Area	Maps	Cartographic	temporal extension	glacier interpretation can be different	hundreds	decadal	19th century
	Remote sensing	Image processing	global archive, fast assessment	difficult debris cover mapping	thousands	sub-decadal	20th century
Volume	Field	Direct mass balance measurement	precise	laborious	dozens	seasonal	20th century
	Remote sensing	Laser and radar profiling	very precise	only point information	hundreds	annual	21st century
	Remote sensing	DEM differencing	large regions covered, overall volume change	only decadal	thousands	decadal	20th century
	Remote sensing	Gravimetry (GRACE)	provides overall mass changes on a global scale	does not resolve individual glaciers	global	seasonal	21st century

32

33

4.3.3 Observed Changes in Glacier Length, Area and Mass

4.3.3.1 Length Changes

Despite their high variability due to different response times and local conditions (see FAQ 4.1), glacier terminus fluctuations provide a largely homogenous signal (Figure 4.9). The available records of cumulative length changes from several hundred glaciers give robust evidence that glaciers retreated strongly worldwide (up to several kilometres) since their last Holocene maximum extent at the end of the LIA (WGMS, 2008), which occurred, with regional differences, between the 17th to 19th century regionally (Rabatel et al., 2008). The general retreat from this rather extreme position was interrupted in several regions by intermediate phases of stability, or even glacier advances around the 1920s and 1970s but local exceptions exist (UNEP, 2007).

The observed cumulative front variations for glaciers of different size reveal a clear overall trend in terminus retreat, but also intermittent advances that are not globally synchronous (Figure 4.9). In all regions, the fluctuations show a typical pattern, with the largest (flatter) glaciers displaying more or less continuous retreat and strong overall changes, medium-sized (steeper) glaciers showing the decadal fluctuations mentioned above, and smaller glaciers showing a high variability along with overall retreat. This general characteristic of glacier response has to be considered when time series of individual glaciers are analysed in climatic terms. The glacier advances in Scandinavia and New Zealand during phases of glacier retreat on a global scale, might be related to regionally different climatic conditions (Chinn et al., 2005; Nesje et al., 2000), while in other regions such as the Karakoram or Svalbard they are also related to dynamical instabilities (surging) of glaciers (e.g., Bolch et al., 2012; Murray et al., 2003; Quincey et al., 2011). For mid-latitude mountain and valley glaciers, typical retreat rates are in the order of 5 to 20 m per year with rates up to 100 m/year for specific glaciers in individual years, or even more in glaciers where the lower part of a tongue lost contact to the main glacier due to steep slopes (FAQ 4.1, Figure 1c). However, some highly sensitive non-calving valley glaciers in Chile have reported retreat rates of more than 300 m/year (Rivera et al., 2012). Glaciers with peculiarities such as calving instabilities can retreat even faster (Pfeffer, 2007), and those with heavily debris-covered tongues much slower (Scherler et al., 2011), than the surrounding ‘normal’ glaciers.

[INSERT FIGURE 4.9 HERE]

Figure 4.9: Selection of cumulative glacier length changes as compiled from long-term in situ measurements in 11 regions. Two of the glaciers in Iceland (Leirufjordjökull and Mulajökull) show surge-type behaviour. Data are from WGMS (2008) and Leclercq et al. (Submitted) for region 5 (Greenland).

4.3.3.2 Area Changes

Relative changes in glacier area over entire mountain ranges are reported for nearly all 19 regions in Table 4.2 (Figure 4.10). Similar area loss rates between 0.1 and 1% yr⁻¹ are found in all investigated regions over the 1960 to 2000 period, including the Arctic and cold/dry continental climatic regions in Asia (Sorg et al., 2012). These rates have increased in some regions over the past 20 years, and are now also found in the -1 to -2% yr⁻¹ range. Strong area losses are reported from mid-latitude (e.g., 2% yr⁻¹ in the Alps) and low-latitude mountain ranges (East Africa, tropical Andes) over the past decade (Paul et al., 2011b; Rabatel, 2012; Thompson et al., 2011c). For even shorter periods and smaller glacier samples, loss rates can even be higher, for example -3.4% yr⁻¹ for the remaining glaciers in Indonesia (Klein and Kincaid, 2006). For some regions area loss rates over the past 20–30 years were also compared over similar decadal time periods and identical samples. These studies found a high regional variability of the annual area loss rates but no general trend towards accelerated area loss (e.g., Davies and Glasser, Submitted; Schmidt and Nusser, 2012).

A large number of studies (e.g., Klein and Kincaid, 2006; Paul et al., 2011a; Pelto, 2010; Thomson et al., 2011, and others given in the caption of Figure 4.10) also report that between two assessments of area, several glaciers completely disappeared in their respective regions. These disappeared glaciers have not been counted systematically, as there is no clear definition of the minimum size of glaciers that should be included, however there is robust evidence that over the past 30 years, globally, more than one hundred glaciers have disappeared completely.

[INSERT FIGURE 4.10 HERE]

Figure 4.10: Mean annual area-loss rates for 14 out of the 19 regions depicted in Figure 4.8. Each line refers to the observed relative area loss from a specific publication and its length is related to the period used for averaging. The publications considered for each subregion (in brackets) are: (1) (Le Bris et al., 2011); (2) (Barrand and Sharp, 2010; Bolch et al., 2010; Debeer and Sharp, 2007; Jiskoot et al., 2009; Thompson et al., 2011a) (3) (Huss et al., 2008); (4) (Dowdeswell et al., 2007; Paul and Kaab, 2005); (7) (Kaab, 2008); (8) (Andreassen et al., 2008; Paul and Andreassen, 2009; Paul et al., 2011a); (10) (Shahgedanova et al., 2010; Shahgedanova et al., 2012); (11) (Abermann et al., 2009; Lambrecht and Kuhn, 2007; Paul et al., 2011b; Paul et al., 2004); (13) (Aizen et al., 2007; Bolch et al., 2010; Cia et al., 2005); (Kutuzov and Shahgedanova, 2009; Li et al., 2006; Narama et al., 2006; Narozhnyi and Zemtsov, 2011; Surazakov et al., 2007; Wang et al., 2009; Ye et al., 2006a; Ye et al., 2006b; Zhang et al., 2012; Zhou et al., 2009); (14) (Bolch, 2007; Khromova et al., 2006; Narama et al., 2006); (15) (Bhambri et al., 2011; Bolch et al., 2008; Kulkarni et al., 2007; Nie et al., 2010; Schmidt and Nusser, 2012); (16) (Cullen et al., 2006; Klein and Kincaid, 2006; Peduzzi et al., 2010; Racoviteanu et al., 2008; Silverio and Jaquet, 2005; Silverio and Jaquet, 2012); (17) (Davies and Glasser, Submitted; Rivera et al., 2005; Rivera et al., 2007; Schneider et al., 2007); (18) (Gjermundsen et al., 2011); (19) (Berthier et al., 2009; Thost and Truffer, 2008). Data compilation by M. Mahrer, University of Zurich.

4.3.3.3 Glacier Volume and Mass Changes

Glacier volume and mass changes in AR4 were extrapolated from surface mass balances of individual glaciers, directly measured with different approaches and uncertainties (Kaser et al., 2006; Lemke et al., 2007). Since AR4, a series of geodetically derived, mainly regional, ice volume changes has widened the focus from surface mass balances to total volume and mass changes that also include calving losses (e.g., Gardner et al., Submitted, and references therein). Although such measurements only cover most recent years, they allow assessment of uncertainties associated to the different methods. In addition, the complete RGI data set of glacier-covered areas (Arendt et al., 2012) has improved spatial extrapolations towards regional and global glacier volume and mass changes (e.g., Cogley, 2009b; Gardner et al., Submitted) as well as modelling with atmospheric input variables (Giesen and Oerlemans, 2012; Hirabayashi et al., submitted; Marzeion et al., submitted). In Figure 4.11 the available values are compiled for each of the 19 regions from 1960 to 2010 if they cover a reasonable percentage of a region's glacier area (sources are listed in the caption of Figure 4.11). A 5–95% confidence envelope is provided. Most studies show results from one method, few merge two and Gardner et al. (Submitted) provide a consensus estimate for 2003–2009 by discussing and merging the results of different methods, including gravimetry from GRACE (Table 4.4). While the mean area specific mass change rates in ($\text{kg m}^{-2} \text{ yr}^{-1}$) in Figure 4.11 are indicative of the climate signal in glacier changes (neglecting that some non-climate effect of ice dynamics is included in some studies), the product of these rates with the respective total glacier area of a region (Figure 4.8) gives its contribution to sea level rise. Regional conversion numbers are given for each region in Figure 4.11.

[INSERT FIGURE 4.11 HERE]

Figure 4.11: Glacier mass change rates in [$\text{kg m}^{-2} \text{ yr}^{-1}$] for the 19 regions from Figure 4.8 and Table 4.2. Regional values are either from airborne and/or satellite repeat topographic mapping (**u**: (Abdalati et al., 2004); **r**: (Arendt et al., 2002); **o**: (Berthier et al., 2010); **v**: (Magnusson et al., 2005); **h**: (Moholdt et al., 2012); **j**: (Moholdt et al., 2010); **i**: (Nuth et al., 2010); **x**: (Paul and Haeberli, 2008); **n**: (Rignot et al., 2003); **s**: Schiefer (Schiefer et al., 2007); **B**: (Willis et al., 2012), from repeated gravity field measurements (GRACE) (**l**: (Chen et al., 2007); **m**: (Ivins et al., 2011); **f**: (Jacob et al., 2012); **p**: (Luthcke et al., 2008); **w**: (Matsuo and Heki, 2010); **y**: (Peltier, 2009); **k**: (Schrama and Wouters, 2011); **q**: (Wu et al., 2010), volume area scaling (**g**: (Glazovsky and Macheret, 2006)); or from extrapolation from single glacier measurements (**b** [spatial interpolation] and **c** [arithmetic mean]: (Cogley, 2009b); (Huss, 2012)), and modelling with atmospheric input variables (**a**: (Marzeion et al., submitted), **d**: (Hock et al., 2009). In addition multiple methods are used by: (**A**: (Gardner et al., 2011), **t**: (Gardner et al., Submitted) and **e**: (Gardner et al., Submitted)). Incomplete regional measurements are up-scaled. Uncertainties bounds show the 90% confidence envelope. Uncertainties not provided by the authors, are assigned a random error of 500 kg m^{-2} for non-elevation difference studies and as a cumulative error of 5 m for elevation change studies. Conversions from ($\text{kg m}^{-2} \text{ yr}^{-1}$) to mm SLE are given for each region below the region names. GRACE estimates of glacier mass change are often not accompanied by estimates of glaciers area that are required for conversion from Gt yr^{-1} to $\text{kg m}^{-2} \text{ yr}^{-1}$. In such cases Gt yr^{-1} values were divided by the total RGI regional glacier area with the exception of **l** and **m** that were divided by the area of the Patagonian Icefields.

Table 4.4: Regional glacier area specific mass change rates ($\text{kg m}^{-2} \text{ yr}^{-1}$) and total regional mass change rates (Gt yr^{-1}) (2003–2009) from merging different methods by Gardner et al. (Submitted).

Nr. Region Name	($\text{kg m}^{-2} \text{ yr}^{-1}$) (Gt yr^{-1})	Nr. Region name	($\text{kg m}^{-2} \text{ yr}^{-1}$) (Gt yr^{-1})
-----------------	---	-----------------	---

1	Alaska	-620 ± 120	-56 ± 11	11	Central Europe	-1080 ± 200	-2 ± 0
2	Western Canada and US	-930 ± 160	-14 ± 2	12	Caucasus and Middle East	-880 ± 100	-1 ± 0
3	Arctic Canada North	-320 ± 40	-34 ± 4	13	Central Asia	-130 ± 300	-8 ± 21
4	Arctic Canada South	-660 ± 100	-27 ± 4	14	South Asia West	-100 ± 70	-3 ± 2
5	Greenland Periphery	-450 ± 110	-40 ± 10	15	South Asia East	-225 ± 120	-5 ± 3
6	Iceland	-950 ± 180	-9 ± 2	16	Low Latitudes	-1090 ± 370	-4 ± 1
7	Svalbard	-160 ± 40	-6 ± 1	17	Southern Andes	-850 ± 240	-28 ± 8
8	Scandinavia	-620 ± 110	-2 ± 0	18	New Zealand	-330 ± 400	0 ± 1
9	Russian Arctic	-190 ± 60	-10 ± 3	19	Antarctic and Subantarctic	-10 ± 100	-1 ± 13
10	North Asia	-630 ± 2200	2 ± 1		Total	-340 ± 40	-251 ± 28

1
2
3 This compilation does not allow to determine the skill and uncertainty of each result. The number of studies,
4 the related uncertainties, and detailed glacier behaviour differ from region to region. Most studies give
5 averages over time periods of up to several decades. Extrapolation from directly measured glaciers (Cogley,
6 2009b; Huss, 2012) and recent modelling from atmospheric input (Marzeion et al., submitted) resolve
7 pentadal (five years) to annual time steps. The highest density and time resolution are available for the Arctic
8 regions [1, 2, 3, 4, 6, 7, 9] and Central Europe [11]. Least coverage and some of the highest uncertainties are
9 in Greenland [5], the Low Latitudes [16], New Zealand [18], and the Antarctic and Subantarctic [19]. As
10 discussed comprehensively in Gardner et al. (Submitted), recent results from gravity field anomalies
11 (GRACE mission) (e.g., Jacob et al., 2012; Matsuo and Heki, 2010) correlate well in regions with large ice
12 cover but deviate considerably from other results in regions with small and scattered ice cover (e.g., Yao et
13 al., 2012, and references therein for the Tibetan Plateau and its surroundings). In Scandinavia [8], Caucasus
14 and Middle East [12], and the Low Latitudes [16] the GRACE related uncertainties exceed the regional panel
15 in Figure 4.11. In Scandinavia [8] even the average value of Jacob et al. (2012) is out of the panel frame.
16 Model results from Hock et al. (Hock et al., 2009), giving an average from 1961 to 2004, deviate from others
17 towards considerably more mass loss in Iceland [6], Svalbard [7] and the Southern Andes [17] and less mass
18 loss in Central Asia [13] and South Asia (West) [14]. The year-by-year modeled time series from Marzeion
19 et al. (submitted) correlate well with the other estimates except in the Low Latitudes [16] which reflects the
20 weak correlation of low latitude glaciers with air temperature (e.g., Sicart et al., 2008) while atmospheric
21 moisture and related terms dominate the glacier variations (e.g., Molg et al., 2008; Vuille et al., 2008). The
22 Marzeion et al. (submitted) results also partly deviate from the general picture, with higher loss rates in
23 Svalbard [7] (recent decade) and the Russian Arctic [9] (1960s), and with lower rates in the Southern Andes
24 [18] (since about 1990).

25
26 Despite the considerable scatter, Figure 4.11 shows likely glacier mass losses in all of the 19 regions over the
27 past five decades, based on a high agreement and an overall robust evidence in most regions. In most regions
28 ice loss has likely increased during the last two decades, with slightly smaller loss during the last pentad in
29 some regions. In Central Europe [11], the increase of loss rates was earliest and strongest. In the Russian
30 Arctic [9] and in the Antarctic and Subantarctic [19] regions the signal is least clear. In several cases, the
31 Gardner et al. (Submitted) (Table 4.4) multi-method values show more negative (Western Canada and US
32 [2], Arctic Canada South [4]), which is also supported by studies of (Fisher et al., 2012) and (Sharp et al.,
33 2011), less negative (Central Asia [13], South Asia [14 and 15], Antarctic and Subantarctic [19]) or even
34 slightly positive (Asia North [10]) specific mass change rates than the other estimates. Particularly in the
35 Karakoram mountains of South Asia West [14], stagnant and even growing glacier volumes and masses have
36 been reported (Kääb et al., 2012, in press), leading to an overall small specific loss rate in this region. Central
37 Asia [13] and South Asia East [15] also show moderate specific loss rates. The recent contributions to sea
38 level rise (obtained by converting the ($\text{kg m}^{-2} \text{yr}^{-1}$) values to (mm SLE) ones in Figure 4.11) are strongest
39 from the regions of Alaska [1], the Canadian Arctic [3 and 4] and Greenland [5] with Central and South Asia
40 together [13-15] following. This also correlates with the estimates from Gardner et al. (Submitted) shown in
41 Table 4.4.

42
43 Studies of recent glacier velocity change (Azam et al., submitted; Heid and Kaab, 2011) and of accumulation
44 areas under present climate conditions (Bahr et al., 2009), indicate that the world's glaciers are currently
45 strongly out of balance with the present climate and thus committed to lose considerable mass in the near
46 future, even without further increasing temperatures. Increasing ice temperatures recorded on a high

1 elevation site the European Alps (Hoelzle et al., 2011; Vincent et al., 2007), as well as the ongoing thinning
2 of cold ice on Storglaciären in northern Sweden (Gusmeroli et al., 2012), support this conclusion.

3 4 **4.3.4 The Glacier's Contribution to Sea Level**

5
6 Globally estimated time series are required to assess the glacier's contribution to sea level change. Four
7 recent studies provide robust evidence of a considerable and continuous contribution but only medium
8 agreement on the rates (Figure 4.12). Cogley (2009b) compiled 4,146 (updated to 4,817) annually directly
9 measured mass budgets from 344 (now 383) glaciers, and 16,383 (19,959) annual values from 754 (983)
10 volume change measurements from an additional 327 (423) glaciers by extending from the data collection of
11 WGMS (2009, and earlier issues), and assembled these data in pentads. By adding volume change data from
12 repeat measurements of surface-elevation changes to the directly measured mass changes from earlier
13 estimates as used in Kaser et al. (2006) and Lemke et al. (2007), the proportion of calving glaciers was
14 increased from 6 (3% marine and 3% lacustrine) to 35% (20% marine and 15% lacustrine): this is more
15 realistic, but possibly still underestimated. An area-weighted global extrapolation was possible after 1961.
16 Earlier estimates allowed only for arithmetically averaging with high uncertainties over a small numbers of
17 glaciers (not shown here).

18
19 Leclercq et al. (updated from, 2011) use length variations from 382 glaciers worldwide (updated from 349)
20 to estimate glacier mass-loss since 1800. The length/mass change conversion is calibrated on mass balance
21 and geodetic observations for the period 1950–2005 from Cogley (2009b). Two extrapolations from the
22 individual glaciers were made for global estimates: the arithmetic mean and an area-weighted estimate. They
23 only give reasonable confidence after 1900 when the number of length variation data series increase from the
24 sparse earlier information. Yet, the first is still shown back to 1800 allowing the view back despite large
25 uncertainties. Uncertainty is estimated from upper and lower bounds of calibration assumptions,
26 cumulatively propagating these backward in time. The time series of Cogley (updated from, 2009b) and
27 Leclercq (updated from, 2011) inherently include ablation at calving fronts (even if not fully), thus giving
28 estimates of the total mass loss from glaciers.

29
30 Marzeion et al. (submitted) calculate monthly mass changes worldwide for automatically delineated
31 individual glaciers based on the RGI. The model is calibrated against measured mass balance time series and
32 validated by cross validation. It is driven by monthly mean temperature and precipitation derived from the
33 gridded climate CRU TS 3.0 dataset (Mitchell and Jones, 2005). Uncertainty estimates are derived by
34 propagating the uncertainties of the mass balance model itself (including uncertainties of the forcing), the
35 surface area measurement, the volume-area scaling relationships, and the representation of dynamic glacier
36 response to volume changes through the entire model system. The uncertainties are accumulated for each
37 glacier relative to the date of surface area measurement, and then accumulated regionally and globally. The
38 model only calculates surface mass balances but reproduces geodetically measured volume changes
39 reasonably well for most glaciers, although less well for calving glaciers.

40
41 Figure 4.12 shows global glacier mass losses including those of the peripheral ice sheet glaciers (Table 4.5)
42 and a best estimate calculated from those time series and for all glaciers including the peripheral glaciers, the
43 peripheral glaciers of the ice sheets, and all glaciers excluding the peripheral glaciers separately, covering
44 time periods of concern for sea level rise (Chapters 3 and 13). Gardner et al. (Submitted) indicate that the
45 time series may show too high loss rates at least in the most recent years (Figure 4.12, a, bottom and Table
46 4.5, last row). For the time being there is no way to validate this, and it is very difficult to know whether a
47 potential over-estimate would also affect the time series back into the 20th century. Both Leclercq et al.
48 (updated from, 2011) and Marzeion et al. (submitted) show high mass loss rates early in the 20th Century
49 and assign these to contributions from glaciers in the Arctic, namely the Greenland periphery and the
50 Russian Arctic. Other studies support these high loss rates (Björk et al., 2012; Box et al., 2009; Zdanowicz et
51 al., 2012; Zeeberg and Forman, 2001). The fact that Leclercq et al. (2011, updated) show the peak of mass
52 loss rates earlier may be a consequence of the bulk adjustment to the complex time lags of glacier tongue
53 position to mass change. Earlier studies of the contribution of glaciers to sea level change on a centennial
54 timescale all give smaller estimates than the new ones (Gregory and Oerlemans, 1998; Kaser et al., 2006;
55 Lemke et al., 2007; Meier, 1984; Oerlemans et al., 2007; Zuo and Oerlemans, 1997). However, these were
56 all impacted by an incomplete glacier inventory, and Zuo and Oerlemans (1997) also underestimated
57 precipitation in most regions (Leclercq et al., 2011).

1
2 **[INSERT FIGURE 4.12 HERE]**

3 **Figure 4.12:** Global cumulative (top graphs) and annual (lower graphs) glacier mass change 1800–2010 and 1960–2010
4 in panel (a) and (b) respectively. The 1986–2005 averages of the different cumulative estimates are all set to zero mm
5 SLE. Estimates are from glacier length variations (updated from, Leclercq et al., 2011), from area-weighted
6 extrapolations of individual directly and geodetically measured glacier mass budgets (updated from, Cogley, 2009b),
7 and from modelling with atmospheric variables as input (Marzeion et al., submitted). Uncertainties are based on
8 comprehensive error analyses in Cogley (2009b) and Marzeion et al. (submitted) and on assumptions about the
9 representativeness of the sampled glaciers in Leclercq et al. (2011). The latter uncertainties are not shown in the lower
10 panels for better clarity. The blue bars show the number of measured single glacier mass balances used in the updated
11 Cogley (2009b) time series. The mean 2003–2009 estimate by Gardner et al. (Submitted) is added to the bottom panel
12 b. Figure drawn by U. Blumthaler, Institute of Meteorology and Geophysics, University of Innsbruck

13
14
15 **Table 4.5:** Average annual rates of global mass loss for different time periods (Ch 13) for all glaciers globally including
16 the peripheral glaciers around the Greenland (GL) and Antarctic (AA) ice sheets; the peripheral glaciers only (PG GL
17 and PG AA); and all glaciers excluding those in the periphery of the ice sheets. For each of the first four time periods
18 two data series are available. The 5–95% confidence level is calculated as uncertainty. The 2003–2009 from Gardner et
19 al. (Gardner et al., 2012) is included for comparison (italics).

	Ref	Including PG		PG GL		PG AA		Excluding PG	
		Gt yr ⁻¹	mm SLE yr ⁻¹	Gt yr ⁻¹	mm SLE yr ⁻¹	Gt yr ⁻¹	mm SLE yr ⁻¹	Gt yr ⁻¹	mm SLE yr ⁻¹
1900* –1990	L+M	313 ± 184	0.86 ± 0.51	63 ± 37	0.17 ± 0.10	48 ± 28	0.13 ± 0.08	203 ± 119	0.56 ± 0.33
1971– 2009	M+C	309 ± 78	0.85 ± 0.21	20 ± 20	0.06 ± 0.06	71 ± 68	0.20 ± 0.19	218 ± 10	0.60 ± 0.03
1993– 2009	M+C	376 ± 69	1.04 ± 0.19	32 ± 9	0.09 ± 0.02	85 ± 75	0.23 ± 0.21	259 ± 15	0.72 ± 0.04
2005– 2009	M+C	371 ± 50	1.02 ± 0.14	45 ± 7	0.13 ± 0.02	64 ± 10	0.18 ± 0.03	262 ± 67	0.72 ± 0.18
2003– 2009	<i>G</i>	<i>251 ± 65</i>	<i>0.69 ± 0.18</i>	<i>40 ± 23</i>	<i>0.11 ± 0.06</i>	<i>1 ± 30</i>	<i>0.00 ± 0.08</i>	<i>210 ± 65</i>	<i>0.58 ± 0.18</i>

20 Notes:

21 L: (Leclercq et al., 2011, updated)

22 M: (Marzeion et al., submitted)

23 C: (updated from, Cogley, 2009b)

24 G: (Gardner et al., Submitted)

25 * M data only start in 1902

26 L + M: the L data for the last three columns (italics) are obtained by the ratio in M of each subgroup to the sum of all
27 glaciers (including PG); In G values for Antarctic and Subantarctic are used instead of PG AA (grey).

28
29
30 There is *robust evidence* and *high agreement* that the sea level contribution rates from glaciers have *very*
31 *likely* gradually increased since about 1985 with a slight decrease in the most recent years. Model studies
32 indicate strongest mass losses during the first half of the 20th Century from then unmeasured regions in the
33 Arctic, particularly the periphery of Greenland. The largest uncertainty in the estimates comes from the
34 Antarctic periphery where most recent estimates show much lower losses than estimated earlier (Table 4.4).
35 The time series seem to over-estimate the global glacier's mass loss at least in recent years. Large glacier
36 basins often end at calving fronts (Figure 4.8), and changes to these only indirectly reflect climate-driven
37 changes (McNabb et al., submitted; Post et al., 2011). In their study, McNabb et al. (submitted) show that 49
38 Alaskan tidewater lost 8.5 Gt yr⁻¹ (69% of their net mass loss) between 1973 and 2000, about half of this
39 coming from Columbia Glacier alone. On a global scale, the contribution from melting at calving fronts and
40 iceberg calving are not fully assessed yet. However, it is *very likely* that mass losses are dominated by
41 negative surface mass budgets (see the Marzeion et al. curve in Figure 4.12). Emerging and growing pro-
42 glacial lakes may hold back melt water from reaching the Ocean on short and intermediate time scales (e.g.,
43 Loriaux and Casassa, 2012).

44
45 **[START BOX 4.1 HERE]**

46
47 **Box 4.1: Interaction of Snow with the Cryosphere**

1 Snow is one component of the cryosphere. Snow sustains ice sheets and glaciers and has strong interactions
2 with all the other cryospheric components, with the exception of sub-sea permafrost. For example, snow can
3 impact the rate of sea ice production, and alter frozen ground through its insulating effect. Snowfall and the
4 persistence of snow cover are strongly dependent on atmospheric temperature and the highly variable
5 precipitation events, and are thus susceptible to climate change.

6
7 For Earth's climate in general, and more specifically, the cryospheric components on which it falls, the two
8 most important physical properties of snow, are its high albedo (reflectivity for solar radiation) and its low
9 thermal conductivity, which is a result of its high air-content and makes it an excellent thermal insulator.
10 Both factors substantially alter the flux of energy between the atmosphere and the material beneath the snow
11 cover. Snow also has a major impact on the total energy balance of the Earth's surface because large regions
12 in the northern hemisphere are covered by seasonal snow.

13
14 The high albedo of snow has a strong impact on the radiative energy balance of all surfaces on which it lies,
15 most of which are normally much less reflective. For example, the albedo of bare glacier or sea ice is
16 typically only 20–30% and hence 70–80% of solar radiation is absorbed at the surface. For ice at the melting
17 point, this energy melts the ice. With a fresh snow cover over ice, the albedo changes to 80% or even higher
18 and melting is greatly reduced. The effect is similar for other land surfaces – bare soil, frozen ground, low-
19 lying vegetation – but here snow cover protects the ground from warming (or cooling, depending on the
20 season) rather than from melting.

21
22 While an insulating snow cover can reduce the growth of sea ice, a heavy snow load, particularly in the
23 Antarctic, often depresses the ice surface below sea level and leads to enhanced snow-ice formation (see
24 FAQ 4.2). Even without flooding, the basal snow layer on Antarctic sea ice tends to be moist and saline
25 because brine is wicked up through the snow cover. In areas of heavily ridged and deformed sea ice, snow
26 redistributed by wind smoothes the ice surface, reducing the ice-air drag coefficient and thus slowing ice
27 drift, and also reducing the turbulent transfer of latent and sensible heat.

28
29 For frozen ground, the insulation characteristic of snow cover can also be significant. If the air above is
30 colder than the material on which it lies, the presence of low-density snow will reduce heat transfer upwards.
31 This could, for example, reduce the seasonal freezing of soil, slow down the freezing of the active-layer, or
32 protect permafrost from cooling. Alternatively, if the air is warmer than the material beneath the snow, heat
33 transfer downwards from the air would be reduced and, a snow cover can reduce the thickness of seasonal
34 soil freeze and protect permafrost from warming. Which process applies depends on the timing of the
35 snowfall, the season, and the duration of the snow cover on the ground.

36
37 For these reasons, the timing of the snowfall and the persistence of snow cover are of major importance.
38 Whereas snow falling on glaciers and ice sheets in summer has a strongly positive (sustaining) effect on the
39 mass budget, an early snow cover over frozen ground can prevent cooling, and even freezing, of the
40 seasonally unfrozen surface layer (the so-called active layer) and potentially contribute to its long-term
41 thawing. During winter, snowfall is the most important source of nourishment for most glaciers but radiative
42 cooling of frozen ground under a thick snow cover is strongly reduced, thereby contributing to an increase in
43 its temperature.

44
45 **[END BOX 4.1 HERE]**

46 47 **4.4 Ice Sheets**

48 49 **4.4.1 Background**

50
51 Today, the vast polar ice sheets in Greenland and Antarctica are losing ice mass as the polar climate becomes
52 warmer. In Greenland, warm summers are extending the zone and intensity of summer melting to higher
53 elevation and have doubled meltwater runoff since the 1980s. In both Greenland and Antarctica, some outlet
54 glaciers and ice streams are accelerating, and their floating extremities (ice shelves) are thinning and even
55 breaking up. As a result of these processes net losses from both ice sheets are increasing. At some locations,
56 glacier acceleration and increased ice-discharge into the ocean is likely due to the presence of warm ocean

1 waters at the ice-ocean boundary that melts submerged ice and changes the terminus geometry (e.g., Motyka
2 et al., 2011; Pritchard et al., 2012).

3 4 **4.4.2 Changes in Mass of Ice Sheets**

5
6 The current state of mass balance of the Greenland and Antarctic ice sheets is discussed here, focusing on
7 improvements made since AR4 in techniques of measurement and understanding of the change (Cazenave et
8 al., 2009; Chen et al., 2011; Lemke et al., 2007).

9 10 **4.4.2.1 Techniques**

11
12 There are broadly three techniques for measuring ice sheet mass balance. All have been applied to both ice
13 sheets by multiple groups, and over time scales ranging from multiple years to decades (Figure 4.13 and
14 Figure 4.14).

15 16 **[INSERT FIGURE 4.13 HERE]**

17 **Figure 4.13:** Temporal pattern of ice loss in Greenland from (a-c) GRACE time-variable gravity in centimetres of water
18 per year for the periods (a) 2002 to 2006, (b) 2006 to 2011 and (c) 2002 to 2011, color coded red (loss) to blue (gain)
19 (updated from, Velicogna, 2009). Circles in c) indicate average ice loss (Gt/yr) from the mass budget (red), GRACE
20 (orange) and ICESat (blue) (Sasgen et al., 2012a; Sasgen et al., 2012b); (d) mean surface mass balance for years
21 1957–2009 from regional atmospheric climate modelling (Ettema et al., 2009); (e) ice velocity from satellite radar
22 interferometry data for years 2007–2009 showing fastest flow in red, fast flow in blue, and slower flow in green and
23 yellow (Rignot and Mouginot, 2012), and (f) ice-thinning rates from ICESat data for years 2003–2008 with thinning in
24 red to thickening in blue (Pritchard et al., 2009).

25 26 **[INSERT FIGURE 4.14 HERE]**

27 **Figure 4.14:** Temporal evolution of ice loss in Antarctica from (a-c) GRACE time-variable gravity in centimetres of
28 water per year for the periods (a) 2002 to 2006, (b) 2006 to 2011 and (c) 2002 to 2011, color coded red (loss) to blue
29 (gain) (adapted from, Velicogna, 2009) Circles in (c) indicate average ice loss (Gt/yr) for years 2003–2008 for the
30 Antarctic Peninsula, West Antarctica and East Antarctica (red = mass budget, orange = GRACE, grey = ICESat)
31 (Shepherd and others, 2012); (d) mean surface mass balance in Antarctica for years 1989–2004 from regional
32 atmospheric climate modelling (van den Broeke et al., 2006); (e) ice sheet velocity for 2007–2009 showing fastest flow
33 in red, fast flow in blue, and slower flow in green and yellow (Rignot et al., 2011a); (f) ice thinning rates from ICESat
34 for years 2003–2008 with thinning in red to thickening in blue (Pritchard et al., 2009).

35 36 **4.4.2.1.1 Mass budget method**

37 The mass budget method relies on estimating the difference between net surface balance (input) and
38 perimeter fluxes (output). It compares two very large numbers, and even small percentage errors in either
39 may result in large errors in total mass balance. However, advances since AR4 have provided increasing
40 reliability.

41
42 For ice discharge, improvements include more complete mapping of perimeter fluxes for both ice sheets
43 (Rignot et al., 2011b) more complete ice-thickness data (Fretwell et al., Submitted; Griggs and Bamber,
44 2011b) and velocity data from satellite radar interferometry (Joughin et al., 2010b; Rignot et al., 2011a).
45 However, incomplete ice thickness mapping still causes uncertainties in ice discharge.

46
47 For surface mass balance, regional atmospheric climate models are increasingly used to produce estimates
48 that are verified using independent in situ data. Surface mass balance in Antarctica (including ice shelves) is
49 estimated at $2418 \pm 181 \text{ Gt yr}^{-1}$ in 1979–2010 (Lenaerts et al., In press; van den Broeke et al., 2006) with
50 interannual variability of 114 Gt yr^{-1} (Figure 4.14). In Antarctica, runoff is negligible but there is ice loss
51 from sub-ice melt processes. Interannual variability in surface mass balance in Greenland is mostly caused
52 by variation in runoff. Surface mass accumulation ranges from 100 to 500 Gt yr^{-1} with an average
53 uncertainty of 40 Gt yr^{-1} (7–40%) (Hanna et al., 2011; van den Broeke et al., 2009) (Figure 4.13). For the
54 past few decades, the reconstructions of surface mass balance by models that rely on calibration against *in*
55 *situ* data (e.g., Hanna et al., 2011), which are sparse particularly in southeast Greenland, exhibit disparities of
56 up to $\sim 200 \text{ Gt yr}^{-1}$ with estimates from climate models such as RACMO2 (van den Broeke et al., 2009)
57 which do not depend on field data. Combining uncorrelated errors in input and output, current mass budget
58 uncertainties are about 180 Gt yr^{-1} in Antarctica and 40 Gt yr^{-1} in Greenland.

4.4.2.1.2 *Repeated altimetry*

Repeated altimetric survey allows measurement of rates of surface-elevation change with time, and after correction for changes in snow density and bed elevation, or if the ice is floating, for tides and sea level, reveals changes in ice sheet mass.

Satellite radar altimetry (SRALT) has been widely used (Thomas et al., 2008b; Wingham et al., 2009) together with laser altimetry from airplanes (Krabill et al., 2002; Thomas et al., 2009) and satellites (Abdalati et al., 2010; Pritchard et al., 2009; Sorensen et al., 2011; Zwally et al., 2011), but with significant challenges. The surface footprint of early SRALT sensors was 20 km, and interpretation over ice sheets with undulating surfaces or significant slopes is complex. Estimates are also affected by surface characteristics, such as wetness, and by wide orbit separation (Thomas et al., 2008b). Errors in surface-elevation change are typically determined from the internal consistency of the measurements, often after iterative removal of surface elevation-change values that exceed some multiple of the local value of their standard deviation; this results in very small error estimates (Zwally et al., 2005). The ESA CryoSat-2 radar altimeter promises to be another valuable tool, although the first release of data is too recent to assess its impact (Wingham et al., 2006b).

Laser altimeters have been used from aircraft for many years, but satellite laser altimetry, available for the first time from NASA's ICESat satellite launched in 2003, provides a major advance in capability since AR4. Laser altimetry is easier to validate and interpret than radar data; the footprint is small (1 m for airborne laser, 60 m for ICESat), and there is negligible penetration into the ice. However, clouds limit spaceborne data acquisition, and accuracy is affected by atmospheric conditions, laser-pointing errors, and data scarcity. ICESat derived surface elevation changes supplemented with differenced ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer) satellite digital elevation models were used for outlet glaciers in southeaster Greenland (Howat et al., 2008) and for the northern Antarctic Peninsula (Shuman et al., 2011). Laser surveys over Greenland yield elevation estimates accurate to 10 cm along survey tracks for airborne platforms (Krabill et al., 1999; Thomas et al., 2009) and 15 cm for ICESat (Siegfried et al., 2011).

4.4.2.1.3 *Temporal variations in Earth gravity field*

Since 2002, the GRACE (Gravity Recovery and Climate Experiment) satellite mission has surveyed Earth's time-variable gravity field. Time-variable gravity provides a direct estimate of the ice-mass change. GRACE data yielded early estimates of 'secular' ice-mass changes over the Greenland and Antarctic ice sheets (Luthcke et al., 2006; Velicogna and Wahr, 2006a, 2006b) and confirmed regions of ice loss in East Greenland and West Antarctica. With extended time series, now more than nine years, GRACE results have lower uncertainties than in AR4. The ice loss signal is also more distinct due to the increased rate of ice sheet loss (e.g., Cazenave et al., 2009; Chen et al., 2009; Velicogna, 2009; Wouters et al., 2008). GRACE ice loss estimates vary among published studies. Some of the difference is caused by the time-variable nature of the signal. Additional differences are caused by: (1) data-centre specific processing, (2) specific methods used to calculate the mass change, and (3) contamination by other signals within the ice sheet (e.g., glacial isostatic adjustment or GIA) or outside the ice sheet (continental hydrology, ocean).

In Antarctica, the GIA signal is of the same order as the ice loss signal, with an uncertainty of 80 Gt yr⁻¹ (Riva et al., 2009; Velicogna, 2009; Velicogna and Wahr, 2006b). The GIA signal is addressed using numerical models, (e.g., Ivins and James, 2005; Paulson et al., 2007; Peltier, 2009). In Greenland, the GIA correction is less than 10% of the GRACE signal with an error of 19 Gt yr⁻¹. However, since the GIA rate is constant over the satellite's lifetime, GIA uncertainty does not affect the estimate of change in the rate of ice mass-loss.

In addition to GRACE, measurements of the elastic response of the crustal deformation shown in GPS measurements of uplift rates confirm increasing rates of ice loss in Greenland (Khan et al., 2010a; Khan et al., 2010b) and Antarctica (Thomas et al., 2011). Analysis of a 34-year time series of Earth's oblateness (J2) by satellite laser ranging also suggests that ice loss from Greenland and Antarctica has progressively dominated the change in oblateness trend since the 1990s (Nerem and Wahr, 2011).

4.4.2.2 *Greenland*

1 There is robust evidence and strong agreement between the methods described above that the Greenland Ice
2 Sheet has been losing ice and contributing to sea level rise over recent years (Sasgen et al., 2012b). Recent
3 GRACE results are in better agreement than in AR4 as discussed above (Baur et al., 2009; Chen et al., 2011;
4 Pritchard et al., 2010; Schrama and Wouters, 2011; Velicogna, 2009; Wu et al., 2010). Altimetry missions
5 report slightly lower losses than other methods (Thomas et al., 2006; Zwally et al., 2011) (Figure 4.13f).

6
7 Figure 4.15 shows the cumulative sea level contribution from the Greenland Ice Sheet over the period 1992–
8 2011 derived from recent studies made by 14 different research groups (Barletta et al., 2012; Baur et al.,
9 2009; Cazenave et al., 2009; Chen et al., 2011; Harig and Simons, Submitted; Pritchard et al., 2010; Rignot
10 et al., 2011c; Sasgen et al., 2012b; Schrama and Wouters, 2011; Slobbe et al., 2009; Sorensen et al., 2011;
11 Velicogna, 2009; Wu et al., 2010; Zwally et al., 2011). These do not include earlier estimates from the same
12 researchers when those have been updated by more recent analyses using extended data. They include
13 estimates made from satellite gravimetry, satellite altimetry and the mass balance method. Details of all
14 studies available for Greenland, and the sub-set of those selected for this assessment are listed in Appendix
15 4.A, Tables 1 and 2.

16
17 The mass balance for each year is estimated as a simple average of all the selected estimates available for
18 that particular year, and Figure 4.15 is an accumulation of these since an arbitrary zero on 31 December
19 1991. The number of estimates available varies with time, with as few as two estimates per year in the 1990s
20 and up to 17 yr⁻¹ after 2003. The cumulative uncertainty in Figure 4.15 is based on the uncertainty cited in
21 the original studies. However, since the annual estimates from different studies often do not overlap within
22 the quoted uncertainties, the error limits used in this assessment are based on the absolute maximum and
23 minimum mass balance estimate for each year within uncertainty ranges cited in the original studies. The
24 cumulative error is weighted by $1/\sqrt{n}$, where n is the number of years accumulated.

25 26 **[INSERT FIGURE 4.15 HERE]**

27 **Figure 4.15:** Cumulative sea level rise contribution (and ice loss equivalent) from Greenland derived from the
28 unweighted annual averages from 14 recent studies (see main text and Appendix 4.A for details).

29
30 Despite year-to-year differences between the various original analyses, the multi-study assessment provides
31 robust evidence that Greenland has lost mass over the last two decades, and that the rate of loss has
32 increased. This increase is also shown in many individual studies (Chen et al., 2011; Rignot et al., 2011c;
33 Velicogna, 2009; Zwally et al., 2011) (Figure 4.13a-c). The total sea level contribution from the Greenland
34 Ice Sheet has been 0.34 mm yr⁻¹ (± 0.06 mm yr⁻¹) over the period 1993–2010, and 0.63 mm yr⁻¹ (± 0.15 mm
35 yr⁻¹) between 2005 and 2010.

36
37 Reconciliation of some of the apparent disparities between the different satellite methods was made by the
38 Ice-sheet Mass Balance Intercomparison Experiment (Shepherd and others, 2012). This combined an
39 ensemble of satellite altimetry, interferometry and gravimetry data, and regional atmospheric climate model
40 output products, for common geographical regions for the time interval 1992–2011. Good agreement was
41 obtained between the estimates from the different satellite methods and, while the uncertainties of any
42 method are sometimes large, the combination of methods considerably improves the overall certainty. For
43 Greenland, Shepherd et al. (2012) estimate an average change in mass for 1992–2011 of -151 ± 48 Gt yr⁻¹
44 (0.42 ± 0.13 mm of sea level rise). For the same period this present assessment, which also averages different
45 techniques but uses the values given in the original studies, estimates a slightly lesser loss of -123 ± 21 Gt
46 yr⁻¹ (0.34 ± 0.06 mm). Shepherd et al. (2012) confirm an increasing mass loss from Greenland, although
47 within cyclical mass balance variations over intermediate (2 to 4 year) periods.

48 49 *4.4.2.2.1 Partitioning of ice loss*

50 The mass budget method shows the partitioning of ice loss from the ice sheet is about 60% surface mass
51 balance (i.e., runoff) and 40% glacier discharge (van den Broeke et al., 2009). Over the last two decades,
52 surface mass balance has decreased as a result of an increase in runoff; whereas ice discharge has increased
53 as a result of enhanced glacier speed.

54 55 *4.4.2.2.2 Surface mass balance*

56 Altimetric measurements of surface height suggest slight inland thickening (Thomas et al., 2006, 2009) that
57 is not confirmed by regional atmospheric climate models (Ettema et al., 2009) or recent ice core data

(Buchardt et al., 2012). Probable changes in accumulation are however exceeded by the increased runoff especially since 2006 (van den Broeke et al., 2009) The total melt area has continued to increase since AR4 and has accelerated in the past few years (Fettweis et al., 2011; Tedesco et al., 2011), with an extreme melt event covering >90% of the ice sheet for a few days in July 2012. The four highest runoff years over the last 140 years occurred since 1995 (Hanna et al., 2011).

4.4.2.2.3 *Regional changes*

There are significant differences in the relative importance of ice-discharge and surface mass balance in various regions of Greenland (Howat et al., 2007; Pritchard et al., 2009; Sasgen et al., 2012b) (van den Broeke et al., 2009) .

Dynamic losses dominate in southeast, central west and northwest Greenland. GRACE results show ice loss was largest in southeast Greenland during 2005 and increased in the northwest after 2007 (Chen et al., 2011; Khan et al., 2010a; Schrama and Wouters, 2011). Subsequent to 2005, ice loss decreased in the southeast. These GRACE results agree with measurements of ice discharge from the major glaciers that confirm the dominance of dynamic losses in these regions (van den Broeke et al., 2009). In particular, a major glacier speed up occurred in central east, southeast and central west Greenland between 1996–2000 (Howat et al., 2008; Rignot and Kanagaratnam, 2006) and in 2001–2006 (Joughin et al., 2010b): in the southeast many glaciers slowed after 2005 (Howat et al., 2007; Howat et al., 2011), with many flow speeds decreasing back towards those of the early 2000s (Murray et al., 2010), although most are still flowing faster and discharging more ice into the ocean than they did in 1996 (Howat et al., 2011; Rignot and Kanagaratnam, 2006).

In the central north, southwest and northeast sectors, changes in surface mass balance dominate. In the northwest, the acceleration in ice loss from 1996–2006 to 2006–2010 was caused by a high accumulation in the late 1990s (Sasgen et al., 2012b).

4.4.2.3 *Antarctica*

Antarctic results from the gravity method are also now more numerous and consistent than in AR4 (Figure 4.14a-c). Methods combining GPS and GRACE at the regional level indicate the Antarctic Peninsula is certainly losing ice (Ivins et al., 2011). In other areas, large uncertainties remain in the global GRACE-GPS solutions (Wu et al., 2010).

Results from the mass budget method have improved significantly since AR4 (Lenaerts et al., 2012; Rignot et al., 2011a; Rignot et al., 2008b; van den Broeke et al., 2006). Reconstructed snowfall from regional atmospheric climate models indicates higher accumulation along the coastal sectors than in prior maps, but little difference in total snowfall. There is no long-term trend in total accumulation over the past few decades (Bromwich et al., 2011; Frezzotti et al., 2012; Lenaerts et al., 2012; Monaghan et al., 2006; van den Broeke et al., 2006). Satellite and airborne laser altimetry indicate that ice volume changes are concentrated on outlet glaciers and ice streams, as illustrated by the strong correspondence between areas of thinning (Figure 4.1314f) and areas of fast flow (Figure 4.14e).

Figure 4.16 shows the cumulative sea level contribution from the Antarctic Ice Sheet over the period 1992–2011 derived from recent studies made by 14 different research groups (Barletta et al., 2012; Cazenave et al., 2009; Chen et al., 2009; Dong-Chen et al., 2009; Horwath and Dietrich, 2009; Ivins et al., 2011; King et al., Submitted; Moore and King, 2008; Rignot et al., 2011c; Shi et al., 2011; Velicogna, 2009; Wingham et al., 2006a; Wu et al., 2010; Zwally et al., 2005). These do not include earlier estimates from the same researchers when those have been updated by more recent analyses using extended data. They include estimates made from satellite gravimetry, satellite altimetry and the mass balance method. Details of all studies available for Antarctica, and the sub-set of those selected for this assessment are listed in Appendix 4.A, Tables 3 and 4. The cumulative curves and associated errors are derived in the same way as those for Figure 4.15 (see above).

[INSERT FIGURE 4.16 HERE]

Figure 4.16: Cumulative sea level rise contribution (and ice loss equivalent) from Antarctica derived from the unweighted annual averages from 14 recent studies (see main text and Appendix 4.A for details).

1 Overall, the ice sheet is very likely currently losing mass. The total sea level contribution from Antarctica
2 has been 0.18 mm yr^{-1} ($\pm 0.09 \text{ mm yr}^{-1}$) over the period 1993–2010, and 0.31 mm yr^{-1} ($\pm 0.16 \text{ mm yr}^{-1}$)
3 between 2005 and 2010.

4
5 A recent intercomparison (Shepherd and others, 2012) for Antarctica, where the GIA signal is less well
6 measured than in Greenland, uses two new GIA models (an updated version of, Ivins and James, 2005;
7 Whitehouse et al., 2012) (for details see, Shepherd and others, 2012). This has the effect of reducing the
8 estimates of East Antarctic ice mass loss from GRACE data. For Antarctica, Shepherd et al. (2012) estimate
9 an average change in mass for 1992–2011 of $-67 \pm 52 \text{ Gt yr}^{-1}$ ($0.18 \pm 0.14 \text{ mm}$ of sea level rise). For the
10 same period this present assessment estimates a loss of $-58 \pm 33 \text{ Gt yr}^{-1}$ ($0.16 \pm 0.09 \text{ mm}$).

11
12 Significantly, the rate of ice loss has almost certainly increased with time over the last two decades (Chen et
13 al., 2009; Rignot et al., 2011c; Velicogna, 2009) (Figure 4.16). For GRACE, this conclusion is independent
14 of the GIA signal, which is constant over the measurement period. From the mass budget method, the
15 increase in loss is certainly caused by an increase in glacier flow-speed in West Antarctica (Joughin et al.,
16 2010a; Rignot, 2008) and the Antarctic Peninsula (Pritchard and Vaughan, 2007; Rott et al., 2011; Scambos
17 et al., 2004). Comparison of GRACE and the mass budget methods indicates an increase in ice loss of 14 ± 2
18 Gt yr^{-1} every year for 1992–2010 versus $21 \pm 2 \text{ Gt yr}^{-1}$ every year for Greenland during the same time period
19 (Rignot et al., 2011c). A recent analysis (Shepherd and others, 2012) showed that the West Antarctic Ice
20 Sheet and the Antarctic Peninsula were losing mass at an increasing rate, but that East Antarctica gained an
21 average of $+21 \pm 43 \text{ Gt yr}^{-1}$ between 1992 and 2011. Zwally and Giovinetto (2011) also estimate a mass gain
22 for East Antarctica ($+16 \text{ Gt yr}^{-1}$ between 1992 and 2001). But their reassessment of Antarctic change to
23 correct for regions of the ice sheet not included in the mass budget method (see Section 4.4.2.1.1) is not
24 supported by the Shepherd et al. (2012) analysis which suggests that the missing regions contribute little
25 change.

26 27 *4.4.2.3.1 Partitioning of ice loss*

28 In the near-absence of surface runoff and long-term change in total snowfall, Antarctic long-term changes in
29 grounded ice mass are almost entirely explained by increased glacier speed. The ice sheet is however
30 strongly influenced by interannual to decadal variability in snowfall which exceeds in magnitude the
31 estimated long term trend. (Rignot et al., 2011c).

32 33 *4.4.2.3.2 Regional changes*

34 The three mass balance methods are in excellent agreement as to the spatial pattern of ice loss (thinning) and
35 gain (thickening) in Antarctica (Figure 4.14). The largest ice losses are located along the northern tip of the
36 Antarctic Peninsula where several ice shelves have broken up during the last two decades and in the
37 Amundsen Sea, in West Antarctica (Figure 4.14c). In the Antarctic Peninsula, precipitation is likely to have
38 increased (Thomas et al., 2008a) but the resulting ice-gains are insufficient to counteract the losses (Cook
39 and Vaughan, 2010; Ivins et al., 2011; Wendt et al., 2010). Changes in the Amundsen Sea region are likely
40 due to the thinning of ice shelves by a warm ocean (Jacobs et al., 2011), which caused grounding line retreat
41 (1 km yr^{-1}) (Joughin et al., 2010a) and glacier thinning (Wingham et al., 2009). Indications of dynamic
42 change are also evident from East Antarctica, primarily Totten Glacier, from GRACE (Chen et al., 2009),
43 SRALT (Wingham et al., 2006a) and satellite radar interferometry (Rignot et al., 2008b). The contribution to
44 the total ice loss from these areas is however small and poorly understood.

45 46 *4.4.2.4 Ice Shelves and Floating Ice Tongues*

47
48 As much as 74% of the ice discharged from the grounded ice sheet in Antarctica passes through ice shelves
49 and floating ice tongues (Bindschadler et al., 2011). Ice shelves help to buttress and restrain flow of the
50 grounded ice (Hulbe et al., 2008; Rignot et al., 2004; Scambos et al., 2004), and so changes in thickness
51 (Pritchard et al., 2012; Shepherd et al., 2003; Shepherd et al., 2010) and extent (Doake and Vaughan, 1991;
52 Scambos et al., 2004) of ice shelves influence current ice sheet change. Indeed, nearly all outlet glaciers and
53 ice streams experiencing high rates of ice loss are flowing into thinning or disintegrated ice shelves
54 (Pritchard et al., 2012), although many of the larger ice shelves exhibit stable conditions (King et al., 2009;
55 Pritchard et al., 2012).

1 Around the Antarctic Peninsula, ice-shelf retreat has been ongoing for several decades (Fricker and Padman,
2 2012), and has continued since AR4 with substantial collapse of a section of Wilkins Ice Shelf (Humbert et
3 al., 2010), which had been retreating since the late-1990s (Scambos et al., 2000). Overall, 7 of 12 ice shelves
4 around the Peninsula have retreated in recent decades with a total loss of 28,000 km², and a continuing rate
5 of loss of around 6,000 km² per decade (Cook and Vaughan, 2010). There is robust evidence that retreat of
6 ice shelves along the Antarctic Peninsula has been related to warming climate, and several studies have been
7 aimed at understanding the threshold for ice-shelf viability which in time may yield a sound predictive
8 capability (e.g., Kuipers Munneke et al., Submitted; Scambos et al., 2000).

9 4.4.2.5 Total Ice Loss from Both Ice Sheets

11 The total ice loss from both ice sheets for the twenty years 1992–2011 (inclusive) has been 3,640 ± 1015 Gt,
12 equivalent to 10.0 ± 2.8 mm of sea level rise. The majority of this ice however has been lost in the second
13 half of the period, and the rate of change has increased steadily with time. Over the last three years (2007–
14 2011) it has been equivalent to 1.0 ± 0.3 mm yr⁻¹ of sea level rise (Figure 4.17, Table 4.6).

17 [INSERT FIGURE 4.17 HERE]

18 **Figure 4.17:** Rate of ice sheet contribution to sea level rise averaged over 5 year periods between 1992 and 2011. These
19 estimates are derived from the data in Figure 4.15 and Figure 4.16.

22 **Table 4.6:** Average of estimates of sea level rise described in Figure 4.15 and Figure 4.16 and listed in Appendix 4.A,
23 Table 4A.1 and Table 4A.3.

Period	Unweighted Average (mm sea level rise yr ⁻¹)	
<i>Greenland</i>		
2005–2010 (6-year)	0.63	±0.15
1993–2010 (18-year)	0.34	±0.06
<i>Antarctica</i>		
2005–2010 (6-year)	0.31	±0.16
1993–2010 (18-year)	0.18	±0.09
<i>Combined</i>		
2005–2010 (6-year)	0.94	±0.31
1993–2010 (18-year)	0.52	±0.15

26 4.4.3 Causes of Changes in Ice Sheets

28 4.4.3.1 Climatic Forcing

30 4.4.3.1.1 Snowfall and surface temperature

31 Ice sheets experience large inter-annual variability in snowfall and local trends may deviate significantly
32 from the long-term trend in integrated snowfall. However, as in AR4, there was very little evidence for long-
33 term change, except for the Antarctic Peninsula (Bromwich et al., 2011; Ettema et al., 2009; Monaghan et
34 al., 2006) (van den Broeke et al., 2009).

36 Warming air temperature will (when above the freezing point) increases the amount of surface melt, and can
37 also increase the moisture bearing capacity of the air, and hence snowfall. Over Greenland, temperature has
38 risen significantly since the early-1990s, reaching values similar to those in the 1930s (Box et al., 2009). The
39 year 2010 was an exceptionally warm year in west Greenland with Nuuk having the warmest year since the
40 start of the temperature record in 1873 (Tedesco et al., 2011). In Antarctica, in response to ozone depletion,
41 the summertime Southern Annular Mode strengthened from the mid-1950s to the mid-1990s (Thompson et
42 al., 2011b). This strengthening has resulted in statistically significant summer warming on the east coast of
43 the northern Antarctic Peninsula (Chapman and Walsh, 2007; Marshall et al., 2006), with extension of
44 summer melt duration (Barrand et al., 2012), while East Antarctica has showed summer cooling (Turner et
45 al., 2005). In contrast, the significant winter warming at Faraday/Vernadsky station on the western Antarctic
46 Peninsula is attributable to a reduction of sea ice extent (Turner et al., 2005).

4.4.3.1.2 Ocean thermal forcing

Interaction between ocean waters and the periphery of large ice sheets very likely plays a major role in present ice sheet changes (Bindschadler, 2006; Pritchard et al., 2012). Ocean waters provide the heat that drives high melt rates beneath ice shelves (Holland and Jenkins, 1999; Jacobs et al., 1992; Pritchard et al., 2012; Rignot and Jacobs, 2002) and at marine-terminating glacier fronts (Holland et al., 2008a; Jacobs et al., 2011; Rignot et al., 2010). The importance of this effect has become increasingly apparent through observations made since AR4.

Ocean circulation delivers warm, salty waters to ice sheets, and variations in wind patterns associated with the North Atlantic Oscillation (Hurrell, 1995; Jacobs et al., 1992) and the Southern Annular Mode (Thompson and Wallace, 2000) are probable drivers of increasing amounts of warm water reaching the ice sheet margins. Limited observations have established that warm waters of sub-tropical origin are present at marine-terminating glaciers in Greenland (Christoffersen et al., 2011; Daniault et al., 2011; Holland et al., 2008a; Myers et al., 2009; Straneo et al., 2010). The presence of warm waters in contact with the ice is a necessary condition for rapid melting but other factors are also important, such as the bathymetry of fjords and ice shelf cavities (Jenkins et al., 2010).

Satellite records and *in situ* observations indicate warming of the Southern Ocean (see Chapter 3) since the 1950s (Gille, 2002; Gille, 2008). This warming is confirmed by data from robotic ocean buoys (Boening et al., 2008) but the observational record remains short and, close to Antarctica, there are only limited observations from ships (Jacobs et al., 2011), short-duration moorings and data from instrumented seals (Charrassin et al., 2008; Costa et al., 2008).

4.4.3.2 Ice Sheet Processes

4.4.3.2.1 Basal lubrication

In many regions close to the Greenland Ice Sheet margin, abundant summer meltwater on the surface of the ice sheet forms large lakes in narrow zones. This surface water can drain to the ice sheet bed, thus increasing basal water pressure, reducing basal friction and increasing ice flow speed. Such conduits are common in southwest and northeast Greenland, but rare in the most rapidly changing southeast and northwest regions (Selmes et al., 2011). This effect can be seen in diurnal flow variations of some land-terminating regions (Das et al., 2008; Shepherd et al., 2009), and after lake-drainage events, when 50–110% speed-up of flow has been observed. However, this effect is temporally and spatially restricted (Das et al., 2008). The summer increase in speed over the annual mean is only ~10–20%, the increase is less at higher elevations (Bartholomew et al., 2011), and observations suggest most lake drainages do not affect ice sheet velocity (Hoffman et al., 2011). Theory and field studies suggest an initial increase in flow rate with increased surface meltwater supply (Bartholomew et al., 2011; Palmer et al., 2011), but if the supply of surface water continues to increase and subglacial drainage becomes more efficient, basal water pressure, and thus basal motion, is reduced (Schoof, 2010; Shannon et al., 2012; Sundal et al., 2011; van de Wal et al., 2008). Overall, basal lubrication appears important in modulating flow in some regions, especially southwest Greenland, but it does not explain recent dramatic regional speed-ups that have resulted in rapid increases in ice loss from calving glaciers.

4.4.3.2.2 Cryo-hydrologic warming

Percolation and refreezing of surface meltwater that drains through the ice column may strongly affect the thermal regime of the ice sheet (Phillips et al., 2010), a process known as cryo-hydrologic warming. This processes could cause near-basal ice to soften and become easier to deform, affecting flow on decadal time scales (Phillips et al., Submitted).

4.4.3.2.3 Ice shelf buttressing

Recent changes in marginal regions of the Greenland and Antarctic ice sheets include some thickening and slowdown but mostly thinning and acceleration (Pritchard et al., 2009; Sorensen et al., 2011) with some glacier speeds increasing two to eightfold (Howat et al., 2007; Joughin et al., 2004; Luckman and Murray, 2005; Rignot and Kanagaratnam, 2006; Rignot et al., 2004; Scambos et al., 2004). Many of the largest and fastest glacier changes appear to be partly in response to ice shelf or floating ice-tongue shrinkage or loss.

1 This glacier response is consistent with classical models of ice shelf buttressing and marine instability
2 proposed 40 years ago (Hughes, 1973; Mercer, 1978; Thomas and Bentley, 1978; Weertman, 1974).

3 4 *4.4.3.2.4 Ice-ocean interaction*

5 Warm waters at depth melt the periphery of ice sheets in Greenland and Antarctica and very likely play a
6 central role in the evolution of the ice sheets. The flux of meltwater is proportional to the product of ocean
7 thermal forcing (difference between ocean temperature and the in-situ freezing point of seawater) and water
8 flow speed at the ice-ocean interface (Holland and Jenkins, 1999). Flow speed may increase with thermal
9 forcing due to greater ice melting increasing the buoyancy of the melt-water plume (Holland et al., 2008b),
10 but observations are not yet sufficient to verify this expectation.

11
12 Since AR4 it has become far more evident that submarine melting can be very large in Greenland (e.g.,
13 Motyka et al., 2011). Melt rates along marine-terminating glacier margins are one to two orders of
14 magnitude greater than for the nearly horizontal faces of ice shelves because of the additional buoyancy
15 forces provided by the discharge of sub-glacial melt water at the glacier base (Jenkins, 2011; Motyka et al.,
16 2003; Straneo et al., Submitted; Xu et al., 2012). In South Greenland, multiple lines of evidence suggest that
17 the acceleration of glaciers from the mid-1990s to mid-2000s was due to the intrusion of ocean waters of
18 sub-tropical origin into glacial fjords (Christoffersen et al., 2011; Holland et al., 2008a; Howat et al., 2008;
19 Motyka et al., 2011; Murray et al., 2010; Rignot and Mouginot, 2012; Straneo et al., 2011; Straneo et al.,
20 2010). The increase in ice melting by the ocean most probably contributed to the reduction of backstress at
21 the glacier fronts and subsequent acceleration (Nick et al., 2009; Nick et al., 2012b; Payne et al., 2004;
22 Schoof, 2007).

23 24 *4.4.3.2.5 Iceberg calving*

25 Calving of icebergs from marine-terminating glaciers and ice-shelves is important in their overall mass
26 balance, but the processes that initiate calving range from seasonal melt-driven processes (Benn et al., 2007),
27 to rare, and hard to predict, break-up events caused by ocean swells and tsunamis (MacAyeal et al., 2006).
28 Some of these processes show strong climate influence, while others do not. Despite arguments of rather
29 limited progress in this area (Pfeffer, 2011), there have been recent advances (Amundson et al., 2010;
30 Blaszczyk et al., 2009; Joughin et al., 2008a; Nick et al., 2010; Otero et al., 2011) and continental-scale ice
31 sheet models currently rely on improved parameterisations (Alley et al., 2008; Pollard and DeConto, 2009).
32 Recently more realistic models have been developed allowing the dependence of calving and climate to be
33 explicitly investigated (Nick et al., 2010; Nick et al., 2012a).

34 35 *4.4.4 Rapid Ice Sheet Changes*

36
37 The estimates of sea level rise in AR4 excluded future rapid dynamical changes in ice flow, and stated that
38 “understanding of these processes is limited and there is no consensus on their magnitude”. Considerable
39 progress has been made since AR4. The processes now thought to be potential causes of rapid changes in ice
40 flow, and new observational evidence that these processes are already underway, are summarised here.
41 “Rapid ice sheet changes” are defined as changes that are of sufficient speed and magnitude to impact on the
42 rate of sea level rise on timescales of several decades or less. A further consideration is whether and under
43 what circumstances any such changes are “irreversible”, i.e., would take several decades to centuries to
44 reverse under a different climate forcing. For example, an effectively irreversible change might be the loss of
45 a significant fraction of the Greenland Ice Sheet, because at its new lower (and therefore warmer) surface
46 elevation, the ice sheet would be able to grow thicker only slowly in a cooler climate (Ridley et al., 2010).

47
48 The importance that warm waters at depth play in melting the periphery of ice sheets in Greenland and
49 Antarctica, and the evolution of these ice sheets has become much clearer since AR4. New observations in
50 Greenland and Antarctica, as well as theoretical advances, show that rapid changes are to be expected in
51 those regions of ice sheets that are grounded well below sea level (Holland et al., 2008a; Joughin and Alley,
52 2011; Motyka et al., 2011; Ross et al., 2012; Schoof, 2007; Young et al., 2011a). Where this ice meets the
53 ocean, warm waters can increase bottom and ice front melting, causing undercutting, higher calving rates,
54 ice-front retreat (Benn et al., 2007; Motyka et al., 2003; Thomas et al., 2011), consequent speed-up and
55 thinning. These processes can occur in tandem with increased surface melting which increases basal motion,
56 ice fracturing and calving rates. Where ice shelves are present, ice melt by the ocean may cause thinning and

1 weakening of the ice shelf as well as migration of the grounding line further inland into the deep basin, with
2 a major impact on buttressing, flow speed and thinning rate (Thomas et al., 2011).

3
4 **[INSERT FIGURE 4.18 HERE]**

5 **Figure 4.18:** Bed topography for Greenland and Antarctica, derived from (Fretwell et al., Submitted; Griggs and
6 Bamber, 2011b) with marine-based parts of the ice sheet highlighted and arrows showing access routes for rapid
7 discharge of marine-based sectors. Figure drawn by P. Fretwell, British Antarctic Survey.

8
9 The influence of the ocean on the ice sheets is controlled by the delivery of heat to the ice sheet margins, in
10 particular to the ocean cavities beneath ice shelf shelves and to calving fronts (Jacobs et al., 2011). This
11 amount of heat delivered is a function of the temperature and salinity of ocean waters, but also of ocean
12 circulation, and the details of the bathymetry on the continental shelves, near glacier fronts and beneath ice
13 shelves (Pritchard et al., 2012). Changes in any of these parameters around the edges of major ice sheets
14 could have a direct and rapid impact on melt rates and potentially calving fluxes (See Chapter 13).

15
16 Ice grounded on a reverse bed-slope, deepening towards the ice sheet interior, is potentially subject to the
17 marine ice sheet instability (Schoof, 2007; Weertman, 1974) (See Chapter 13, Box 13.x). Much of the bed of
18 the West Antarctic Ice Sheet (WAIS) lies below sea level and on a reverse bed-slope, with basins extending
19 to depths greater than 2 km (Figure 4.18). The marine parts of the WAIS contain at ~3.4 m of equivalent sea
20 level rise (Fretwell et al., Submitted) and a variety of evidence strongly suggests that the ice sheet has been
21 much smaller than present in the last 1 million years, during periods with temperatures similar to those
22 predicted in the next century (Kopp et al., 2009). Potentially unstable marine ice sheets also exist in East
23 Antarctica, e.g., in Wilkes Land (Pritchard et al., 2009; Young et al., 2011a), and these contain more ice than
24 WAIS (9 m sea level equivalent for Wilkes Land). The Totten Glacier, Cook Ice Shelf and Denman Glacier
25 in East Antarctica, are showing signs of dynamic thinning at present. In northern Greenland, ice is also
26 grounded below sea level, with reverse slopes (Figure 4.18) (Joughin et al., 1999). In north Greenland,
27 marine sectors have not yet shown significant thinning.

28
29 Observations since AR4 confirm that rapid changes are occurring at the marine margins of ice sheets, and
30 that these changes can quickly penetrate hundreds of kilometres inland (Pritchard et al., 2009) (Joughin et al.,
31 2010b). Collapse of floating ice shelves on the Antarctic Peninsula has resulted in speeding up of tributary
32 glaciers of 300–800% (Rignot et al., 2004). This speed-up in turn has drawn down ice in the interior,
33 accelerating the loss of grounded ice previously buttressed by the ice shelves. The glaciers on this peninsula
34 contain only a few centimetres of equivalent sea level, however, similar processes acting on the larger ice
35 shelves further south could lead to rapid loss of ice from the West Antarctic Ice Sheet. The Amundsen Sea
36 sector of West Antarctica is grounded significantly below sea level and is the region changing most rapidly
37 at present. As a result of grounding line retreat, very likely caused by the intrusion of warm ocean water into
38 the sub-ice shelf cavity (Jenkins et al., 2010), Pine Island Glacier has sped up 73% since 1974 (Rignot, 2008)
39 and the floating ice tongue of this glacier thinned throughout 1995–2008 at increasing rates (Wingham et al.,
40 2009). The neighbouring Thwaites, Smith and Kohler glaciers are also speeding-up and thinning (Figure
41 4.14).

42
43 Similarly, in Greenland, the recent rapid retreat of Jakobshavn Isbrae was very likely caused by the intrusion
44 of warm ocean water beneath the floating ice tongue (Holland et al., 2008a; Motyka et al., 2011) combined
45 with other factors, such as weakening of the floating mixture of sea ice, iceberg debris and blown snow
46 within ice rifts (Amundson et al., 2010; Joughin et al., 2008b). It is likely that recent variations in South-East
47 Greenland's glaciers have been caused by the intrusion of warm waters of sub-tropical origin. Since AR4 it
48 has become clear that the mid-2000s speed up of South-East Greenland glaciers, which caused a doubling of
49 ice loss from the Greenland ice sheet (Howat et al., 2008; Luthcke et al., 2006; Rignot and Kanagaratnam,
50 2006; Wouters et al., 2008), was a pulse which was followed by a partial slow down (Howat et al., 2008;
51 Murray et al., 2010).

52
53 In contrast to the rapidly changing marine margins of the ice sheets, the land-terminating regions of the
54 Greenland Ice Sheet are changing more slowly, and those changes are largely explained by changes in the
55 input of snow and loss of meltwater (Sole et al., 2011). Surface meltwater, while abundant on the Greenland
56 Ice Sheet, does not seem to be driving significant changes in basal lubrication that impact on ice sheet flow
57 (Joughin et al., 2008b; Selmes et al., 2011; Sundal et al., 2011).

1
2 The ice shelves and glaciers on the Antarctic Peninsula have continued to experience irreversible changes,
3 coincident with air temperatures at some stations rising at four to six times the global average rate (Vaughan
4 et al., 2003) and with warm Circumpolar Deep Water becoming widespread on the western continental shelf
5 (Martinson et al., 2008). The 2002 collapse of the Larsen B Ice Shelf has been unprecedented in the last
6 10,000 years (Domack et al., 2005) and is irreversible: even if iceberg calving were to cease entirely,
7 regrowth of the Larsen B ice shelf to its pre-collapse state would take centuries based on the ice-shelf speed
8 and length prior to its collapse (Rignot et al., 2004).

9
10 In contrast, in Greenland changes do not yet appear irreversible. For example, the breakup of the floating
11 tongue of Jakobshavn Isbrae and consequent loss of buttressing has increased ice flow speeds and discharge
12 from the ice sheet, but Jakobshavn Isbrae has undergone significant margin changes over the last ~8,000
13 years which have been both more and less extensive than the recent ones (Young et al., 2011b). A second
14 example is Helheim Glacier in south-east Greenland which accelerated, retreated and increased its calving
15 flux during the period 2002–2005 (Andresen et al., 2012; Howat et al., 2011). However, evidence shows its
16 calving flux similarly increased during the late 1930s - early 1940s (Andresen et al., 2012): an episode from
17 which the glacier recovered and readvanced (Joughin et al., 2008b).

18
19 Despite many new observations that demonstrate changes can happen more rapidly than previously thought
20 together with strong evidence that ice-ocean interactions are the likely key to future decadal changes, there is
21 still an incomplete understanding of the processes that control the evolution of ice sheets in a warming
22 climate. Longer observational records, especially including observation in ocean waters beneath ice shelves
23 and in front of glaciers, would improve this understanding. At present, there is no indication of a slowdown
24 in the mass loss of ice sheets; instead observations suggest an ongoing increase in mass loss from ice sheets
25 with time.

26 27 **4.5 Seasonal Snow and Freshwater Ice Cover**

28 29 **4.5.1 Background**

30
31 Snowfall is a component of total precipitation and, in that context, is discussed in Chapter 2 (See Section
32 2.3.1.3); here we discuss accumulated snow as a climatological indicator. Snow is measured using a variety
33 of instruments and techniques, and reported using several quantitative metrics, including: snow cover extent
34 (SCE), the seasonal sum of daily snowfall, snow depth (SD), number of days with snow exceeding a
35 threshold depth, or snow water equivalent (SWE).

36
37 Long-duration, consistent records of snow are rare owing to many challenges in making accurate and
38 representative measurements. While weather stations in snowy inhabited areas often report snow depth,
39 records of snowfall are often patchy or use techniques that change over time (e.g., Kunkel et al., 2007),
40 except in certain parts of the European Alps. The density of stations and the choice of metric also varies
41 considerably from country to country. The longest satellite-based record of SCE is the visible-wavelength
42 weekly product of the National Oceanic and Atmospheric Administration (NOAA) dating to 1966 (Robinson
43 et al., 1993), but this only covers the Northern Hemisphere (NH). Satellite mapping of snow depth and SWE
44 has lower accuracy than SCE, especially in mountainous and heavily vegetated areas. Measurement
45 challenges are particularly acute in the Southern Hemisphere (SH), where snow rarely accumulates outside
46 mountainous areas, and where satellite-based mapping of snow began only in 1978. Indeed, with hardly any
47 inhabited snowy areas in the SH, only 10 long-duration records continue to recent times: six in the central
48 Andes and four in southeast Australia.

49 50 **4.5.2 Hemispheric View**

51
52 By blending *in situ* and satellite records, Brown and Robinson (2011) have updated a key indicator of
53 climate change, namely the time series of NH SCE (Figure 4.19). This time series shows significant
54 reductions over the past 90 years with most of the reductions occurring in the 1980s, and is an improvement
55 over that presented in AR4 in several ways, not least because the uncertainty estimates are explicitly derived
56 through the statistical analysis of multiple datasets. Snow cover decreases are largest in the spring period and
57 the rate of decrease increases with latitude in response to greater potential for albedo feedbacks (Dery and

1 Brown, 2007). June SCE loss is observed to be accelerating over the past decade (-21.5% loss per decade
2 over the 1979–2012 period) and exceeds the loss rate for minimum sea ice extent and CMIP5 model
3 projections (Derksen and Brown, in press). Averaged March and April NH SCE was around 8% lower (7
4 Mkm^2) over the period 1970–2010 than over the period 1922–1970. Viewed another way, the NOAA SCE
5 data indicate that, owing to earlier spring snowmelt, the duration of the snow season averaged over NH grid
6 points declined by 5.3 days per decade since winter 1972–1973 (Choi et al., 2010). Since AR4, the rate of
7 reductions in June SCE – both absolute and relative - has surpassed the rate of reduction of March–April SCE
8 (Figure 4.19).

9
10 In North America, Dyer and Mote (2006) used a gridded dataset of snow depth derived from observations for
11 1960–2000, finding minimal change in early winter and regional decreases beginning in late January. Over
12 Eurasia, in situ data show significant increases in winter snow accumulation but a shorter snowmelt season
13 (Bulygina et al., 2009). From analysis of passive microwave satellite data since 1979, significant trends
14 toward a shortening of the snowmelt season have been identified over much of Eurasia (Takala et al., 2009)
15 and the pan-Arctic region (Tedesco et al., 2009), with a trend toward earlier melt of about $+0.5$ days a^{-1} for
16 the beginning of the melt season, and about -1 day a^{-1} for the end of the melt season.

17 [INSERT FIGURE 4.19 HERE]

18 **Figure 4.19:** March–April NH snow cover extent (SCE, circles) over the period of available data, shown with the 13-
19 term filtered running mean and 95% confidence interval; and June SCE (x's, from satellite data alone). The width of the
20 smoothed confidence interval is also influenced by the interannual variability in SCE. Updated, from Brown and
21 Robinson (2011). For both time series the anomalies are calculated relative to the 1971–2000 mean.
22

23
24 The correlation between spring temperature and SCE (Figure 4.20) demonstrates that trends in spring SCE
25 are linked to rising temperature, and for a well-understood reason. The spring snow cover–albedo feedback
26 contributes substantially to the hemispheric response to rising greenhouse gases and provides a useful test of
27 GCMs (Fernandes et al., 2009) (see also Chapter 9). Indeed, the observed declines in land snow cover and
28 sea ice have contributed roughly the same amount to reductions in the surface energy fluxes, and the albedo
29 feedback of the NH cryosphere is likely in the range $0.3\text{--}1.1$ $\text{W m}^{-2} \text{K}^{-1}$ (Flanner et al., 2011). Brown et al.
30 (2010) used satellite, reanalyses and in situ observations to document variability and trend in Arctic spring
31 (May–June) SCE over the 1967–2008 period. In June, with Arctic albedo feedback at a maximum, SCE
32 decreased 46% and air temperature explains 56% of the variability; SCE and sea ice extent in June are both
33 significantly correlated to air temperature and decreased by similar amounts.

34 [INSERT FIGURE 4.20 HERE]

35 **Figure 4.20:** Relationship between NH April SCE and corresponding land area air temperature anomalies over 40°N –
36 60°N from the CRU dataset (Jones et al., 2012). Air temperature explains 48.7% of the variance. From Brown and
37 Robinson (2011).
38

39
40 For the SH, as noted above, there are no corresponding visible-wavelength satellite records, but microwave
41 data date from 1979. Foster et al. (2009) presented the first satellite study of variability and trends in any
42 measure of snow for South America, in this case SWE from microwave data. They focused on the May–
43 September period and noted large year-to-year variability and some lower frequency variability—the July
44 with most extensive snow cover had almost 6 times as much as the July with the least extensive snow
45 cover—but no trends.

46 4.5.3 Trends from In Situ Measurements

47
48 AR4 stimulated a review paper (Brown and Mote, 2009) that synthesized modelling results as well as
49 observations from many countries. They showed that decreases in various metrics of snow are most likely to
50 be observed in spring and at locations where temperatures are close to the freezing point, where changes in
51 temperature are most effective at reducing snow accumulation, increasing snowmelt, or both. However,
52 unravelling the competing effects of rising temperatures and changing precipitation remains an important
53 challenge in understanding and interpreting observed changes. Figure 4.21 shows a compilation of many
54 published trends observed at individual locations; data were obtained either from tables in the published
55 papers, or directly from the author, in some cases including updates to the published datasets. The figure
56 shows that in most studies, a majority of sites experienced declines during the varying periods of record, and
57

1 where data on site mean temperature or elevation were available, warmer/lower sites (red circles) were more
2 likely to experience declines.

3
4 Some in situ studies were not included in Figure 4.21; for example, because records were too short or too
5 few, or the study did not include relative changes. Ke et al. (2009) discussed snowfall trends by month at 25
6 stations in Qinghai province, China over 1957–2007; for annual mean snowfall; 5 stations showed
7 significant decreases, 2 showed significant increases, and 18 had insignificant changes. The most significant
8 trends were in May, with 11 decreases and no increases. Marty and Meister (2012) noted changes at high-
9 elevation (>2200 m) sites in the European Alps consistent with Figure 4.21: no change in SD in midwinter,
10 shortening of SCD in spring, and reduction in spring SWE and SD coincident with warming. For the
11 Pyrenees, Lopez-Moreno and Vicente-Serrano (2007) derived proxy SD for 106 sites since 1950 from actual
12 SD measurements since 1985 and weather measurements; they noted declines in spring SD that were related
13 to changes in atmospheric circulation. In the SH, six records in the Andes mostly show increases in
14 maximum SWE and four in Australia all show decreases in spring SWE over their respective periods of
15 record (Brown and Mote, 2009, and references therein) .

16 [INSERT FIGURE 4.21 HERE]

17 **Figure 4.21:** Summary of station trends in metrics of snow that, based on the work of Brown and Mote (2009), are (top
18 half) more reflective of mid-winter conditions and (bottom half) more reflective of spring conditions. See text for
19 definitions of abbreviations. Where symbols are circles, the quantity plotted is the percentage change of a linear fit
20 divided by the number of years of the fit. For the Bulygina study, the quantity plotted is the trend in cm a^{-1} (top) and #
21 days a^{-1} (bottom). Solid circles in the Skaugen study were statistically significant. Asterisks in the ‘N’ column
22 emphasize that Christy 2011 combined records from over 500 stations into 18 regions; none of the trends was
23 statistically significant. He judged time series from some regions to be unsuitable for statistical analysis and these are
24 indicated here by an ‘x’. For studies with more than 50 sites, the median, upper and lower quartiles are shown with
25 vertical lines. In a few cases, some trends lie beyond the edges of the graph; these are indicated by a numeral at the
26 corresponding edge of the graph, e.g., 2 sites $>2\% \text{ a}^{-1}$ for the Ishizaka study. Colours indicate temperature or, where
27 indicated, elevation using the lowest and highest site to set the colour scale. Note the prevalence of negative trends at
28 lower/warmer sites, especially in spring.

29 4.5.4 Changes in Snow Albedo

30
31 In addition to reductions in snow cover extent, which will reduce the mean reflectivity of particular regions,
32 the reflectivity (albedo) of the snow itself may also be changing in response to human activities. There are
33 two related causes of albedo change (Flanner et al., 2007): 1) darker snow grains that result from increased
34 combustion of both fossil fuels and northern forests, and 2) accelerated snow metamorphosis as a result of
35 higher temperatures. In addition, dust and vegetation can change snow albedo.

36
37 Unfortunately, there are extremely limited data on the changes of albedo over time, and we must rely instead
38 on analyses from ice cores, direct recent observations, and modelling. Flanner et al. (2007), using a detailed
39 snow radiative model coupled to a GCM and estimates of biomass burning in years with low (2001) and high
40 (1998) incidence of Arctic wildfire, estimated that the human-induced radiative forcing by interactions
41 between black carbon and snow cover is roughly 0.05 W m^{-2} , of which 80% is from fossil fuels. However,
42 spatially comprehensive surveys of impurities in Arctic snow in the late-2000s and mid-1980s suggested that
43 impurities decreased between those two periods (Doherty et al., 2010) and hence albedo changes have
44 probably not made a significant contribution to recent reductions in Arctic ice and snow.

45 4.5.5 River and Lake Ice

46
47 Many long-term observations of freshwater ice have been discontinued in recent decades (Prowse et al.,
48 2011). In the case of long-term lake and river sites in the Northern Hemisphere with ice-phenology records
49 longer than 100 years, Magnuson et al. (2000) reported that 38 of the 39 time series (1846–1995) showed
50 either later freeze-up (15 sites averaging $+0.06 \text{ days a}^{-1}$) or earlier break-up (24 sites averaging -0.058 days
51 a^{-1}), thus resulting in an average reduction in ice duration of 0.12 days a^{-1} . A subsequent analysis by (Benson
52 et al., 2012) for two periods (1855–2004; 1905–2004) found a continuing trend in Northern Hemisphere
53 lakes toward later freeze, earlier breakup, and shorter duration, with breakup overall changing more rapidly
54 than the freeze-up. Trends in the 150-year period were steeper than in the 100-year period but the most rapid
55 changes occurred in the most recent 30-year period with freeze-up 0.16 days a^{-1} later, breakup 0.19 days a^{-1}

1 earlier, and ice duration 0.43 days a^{-1} shorter. This recent period of most pronounced change was attributed
2 to more rapid increases in air temperature and an increasing sensitivity of lake and river ice, especially in
3 temperate regions, to rises in air temperature (Weyhenmeyer et al., 2011).
4

5 Although changes in timing of both ice break-up and freeze-up tends to be more sensitive to variations in air
6 temperature at lower latitudes than at higher latitudes (Livingstone et al., 2010), data obtained by remote
7 sensing of Canadian lakes (Latifovic and Pouliot, 2007) indicate that very high-latitude lakes appear to be
8 experiencing more rapid reductions in ice cover than those at lower latitudes. Specifically, while the majority
9 of all sites showed earlier break-up and delayed freeze-up (averaging -0.18 and $+0.12 \text{ days a}^{-1}$, respectively)
10 for the period 1950s to 2004, as well as faster rates (to averages of $-0.23 \text{ days a}^{-1}$ and $+0.16 \text{ days a}^{-1}$) for the
11 1970–2004 period, the most rapid rates of change (-0.99 d/a and $+0.76 \text{ days a}^{-1}$) occurred in six high-latitude
12 lakes (primarily on the Canadian Archipelago) for the even more recent period of 1985 to 2004. This
13 translates into an ice-cover reduction rate of 1.75 days a^{-1} , or about 4.5 times that found for the more
14 southern parts of Canada for the most rapid depletion period of 1970 to 2004. The degree to which this
15 reflects the more recent or higher-latitude warming, or potential differences in observational techniques, is
16 unclear (Prowse and Brown, 2010).
17

18 Studies of changes in river ice have used both disparate data and time intervals, ranging in duration from
19 multi-decade to over two centuries. Beltaos and Prowse (2009), summarizing most available information for
20 northern rivers, noted an almost universal trend towards earlier break-up dates but considerable spatial
21 variability in those for freeze-up, and noted too that changes were often more pronounced during the last few
22 decades of the twentieth century. They note that the 20th Century increase in mean air temperature of 2°C –
23 3°C in spring and autumn has produced in many areas a change of about 10 to 15 days toward earlier break-
24 up and later freeze-up, although the relationship with air temperatures is complicated by the roles of snow
25 accumulation and spring runoff. In summary, the limited studies available for freshwater ice indicate that
26 winter ice duration is contracting, with delays in autumn freeze-up proceeding more slowly than advances in
27 spring break-up, and there is some evidence of recent acceleration in both. The seasonal contrast in rates is
28 not as strong as for snow cover, which shows no trends in autumn and large trends in spring and early
29 summer.
30

31 **4.6 Frozen Ground**

32 **4.6.1 Background**

33 Frozen ground is a product of cold weather and climate and can be diurnal, seasonal, or perennial, but where
34 the ground is perennially frozen, and remains at or below 0°C for at least two consecutive years, it is called
35 permafrost (van Everdingen, 1998a). Permafrost occurs over land, sometimes referring to terrestrial
36 permafrost, and under oceans, often referring to subsea permafrost. In the following text, permafrost refers to
37 terrestrial permafrost unless specified. Changes in permafrost temperature and extent are sensitive indicators
38 of climate change (Osterkamp, 2007). The seasonal freezing and thawing of frozen ground, is coupled to the
39 land-surface energy and moisture balances, hence to the atmospheric system, and thus climate. Dramatic
40 changes in landscapes, ecosystems and hydrological processes can occur when permafrost degrades (Gruber
41 and Haeberli, 2007; Jorgenson et al., 2006; White et al., 2007). Furthermore, permafrost contains
42 considerable quantities of carbon, roughly twice the amount of carbon currently in the atmosphere (Tarnocai
43 et al., 2009). Therefore, permafrost thawing, which increases organic matter in the active layer and newly-
44 developed taliks, exposes frozen carbon to microbial degradation, releasing CO_2 and CH_4 into the
45 atmosphere (Schaefer et al., 2011; Schuur et al., 2009; Zimov et al., 2006) (for a detailed assessment of this
46 issue, see Chapter 6). Permafrost degradation also affects the lives of people, both in northern and high-
47 mountain areas, through dramatic changes in landscape, vegetation and impacts on infrastructure.
48
49
50

51 **4.6.2 Changes in Permafrost**

52 **4.6.2.1 Permafrost Temperature**

53 The temperature of permafrost is the key parameter that determines its physical and thermal state. In this
54 section, permafrost temperatures usually refer to mean annual temperatures at certain depths below the
55 permafrost table unless specified. Mean annual ground temperatures refer to the temperature at the depth of
56
57

1 zero annual amplitude. The depth of zero annual amplitude varies from several meters to 20 m depending on
 2 mainly annual amplitude of ground surface temperatures and soil properties. However, it has been reported
 3 that the difference between the mean annual permafrost temperature above the depth of zero amplitude and
 4 the mean annual ground temperature is usually very small (Romanovsky et al., 2010b). In this case, mean
 5 annual temperatures within 20 m depth from the permafrost table are comparable. In the Antarctic,
 6 permafrost temperature has been observed as low as -23.6°C (Vieira et al., 2010), but in the northern
 7 hemisphere, it ranges from -15°C to within a few tenths of a degree from the freezing point (Figure 4.22)
 8 (Romanovsky et al., 2010a). Permafrost temperatures are usually lowest in high Arctic regions and gradually
 9 increase southwards in the northern hemisphere, but substantial difference does occur at the same latitude.
 10 For example, due to the effect of warm ocean currents, the southern limit of permafrost is farther north
 11 (Brown et al., 1998), and permafrost temperature is higher in Scandinavia, and north-western Russia, than it
 12 is in Arctic regions of Siberia and North America (McBean et al., 2005). The southern limit of permafrost
 13 refers to the southernmost limit of the occurrence of permafrost over a region. Permafrost can occur at high
 14 altitude as far south as 26°N in the Himalayas (Brown et al., 1998). Site-specific factors, such as slope aspect,
 15 snow cover, vegetation cover, soil type and moisture content control permafrost distribution and temperature.

16
 17 In Russia, permafrost temperature measurements reach back to the early-1930s (Romanovsky et al., 2010b),
 18 in North America to the late-1940s (Brewer, 1958), and in China to the early-1960s (Zhou et al., 2000).
 19 Systematic measurements, however, mostly began in the late-1970s and early-1980s (Osterkamp, 2007;
 20 Smith et al., 2010; Zhou et al., 2000).

21 [INSERT FIGURE 4.22 HERE]

22 **Figure 4.22:** Time series of mean annual ground temperatures at depths between 10 and 20 m for boreholes throughout
 23 the circumpolar northern permafrost regions (Romanovsky et al., 2010a). Data sources for North American, Russian
 24 and Nordic sites are Smith et al. (2010), Romanovsky et al. (2010b) and Christiansen et al. (2010), respectively. **C**
 25 Canadian site; **A** Alaskan site; **R** Russian site. The Svalbard site is Janssonhaugen (PACE-10) (Isaksen et al., 2007).
 26 Measurement depth for Russian boreholes and 85–8A is 10 m, Gulkana, Oldman and Alert are 15 m, and 20 m for all
 27 other boreholes. Borehole locations are: ZS-124 – 67.48°N 063.48°E ; 85-8A – 61.68°N 121.18°W ; Gulkana – 62.28°N
 28 145.58°W ; YA-1 – 67.58°N 648°E ; Oldman – 66.48°N 150.68°W ; Happy Valley – 69.18°N 148.88°W ; Svalbard –
 29 78.28°N 016.58°E ; Deadhorse – 70.28°N 148.58°W ; West Dock – 70.48°N 148.58°W ; Alert – 82.58°N 062.48°W .

30
 31
 32 Permafrost temperatures have generally increased during the past three decades, although at some sites, they
 33 show little change, or slight decrease (Figure 4.22; Table 4.7). In this regard, it is important to discriminate
 34 *cold permafrost*, with mean annual ground temperature below -2°C at depths from 10 to 15 m (Cheng and
 35 Wu, 2007; Smith et al., 2010; Wu and Zhang, 2010) from *warm permafrost* or those with mean temperature
 36 of -2°C or higher at the depth of zero annual amplitude (Smith et al., 2010). Generally speaking, warm
 37 permafrost is mostly found in the discontinuous permafrost zone, while cold permafrost exists in the
 38 continuous permafrost zone although it is possible that cold permafrost may also occur in the discontinuous
 39 permafrost zone (Romanovsky et al., 2010a). Overall, permafrost temperature increases are generally greater
 40 in cold permafrost than in warm permafrost, especially if the permafrost is ice rich, due to latent heat effect
 41 (Riseborough, 1990; Romanovsky et al., 2010a). Temperatures of cold permafrost have increased by $2.0 -$
 42 3.0°C during the last three decades (Table 4.7). The majority of warming occurred between the early-1980s
 43 and the late-1990s, with small changes during the 2000s. Temperatures of warm permafrost also increased,
 44 but generally by less than 1.0°C . Near-isothermal conditions of warm permafrost are often observed in
 45 mountain permafrost regions such as the European Alps (Noetzli and Vonder Muehll, 2010), Scandinavia
 46 (Christiansen et al., 2010), the Western Cordillera of North America (Lewkowicz et al., 2011; Smith et al.,
 47 2010), the Qinghai-Tibetan Plateau (Cheng and Wu, 2007; Wu et al., 2012; Wu and Zhang, 2008; Zhao et
 48 al., 2010) and in the southern margins of discontinuous permafrost regions at high latitudes (Romanovsky et
 49 al., 2010b; Smith et al., 2010). In these sites, permafrost temperatures have shown little or no change,
 50 indicating that permafrost is thawing internally (Riseborough, 1990). Episodic cooling as part of temperature
 51 fluctuations has been observed but is usually been short-lived and controlled by site-specific conditions
 52 (Marchenko et al., 2007; Noetzli and Vonder Muehll, 2010; Wu and Zhang, 2008; Zhao et al., 2010).

53
 54
 55 **Table 4.7:** Permafrost temperatures during the International Polar Year (2007–2009) and their recent changes.

Region	Permafrost Temperature During IPY ($^{\circ}\text{C}$)	Permafrost Temperature Change ($^{\circ}\text{C}$)	Depth (m)	Period of Record	Source
--------	--	--	-----------	------------------	--------

<i>North America</i>					
Northern Alaska,	−5.0 to −10.0	1.9–3.1	10–20	Early-1980s–2009	Osterkamp, 2007; Smith et al., 2010
Mackenzie Delta	−5.0 to −8.0	1.0–2.0	12–20	Late-1960s–2009	Burn and Kokelj, 2009; Burn and Zhang, 2009; Smith et al., 2010
Canadian Arctic Archipelago	−10.0 to −14.9	2.0–3.0	12–15	1978–2008	Smith et al., 2010
Interior of Alaska,	0.0 to −5.0	0.0–2.0	15–20	1985–2009	Osterkamp, 2008
Central and Southern Mackenzie Valley	−0.3 to −5.0	0.0–0.5	10–12	1984–2008	Burn and Kokelj, 2009; Smith et al., 2010
Northern Quebec	>−2.6	0.0–1.0	12–18	1989–2008	Allard et al., 1995; Smith et al., 2010
<i>Europe</i>					
Russian European North	−0.1 to −4.1	0.3–2.0	8–22	1971–2010	Malkova, 2008; Oberman, 2008; Oberman, 2012; Romanovsky et al., 2010b
Nordic Countries	−0.1 to −5.6	0.0–1.0	2–15	1999–2009	Christiansen et al., 2010; Isaksen et al., 2011
<i>Northern and Central Asia</i>					
Northern Yakutia	−4.3 to −10.8	0.5–1.5	14–25	Early-1950s–2009	Romanovsky et al., 2010b
Trans-Baykal region	−4.7 to −5.1	0.5–0.8	19–20	Late-1980s–2009	Romanovsky et al., 2010b
Qinghai-Xizang Plateau	−0.2 to −3.4	0.2–0.7	6	1996–2010	Cheng and Wu, 2007; Li et al., 2008; Wu and Zhang, 2008; Zhao et al., 2010
Tian Shan	−0.4 to −1.1	0.3–0.9	10–25	1974–2009	Marchenko et al., 2007; Zhao et al., 2010
Mongolia	0.0 to < −2.0	0.2–0.6	10–15	1970–2009	Ishikawa et al., 2012; Sharkhuu et al., 2007; Zhao et al., 2010
<i>Others</i>					
Antarctic	−8.3 to −23.6	–	2–15	2007–2009	Vieira et al., 2010
East Greenland	−8.1	–	3.25	2008–2009	Christiansen et al., 2010

Permafrost warming is mainly in response to increased air temperature and changing snow cover (see Box 4.1). In cold permafrost regions, especially in tundra regions with low ground ice content such as bedrocks, where permafrost warming rates have been fastest, changes in snow cover conditions may play an important role (Smith et al., 2010; Zhang, 2005). In forested areas with relatively warm permafrost, especially ice-rich warm permafrost, changes in permafrost temperature are relatively small due to the effects of the surface buffer layer and latent heat (Isaksen et al., 2011; Riseborough, 1990; Romanovsky et al., 2010a).

4.6.2.2 Permafrost Degradation

Permafrost degradation refers to any decrease in thickness and/or areal extent. In particular, the degradation may be manifested by the thickening of the active layer, or top-down or bottom-up thawing, talik development (areas of unfrozen ground within permafrost), or the poleward migration of permafrost boundaries. Permafrost degradation can be identified through geomorphologic indicators including thermokarst terrain (Jorgenson et al., 2006), expansion of thaw lakes (Sannel and Kuhry, 2011), active-layer detachment slides along slopes, and rock falls (Raveland et al., 2010), destabilized rock glaciers (Delaloye et al., 2011). Although, most permafrost has been degrading since the Little Ice Age (Halsey et al., 1995), the trend has been relatively modest until the past two decades when acceleration of degradation has been observed (Romanovsky et al., 2010b).

Significant permafrost degradation has been reported in the Russian European North. Permafrost with thickness of 10 to 15 m completely thawed in the period 1975–2005 in the Vorkuta area (Oberman, 2008), while the southern permafrost boundary moved north by about 80 km and the boundary of continuous

1 permafrost has moved north by 15–50 km (Oberman, 2008). Taliks have also developed in relatively thick
2 permafrost during the past several decades. In the Vorkuta region, the thickness of existing closed taliks
3 increased by 0.6 to 6.7 m over the past 30 years (Romanovsky et al., 2010b). Permafrost thawing and talik
4 formation is occurring in the Nadym and Urengoy regions in north western Russian (Drozdo et al., 2010).
5 Long-term permafrost thawing has been reported around the city of Yakutsk, but this was due mainly to
6 forest fires or human disturbance (Fedorov and Konstantinov, 2008). Permafrost degradation has also been
7 reported on the Qinghai-Xizang (Tibet) Plateau (Cheng and Wu, 2007; Li et al., 2008).

8
9 Permafrost degradation has caused erosion and an accelerated retreat of many Arctic coasts in recent years
10 (Jones et al., 2009). This implies a transformation of some cold terrestrial permafrost that is immersed in
11 seawater. Such cold permafrost immediately starts to degrade under the influence of both thermal and
12 chemical impact of overlying sea water (Rachold et al., 2007) and geothermal heat flux (Romanovskii et al.,
13 2004). Subsea permafrost degradation rates (from above) have been estimated to be 1–20 cm a⁻¹ on the East
14 Siberian Shelf (Overduin et al., 2007) and 1–4 cm a⁻¹ in the Alaskan Chukchi Sea (Overduin et al.,
15 submitted). Similar impacts arise for permafrost beneath new thaw lakes, the number and area of which is
16 increasing (Sannel and Kuhry, 2011; van Huissteden et al., 2011). In northern Alaska, estimates of
17 permafrost thawing under thaw lakes are in the range 0.9–1.7 cm a⁻¹ (Ling and Zhang, 2003), which is
18 consistent with the rate of subsea permafrost thawing.

19
20 During recent years, destabilized rock glaciers have received increased attention by researchers. Rock glacier
21 refers to a mass of rock fragments and finer material, on a slope, that contains either interstitial ice or an ice
22 core and shows evidence of past or present movement (van Everdingen, 1998b). Time series acquired during
23 recent decades by terrestrial surveys indicate dramatic speed-up of some rock glaciers as well as seasonal
24 velocity changes related with ground temperatures (Bodin et al., 2009; Delaloye et al., 2011; Noetzli and
25 Vonder Muehll, 2010; Schoeneich et al., 2010). Photo comparison and photogrammetry indicates an
26 increased activity and collapse-like features on some rock glaciers (Roer et al., 2008). The clear relationship
27 between mean annual air temperature at the rock glacier front and rock glacier velocity points to a
28 temperature dependence and thus, a plausible causal connection to climate (Kaab et al., 2007). Strong
29 surface lowering of rock glaciers has been reported in the Andes (Bodin et al., 2010), indicating melting of
30 ground ice in rock glaciers and permafrost degrading.

31
32 Many rock fall events have originated from permafrost slopes during recent years (Ravanel and Deline,
33 2011; Ravanel et al., 2010). Increasing evidence based on exposed ice and on event statistics supports the
34 hypothesis that this is in part due to thaw of permafrost on steep slopes (Gruber and Haeberli, 2007).

35 36 **4.6.3 Subsea Permafrost**

37
38 Subsea permafrost is similar to its terrestrial counterpart, but lies beneath the coastal seas and the ocean.
39 Subsea permafrost in the Arctic is generally relict terrestrial permafrost (Vigdorichik, 1980), inundated after
40 the last glaciation and now degrading under the overlying shelf sea. Permafrost may, however, also form
41 when the sea is shallow, permitting sediment freezing through bottom-fast ice in the winter (Solomon et al.,
42 2008). One 76-year time series of bottom water temperature in the Laptev Sea (Dmitrenko et al., 2011)
43 showed warming of 2.1°C since 1985 in the near-shore zone (<10 m water depth), as lengthening summers
44 reduced sea ice extent and increased solar heating. Increasing permafrost degradation rates due to warming
45 Arctic shelf bottom water have not been directly observed.

46
47 Circum-arctic subsea permafrost is regarded as a substantial reservoir and/or a confining layer for gas
48 hydrates (O'Connor et al., 2010), roughly estimated as containing 2–65 Pg of CH₄ hydrate (McGuire et al.,
49 2009). The gas hydrate stability zone (GSHZ) relies on subsea permafrost, so degradation of subsea
50 permafrost could destabilize the GSHZ (Romanovskii et al., 2005) and potentially release CH₄ into the
51 atmosphere. Observations of gas trapped in subsea permafrost on the East Siberian Shelf (Shakhova et al.,
52 2010a) and high methane concentrations in water-column and air above (Shakhova et al., 2010a; Shakhova
53 et al., 2010b) have led to the suggestion that thawed zones act as pathways through permafrost for gas
54 release. Much current debate focusses on whether or not measured methane emission rates from Arctic shelf
55 seas represent an increase (Petrenko et al., 2010), thus a link between changes in subsea permafrost and
56 climate change remains speculative.

4.6.4 Changes in Seasonally-Frozen Ground

Seasonally-frozen ground is a soil layer that freezes and thaws annually, which may, or may not, overlies terrestrial permafrost, and also includes portions of the Arctic seabed that freeze in winter.

4.6.4.1 Changes in Active-Layer Thickness

The active layer is that portion of the soil above permafrost that thaws in summer and re-freezes in winter. Observations have revealed a general positive trend in active-layer thickness (ALT) of discontinuous permafrost regions at high latitudes. Based on measurements from the International Permafrost Association (IPA) Circumpolar Active Layer Monitoring (CALM) program, active-layer thickening has been observed since the 1970s and has accelerated since 1995 in northern Europe (Akerman and Johansson, 2008; Callaghan et al., 2010), and on Svalbard and Greenland since the late-1990s (Christiansen et al., 2010). ALT has increased significantly in the Russian European North (Mazhitova, 2008), East Siberia (Fyodorov-Davydov et al., 2008), and Chukotka (Zamolodchikov, 2008) since the mid-1990s (Figure 4.23). ALT has been observed over discontinuous permafrost regions in the interior of Alaska during the past two decades with a slight positive trend (Figure 4.23b). Increase of 8 cm in ALT between 1983 and 2008 has been observed in the northern portion of the Mackenzie River Valley (Burn and Kokelj, 2009) with large inter-annual variability over the region (Smith et al., 2009). ALT has increased since the mid-1990s in the eastern portion of the Canadian Arctic with the largest increase occurring in the bedrock of the discontinuous permafrost zone (Smith et al., 2010).

[INSERT FIGURE 4.23 HERE]

Figure 4.23: Locations for the Circumpolar Active Layer Monitoring (CALM) sites (Brown et al., 2000) and the Russian Hydrometeorological Stations (Frauenfeld et al., 2004; Zhang et al., 2005) (a), and changes in active layer thickness in Northern America (b), Northern Asia (c), European North (d), and Siberia (e). ALT data for Northern America, Northern Asia and European North are obtained from the International Permafrost Association (IPA) CALM website at <http://www.udel.edu/Geography/calm/about/permafrost.html>. Data for Siberia stations are obtained from the Russian Hydrometeorological Stations (RHM). Figure 4.23 (e) shows a composite of ALT changes extracted from RHM stations soil temperature data. The number of RHM stations has expanded from 31 stations as reported from Frauenfeld et al. (2004) and Zhang et al. (2005) to 44 stations and the time series has extended from 1990 to 2008. The red star represents the mean composite value, the shaded area indicates the standard deviation, and the line is the trend.

Over the Qinghai-Tibetan Plateau, ALT has increased of about 7.8 cm yr^{-1} over a period from 1995 through 2010 (Wu and Zhang, 2010; Zhao et al., 2010). Rates of up to 4.0 cm yr^{-1} were observed in Mongolian sites characterized by warm permafrost during the past decade (Sharkhuu et al., 2007). A clear trend of increasing ALT was also detected in Tian Shan (Marchenko et al., 2007; Zhao et al., 2010), and in the European Alps, changes in ALT were largest in response to years with hot summers, although a strong dependence on surface and subsurface characteristics was noted (Noetzli and Vonder Muehll, 2010).

Changes in ALT on the Alaskan North Slope, displayed no trend from 1993–2010 (Shiklomanov et al., 2010; Streletskiy et al., 2008), with similar results noted in the Mackenzie Valley (Smith et al., 2009) and in the West Siberia (Vasiliev et al., 2008) since the mid-1990s (Figure 4.23b). Low rates of increase or no change in ALT increase occurred in shallow active-layer areas over ice-rich and cold permafrost on the Qinghai-Tibetan Plateau (Zhao et al., 2010). Little or no observed trend in ALT change may be in part explained by observed surface subsidence. Thaw penetration into ice-rich permafrost at the base of the active layer is often accompanied by loss of volume due to thaw consolidation manifested as a ground surface subsidence. Results from ground-based measurements at selected sites on the North Slope of Alaska indicate 11–13 cm in surface subsidence over the period 2001–2006 (Streletskiy et al., 2008), 4–10 cm from 2003 to 2005 in the Brooks Range (Overduin and Kane, 2006), and up to 20 cm in the Russian European North (Mazhitova and Kaverin, 2007).

Subsidence has also been identified using space-borne interferometric synthetic aperture radar (InSAR) data. Liu et al. (2010) detected surface deformation over permafrost on the North Slope of Alaska during the 1992–2000 thaw seasons and a long-term surface subsidence of 1–4 cm per decade. These results could explain why in situ measurements at some locations reveal negligible trends in ALT changes during the past two decades, despite the fact that atmospheric and permafrost temperatures increased during that time.

4.6.4.2 Changes in Seasonally Frozen Ground in Areas not Underlain by Permafrost

An estimate based on monthly mean soil temperatures from 387 stations across Russia, suggested that the thickness of seasonally frozen ground decreased by about 0.32 m in the period 1930–2000 (Figure 4.24) (Frauenfeld and Zhang, 2011). Inter-decadal variability was such that no trend could be identified late-1960s, after which seasonal freeze depths decreased significantly until the early-1990s. From then, until about 2008, no further change was evident. Such changes are closely linked with the freezing index, but also mean annual air temperatures, and snow depth (Frauenfeld and Zhang, 2011).

Thickness of seasonally frozen ground in western China decreased by 20–40 cm since the early-1960s (Li et al., 2008). Evidence from the satellite record indicates that the onset dates of spring thaw advanced by 14 days, while the autumn freeze date was delayed by 10 days on the Qinghai-Xizang (Tibetan) Plateau from 1988 through 2007 (Li et al., 2012)

[INSERT FIGURE 4.24 HERE]

Figure 4.24: Top: distribution of hydrometeorological stations across Russia; bottom: changes in thickness of seasonally frozen ground (From, Frauenfeld and Zhang, 2011).

[START FAQ 4.1 HERE]

FAQ 4.1: Are Glaciers in Mountain Regions Disappearing?

In many mountain ranges around the world, glaciers are disappearing in response to the atmospheric temperature increases of past decades. In the Swiss Alps, for example, more than one hundred mostly very small glaciers disappeared between about 1850 and the 1970s. Disappearing glaciers have also been reported from the Austrian Alps, the Pyrenees, the Jotunheimen region in Norway, the North Cascades in the USA, the Tibetan Plateau, Irian Jaya in Indonesia, and Bolivia. If atmospheric warming continues through the 21st century, many more glaciers will inevitably disappear. It is also likely that some mountain ranges will lose most, if not all, of their glaciers.

In all mountain regions where glaciers exist today, glacier volume has decreased considerably over the past 150 years. Since that time, many small glaciers have disappeared. With some local exceptions, glacier shrinkage was globally widespread and particularly strong during the 1940s and since the 1980s. However, there were phases of relative stability during the 1890s, 1920s and 1970s. Conventional ground measurements—and increasingly, airborne and satellite measurements—offer robust evidence that, in most mountain regions, the rate of glacier shrinkage was higher over the past two decades than previously, and that glaciers continue to shrink. This picture is the same in most glacierised mountain regions of the world, apart from a few exceptional regions subject to special local conditions, such as the western coast of New Zealand, and the Karakoram in Asia.

It takes several decades for a glacier to adjust its extent to a change in climate, so most glaciers are currently larger than they would be if they were in balance with current climatic conditions. Because the time lag for the adjustment increases with glacier size, larger glaciers will continue to shrink over the next few decades, even if temperatures stabilise. Smaller glaciers will also continue to shrink, but they will adjust their extent faster.

Many factors influence the future development of each glacier, and whether it will eventually disappear: for instance, its size, slope, elevation range, distribution of area with elevation, and its surface characteristics, such as whether or not a glacier tongue is debris covered. Those factors vary substantially from region to region, and also between neighbouring glaciers. External factors, such as the surrounding topography and the climatic regime, are also important for future glacier evolution. In detail and over shorter time scales, each glacier responds differently to climate change.

Over time scales longer than 50 years, the response is more coherent, and robust modelling approaches can be applied to determine long-term trends in glacier development. Such models are built on an understanding of the basic physical principles. For example, an increase in local mean air temperature, with no change in precipitation, will cause an upward shift of the equilibrium line altitude (ELA) by about 150 m for each

1 degree Celsius of atmospheric warming. Such a shift and its consequences for glaciers of different size is
2 illustrated in FAQ 4.1, Figure 1.

3
4 As the ELA shifts upwards, the accumulation area of the glacier shrinks (FAQ 4.1, Figure 1a). At the same
5 time, the ablation area expands, thus increasing the area over which ice is lost through melt (FAQ 4.1, Figure
6 1b). This imbalance between accumulation and ablation results in an overall loss of ice from the glacier.
7 After several years, the glacier front retreats, and the ablation area decreases in size until the glacier has
8 adjusted its extent to the new climatic conditions (FAQ 4.1, Figure 1c). Where climate change is sufficiently
9 strong to raise the ELA persistently above the highest point on a glacier (FAQ 4.1, Figure 1b, right), the
10 glacier will eventually disappear entirely (FAQ 4.1, Figure 1c, right). Higher glaciers, which still have an
11 accumulation area under these conditions, will shrink, but not disappear (FAQ 4.1, Figure 1c, left and
12 middle). A large valley glacier might lose much of its tongue, probably leaving a lake in its place (FAQ 4.1,
13 Figure 1c, left). Besides air temperature, changes in the quantity and seasonality of precipitation influence
14 the ELA as well.

15
16 Many observations have confirmed that different glacier types indeed respond differently to recent climate
17 change. For example, the flat, low-lying tongues of large valley glaciers (such as in Alaska, Canada or the
18 Alps) currently show the strongest mass losses, largely independent of aspect, shading or debris cover. This
19 type of glacier is slow in adjusting its extent to new climatic conditions and mainly reacts by thinning
20 without substantial terminus retreat. In contrast, smaller mountain glaciers with more or less constant slopes
21 adjust more quickly to the new climate by changing the size of their ablation area more constantly. (FAQ
22 4.1, Figure 1c, middle).

23
24 The long-term response of most glacier types can be determined very well with the approach illustrated in
25 FAQ 4.1, Figure 1. However, modelling the short-term response, or more complex glacier types (heavily
26 debris-covered, surging, or calving into water) is difficult, and requires detailed knowledge of further glacier
27 characteristics, such as mass balance sensitivity, ice thickness distribution, and internal hydraulics. For the
28 majority of glaciers world-wide, these are largely unknown, and the response to climate change is thus
29 difficult to model.

30
31 The Karakoram-Himalaya mountain range, for instance, has a large variety of glacier types and climatic
32 conditions, and glacier characteristics are still very poorly known. This makes determining their future
33 evolution particularly uncertain. However, gaps in knowledge are expected to decrease substantially in
34 coming years, thanks to increased use of satellite data (e.g., to compile glacier inventories or derive flow
35 velocities) and extension of the ground-based measurement network.

36
37 In summary, the fate of glaciers in the mountain regions of the world will be highly variable, depending on
38 both their specific characteristics and future climate conditions. Some individual glaciers will disappear,
39 while others will lose most of their low-lying portions. Currently, glaciers are determined to disappear where
40 the ELA is already above the highest glacier elevation. In the future, glaciers will also disappear in regions
41 where the ELA will rise above that elevation as a consequence of ongoing climate change.

42 43 **[INSERT FAQ 4.1, FIGURE 1 HERE]**

44 **FAQ 4.1, Figure 1:** Schematic of three types of glaciers located at different elevations, and their response to an upward
45 shift of the equilibrium line altitude (ELA). a) For a given climate, the ELA has a specific altitude (ELA1), and all
46 glaciers have a specific size. b) Due to a temperature increase, the ELA shifts upwards to a new altitude (ELA2),
47 initially resulting in reduced accumulation and larger ablation areas for all glaciers. c) After glacier size has adjusted to
48 the new ELA, the valley glacier (left) has lost its tongue and the small glacier (right) has disappeared entirely.

49 50 **[END FAQ 4.1 HERE]**

51 52 **[START FAQ 4.2 HERE]**

53 54 **FAQ 4.2: How is Sea Ice Changing in the Arctic and Antarctic?**

55
56 *The sea ice covers on the Arctic Ocean and on the Southern Ocean around Antarctica have quite different*
57 *characteristics, and are showing different changes with time. Over the last 32 years, there has been a*
58 *significant trend of -3.9% per decade in the annual average extent of sea ice in the Arctic. The average*

1 *winter thickness of Arctic Ocean sea ice has thinned by 1.8 m between 1978 and 2008, and the total volume*
2 *(mass) of Arctic sea ice has decreased significantly at all times of year. The more rapid decrease in the*
3 *extent of sea ice at the summer minimum is a consequence of these trends. In contrast, over the same 32-year*
4 *period, the total extent of Antarctic sea ice show a small but statistically significant increase of 1.4% per*
5 *decade, but there are strong regional differences in the changes around the Antarctic. Measurements of*
6 *Antarctic sea ice thickness are too few to be able to judge whether its total volume (mass) is decreasing,*
7 *steady, or increasing.*

8
9 A large part of the total Arctic sea ice cover lies above 60°N (FAQ 4.2, Figure 1) and is surrounded by land
10 to the south with openings to the Canadian Arctic Archipelago, and the Bering, the Barents, and Greenland
11 seas. A fraction of the ice within the Arctic basin survives for several seasons, growing in thickness by
12 freezing of seawater at the base and by deformation (ridging and rafting). Seasonal sea ice grows to only
13 about 2 m in thickness but sea ice that is more than one year old (multiyear ice) can be several metres
14 thicker. Arctic sea ice drifts within the basin, driven by wind and ocean currents: the mean drift pattern is
15 dominated by a clockwise circulation pattern in the western Arctic and a Transpolar Drift Stream that
16 transports Siberian sea ice across the Arctic and exports it from the basin through the Fram Strait.

17
18 Satellites with the capability to distinguish ice and open water have provided a picture of the changes of the
19 sea ice cover. Since 1979, the annual average extent of ice in the Arctic has decreased by 3.9% per decade.
20 The decline in extent at the end of summer (in late-September) has been even greater at 12% per decade,
21 reaching a record minimum in 2007. The decadal average extent of the September minimum Arctic ice cover
22 has decreased for each decade since satellite records began. Submarine and satellite records suggest that the
23 thickness of Arctic ice, and hence the total volume, is also decreasing. This is occurring because of loss of
24 the thicker multiyear ice: approximately 17% of this type of sea ice per decade has been lost to melt and
25 export out of the basin since 1979 and 40% since 1999. While the area of Arctic sea ice coverage can
26 fluctuate from year to year because of variable seasonal production, the proportion of thick multiyear ice,
27 and the total sea ice volume, can only recover slowly.

28
29 Unlike the Arctic, the sea ice cover around Antarctica is constrained to latitudes north of 78°S because of the
30 presence of the continental land mass. The Antarctic sea ice cover is largely seasonal, with an average
31 thickness of only about 1 m at the time of maximum extent in September. Only a small fraction of the ice
32 cover survives the summer minimum in February, and very little Antarctic sea ice is more than two years
33 old. The ice edge is exposed to the open ocean and the snowfall rate over Antarctic sea ice is higher than in
34 the Arctic. When the snow load from snowfall is sufficient to depress the ice surface below sea level,
35 seawater infiltrates the base of the snow pack and snow-ice is formed when the resultant slush freezes.
36 Consequently, snow-to-ice conversion (as well as basal freezing as in the Arctic) contributes to the seasonal
37 growth in ice thickness and total ice volume in the Antarctic. Snow-ice formation is sensitive to changes in
38 precipitation and thus changes in regional climate. The consequence of changes in precipitation on Antarctic
39 sea ice thickness and volume remains a focus for research.

40
41 Unconstrained by land boundaries, the latitudinal extent of the Antarctic sea ice cover is highly variable.
42 Near the Antarctic coast, sea ice drift is predominantly from east to west, but further north, it is from west to
43 east and highly divergent. Distinct clockwise circulation patterns that transport ice northward can be found in
44 the Weddell and Ross Seas, while the circulation is more variable around East Antarctica. The northward
45 extent of the sea ice cover is controlled in part by the divergent drift that is conducive in winter months to
46 new ice formation in persistent open water areas (polynyas) along the coastlines. These zones of ice
47 formation result in saltier and thus denser ocean water and become one of the primary sources of the deepest
48 water found in the global oceans.

49
50 Over the 32-year satellite record, the increase of 1.4% per decade in total extent of Antarctic sea ice is small
51 compared to natural variability. However, there are large regional differences in trends with decreases seen
52 in the Bellingshausen and Amundsen seas, but a significant increase in sea ice extent in the Ross Sea that
53 dominates the overall trend. Whether the small overall increase in Antarctic sea ice extent is meaningful as
54 an indicator of climate is uncertain because the extent varies so much from year to year and from place to
55 place around the continent. Without better ice thickness and ice volume estimates, it is difficult to
56 characterize how Antarctic sea ice cover is responding to changing climate, or which climate parameters are
57 most influential.

1
2 There are large differences in the physical environment and processes that affect the state of Arctic and
3 Antarctic sea ice cover and contribute to their dissimilar responses to climate change. The long, and
4 unbroken, record of satellite observations have provided a clear picture of the decline of the Arctic sea ice
5 cover, but available evidence precludes us from making robust statements about overall changes in Antarctic
6 sea ice and their causes.

7
8 **[INSERT FAQ 4.2, FIGURE 1 HERE]**

9 **FAQ 4.2, Figure 1:** The mean circulation pattern of sea ice and the decadal trends (%) in annual average ice extent in
10 different sectors of the Arctic and Antarctic. The average sea ice cover for the period 1979 through May 2011, from
11 satellite observations, at maximum (minimum) extent is shown as light (dark) grey shading.

12
13 **[END FAQ 4.2 HERE]**

14 15 **4.7 Synthesis**

16
17 The cryosphere has been undergoing substantial changes during the last few decades. In situ and shorter-term
18 satellite observations show that practically all elements of the cryosphere are showing signals consistent with
19 surface warming (Figure 4.25). The magnitude of the change, however, depends on location with some
20 regions showing dramatic changes while other regions show only minor, and sometimes opposite, changes.
21 In the Arctic, among the most visible warming signals during the period 1979 to 2011, was the rapid decline
22 (-15% per decade) of the extent of the thick component of the sea ice cover (multiyear ice). Overall, the
23 Arctic ice cover thinned by 48% since 1980, with a decrease in total ice volume. The total Antarctic sea ice
24 extent, on the other hand, is increasing at the rate of about 1.4% per decade. This slight positive trend
25 appears counter-intuitive, but in part it reflects the large regional variability in the observed trends of global
26 surface temperature. The global extent of sea ice cover has an overall negative trend of -1.4% per decade.

27
28 The assessment of changes in the volume of glaciers in AR4 was based on data from an estimated 42% of the
29 world's glaciers. Since AR4, a new globally complete vector dataset of glacier outlines has been compiled
30 and has made it possible to provide a more complete evaluation of global glacier mass and its changes
31 through time. Overall, the volume of glaciers has declined considerably with the magnitude of the changes
32 varying regionally. In recent years, the most drastic changes have occurred in Alaska, the Canadian
33 Archipelago, the European Alps and the Southern Andes including Patagonia. Most of the worldwide
34 glaciers are declining, and some have already completely disappeared.

35
36 The contribution to sea level rise from the Greenland and Antarctic ice sheets over the period 1993–2010,
37 has been about 6 mm and 3 mm, respectively. The reliability of observations of ice loss from the ice sheets
38 of Greenland and Antarctica has been enhanced with the introduction of advanced techniques including those
39 using more precise and more reliable satellite sensors (e.g., GRACE; AMSR-E, ICESat). The uncertainties in
40 the estimates have been significantly reduced and although some disagreements remain between different
41 techniques and analyses, the trends are broadly consistent with the ice loss over the period 1993–2010
42 averaging $123 \pm 22 \text{ Gt yr}^{-1}$ from Greenland, and $64 \pm 33 \text{ Gt yr}^{-1}$ from Antarctica. The ice loss for the later
43 period, 2005–2010, averaged $228 \pm 54 \text{ Gt yr}^{-1}$ from Greenland and $112 \pm 58 \text{ Gt yr}^{-1}$ from Antarctica.

44
45 The global extent of snow cover and its thickness have been observed using a combination of in situ and
46 satellite data to have decreased significantly during the period 1922–2010. The trends in snow cover also
47 vary considerably in different regions and with different seasons. The steepest declines have been observed
48 during spring (8% during the period 1922–2010). Trends in extent and thickness, are however, sometimes
49 positive, especially at high elevations where some cooling has been observed.

50
51 Permafrost continues to experience warming and thawing due to increasing surface temperature and changes
52 in snow cover. Permafrost temperatures have increased up to 3°C during the last 30 to 40 years, in some
53 regions of Northern Alaska, Canadian Archipelago, Eastern Russia, and Qinghai-Xizang (Tibet) Plateau. The
54 southern limit of permafrost in European North of Russia has moved northward and new taliks have
55 developed during the last 30 years. At the same time, active layer thickness (ALT) has increased by up to 90
56 cm. The areal extent of seasonally frozen ground (SFG) continues to decrease, and across the Eurasian
57 continent, its thickness has, on average, declined by as much as 30 cm since the 1930s.

1

2 The overall consistency of the changes observed in the various components of the cryosphere (Figure 4.25),
3 suggests the existence of a common driving force that affects each of the elements. Regional differences in
4 the magnitude and direction of the signals are apparent, but these not unexpected when we consider the large
5 variability and complexity of atmospheric and oceanic circulations. There are still gaps in measurements and
6 many processes that are not well understood, especially in the Antarctic, but it is very likely that around the
7 Arctic, cryospheric changes will lead to a different environment and ecology.

8

9 **[INSERT FIGURE 4.25 HERE]**

10 **Figure 4.25:** Schematic summary of the observed variations in the cryosphere. The insert figure summarises our
11 assessment of the contribution of ice loss from the ice sheets of Greenland and Antarctica to global sea level rise,
12 together the contribution from all glaciers except those in the periphery of the ice sheets (Section 4.3.4).

13

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References

- 1
2
3 Abdalati, W., et al., 2004: Elevation changes of ice caps in the Canadian Arctic Archipelago. *J. Geophys. Res.-Earth Surf.*, **109**, 11. doi:F04007, 10.1029/2003jf000045
- 4
5 Abdalati, W., et al., 2010: The ICESat-2 Laser Altimetry Mission. *Proceedings of the Ieee*, **98**, 735-751.
6 doi:10.1109/jproc.2009.2034765
- 7 Abermann, J., A. Lambrecht, A. Fischer, and M. Kuhn, 2009: Quantifying changes and trends in glacier area and
8 volume in the Austrian Otztal Alps (1969-1997-2006). *Cryosphere*, **3**, 205-215
- 9 Aizen, V. B., V. A. Kuzmichenok, A. B. Surazakov, and E. M. Aizen, 2007: Glacier changes in the Tien Shan as
10 determined from topographic and remotely sensed data. *Glob. Planet. Change*, **56**, 328-340.
11 doi:10.1016/j.gloplacha.2006.07.016
- 12 Akerman, H. J., and M. Johansson, 2008: Thawing permafrost and thicker active layers in sub-arctic Sweden.
13 *Permafrost and Periglacial Processes*, **19**, 279-292. doi:10.1002/ppp.626
- 14 Allard, M., B. L. Wang, and J. A. Pilon, 1995: Recent cooling along the southern shore of the Hudson Strait, Quebec,
15 Canada, documented from permafrost temperature-measurements. *Arc.Alp.Res.*, **27**, 157-166.
16 doi:10.2307/1551897
- 17 Alley, R. B., et al., 2008: A Simple Law for Ice-Shelf Calving. *Science*, **322**, 1344-1344. doi:10.1126/science.1162543
- 18 Amundson, J. M., M. Fahnstock, M. Truffer, J. Brown, M. P. Luthi, and R. J. Motyka, 2010: Ice melange dynamics
19 and implications for terminus stability, Jakobshavn Isbrae Greenland. *J. Geophys. Res.-Earth Surf.*, **115**, 12.
20 doi:F01005, 10.1029/2009jf001405
- 21 Andreassen, L. M., F. Paul, A. Kaab, and J. E. Hausberg, 2008: Landsat-derived glacier inventory for Jotunheimen,
22 Norway, and deduced glacier changes since the 1930s. *Cryosphere*, **2**, 131-145
- 23 Andresen, C. S., et al., 2012: Rapid response of Helheim Glacier in Greenland to climate variability over the past
24 century. *Nature Geoscience*, **5**. doi:10.1038/ngeo1349
- 25 Arendt, A., et al., 2012: Randolph Glacier Inventory [v2.0]: A Dataset of Global Glacier Outlines, 1-32 pp.
- 26 Arendt, A. A., K. A. Echelmeyer, W. D. Harrison, C. S. Lingle, and V. B. Valentine, 2002: Rapid wastage of Alaska
27 glaciers and their contribution to rising sea level. *Science*, **297**, 382-386
- 28 Azam, M. F., et al., submitted: Imbalance between climate and glacier: a comparison of ice fluxes and mass balance
29 measurements on Chota Shigi Glacier, Western Himalaya, India. *Journal of Glaciology*
- 30 Bahr, D. B., M. F. Meier, and S. D. Peckham, 1997: The physical basis of glacier volume-area scaling. *Journal of*
31 *Geophysical Research-Solid Earth*, **102**, 20355-20362
- 32 Bahr, D. B., M. Dyurgerov, and M. F. Meier, 2009: Sea level rise from glaciers and ice caps: A lower bound. *Geophys.*
33 *Res. Lett.*, **36**, 4. doi:L03501, 10.1029/2008gl036309
- 34 Barletta, V. R., L. S. Sorensen, and R. Forsberg, 2012: Variability of mass changes at basin scale for Greenland and
35 Antarctica. *The Cryosphere*
- 36 Barrand, N., D. G. Vaughan, N. Steiner, M. Tedesco, P. Kuipers Munneke, M. R. van den Broeke, and J. S. Hosking,
37 2012: Trends in Antarctic Peninsula surface melting conditions from observations and regional climate
38 modelling. *J. Geophys. Res.*
- 39 Barrand, N. E., and M. J. Sharp, 2010: Sustained rapid shrinkage of Yukon glaciers since the 1957-1958 International
40 Geophysical Year. *Geophys. Res. Lett.*, **37**. doi:10.1029/2009gl042030
- 41 Barry, R. G., R. E. Moritz, and J. C. Rogers, 1979: FAST ICE REGIMES OF THE BEAUFORT AND CHUKCHI SEA
42 COASTS, ALASKA. *Cold Reg.Sci.Tech.*, **1**, 129-152. doi:10.1016/0165-232x(79)90006-5
- 43 Bartholomew, I. D., P. Nienow, A. Sole, D. Mair, T. Cowton, M. A. King, and S. Palmer, 2011: Seasonal variations in
44 Greenland Ice Sheet motion: Inland extent and behaviour at higher elevations. *Earth and Planetary Science*
45 *Letters*, **307**, 271-278. doi:10.1016/j.epsl.2011.04.014
- 46 Baur, O., M. Kuhn, and W. E. Featherstone, 2009: GRACE-derived ice-mass variations over Greenland by accounting
47 for leakage effects. *Journal of Geophysical Research-Solid Earth*, **114**, 13. doi:B06407, 10.1029/2008jb006239
- 48 Belchansky, G. I., D. C. Douglas, and N. G. Platonov, 2004: Duration of the Arctic Sea ice melt season: Regional and
49 interannual variability, 1979-2001. *J.Clim.*, **17**, 67-80. doi:10.1175/1520-0442(2004)017<0067:dotasi>2.0.co;2
- 50 Beltaos, S., and T. Prowse, 2009: River-ice hydrology in a shrinking cryosphere. *Hydrological Processes*, **23**, 122-144.
51 doi:10.1002/hyp.7165
- 52 Benn, D. I., C. R. Warren, and R. H. Mottram, 2007: Calving processes and the dynamics of calving glaciers. *Earth-*
53 *Science Reviews*, **82**, 143-179. doi:10.1016/j.earscirev.2007.02.002
- 54 Benson, B. J., et al., 2012: Extreme events, trends, and variability in Northern Hemisphere lake-ice phenology (1855-
55 2005). *Climatic Change*, **112**, 299-323. doi:10.1007/s10584-011-0212-8
- 56 Berthier, E., R. Le Bris, L. Mabileau, L. Testut, and F. Remy, 2009: Ice wastage on the Kerguelen Islands (49 degrees
57 S, 69 degrees E) between 1963 and 2006. *J. Geophys. Res.-Earth Surf.*, **114**, 11. doi:F03005,
58 10.1029/2008jf001192
- 59 Berthier, E., E. Schiefer, G. K. C. Clarke, B. Menounos, and F. Remy, 2010: Contribution of Alaskan glaciers to sea
60 level rise derived from satellite imagery. *Nature Geoscience*, **3**, 92-95. doi:10.1038/ngeo737
- 61 Bhambri, R., T. Bolch, R. K. Chaujar, and S. C. Kulshreshtha, 2011: Glacier changes in the Garhwal Himalaya, India,
62 from 1968 to 2006 based on remote sensing. *Journal of Glaciology*, **57**

- 1 Bindschadler, R., 2006: Climate change - Hitting the ice sheets where it hurts. *Science*, **311**, 1720-1721.
2 doi:10.1126/science.1125226
- 3 Bindschadler, R., et al., 2011: Getting around Antarctica: new high-resolution mappings of the grounded and freely-
4 floating boundaries of the Antarctic ice sheet created for the International Polar Year. *Cryosphere*, **5**, 569-588.
5 doi:10.5194/tc-5-569-2011
- 6 Bjørk, A. A., et al., 2012: An aerial view of 80 years of climate-related glacier fluctuations in southeast Greenland.
7 *Nature Geoscience*, **published online first**
- 8 Blaszczyk, M., J. A. Jania, and J. O. Hagen, 2009: Tidewater glaciers of Svalbard: Recent changes and estimates of
9 calving fluxes. *Pol. Polar. Res.*, **30**, 85-142
- 10 Bliss, A., R. Hock, and J. G. Cogley, Submitted: A new inventory of mountain glaciers and ice caps for the Antarctic
11 periphery. *XXX*
- 12 Bodin, X., F. Rojas, and A. Brenning, 2010: Status and evolution of the cryosphere in the Andes of Santiago (Chile,
13 33.5 degrees S.). *Geomorphology*, **118**, 453-464. doi:10.1016/j.geomorph.2010.02.016
- 14 Bodin, X., et al., 2009: Two Decades of Responses (1986-2006) to Climate by the Laurichard Rock Glacier, French
15 Alps. *Permafrost and Periglacial Processes*, **20**, 331-344. doi:10.1002/ppp.665
- 16 Boening, C. W., A. Dispert, M. Visbeck, S. R. Rintoul, and F. U. Schwarzkopf, 2008: The response of the Antarctic
17 Circumpolar Current to recent climate change. *Nature Geoscience*, **1**, 864-869. doi:10.1038/ngeo362
- 18 Bolch, T., 2007: Climate change and glacier retreat in northern Tien Shan (Kazakhstan/Kyrgyzstan) using remote
19 sensing data. *Glob. Planet. Change*, **56**, 1-12. doi:10.1016/j.gloplacha.2006.07.009
- 20 Bolch, T., B. Menounos, and R. Wheate, 2010: Landsat-based inventory of glaciers in western Canada, 1985-2005.
21 *Remote Sensing of Environment*, **114**, 127-137. doi:10.1016/j.rse.2009.08.015
- 22 Bolch, T., M. Buchroithner, T. Pieczonka, and A. Kunert, 2008: Planimetric and volumetric glacier changes in the
23 Khumbu Himal, Nepal, since 1962 using Corona, Landsat TM and ASTER data. *Journal of Glaciology*, **54**, 592-
24 600. doi:10.3189/002214308786570782
- 25 Bolch, T., et al., 2012: The State and Fate of Himalayan Glaciers. *Science*, **336**, 310-314. doi:10.1126/science.1215828
- 26 Box, J. E., L. Yang, D. H. Bromwich, and L. S. Bai, 2009: Greenland Ice Sheet Surface Air Temperature Variability:
27 1840-2007. *J. Clim.*, **22**, 4029-4049. doi:10.1175/2009jcli2816.1
- 28 Brewer, M. C., 1958: Some results of geothermal investigations of permafrost. 19-26.
- 29 Bromwich, D. H., J. P. Nicolas, and A. J. Monaghan, 2011: An Assessment of Precipitation Changes over Antarctica
30 and the Southern Ocean since 1989 in Contemporary Global Reanalyses. *J. Clim.*, **24**, 4189-4209.
31 doi:10.1175/2011jcli4074.1
- 32 Brown, J., K. M. Hinkel, and F. E. Nelson, 2000: Circumpolar Active Layer Monitoring (CALM) program: Research
33 designs and initial results. *Polar Geography*, **24**, 163-258
- 34 Brown, J., O. J. Ferrians Jr., J. A. Heginbottom, and E. S. Melnikov, 1998: *Circum-Arctic Map of Permafrost and Land*
35 *Ice Conditions (revised February 2001)*. National Snow and Ice Data Center / World Data Center for
36 Glaciology.
- 37 Brown, R., C. Derksen, and L. B. Wang, 2010: A multi-data set analysis of variability and change in Arctic spring snow
38 cover extent, 1967-2008. *Journal of Geophysical Research-Atmospheres*, **115**. doi:10.1029/2010jd013975
- 39 Brown, R. D., and P. Cote, 1992: INTERANNUAL VARIABILITY OF LANDFAST ICE THICKNESS IN THE
40 CANADIAN HIGH ARCTIC, 1950-89. *Arctic*, **45**, 273-284
- 41 Brown, R. D., and P. W. Mote, 2009: The Response of Northern Hemisphere Snow Cover to a Changing Climate.
42 *J. Clim.*, **22**, 2124-2145. doi:10.1175/2008jcli2665.1
- 43 Brown, R. D., and D. A. Robinson, 2011: Northern Hemisphere spring snow cover variability and change over 1922-
44 2010 including an assessment of uncertainty. *The Cryosphere*, **5**, 219-229. doi:doi:10.5194/tc-5-219-2011
- 45 Buchardt, S. L., H. B. Clausen, B. M. Vinther, and D. Dahl-Jensen, 2012: Investigating the past and recent delta18O-
46 accumulation relationship seen in Greenland ice cores. *Climate of the Past*
- 47 Bulygina, O. N., V. N. Razuvaev, and N. N. Korshunova, 2009: Changes in snow cover over Northern Eurasia in the
48 last few decades. *Environmental Research Letters*, **4**. doi:10.1088/1748-9326/4/4/045026
- 49 Burn, C. R., and S. V. Kokelj, 2009: The Environment and Permafrost of the Mackenzie Delta Area. *Permafrost and*
50 *Periglacial Processes*, **20**, 83-105. doi:10.1002/ppp.655
- 51 Burn, C. R., and Y. Zhang, 2009: Permafrost and climate change at Herschel Island (Qikiqtaruk), Yukon Territory,
52 Canada. *J. Geophys. Res.-Earth Surf.*, **114**. doi:10.1029/2008jf001087
- 53 Callaghan, T. V., F. Bergholm, T. R. Christensen, C. Jonasson, U. Kokfelt, and M. Johansson, 2010: A new climate era
54 in the sub-Arctic: Accelerating climate changes and multiple impacts. *Geophys. Res. Lett.*, **37**.
55 doi:10.1029/2009gl042064
- 56 Cavalieri, D. J., and C. L. Parkinson, 2012: Arctic sea ice variability and trends, 1979-2010. *The Cryosphere*, **6**, 871-
57 880
- 58 Cazenave, A., et al., 2009: Sea level budget over 2003-2008: A reevaluation from GRACE space gravimetry, satellite
59 altimetry and Argo. *Glob. Planet. Change*, **65**, 83-88. doi:10.1016/j.gloplacha.2008.10.004
- 60 Chapman, W. L., and J. E. Walsh, 2007: A synthesis of Antarctic temperatures. *J. Clim.*, **20**, 4096-4117.
61 doi:10.1175/jcli4236.1

- 1 Charrassin, J. B., et al., 2008: Southern Ocean frontal structure and sea-ice formation rates revealed by elephant seals.
2 *Proceedings of the National Academy of Sciences of the United States of America*, **105**, 11634-11639.
3 doi:10.1073/pnas.0800790105
- 4 Chen, J. L., C. R. Wilson, and B. D. Tapley, 2006: Satellite gravity measurements confirm accelerated melting of
5 Greenland ice sheet. *Science*, **313**, 1958-1960
- 6 ———, 2011: Interannual variability of Greenland ice losses from satellite gravimetry. *Journal of Geophysical Research-*
7 *Solid Earth*, **116**, 11. doi:B07406, 10.1029/2010jb007789
- 8 Chen, J. L., C. R. Wilson, D. Blankenship, and B. D. Tapley, 2009: Accelerated Antarctic ice loss from satellite gravity
9 measurements. *Nature Geoscience*, **2**, 859-862. doi:10.1038/ngeo694
- 10 Chen, J. L., C. R. Wilson, B. D. Tapley, D. D. Blankenship, and E. R. Ivins, 2007: Patagonia icefield melting observed
11 by gravity recovery and climate experiment (GRACE). *Geophys. Res. Lett.*, **34**, 6. doi:L22501,
12 10.1029/2007gl031871
- 13 Cheng, G. D., and T. H. Wu, 2007: Responses of permafrost to climate change and their environmental significance,
14 Qinghai-Tibet Plateau. *J. Geophys. Res.-Earth Surf.*, **112**. doi:10.1029/2006jf000631
- 15 Chinn, T., S. Winkler, M. J. Salinger, and N. Haakensen, 2005: Srecent glacier advances in Norway and New Zealand:
16 A comparison of their glaciological and meteorological causes. *Geografiska Annaler Series a-Physical*
17 *Geography*, **87A**. doi:10.1111/j.0435-3676.2005.00249.x
- 18 Choi, G., D. A. Robinson, and S. Kang, 2010: Changing Northern Hemisphere Snow Seasons. *J. Clim.*, **23**, 5305-5310.
19 doi:10.1175/2010jcli3644.1
- 20 Christiansen, H. H., et al., 2010: The Thermal State of Permafrost in the Nordic Area during the International Polar
21 Year 2007-2009. *Permafrost and Periglacial Processes*, **21**, 156-181. doi:10.1002/ppp.687
- 22 Christoffersen, P., et al., 2011: Warming of waters in an East Greenland fjord prior to glacier retreat: mechanisms and
23 connection to large-scale atmospheric conditions. *Cryosphere*, **5**, 701-714. doi:10.5194/tc-5-701-2011
- 24 Cia, J. C., A. J. Andres, M. A. S. Sanchez, J. C. Novau, and J. I. L. Moreno, 2005: Responses to climatic changes since
25 the little ice age on maladeta glacier (Central pyrenees). *Geomorphology*, **68**, 167-182.
26 doi:10.1016/j.geomorph.2004.11.012
- 27 Citterio, M., F. Paul, A. P. Ahlstrom, H. F. Jepsen, and A. Weidick, 2009: Remote sensing of glacier change in West
28 Greenland: accounting for the occurrence of surge-type glaciers. *Annals of Glaciology*, **50**
- 29 Clarke, G. K. C., F. S. Anslow, A. H. Jarosch, V. Radić, B. Menounos, T. Bolch, and E. Berthier, submitted: Ice
30 volume and subglacial topography for western Canadian glaciers from mass balance fields, thinning rates, and a
31 bed stress model. *J. Clim.*
- 32 Cogley, J. G., 2009a: A more complete version of the World Glacier Inventory. *Annals of Glaciology*, **50**, 32-38
33 ———, 2009b: Geodetic and direct mass-balance measurements: comparison and joint analysis. *Annals of Glaciology*,
34 **50**, 96-100
- 35 Cogley, J. G., et al., 2011: *Glossary of Glacier Mass Balance and Related Terms*. UNESCO-IHP.
- 36 Comiso, J. C., 2002: A rapidly declining perennial sea ice cover in the Arctic. *Geophys. Res. Lett.*, **29**.
37 doi:10.1029/2002gl015650
- 38 ———, 2011a: Large decadal decline in the Arctic multiyear ice cover. *J. Clim.*, **5**, 1176-1193. doi:doi:10.1175/JCLI-D-
39 11-00113.1
- 40 ———, 2011b: Large decadal decline in the Arctic multiyear ice cover. *Journal of Climate*. doi:doi:10.1175/JCLI-D-11-
41 00113.1
- 42 Comiso, J. C., and F. Nishio, 2008: Trends in the sea ice cover using enhanced and compatible AMSR-E, SSM/I, and
43 SMMR data. *Journal of Geophysical Research-Oceans*, **113**. doi:C02s07, 10.1029/2007jc004257
- 44 Comiso, J. C., C. L. Parkinson, R. Gersten, and L. Stock, 2008: Accelerated decline in the Arctic Sea ice cover.
45 *Geophys. Res. Lett.*, **35**. doi:Artn L01703
46 Doi 10.1029/2007gl031972
- 47 Comiso, J. C., R. Kwok, S. Martin, and A. L. Gordon, 2011: Variability and trends in sea ice extent and ice production
48 in the Ross Sea. *Journal of Geophysical Research-Oceans*, **116**. doi:10.1029/2010jc006391
- 49 Cook, A. J., and D. G. Vaughan, 2010: Overview of areal changes of the ice shelves on the Antarctic Peninsula over the
50 past 50 years. *Cryosphere*, **4**, 77-98. doi:Doi 10.5194/Tc-4-77-2010
- 51 Costa, D. P., J. M. Klinck, E. E. Hofmann, M. S. Dinniman, and J. M. Burns, 2008: Upper ocean variability in west
52 Antarctic Peninsula continental shelf waters as measured using instrumented seals. *Deep-Sea Research Part II-*
53 *Topical Studies in Oceanography*, **55**, 323-337. doi:10.1016/j.dsr2.2007.11.003
- 54 Cullen, N. J., T. Molg, G. Kaser, K. Hussein, K. Steffen, and D. R. Hardy, 2006: Kilimanjaro Glaciers: Recent areal
55 extent from satellite data and new interpretation of observed 20th century retreat rates. *Geophys. Res. Lett.*, **33**, 6.
56 doi:L16502, 10.1029/2006gl027084
- 57 Daniault, N., H. Mercier, and P. Lherminier, 2011: The 1992-2009 transport variability of the East Greenland-Irminger
58 Current at 60 degrees N. *Geophys. Res. Lett.*, **38**, 4. doi:L07601, 10.1029/2011gl046863
- 59 Das, S. B., I. Joughin, M. D. Behn, I. M. Howat, M. A. King, D. Lizarralde, and M. P. Bhatia, 2008: Fracture
60 propagation to the base of the Greenland Ice Sheet during supraglacial lake drainage. *Science*, **320**, 778-781.
61 doi:10.1126/science.1153360
- 62 Davies, B. J., and N. F. Glasser, Submitted: Accelerating recession of Patagonian glaciers from the “Little Ice Age” (c.
63 AD 1870) to 2011

- 1 Davis, C. H., Y. H. Li, J. R. McConnell, M. M. Frey, and E. Hanna, 2005: Snowfall-driven growth in East Antarctic ice
2 sheet mitigates recent sea level rise. *Science*, **308**, 1898-1901. doi:10.1126/science.1110662
- 3 Debeer, C. M., and M. J. Sharp, 2007: Recent changes in glacier area and volume within the southern Canadian
4 Cordillera. *Annals of Glaciology, Vol 46, 2007*, **46**, 215-221. doi:10.3189/172756407782871710
- 5 DeConto, R. M., and D. Pollard, 2003: Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO₂.
6 *Nature*, **421**, 245-249
- 7 Delaloye, R., et al., cited 2011: Recent interannual variations of rock glacier creep in the European Alps. [Available
8 online at <http://www.zora.uzh.ch/7031/>.]
- 9 DeLiberty, T. L., C. A. Geiger, S. F. Ackley, A. P. Worby, and M. L. Van Woert, 2011: Estimating the annual cycle of
10 sea-ice thickness and volume in the Ross Sea. *Deep-Sea Research Part II-Topical Studies in Oceanography*, **58**,
11 1250-1260. doi:10.1016/j.dsr2.2010.12.005
- 12 Derksen, C., and R. Brown, in press: Spring snow cover extent reductions in the 2008-2012 period exceeding climate
13 model projections. *Geophys. Res. Lett.*
- 14 Dery, S. J., and R. D. Brown, 2007: Recent Northern Hemisphere snow cover extent trends and implications for the
15 snow-albedo feedback. *Geophys. Res. Lett.*, **34**, 6. doi:L22504, 10.1029/2007gl031474
- 16 Diolaiuti, G., D. Bocchiola, C. D'Agata, and C. Smiraglia, 2012: Evidence of climate change impact upon glaciers'
17 recession within the Italian Alps - The case of Lombardy glaciers. *Theoretical Applied Climatology*, **109**, 429-
18 445
- 19 Dmitrenko, I. A., et al., 2011: Recent changes in shelf hydrography in the Siberian Arctic: Potential for subsea
20 permafrost instability. *Journal of Geophysical Research-Oceans*, **116**, 10. doi:C10027, 10.1029/2011jc007218
- 21 Doake, C. S. M., and D. G. Vaughan, 1991: RAPID DISINTEGRATION OF THE WORDIE ICE SHELF IN
22 RESPONSE TO ATMOSPHERIC WARMING. *Nature*, **350**, 328-330. doi:10.1038/350328a0
- 23 Doherty, S. J., S. G. Warren, T. C. Grenfell, A. D. Clarke, and R. E. Brandt, 2010: Light-absorbing impurities in Arctic
24 snow. *Atmospheric Chemistry and Physics*, **10**, 11647-11680. doi:10.5194/acp-10-11647-2010
- 25 Domack, E., et al., 2005: Stability of the Larsen B ice shelf on the Antarctic Peninsula during the Holocene epoch.
26 *Nature*, **436**, 681-685
- 27 Dong-Chen, E., Y.-D. Yang, and D.-B. Chao, 2009: The sea level change from the Antarctic ice sheet based on
28 GRACE. *Chinese Journal of Geophysics-Chinese Edition*, **52**, 2222-2228. doi:10.3969/j.issn.0001-
29 5733.2009.09.005
- 30 Dowdeswell, E. K., J. A. Dowdeswell, and F. Cawkwell, 2007: On the glaciers of Bylot Island, Nunavut, Arctic
31 Canada. *Arct. Antarct. Alp. Res.*, **39**. doi:10.1657/1523-0430(05-123)
- 32 Drobot, S. D., and M. R. Anderson, 2001: An improved method for determining snowmelt onset dates over Arctic sea
33 ice using scanning multichannel microwave radiometer and Special Sensor Microwave/Imager data. *Journal of*
34 *Geophysical Research-Atmospheres*, **106**, 24033-24049. doi:10.1029/2000jd000171
- 35 Drozdov, D. S., N. G. Ukraintseva, A. M. Tsarev, and S. N. Chekrygina, 2010: Changes in the temperature field and in
36 the state of the geosystems within the territory of the Urengoy field during the last 35 years (1974-2008). *Earth's*
37 *Cryosphere*, **14**, 22-31
- 38 Druckenmiller, M. L., H. Eicken, M. A. Johnson, D. J. Pringle, and C. C. Williams, 2009: Toward an integrated coastal
39 sea-ice observatory: System components and a case study at Barrow, Alaska. *Cold Reg. Sci. Tech.*, **56**, 61-72.
40 doi:10.1016/j.coldregions.2008.12.003
- 41 Drucker, R., S. Martin, and R. Kwok, 2011: Sea ice production and export from coastal polynyas in the Weddell and
42 Ross Seas. *Geophys. Res. Lett.*, **38**, 4. doi:L17502, 10.1029/2011gl048668
- 43 Dyer, J. L., and T. L. Mote, 2006: Spatial variability and trends in observed snow depth over North America. *Geophys.*
44 *Res. Lett.*, **33**. doi:10.1029/2006gl027258
- 45 Ettema, J., M. R. van den Broeke, E. van Meijgaard, W. J. van de Berg, J. L. Bamber, J. E. Box, and R. C. Bales, 2009:
46 Higher surface mass balance of the Greenland ice sheet revealed by high-resolution climate modeling. *Geophys.*
47 *Res. Lett.*, **36**. doi:L12501, 10.1029/2009gl038110
- 48 Farinotti, D., M. Huss, A. Bauder, M. Funk, and M. Truffer, 2009: A method to estimate the ice volume and ice-
49 thickness distribution of alpine glaciers. *Journal of Glaciology*, **55**, 422-430
- 50 Fauria, M. M., et al., 2010: Unprecedented low twentieth century winter sea ice extent in the Western Nordic Seas since
51 AD 1200. *Climate Dynamics*, **34**, 781-795. doi:10.1007/s00382-009-0610-z
- 52 Fedorov, A. N., and P. Y. Konstantinov, 2008: Recent changes in ground temperature and the effect on permafrost
53 landscapes in Central Yakutia. *9th International Conference on Permafrost*, Institute of Northern Engineering,
54 University of Alaska, Fairbanks, 433-438.
- 55 Fernandes, R., H. X. Zhao, X. J. Wang, J. Key, X. Qu, and A. Hall, 2009: Controls on Northern Hemisphere snow
56 albedo feedback quantified using satellite Earth observations. *Geophys. Res. Lett.*, **36**.
57 doi:10.1029/2009gl040057
- 58 Fettweis, X., M. Tedesco, M. R. van de Broeke, and J. Ettema, 2011: Melting trends over the Greenland ice sheet
59 (1958-2009) from spaceborne microwave data and regional climate models. *The Cryosphere*, **5**, 359-375.
60 doi:10.5194/tc-5-359-2011
- 61 Fisher, D., J. Zheng, D. Burgess, C. Zdanowicz, C. Kinnard, M. Sharp, and J. Bourgeois, 2012: Recent melt rates of
62 Canadian arctic ice caps are the highest in four millennia. *Glob. Planet. Change*, **84-85**.
63 doi:10.1016/j.gloplacha.2011.06.005

- 1 Flanner, M. G., C. S. Zender, J. T. Randerson, and P. J. Rasch, 2007: Present-day climate forcing and response from
2 black carbon in snow. *Journal of Geophysical Research-Atmospheres*, **112**. doi:10.1029/2006jd008003
- 3 Flanner, M. G., K. M. Shell, M. Barlage, D. K. Perovich, and M. A. Tschudi, 2011: Radiative forcing and albedo
4 feedback from the Northern Hemisphere cryosphere between 1979 and 2008. *Nature Geoscience*, **4**, 151-155.
5 doi:doi:10.1038/ngeo1062
- 6 Foster, J. L., D. K. Hall, R. E. J. Kelly, and L. Chiu, 2009: Seasonal snow extent and snow mass in South America
7 using SMMR and SSM/I passive microwave data (1979-2006). *Remote Sensing of Environment*, **113**, 291-305.
8 doi:10.1016/j.rse.2008.09.010
- 9 Fraser, A. D., R. A. Massom, K. J. Michael, B. K. Galton-Fenzi, and J. L. Lieser, in press: East Antarctic landfast sea
10 ice distribution and variability, 2000-2008. *J.Clim.* doi:doi: 10.1175/JCLI-D-10-05032.1
- 11 Frauenfeld, O. W., and T. J. Zhang, 2011: An observational 71-year history of seasonally frozen ground changes in the
12 Eurasian high latitudes. *Environmental Research Letters*, **6**. doi:10.1088/1748-9326/6/4/044024
- 13 Frauenfeld, O. W., T. J. Zhang, R. G. Barry, and D. Gilichinsky, 2004: Interdecadal changes in seasonal freeze and
14 thaw depths in Russia. *Journal of Geophysical Research-Atmospheres*, **109**. doi:10.1029/2003jd004245
- 15 Fretwell, P. T., et al., Submitted: Bedmap2: improved ice bed, surface and thickness datasets for Antarctica. *The
16 Cryosphere*
- 17 Frezzotti, M., C. Scarchilli, S. Becagli, M. Proposito, and S. Urbini, 2012: A synthesis of the Antarctic Surface Mass
18 Balance during the last eight centuries. *The Cryosphere*
- 19 Fricker, H. A., and L. Padman, 2012: Thirty years of elevation change on Antarctic Peninsula ice shelves from
20 multimission satellite radar altimetry. *Journal of Geophysical Research-Oceans*, **117**. doi:10.1029/2011jc007126
- 21 Fyodorov-Davydov, D. G., A. L. Kholodov, V. E. Ostroumov, G. N. Kraev, V. A. Sorokovikov, S. P. Davudov, and A.
22 A. Merekalova, 2008: Seasonal Thaw of Soils in the North Yakutian Ecosystems. *9th International Conference
23 on Permafrost*, Institute of Northern Engineering, University of Alaska, Fairbanks, 481-486.
- 24 Gardner, A. S., et al., 2011: Sharply increased mass loss from glaciers and ice caps in the Canadian Arctic Archipelago.
25 *Nature*, **473**, 357-360. doi:10.1038/nature10089
- 26 Gardner, A. S., et al., 2012: Narrowing the gap: A consensus estimate of glacier mass wastage., submitted
- 27 Gardner, A. S., et al., Submitted: Narrowing the gap: A consensus estimate of glacier mass wastage
- 28 Gerland, S., A. H. H. Renner, F. Godtlielsen, D. Divine, and T. B. Loyning, 2008: Decrease of sea ice thickness at
29 Hopen, Barents Sea, during 1966-2007. *Geophys. Res. Lett.*, **35**. doi:10.1029/2007gl032716
- 30 Giesen, R. H., and J. Oerlemans, 2012: Global application of a surface mass balance model using gridded climate data.
31 *The Cryosphere Discussions*, **6**, 1445-1490. doi:doi:10.5194/tcd-6-1445-2012
- 32 Giles, A. B., R. A. Massom, and V. I. Lytle, 2008a: Fast-ice distribution in East Antarctica during 1997 and 1999
33 determined using RADARSAT data. *Journal of Geophysical Research-Oceans*, **113**, 15. doi:C02s14,
34 10.1029/2007jc004139
- 35 Giles, K. A., S. W. Laxon, and A. L. Ridout, 2008b: Circumpolar thinning of Arctic sea ice following the 2007 record
36 ice extent minimum. *Geophys. Res. Lett.*, **35**. doi:10.1029/2008gl035710
- 37 Gille, S. T., 2002: Warming of the Southern Ocean since the 1950s. *Science*, **295**, 1275-1277
- 38 Gille, S. T., 2008: Decadal-scale temperature trends in the Southern Hemisphere ocean. *J.Clim.*, **21**, 4749-4765.
39 doi:10.1175/2008jcli2131.1
- 40 Gjermundsen, E. F., R. Mathieu, A. Kaab, T. Chinn, B. Fitzharris, and J. O. Hagen, 2011: Assessment of multispectral
41 glacier mapping methods and derivation of glacier area changes, 1978-2002, in the central Southern Alps, New
42 Zealand, from ASTER satellite data, field survey and existing inventory data. *Journal of Glaciology*, **57**
- 43 Glazovsky, A., and Y. Macheret, 2006: Eurasian Arctic. *Glaciation in north and central Eurasia in present time*, V. M.
44 Kotlyakov, Ed., Nauka.
- 45 Gloersen, P., W. Campbell, D. Cavalieri, J. Comiso, C. Parkinson, and H. J. Zwally, 1992: Arctic and Antarctic sea ice:
46 satellite passive microwave observations and analysis.
- 47 Gordon, A. L., M. Visbeck, and J. C. Comiso, 2007: A possible link between the Weddell Polynya and the Southern
48 Annular Mode. *J.Clim.*, **20**, 2558-2571. doi:10.1175/jcli4046.1
- 49 Gregory, J. M., and J. Oerlemans, 1998: Simulated future sea level rise due to glacier melt based on regionally and
50 seasonally resolved temperature changes. *Nature*, **391**, 474-476
- 51 Griggs, J., and J. L. Bamber, 2011a: Antarctic ice-shelf thickness from satellite radar altimetry. *Journal of Glaciology*,
52 **57**, 485-498
- 53 Griggs, J. A., and J. L. Bamber, 2011b: Updated bedrock elevation model for Greenland. European Union Framework 7
54 Programme ice2sea.
- 55 Grinsted, A., submitted: An estimate of global glacier volume. *The Cryosphere*
- 56 Gruber, S., 2012: Derivation and analysis of a high-resolution estimate of global permafrost zonation. *The Cryosphere*,
57 **6**, 221-233
- 58 Gruber, S., and W. Haeberli, 2007: Permafrost in steep bedrock slopes and its temperature-related destabilization
59 following climate change. *J. Geophys. Res.-Earth Surf.*, **112**, 10. doi:F02s18, 10.1029/2006jf000547
- 60 Gunter, B., et al., 2009: A comparison of coincident GRACE and ICESat data over Antarctica. *Journal of Geodesy*, **83**,
61 1051-1060. doi:10.1007/s00190-009-0323-4
- 62 Gusmeroli, A., P. Jansson, R. Pettersson, and T. Murray, 2012: Twenty years of cold surface layer thinning at
63 Storglaciären, sub-Arctic Sweden, 1989-2009. *Journal of Glaciology*, **58**. doi:10.3189/2012JoG11J018

- 1 Haas, C., S. Hendricks, H. Eicken, and A. Herber, 2010: Synoptic airborne thickness surveys reveal state of Arctic sea
2 ice cover. *Geophys. Res. Lett.*, **37**. doi:10.1029/2010gl042652
- 3 Haas, C., A. Pfaffling, S. Hendricks, L. Rabenstein, J. L. Etienne, and I. Rigor, 2008: Reduced ice thickness in Arctic
4 Transpolar Drift favors rapid ice retreat. *Geophys. Res. Lett.*, **35**. doi:10.1029/2008gl034457
- 5 Halsey, L. A., D. H. Vitt, and S. C. Zoltai, 1995: Disequilibrium response of permafrost in boreal continental western
6 Canada to climate-change. *Climatic Change*, **30**, 57-73. doi:10.1007/bf01093225
- 7 Hanna, E., et al., 2011: Greenland Ice Sheet surface mass balance 1870 to 2010 based on Twentieth Century Reanalysis,
8 and links with global climate forcing. *Journal of Geophysical Research-Atmospheres*, **116**.
9 doi:10.1029/2011jd016387
- 10 Harig, C., and F. J. Simons, Submitted: Mapping Greenland's mass loss in space and time. *PNAS*
- 11 Heid, T., and A. Kaab, 2011: Worldwide widespread decadal-scale decrease of glacier speed revealed using repeat
12 optical satellite images. *The Cryosphere Discussions*, **5**
- 13 Helsen, M. M., et al., 2008: Elevation changes in Antarctica mainly determined by accumulation variability. *Science*,
14 **320**, 1626-1629. doi:10.1126/science.1153894
- 15 Hirabayashi, Y., ., Y. Zhang, S. Watanabe, S. Koirala, and S. Kanae, submitted: Development of a global glacier model
16 HYOGA2 1 for water resources assessment. *Hydrological Research Letters*
- 17 Hock, R., M. de Woul, V. Radic, and M. Dyurgerov, 2009: Mountain glaciers and ice caps around Antarctica make a
18 large sea level rise contribution. *Geophys. Res. Lett.*, **36**. doi:10.1029/2008gl037020
- 19 Hoelzle, M., G. Darms, M. P. Lüthi, and S. Suter, 2011: Evidence of accelerated englacial warming in the Monte Rosa
20 area, Switzerland/Italy. *The Cryosphere*, **5**, 231-243. doi:10.5194/tc-5-231-2011
- 21 Hoffman, M. J., G. A. Catania, T. A. Neumann, L. C. Andrews, and J. A. Rumrill, 2011: Links between acceleration,
22 melting, and supraglacial lake drainage of the western Greenland Ice Sheet. *J. Geophys. Res.-Earth Surf.*, **116**.
23 doi:10.1029/2010jf001934
- 24 Holland, D. M., and A. Jenkins, 1999: Modeling thermodynamic ice-ocean interactions at the base of an ice shelf.
25 *Journal of Physical Oceanography*, **29**, 1787-1800. doi:10.1175/1520-0485(1999)029<1787:mtioia>2.0.co;2
- 26 Holland, D. M., R. H. Thomas, B. De Young, M. H. Ribergaard, and B. Lyberth, 2008a: Acceleration of Jakobshavn
27 Isbrae triggered by warm subsurface ocean waters. *Nature Geoscience*, **1**, 659-664. doi:10.1038/ngeo316
- 28 Holland, P. R., and R. Kwok, Submitted: Wind-driven trends in Antarctic sea ice motion. *Nature Geoscience*
- 29 Holland, P. R., A. Jenkins, and D. M. Holland, 2008b: The response of ice shelf basal melting to variations in ocean
30 temperature. *J. Clim.*, **21**, 2558-2572. doi:10.1175/2007jcli1909.1
- 31 Holzhauser, H., M. Magny, and H. J. Zumbuhl, 2005: Glacier and lake-level variations in west-central Europe over the
32 last 3500 years. *Holocene*, **15**, 789-801. doi:10.1191/0959683605hl853ra
- 33 Horwath, M., and R. Dietrich, 2009: Signal and error in mass change inferences from GRACE: the case of Antarctica.
34 *Geophysical Journal International*, **177**, 849-864. doi:10.1111/j.1365-246X.2009.04139.x
- 35 Howat, I. M., I. Joughin, and T. A. Scambos, 2007: Rapid changes in ice discharge from Greenland outlet glaciers.
36 *Science*, **315**, 1559-1561
- 37 Howat, I. M., I. Joughin, M. Fahnestock, B. E. Smith, and T. A. Scambos, 2008: Synchronous retreat and acceleration
38 of southeast Greenland outlet glaciers 2000-06: ice dynamics and coupling to climate. *Journal of Glaciology*, **54**,
39 646-660
- 40 Howat, I. M., Y. Ahn, I. Joughin, M. R. van den Broeke, J. T. M. Lenaerts, and B. Smith, 2011: Mass balance of
41 Greenland's three largest outlet glaciers, 2000-2010. *Geophys. Res. Lett.*, **38**, 5. doi:L12501,
42 10.1029/2011gl047565
- 43 Hudson, S. R., 2011: Estimating the global radiative impact of the sea ice-albedo feedback in the Arctic. *Journal of*
44 *Geophysical Research-Atmospheres*, **116**. doi:10.1029/2011jd015804
- 45 Hughes, T. J., 1973: Is the West Antarctic ice sheet disintegrating? *J. Geophys. Res.*
- 46 Hulbe, C. L., T. A. Scambos, T. Youngberg, and A. K. Lamb, 2008: Patterns of glacier response to disintegration of the
47 Larsen B ice shelf, Antarctic Peninsula. *Glob. Planet. Change*, **63**, 1-8. doi:10.1016/j.gloplacha.2008.04.001
- 48 Humbert, A., et al., 2010: Deformation and failure of the ice bridge on the Wilkins Ice Shelf, Antarctica. *Annals of*
49 *Glaciology*, **51**, 49-55. doi:Doi:10.3189/172756410791392709
- 50 Hurrell, J. W., 1995: DECADAL TRENDS IN THE NORTH-ATLANTIC OSCILLATION - REGIONAL
51 TEMPERATURES AND PRECIPITATION. *Science*, **269**, 676-679. doi:10.1126/science.269.5224.676
- 52 Huss, M., 2012: Extrapolating glacier mass balance to the mountain-range scale: the European Alps 1900-2100. *The*
53 *Cryosphere*, **6**
- 54 Huss, M., and D. Farinotti, Submitted: Distributed ice thickness and volume of 180,000 glaciers around the globe
- 55 Huss, M., R. Stockil, G. Kappenberger, and H. Blatter, 2008: Temporal and spatial changes of Laika Glacier, Canadian
56 Arctic, since 1959, inferred from satellite remote sensing and mass-balance modelling. *Journal of Glaciology*,
57 **54**, 857-866
- 58 Isaksen, K., J. L. Sollid, P. Holmlund, and C. Harris, 2007: Recent warming of mountain permafrost in Svalbard and
59 Scandinavia. *J. Geophys. Res.-Earth Surf.*, **112**. doi:10.1029/2006jf000522
- 60 Isaksen, K., et al., 2011: Degrading mountain permafrost in southern Norway: spatial and temporal variability of mean
61 ground temperatures, 1999-2009. *Permafrost and Periglacial Processes*, doi:10.1002/ppp.1728

- 1 Ishikawa, M., N. Sharkhuu, Y. Jambaljav, G. Davaa, K. Yoshikawa, and T. Ohata, 2012: Thermal state of Mongolian
2 permafrost. *Tenth International Conference on Permafrost*, Salekhard, Yamel-nenets Autonomous District,
3 Russia, 173-178.
- 4 Ivins, E. R., and T. S. James, 2005: Antarctic glacial isostatic adjustment: a new assessment. *Antarct. Sci.*, **17**, 541-553.
5 doi:10.1017/s0954102005002968
- 6 Ivins, E. R., E. Rignot, X. P. Wu, T. S. James, and G. Casassa, 2005: *Ice mass balance and Antarctic gravity change:
7 Satellite and terrestrial perspectives*. 3-12 pp.
- 8 Ivins, E. R., M. M. Watkins, D. N. Yuan, R. Dietrich, G. Casassa, and A. Rulke, 2011: On-land ice loss and glacial
9 isostatic adjustment at the Drake Passage: 2003-2009. *Journal of Geophysical Research-Solid Earth*, **116**, 24.
10 doi:10.1029/2010jb007607
- 11 Jacob, T., J. Wahr, W. T. Pfeffer, and S. Swenson, 2012: Recent contributions of glaciers and ice caps to sea level rise.
12 *Nature*, **482**. doi:10.1038/nature10847
- 13 Jacobs, S. S., A. Jenkins, C. F. Giulivi, and P. Dutrieux, 2011: Stronger ocean circulation and increased melting under
14 Pine Island Glacier ice shelf. *Nature Geoscience*, **4**, 519-523. doi:10.1038/ngeo1188
- 15 Jacobs, S. S., H. H. Helmer, C. S. M. Doake, A. Jenkins, and R. M. Frolich, 1992: MELTING OF ICE SHELVES AND
16 THE MASS BALANCE OF ANTARCTICA. *Journal of Glaciology*, **38**, 375-387
- 17 Jenkins, A., 2011: Convection-driven melting near the grounding lines of ice shelves and tidewater glaciers. *J.Clim.*,
18 doi: 10.1175/JPO-D-1111-1103.1171
- 19 Jenkins, A., P. Dutrieux, S. S. Jacobs, S. D. McPhail, J. R. Perrett, A. T. Webb, and D. White, 2010: Observations
20 beneath Pine Island Glacier in West Antarctica and implications for its retreat. *Nature Geoscience*, **3**, 468-472.
21 doi:10.1038/ngeo890
- 22 Jia, L. L., H. S. Wang, and L. W. Xiang, 2009: Effect of glacio-static adjustment on the estimate of ice mass balance
23 over Antarctic and uncertainties. *Chinese Journal of Geophysics*, **54**, 1466-1477
- 24 Jiskoot, H., C. J. Curran, D. L. Tessler, and L. R. Shenton, 2009: Changes in Clemenceau Icefield and Chaba Group
25 glaciers, Canada, related to hypsometry, tributary detachment, length-slope and area-aspect relations. *Annals of
26 Glaciology*, **50**, 133-143
- 27 Johannessen, O. M., E. V. Shalina, and M. W. Miles, 1999: Satellite evidence for an Arctic sea ice cover in
28 transformation. *Science*, **286**, 1937-1939. doi:10.1126/science.286.5446.1937
- 29 Jones, B. M., C. D. Arp, M. T. Jorgenson, K. M. Hinkel, J. A. Schmutz, and P. L. Flint, 2009: Increase in the rate and
30 uniformity of coastline erosion in Arctic Alaska. *Geophys. Res. Lett.*, **36**, 5. doi:L03503, 10.1029/2008gl036205
- 31 Jones, P. D., D. H. Lister, T. J. Osborn, C. Harpham, M. Salmon, and C. P. Morice, 2012: Hemispheric and large-scale
32 land-surface air temperature variations: An extensive revision and an update to 2010. *Journal of Geophysical
33 Research-Atmospheres*, **117**. doi:10.1029/2011jd017139
- 34 Jorgenson, M. T., Y. L. Shur, and E. R. Pullman, 2006: Abrupt increase in permafrost degradation in Arctic Alaska.
35 *Geophys. Res. Lett.*, **33**, 4. doi:L02503, 10.1029/2005gl024960
- 36 Joughin, I., and R. B. Alley, 2011: Stability of the West Antarctic ice sheet in a warming world. *Nature Geoscience*, **4**.
37 doi:10.1038/ngeo1194
- 38 Joughin, I., W. Abdalati, and M. Fahnestock, 2004: Large fluctuations in speed on Greenland's Jakobshavn Isbrae
39 glacier. *Nature*, **432**, 608-610
- 40 Joughin, I., B. E. Smith, and D. M. Holland, 2010a: Sensitivity of 21st century sea level to ocean-induced thinning of
41 Pine Island Glacier, Antarctica. *Geophys. Res. Lett.*, **37**. doi:10.1029/2010gl044819
- 42 Joughin, I., M. Fahnestock, R. Kwok, P. Gogineni, and C. Allen, 1999: Ice flow of Humboldt, Petermann and Ryder
43 Gletscher, northern Greenland. *Journal of Glaciology*, **45**. doi:10.3189/002214399793377284
- 44 Joughin, I., B. E. Smith, I. M. Howat, T. Scambos, and T. Moon, 2010b: Greenland flow variability from ice-sheet-wide
45 velocity mapping. *Journal of Glaciology*, **56**, 415-430
- 46 Joughin, I., S. B. Das, M. A. King, B. E. Smith, I. M. Howat, and T. Moon, 2008a: Seasonal Speedup Along the
47 Western Flank of the Greenland Ice Sheet. *Science*, **320**, 781-783
- 48 Joughin, I., et al., 2008b: Ice-front variation and tidewater behavior on Helheim and Kangerdlugssuaq Glaciers,
49 Greenland. *J. Geophys. Res.-Earth Surf.*, **113**. doi:F01004, 10.1029/2007jf000837
- 50 Kaab, A., 2008: Glacier Volume Changes Using ASTER Satellite Stereo and ICESat GLAS Laser Altimetry. A Test
51 Study on Edgeoya, Eastern Svalbard. *Ieee Transactions on Geoscience and Remote Sensing*, **46**, 2823-2830.
52 doi:10.1109/tgrs.2008.2000627
- 53 Kaab, A., R. Frauenfelder, and I. Roer, 2007: On the response of rockglacier creep to surface temperature increase.
54 *Glob. Planet. Change*, **56**, 172-187. doi:10.1016/j.gloplacha.2006.07.005
- 55 Kääb, A., E. Berthier, C. Nuth, J. Gardelle, and Y. Arnaud, 2012 (in Press): Contrasting patterns of early 21st century
56 glacier mass change in the Hindu Kush - Karakoram - Himalaya. *Nature*
- 57 Kaser, G., J. G. Cogley, M. B. Dyurgerov, M. F. Meier, and A. Ohmura, 2006: Mass balance of glaciers and ice caps:
58 Consensus estimates for 1961-2004. *Geophys. Res. Lett.*, **33**. doi:L19501, 10.1029/2006gl027511
- 59 Ke, C. Q., T. Yu, K. Yu, G. D. Tang, and L. King, 2009: Snowfall trends and variability in Qinghai, China. *Theoretical
60 and Applied Climatology*, **98**, 251-258. doi:10.1007/s00704-009-0105-1
- 61 Khan, S. A., J. Wahr, M. Bevis, I. Velicogna, and E. Kendrick, 2010a: Spread of ice mass loss into northwest
62 Greenland observed by GRACE and GPS. *Geophys. Res. Lett.*, **37**. doi:L06501, 10.1029/2010gl042460

- 1 Khan, S. A., L. Liu, J. Wahr, I. Howat, I. Joughin, T. van Dam, and K. Fleming, 2010b: GPS measurements of crustal
2 uplift near Jakobshavn Isbrae due to glacial ice mass loss. *Journal of Geophysical Research-Solid Earth*, **115**,
3 13. doi:B09405, 10.1029/2010jb007490
- 4 Khromova, T. E., G. B. Osipova, D. G. Tsvetkov, M. B. Dyurgerov, and R. G. Barry, 2006: Changes in glacier extent in
5 the eastern Pamir, Central Asia, determined from historical data and ASTER imagery. *Remote Sensing of*
6 *Environment*, **102**, 24-32. doi:10.1016/j.rse.2006.01.019
- 7 King, M. A., Rory J. Bingham, Phil Moore, Pippa L. Whitehouse, M. J. Bentley, and G. A. Milne, Submitted: Lower
8 Antarctic sea level contribution from GRACE using a new glacial isostatic adjustment model. *Nature*
- 9 King, M. A., et al., 2009: A 4-decade record of elevation change of the Amery Ice Shelf, East Antarctica. *J. Geophys.*
10 *Res.-Earth Surf.*, **114**. doi:F01010, 10.1029/2008jf001094
- 11 Kinnard, C., C. M. Zdanowicz, D. A. Fisher, E. Isaksson, A. de Vernal, and L. G. Thompson, 2011: Reconstructed
12 changes in Arctic sea ice over the past 1,450 years. *Nature*, **479**, 509-U231. doi:10.1038/nature10581
- 13 Klein, A. G., and J. L. Kincaid, 2006: Retreat of glaciers on Puncak Jaya, Irian Jaya, determined from 2000 and 2002
14 IKONOS satellite images. *Journal of Glaciology*, **52**, 65-79. doi:10.3189/172756506781828818
- 15 Kopp, R. E., F. J. Simons, J. X. Mitrovica, A. C. Maloof, and M. Oppenheimer, 2009: Probabilistic assessment of sea
16 level during the last interglacial stage. *Nature*, **462**, 863-867. doi:10.1038/nature08686
- 17 Kozlovsky, A. M., Y. L. Nazintsev, V. I. Fedotov, and N. V. Cherepanov, 1977: Fast ice of the Eastern Antarctic (in
18 Russian). *Proceedings of the Soviet Antarctic Expedition*, **63**, 1-129
- 19 Krabill, W., et al., 1999: Rapid thinning of parts of the southern Greenland ice sheet. *Science*, **283**, 1522-1524.
20 doi:10.1126/science.283.5407.1522
- 21 Krabill, W. B., et al., 2002: Aircraft laser altimetry measurement of elevation changes of the greenland ice sheet:
22 technique and accuracy assessment. *Journal of Geodynamics*, **34**, 357-376. doi:10.1016/s0264-3707(02)00040-6
- 23 Kuipers Munneke, P., S. R. M. Ligtenberg, M. R. van den Broeke, and D. G. Vaughan, Submitted: Firn air depletion as
24 a trigger for Antarctic ice-shelf collapse. *Science*
- 25 Kulkarni, A. V., I. M. Bahuguna, B. P. Rathore, S. K. Singh, S. S. Randhawa, R. K. Sood, and S. Dhar, 2007: Glacial
26 retreat in Himalaya using Indian Remote Sensing satellite data. *Curr. Sci.*, **92**, 69-74
- 27 Kunkel, K. E., M. A. Palecki, K. G. Hubbard, D. A. Robinson, K. T. Redmond, and D. R. Easterling, 2007: Trend
28 identification in twentieth-century US snowfall: The challenges. *J. Atmos. Ocean. Technol.*, **24**, 64-73.
29 doi:10.1175/jtech2017.1
- 30 Kutuzov, S., and M. Shahgedanova, 2009: Glacier retreat and climatic variability in the eastern Terskey-Alatoo, inner
31 Tien Shan between the middle of the 19th century and beginning of the 21st century. *Glob. Planet. Change*, **69**,
32 59-70. doi:10.1016/j.gloplacha.2009.07.001
- 33 Kwok, R., 2004: Annual cycles of multiyear sea ice coverage of the Arctic Ocean: 1999-2003. *Journal of Geophysical*
34 *Research-Oceans*, **109**. doi:10.1029/2003jc002238
- 35 ———, 2005: Variability of Nares Strait ice flux. *Geophys. Res. Lett.*, **32**. doi:L24502, 10.1029/2005gl024768
- 36 ———, 2007: Near zero replenishment of the Arctic multiyear sea ice cover at the end of 2005 summer. *Geophys. Res.*
37 *Lett.*, **34**. doi:L05501, 10.1029/2006gl028737
- 38 ———, 2009: Outflow of Arctic Ocean Sea Ice into the Greenland and Barents Seas: 1979-2007. *J. Clim.*, **22**, 2438-2457.
39 doi:10.1175/2008jcli2819.1
- 40 Kwok, R., and D. A. Rothrock, 1999: Variability of Fram Strait ice flux and North Atlantic Oscillation. *Journal of*
41 *Geophysical Research-Oceans*, **104**, 5177-5189. doi:10.1029/1998jc900103
- 42 ———, 2009: Decline in Arctic sea ice thickness from submarine and ICESat records: 1958-2008. *Geophys. Res. Lett.*,
43 **36**. doi:L15501, 10.1029/2009gl039035
- 44 Kwok, R., and G. F. Cunningham, 2010: Contribution of melt in the Beaufort Sea to the decline in Arctic multiyear sea
45 ice coverage: 1993-2009. *Geophys. Res. Lett.*, **37**. doi:L20501, 10.1029/2010gl044678
- 46 Kwok, R., G. F. Cunningham, M. Wensnahan, I. Rigor, H. J. Zwally, and D. Yi, 2009: Thinning and volume loss of the
47 Arctic Ocean sea ice cover: 2003-2008. *Journal of Geophysical Research-Oceans*, **114**. doi:C07005,
48 10.1029/2009jc005312
- 49 Lambrecht, A., and M. Kuhn, 2007: Glacier changes in the Austrian Alps during the last three decades, derived from
50 the new Austrian glacier inventory. *Annals of Glaciology, Vol 46, 2007*, **46**, 177-184.
51 doi:10.3189/172756407782871341
- 52 Larsen, C. F., R. J. Motyka, J. T. Freymueller, K. A. Echelmeyer, and E. R. Ivins, 2005: Rapid viscoelastic uplift in
53 southeast Alaska caused by post-Little Ice Age glacial retreat. *Earth and Planetary Science Letters*, **237**.
54 doi:10.1016/j.epsl.2005.06.032
- 55 Latifovic, R., and D. Pouliot, 2007: Analysis of climate change impacts on lake ice phenology in Canada using the
56 historical satellite data record. *Remote Sensing of Environment*, **106**, 492-507. doi:10.1016/j.rse.2006.09.015
- 57 Laxon, S., N. Peacock, and D. Smith, 2003: High interannual variability of sea ice thickness in the Arctic region.
58 *Nature*, **425**, 947-950. doi:10.1038/nature02050
- 59 Le Bris, R., F. Paul, H. Frey, and T. Bolch, 2011: A new satellite-derived glacier inventory for western Alaska. *Annals*
60 *of Glaciology*, **52**
- 61 Leclercq, P. W., and J. Oerlemans, 2012: Global and hemispheric temperature reconstruction from glacier length
62 fluctuations. *Climate Dynamics*, **38**, 1065-1079. doi:10.1007/s00382-011-1145-7

- 1 Leclercq, P. W., J. Oerlemans, and J. G. Cogley, 2011: Estimating the Glacier Contribution to Sea level Rise for the
2 Period 1800-2005. *Surveys in Geophysics*, **32**, 519-535. doi:10.1007/s10712-011-9121-7
- 3 Leclercq, P. W., A. Weidick, F. Paul, T. Bolch, M. Citterio, and Oerlemans, J., Submitted: Historical glacier length
4 changes in West Greenland. *The Cryosphere*, submitted
- 5 Lemke, P., et al., 2007: Observations: Changes in Snow, Ice and Frozen Ground. *Climate Change 2007: The Physical
6 Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental
7 Panel on Climate Change*, Cambridge University Press.
- 8 Lenaerts, J. T. M., M. R. van den Broeke, W. J. van de Berg, E. van Meijgaard, and P. Kuipers Munneke, 2012: A new,
9 high resolution surface mass balance map of Antarctica (1979-2010) based on regional climate modeling.
10 *Geophys. Res. Lett.*, **39**, 1-5. doi:doi:10.1029/2011GL050713
- 11 Lenaerts, J. T. M., M. R. van den Broeke, W. J. van de Berg, E. van Meijgaard, and P. Kuipers Munneke, In press: A
12 new, high-resolution surface mass balance map of Antarctica (1989-2009) based on regional atmospheric climate
13 modeling. *Geophys. Res. Lett.*
- 14 Lewkowicz, A. G., B. Etzelmuller, and S. L. Smith, 2011: Characteristics of Discontinuous Permafrost based on
15 Ground Temperature Measurements and Electrical Resistivity Tomography, Southern Yukon, Canada.
16 *Permafrost and Periglacial Processes*, **22**, 320-342. doi:10.1002/ppp.703
- 17 Leysinger Vieli, G., and G. H. Gudmundsson, 2004: On estimating length fluctuations of glaciers caused by changes in
18 climatic forcing. *J. Geophys. Res.-Earth Surf.*, **109**, 14. doi:F01007, 10.1029/2003jf000027
- 19 Li, B. L., A. X. Zhu, Y. C. Zhang, T. Pei, C. Z. Qin, and C. H. Zhou, 2006: Glacier change over the past four decades in
20 the middle Chinese Tien Shan. *Journal of Glaciology*, **52**, 425-432. doi:10.3189/172756506781828557
- 21 Li, X., R. Jin, X. D. Pan, T. J. Zhang, and J. W. Guo, 2012: Changes in the near-surface soil freeze-thaw cycle on the
22 Qinghai-Tibetan Plateau. *International Journal of Applied Earth Observation and Geoinformation*, **17**, 33-42.
23 doi:10.1016/j.jag.2011.12.002
- 24 Li, X., et al., 2008: Cryospheric change in China. *Glob. Planet. Change*, **62**, 210-218.
25 doi:10.1016/j.gloplacha.2008.02.001
- 26 Ling, F., and T. Zhang, 2003: Numerical simulation of permafrost thermal regime and talik development under shallow
27 thaw lakes on the Alaskan Arctic Coastal Plain. *Journal of Geophysical Research-Atmospheres*, **108**, 11.
28 doi:4511, 10.1029/2002jd003014
- 29 Linsbauer, A., F. Paul, and W. Haeberli, 2012: Modeling glacier thickness distribution and bed topography over entire
30 mountain ranges with GlabTop: A new, fast and robust approach. *Journal of Geophysical Research - Earth
31 Surface*, **117**, F03007
- 32 Liu, L., T. Zhang, and J. Wahr, 2010: InSAR measurements of surface deformation over permafrost on the North Slope
33 of Alaska. *J. Geophys. Res.-Earth Surf.*, **115**. doi:10.1029/2009jf001547
- 34 Livingstone, D. M., R. Adrian, T. Blencker, G. George, and G. A. Weyhenmeyer, 2010: Chapter 4: Lake Ice Phenology.
35 *The Impact of Climate Change on European Lakes*, D. G. George, Ed., 51-61, doi:10.1007.1978-1090-1481-
36 2945-1004_1004.
- 37 Lopez-Moreno, J. I., and S. M. Vicente-Serrano, 2007: Atmospheric circulation influence on the interannual variability
38 of snow pack in the Spanish Pyrenees during the second half of the 20th century. *Nordic Hydrology*, **38**, 33-44.
39 doi:10.2166/nh.2007.030
- 40 Lopez, P., P. Chevallier, V. Favier, B. Pouyaud, F. Ordenes, and J. Oerlemans, 2010: A regional view of fluctuations in
41 glacier length in southern South America. *Glob. Planet. Change*, **71**, 85-108.
42 doi:10.1016/j.gloplacha.2009.12.009
- 43 Loriaux, T., and G. Casassa, 2012: Evolution of glacial lakes from the Northern Patagonia Icefield and terrestrial water
44 storage in a sea level rise context. *Journal of Glaciology*
- 45 Luckman, A., and T. Murray, 2005: Seasonal variation in velocity before retreat of Jakobshavn Isbrae, Greenland.
46 *Geophys. Res. Lett.*, **32**, 4. doi:L08501, 10.1029/2005gl022519
- 47 Luethi, M. P., A. Bauder, and M. Funk, 2010: Volume change reconstruction of Swiss glaciers from length change data.
48 *J. Geophys. Res.-Earth Surf.*, **115**. doi:F04022, 10.1029/2010jf001695
- 49 Luthcke, S. B., A. A. Arendt, D. D. Rowlands, J. J. McCarthy, and C. F. Larsen, 2008: Recent glacier mass changes in
50 the Gulf of Alaska region from GRACE mascon solutions. *Journal of Glaciology*, **54**, 767-777
- 51 Luthcke, S. B., et al., 2006: Recent Greenland ice mass loss by drainage system from satellite gravity observations.
52 *Science*, **314**, 1286-1289. doi:10.1126/science.1130776
- 53 MacAyeal, D. R., et al., 2006: Transoceanic wave propagation links iceberg calving margins of Antarctica with storms
54 in tropics and Northern Hemisphere. *Geophys. Res. Lett.*, **33**, 4. doi:L17502, 10.1029/2006gl027235
- 55 Machguth, H., F. Paul, S. Kotlarski, and M. Hoelzle, 2009: Calculating distributed glacier mass balance for the Swiss
56 Alps from regional climate model output: A methodical description and interpretation of the results. *Journal of
57 Geophysical Research-Atmospheres*, **114**. doi:10.1029/2009jd011775
- 58 Magnuson, J. J., et al., 2000: Historical trends in lake and river ice cover in the Northern Hemisphere. *Science*, **289**,
59 1743-1746. doi:10.1126/science.289.5485.1743
- 60 Magnusson, E., H. Bjornsson, J. Dall, and F. Palsson, 2005: Volume changes of Vatnajokull ice cap, Iceland, due to
61 surface mass balance, ice flow, and subglacial melting at geothermal areas. *Geophys. Res. Lett.*, **32**, 4.
62 doi:L05504, 10.1029/2004gl021615

- 1 Mahoney, A., H. Eicken, and L. Shapiro, 2007: How fast is landfast sea ice? A study of the attachment and detachment
2 of nearshore ice at Barrow, Alaska. *Cold Reg. Sci. Tech.*, **47**, 233-255. doi:10.1016/j.coldregions.2006.09.005
- 3 Malkova, G. V., 2008: The last twenty-five years of changes in permafrost temperature of the European Russian Arctic.
4 *Ninth International Conference on Permafrost*, Institute of Northern Engineering, University of Alaska
5 Fairbanks, 1119-1124.
- 6 Marchenko, S. S., A. P. Gorbunov, and V. E. Romanovsky, 2007: Permafrost warming in the Tien Shan Mountains,
7 Central Asia. *Glob. Planet. Change*, **56**, 311-327. doi:10.1016/j.glopacha.2006.07.023
- 8 Markus, T., and D. J. Cavalieri, 2000: An enhancement of the NASA Team sea ice algorithm. *Ieee Transactions on*
9 *Geoscience and Remote Sensing*, **38**, 1387-1398. doi:10.1109/36.843033
- 10 Markus, T., J. C. Stroeve, and J. Miller, 2009: Recent changes in Arctic sea ice melt onset, freezeup, and melt season
11 length. *Journal of Geophysical Research-Oceans*, **114**. doi:10.1029/2009jc005436
- 12 Marshall, G. J., A. Orr, N. P. M. van Lipzig, and J. C. King, 2006: The Impact of a Changing Southern Hemisphere
13 Annular Mode on Antarctic Peninsula Summer Temperatures. *J. Clim.*, **19**, 5388--5404
- 14 Martinson, D. G., S. E. Stammerjohn, R. A. Iannuzzi, R. C. Smith, and M. Vernet, 2008: Western Antarctic Peninsula
15 physical oceanography and spatio-temporal variability. *Deep-Sea Research Part II-Topical Studies in*
16 *Oceanography*, **55**, 1964-1987. doi:10.1016/j.dsr2.2008.04.038
- 17 Marty, C., and R. Meister, 2012: Long-term snow and weather observations at Weissfluhjoch and its relation to other
18 high-altitude observatories in the Alps. *Theoretical and Applied Climatology*. doi:doi: 10.1007/s00704-012-
19 0584-3
- 20 Marzeion, B., A. Jarosch, and M. Hofer, submitted: Contribution of Mountain Glaciers to Sea Level Rise during the
21 20th and 21st Centuries. *The Cryosphere*
- 22 Masiokas, M. H., A. Rivera, L. E. Espizua, R. Villalba, S. Delgado, and J. C. Aravena, 2009: Glacier fluctuations in
23 extratropical South America during the past 1000 years. *Palaeogeography Palaeoclimatology Palaeoecology*,
24 **281**, 242-268. doi:10.1016/j.palaeo.2009.08.006
- 25 Maslanik, J. A., C. Fowler, J. Stroeve, S. Drobot, J. Zwally, D. Yi, and W. Emery, 2007: A younger, thinner Arctic ice
26 cover: Increased potential for rapid, extensive sea-ice loss. *Geophys. Res. Lett.*, **34**. doi:10.1029/2007gl032043
- 27 Massom, R. A., P. Reid, B. Raymond, S. Stammerjohn, A. D. Fraser, and S. Ushio, Submitted: Change and variability
28 in East Antarctic sea ice seasonality, 1979-2010. *PLoS ONE*
- 29 Massom, R. A., et al., 2008: West Antarctic Peninsula sea ice in 2005: Extreme ice compaction and ice edge retreat due
30 to strong anomaly with respect to climate. *Journal of Geophysical Research-Oceans*, **113**, 23. doi:C02s20,
31 10.1029/2007jc004239
- 32 Matsuo, K., and K. Heki, 2010: Time-variable ice loss in Asian high mountains from satellite gravimetry. *Earth and*
33 *Planetary Science Letters*, **290**, 30-36. doi:10.1016/j.epsl.2009.11.053
- 34 Mazhitova, G. G., 2008: Soil temperature regimes in the discontinuous permafrost zone in the east European Russian
35 Arctic. *Eurasian Soil Science*, **41**, 48-62. doi:10.1134/s1064229308010067
- 36 Mazhitova, G. G., and D. A. Kaverin, 2007: Thaw depth dynamics and soil surface subsidence at a Circumpolar Active
37 Layer Monitoring (CALM) site in the East European Russian Arctic. *Kriosfera Zemli*, vol. **XI**, N, 20-30
- 38 McBean, G., et al., 2005: Arctic Climate: Past and Present. *Arctic Climate Impact Assessment*, Cambridge University
39 Press, 21-60.
- 40 McGuire, A. D., et al., 2009: Sensitivity of the carbon cycle in the Arctic to climate change. *Ecol. Monogr.*, **79**, 523-
41 555. doi:10.1890/08-2025.1
- 42 McNabb, R. W., R. Hock, and A. C. Beedlow, submitted: A first order regional estimate of frontal ablation at Alaska
43 tidewater glaciers. *Journal of Glaciology*
- 44 Meier, M. F., 1984: Contribution of small glaciers to global sea level. *Science*, **226**, 1418-1421
- 45 Mercer, J. H., 1978: WEST ANTARCTIC ICE SHEET AND CO2 GREENHOUSE EFFECT - THREAT OF
46 DISASTER. *Nature*, **271**, 321-325. doi:10.1038/271321a0
- 47 Mitchell, T. D., and P. D. Jones, 2005: An improved method of constructing a database of monthly climate observations
48 and associated high-resolution grids. *International Journal of Climatology*, **25**. doi:10.1002/joc.1181
- 49 Moholdt, G., B. Wouters, and A. S. Gardner, 2012: Recent contribution to sea level rise from glaciers and ice caps in
50 the Russian High Arctic. *Geophys. Res. Lett.*, **39**. doi:10.1029/2012GL051466, 2012
- 51 Moholdt, G., C. Nuth, J. O. Hagen, and J. Kohler, 2010: Recent elevation changes of Svalbard glaciers derived from
52 ICESat laser altimetry. *Remote Sensing of Environment*, **114**, 2756-2767. doi:10.1016/j.rse.2010.06.008
- 53 Molg, T., N. J. Cullen, D. R. Hardy, G. Kaser, and L. Klok, 2008: Mass balance of a slope glacier on Kilimanjaro and
54 its sensitivity to climate. *International Journal of Climatology*, **28**, 881-892. doi:10.1002/joc.1589
- 55 Molnia, B. F., 2007: Late nineteenth to early twenty-first century behavior of Alaskan glaciers as indicators of changing
56 regional climate. *Glob. Planet. Change*, **56**. doi:10.1016/j.glopacha.2006.07.011
- 57 Monaghan, A. J., D. H. Bromwich, and S. H. Wang, 2006: Recent trends in Antarctic snow accumulation from Polar
58 MM5 simulations. *Philosophical Transactions of the Royal Society a-Mathematical Physical and Engineering*
59 *Sciences*, **364**, 1683-1708. doi:10.1098/rsta.2006.1795
- 60 Moore, P., and M. A. King, 2008: Antarctic ice mass balance estimates from GRACE: Tidal aliasing effects. *J.*
61 *Geophys. Res.-Earth Surf.*, **113**. doi:F02005, 10.1029/2007jf000871
- 62 Motyka, R. J., L. Hunter, K. A. Echelmeyer, and C. Connor, 2003: Submarine melting at the terminus of a temperate
63 tidewater glacier, LeConte Glacier, Alaska, USA. *Annals of Glaciology, Vol 36*, **36**, 57-65

- 1 Motyka, R. J., M. Truffer, M. Fahnestock, J. Mortensen, S. Rysgaard, and I. Howat, 2011: Submarine melting of the
2 1985 Jakobshavn Isbrae floating tongue and the triggering of the current retreat. *J. Geophys. Res.-Earth Surf.*,
3 **116**. doi:10.1029/2009jf001632
- 4 Murray, T., T. Strozzi, A. Luckman, H. Jiskoot, and P. Christakos, 2003: Is there a single surge mechanism? Contrasts
5 in dynamics between glacier surges in Svalbard and other regions. *Journal of Geophysical Research-Solid Earth*,
6 **108**. doi:10.1029/2002jb001906
- 7 Murray, T., et al., 2010: Ocean regulation hypothesis for glacier dynamics in southeast Greenland and implications for
8 ice sheet mass changes. *J. Geophys. Res.-Earth Surf.*, **115**. doi:F03026, 10.1029/2009jf001522
- 9 Myers, P. G., C. Donnelly, and M. H. Ribergaard, 2009: Structure and variability of the West Greenland Current in
10 Summer derived from 6 repeat standard sections. *Progress in Oceanography*, **80**, 93-112.
11 doi:10.1016/j.pocean.2008.12.003
- 12 Narama, C., Y. Shimamura, D. Nakayama, and K. Abdrakhmatov, 2006: Recent changes of glacier coverage in the
13 western Terskey-Alatau range, Kyrgyz Republic, using Corona and Landsat. *Annals of Glaciology, Vol 43, 2006*,
14 **43**, 223-229. doi:10.3189/172756406781812195
- 15 Narozhnyi, Y., and V. Zemtsov, 2011: Current State of the Altai Glaciers (Russia) and Trends Over the Period of
16 Instrumental Observations 1952-2008. *Ambio*, **40**. doi:10.1007/s13280-011-0166-0
- 17 Nerem, R. S., and J. Wahr, 2011: Recent changes in the Earth's oblateness driven by Greenland and Antarctic ice mass
18 loss. *Geophys. Res. Lett.*, **38**, 6. doi:L13501, 10.1029/2011gl047879
- 19 Nesje, A., O. Lie, and S. O. Dahl, 2000: Is the North Atlantic Oscillation reflected in Scandinavian glacier mass balance
20 records? *Journal of Quaternary Science*, **15**. doi:10.1002/1099-1417(200009)15:6<587::aid-jqs533>3.0.co;2-2
- 21 Nghiem, S. V., I. G. Rigor, D. K. Perovich, P. Clemente-Colon, J. W. Weatherly, and G. Neumann, 2007: Rapid
22 reduction of Arctic perennial sea ice. *Geophys. Res. Lett.*, **34**, 6. doi:L19504, 10.1029/2007gl031138
- 23 Nick, F. M., A. Vieli, I. M. Howat, and I. Joughin, 2009: Large-scale changes in Greenland outlet glacier dynamics
24 triggered at the terminus. *Nature Geoscience*, **2**, 110-114. doi:10.1038/ngeo394
- 25 Nick, F. M., C. J. van der Veen, A. Vieli, and D. Benn, 2010: A physically based calving model applied to marine outlet
26 glaciers and implications for their dynamics. *Journal of Glaciology*, **56**, 781-794
- 27 Nick, F. M., et al., 2012a: Future sea level rise from Greenland's major outlet glaciers in a warming climate. *Nature*
- 28 Nick, F. M., et al., 2012b: The response of Petermann Glacier to shelf retreat and its future stability in context of
29 atmospheric and oceanic warming. *Journal of Glaciology*, **58**, 229-239. doi:10.3189/2012JoG11J242
- 30 Nie, Y., Y. L. Zhang, L. S. Liu, and J. P. Zhang, 2010: Glacial change in the vicinity of Mt. Qomolangma (Everest),
31 central high Himalayas since 1976. *Journal of Geographical Sciences*, **20**, 667-686. doi:10.1007/s11442-010-
32 0803-8
- 33 Noetzli, J., and D. Vonder Muehll, 2010: Permafrost in Switzerland 2006/2007 and 2007/2008. Cryospheric
34 Commission of the Swiss Academy of Sciences.
- 35 Nussbaumer, S. U., A. Nesje, and H. J. Zumbuhl, 2011: Historical glacier fluctuations of Jostedalbreen and Folgefonna
36 (southern Norway) reassessed by new pictorial and written evidence. *Holocene*, **21**, 455-471.
37 doi:10.1177/0959683610385728
- 38 Nuth, C., G. Moholdt, J. Kohler, J. O. Hagen, and A. Kaab, 2010: Svalbard glacier elevation changes and contribution
39 to sea level rise. *J. Geophys. Res.-Earth Surf.*, **115**, 16. doi:F01008, 10.1029/2008jf001223
- 40 O'Connor, F. M., et al., 2010: Possible role of wetlands, permafrost, and methane hydrates in the methane cycle under
41 future climate change: a review. *Reviews of Geophysics*, **48**, 33. doi:10.1029/2010rg000326
- 42 Oberman, N. G., 2008: Contemporary Permafrost Degradation of Northern European Russia. *Ninth International*
43 *Conference on Permafrost*, Institute of Northern Engineering, University of Alaska, Fairbanks, 1305-1310.
- 44 Oberman, N. G., 2012: Long-term temperature regime of the Northeast European permafrost region during
45 contemporary climate warming. *Tenth International Conference on Permafrost*, Salekhard, Yamel-Nenets
46 Autonomous District, Russia, 287-291.
- 47 Oerlemans, J., M. Dyurgerov, and R. de Wal, 2007: Reconstructing the glacier contribution to sea level rise back to
48 1850. *Cryosphere*, **1**, 59-65
- 49 Osterkamp, T. E., 2001: Sub-sea permafrost. *Encyclopedia of Ocean Sciences*, J. H. Steele, S. A. Thorpe, and K. K.
50 Turekian, Eds., Elsevier, 2902-2912.
- 51 ———, 2007: Characteristics of the recent warming of permafrost in Alaska. *J. Geophys. Res.-Earth Surf.*, **112**, 10.
52 doi:F02s02, 10.1029/2006jf000578
- 53 ———, 2008: Thermal State of Permafrost in Alaska During the Fourth Quarter of the Twentieth Century. *Ninth Int.*
54 *Conf. on Permafrost, June 29- July 3, 2008*, Univ. of Alaska, Fairbanks, Alaska, Institute of Northern
55 Engineering, 1333-1338.
- 56 Otero, J., F. J. Navarro, J. J. Lapazaran, M. Grabiec, D. Puczko, C. Molina, and A. Vieli, 2011: Modelling the seasonal
57 and long-term variations of the calving front position of Hansbreen. *Journal of Glaciology*
- 58 Overduin, P. P., and D. L. Kane, 2006: Frost boils and soil ice content: Field observations. *Permafrost and Periglacial*
59 *Processes*, **17**, 291-307. doi:10.1002/ppp.567
- 60 Overduin, P. P., H.-W. Hubberten, V. Rachold, N. Romanovskii, M. N. Grigoriev, and M. Kasymkaya, 2007:
61 Evolution and degradation of coastal and offshore permafrost in the Laptev and East Siberian Seas during the
62 last climatic cycle. *GSA Special Papers*, 97-111.

- 1 Overduin, P. P., S. Westermann, K. Yoshikawa, T. Haberlau, V. Romanovsky, and S. Wetterich, submitted: Geoelectric
2 observations of the degradation of near-shore submarine permafrost at Barrow (Alaskan Beaufort Sea). *J.*
3 *Geophys. Res.*
- 4 Palmer, S., A. Shepherd, P. Nienow, and I. Joughin, 2011: Seasonal speedup of the Greenland Ice Sheet linked to
5 routing of surface water. *Earth and Planetary Science Letters*, **302**, 423-428. doi:10.1016/j.epsl.2010.12.037
- 6 Parkinson, C. L., 2002: Trends in the length of the Southern Ocean sea-ice season, 1979-99. *Annals of Glaciology, Vol*
7 *34, 2002*, **34**, 435-440. doi:10.3189/172756402781817482
- 8 Parkinson, C. L., and J. C. Comiso, 2008: Antarctic sea ice parameters from AMSR-E data using two techniques and
9 comparisons with sea ice from SSM/I. *Journal of Geophysical Research-Oceans*, **113**. doi:Artn C02s06,
10 10.1029/2007jc004253
- 11 Parkinson, C. L., and D. J. Cavalieri, Submitted: Antarctic Sea Ice Variability and Trends, 1979-2010
- 12 Paul, F., and A. Kaab, 2005: Perspectives on the production of a glacier inventory from multispectral satellite data in
13 Arctic Canada: Cumberland Peninsula, Baffin Island. *Annals of Glaciology, Vol 42, 2005*, **42**, 59-66.
14 doi:10.3189/172756405781813087
- 15 Paul, F., and W. Haeberli, 2008: Spatial variability of glacier elevation changes in the Swiss Alps obtained from two
16 digital elevation models. *Geophys. Res. Lett.*, **35**, 5. doi:L21502, 10.1029/2008gl034718
- 17 Paul, F., and L. M. Andreassen, 2009: A new glacier inventory for the Svartisen region, Norway, from Landsat ETM
18 plus data: challenges and change assessment. *Journal of Glaciology*, **55**, 607-618
- 19 Paul, F., A. Kaab, and W. Haeberli, 2007: Recent glacier changes in the Alps observed by satellite: Consequences for
20 future monitoring strategies. *Glob. Planet. Change*, **56**, 111-122. doi:10.1016/j.gloplacha.2006.07.007
- 21 Paul, F., L. M. Andreassen, and S. H. Winsvold, 2011a: A new glacier inventory for the Jostedalbreen region, Norway,
22 from Landsat TM scenes of 2006 and changes since 1966. *Annals of Glaciology*, **52**
- 23 Paul, F., H. Frey, and R. Le Bris, 2011b: A new glacier inventory for the European Alps from Landsat TM scenes of
24 2003: challenges and results. *Annals of Glaciology*, **52**, 144-152
- 25 Paul, F., A. Kaab, M. Maisch, T. Kellenberger, and W. Haeberli, 2004: Rapid disintegration of Alpine glaciers observed
26 with satellite data. *Geophys. Res. Lett.*, **31**, 4. doi:L21402, 10.1029/2004gl020816
- 27 Paulson, A., S. J. Zhong, and J. Wahr, 2007: Inference of mantle viscosity from GRACE and relative sea level data.
28 *Geophysical Journal International*, **171**, 497-508. doi:10.1111/j.1365-246X.2007.03556.x
- 29 Payne, A. J., A. Vieli, A. P. Shepherd, D. J. Wingham, and E. Rignot, 2004: Recent dramatic thinning of largest West
30 Antarctic ice stream triggered by oceans. *Geophys. Res. Lett.*, **31**. doi:10.1029/2004gl021284
- 31 Peduzzi, P., C. Herold, and W. Silverio, 2010: Assessing high altitude glacier thickness, volume and area changes using
32 field, GIS and remote sensing techniques: the case of Nevado Coropuna (Peru). *Cryosphere*, **4**, 313-323.
33 doi:10.5194/tc-4-313-2010
- 34 Peltier, W. R., 2009: Closure of the budget of global sea level rise over the GRACE era: the importance and magnitudes
35 of the required corrections for global glacial isostatic adjustment. *Quat. Sci. Rev.*, **28**, 1658-1674.
36 doi:10.1016/j.quascirev.2009.04.004
- 37 Pelto, M. S., 2010: Forecasting temperate alpine glacier survival from accumulation zone observations. *Cryosphere*, **4**,
38 67-75
- 39 Perovich, D. K., J. A. Richter-Menge, K. F. Jones, and B. Light, 2008: Sunlight, water, and ice: Extreme Arctic sea ice
40 melt during the summer of 2007. *Geophys. Res. Lett.*, **35**. doi:10.1029/2008gl034007
- 41 Petrenko, V. V., et al., 2010: Methane from the East Siberian Arctic Shelf. *Science*, **329**, 1146-1147
- 42 Pfeffer, W. T., 2007: A simple mechanism for irreversible tidewater glacier retreat. *J. Geophys. Res.-Earth Surf.*, **112**.
43 doi:10.1029/2006jfr000590
- 44 ———, 2011: Land Ice and Sea Level Rise A Thirty-Year Perspective. *Oceanography*, **24**, 94-111
- 45 Phillips, T., H. Rajaram, and K. Steffen, 2010: Cryo-hydrologic warming: A potential mechanism for rapid thermal
46 response of ice sheets. *Geophys. Res. Lett.*, **37**. doi:10.1029/2010gl044397
- 47 Phillips, T., W. Colgan, H. Rajaram, and K. Steffen, Submitted: Evaluation of cryo-hydrologic warming as an
48 explanation for increased ice velocities near the equilibrium line, Southwest Greenland. *J. Geophys. Res.*
- 49 Pollard, D., and R. M. DeConto, 2009: Modelling West Antarctic ice sheet growth and collapse through the past five
50 million years. *Nature*, **458**, 329-333. doi:10.1038/nature07809
- 51 Polyakov, I. V., et al., 2003: Long-term ice variability in Arctic marginal seas. *J. Clim.*, **16**, 2078-2085.
52 doi:10.1175/1520-0442(2003)016<2078:liviam>2.0.co;2
- 53 Polyakov, I. V., et al., 2010: Arctic Ocean Warming Contributes to Reduced Polar Ice Cap. *Journal of Physical*
54 *Oceanography*, **40**. doi:10.1175/2010jpo4339.1
- 55 Post, A., S. O'Neel, R. J. Motyka, and G. Streveler, 2011: A Complex Relationship Between Calving Glaciers and
56 Climate. *EOS*, **92**, 305-306
- 57 Pritchard, H. D., and D. G. Vaughan, 2007: Widespread acceleration of tidewater glaciers on the Antarctic Peninsula. *J.*
58 *Geophys. Res.-Earth Surf.*, **112**. doi:F03s29, 10.1029/2006jfr000597
- 59 Pritchard, H. D., S. B. Luthcke, and A. H. Fleming, 2010: Understanding ice-sheet mass balance: progress in satellite
60 altimetry and gravimetry. *Journal of Glaciology*, **56**, 1151-1161
- 61 Pritchard, H. D., R. J. Arthern, D. G. Vaughan, and L. A. Edwards, 2009: Extensive dynamic thinning on the margins of
62 the Greenland and Antarctic ice sheets. *Nature*, **461**, 971-975. doi:doi 10.1038/nature08471

- 1 Pritchard, H. D., S. R. M. Ligtenberg, H. A. Fricker, D. G. Vaughan, M. R. van den Broeke, and L. Padman, 2012:
2 Antarctic ice loss driven by ice-shelf melt. *Nature*, **484**, 502-505. doi:10.1038/nature10968
- 3 Prowse, T., et al., 2011: Arctic Freshwater Ice and Its Climatic Role. *Ambio*, **40**, 46-52
- 4 Prowse, T. D., and K. Brown, 2010: Hydro-ecological effects of changing Arctic river and lake ice covers: a review.
5 *Hydrol. Res.*, **41**, 454-461. doi:10.2166/nh.2010.142
- 6 Quincey, D. J., M. Braun, N. F. Glasser, M. P. Bishop, K. Hewitt, and A. Luckman, 2011: Karakoram glacier surge
7 dynamics. *Geophys. Res. Lett.*, **38**. doi:10.1029/2011gl049004
- 8 Rabatel, A., 2012: Review article of the current state of glaciers in the tropical Andes: a multi-century perspective on
9 glacier evolution and climate change. . *The Cryosphere*, submitted
- 10 Rabatel, A., B. Francou, V. Jomelli, P. Naveau, and D. Grancher, 2008: A chronology of the Little Ice Age in the
11 tropical Andes of Bolivia (16 degrees S) and its implications for climate reconstruction. . **70**, 198-212.
12 doi:10.1016/j.yqres.2008.02.012
- 13 Rachold, V., et al., 2007: Near-shore Arctic Subsea Permafrost in Transition. *EOS Transactions of the American*
14 *Geophysical Union*, **88**, 149-156
- 15 Racoviteanu, A. E., Y. Arnaud, M. W. Williams, and J. Ordonez, 2008: Decadal changes in glacier parameters in the
16 Cordillera Blanca, Peru, derived from remote sensing. *Journal of Glaciology*, **54**, 499-510.
17 doi:10.3189/002214308785836922
- 18 Radic, V., and R. Hock, 2010: Regional and global volumes of glaciers derived from statistical upscaling of glacier
19 inventory data. *J. Geophys. Res.-Earth Surf.*, **115**. doi:10.1029/2009jf001373
- 20 Radic, V., A. Bliss, A. C. Beedlow, R. Hock, E. Miles, and J. G. Cogley, submitted: Regional and global projections of
21 the 21st century glacier mass changes in response to climate scenarios from GCMs. *Climate Dynamics*
- 22 Ramillien, G., A. Lombard, A. Cazenave, E. R. Ivins, M. Llubes, F. Remy, and R. Biancale, 2006: Interannual
23 variations of the mass balance of the Antarctica and Greenland ice sheets from GRACE. *Glob. Planet. Change*,
24 **53**, 198-208
- 25 Rampal, P., J. Weiss, and D. Marsan, 2009: Positive trend in the mean speed and deformation rate of Arctic sea ice,
26 1979-2007. *Journal of Geophysical Research-Oceans*, **114**. doi:10.1029/2008jc005066
- 27 Rastner, P., T. Bolch, N. Molg, H. Machguth, and F. Paul, 2012: The first complete glacier inventory for entire
28 Greenland. *The Cryosphere*
- 29 Ravel, L., and P. Deline, 2011: Climate influence on rockfalls in high-Alpine steep rockwalls: The north side of the
30 Aiguilles de Chamonix (Mont Blanc massif) since the end of the 'Little Ice Age'. *Holocene*, **21**, 357-365.
31 doi:10.1177/0959683610374887
- 32 Ravel, L., F. Allignol, P. Deline, S. Gruber, and M. Ravello, 2010: Rock falls in the Mont Blanc Massif in 2007 and
33 2008. *Landslides*, **7**, 493-501. doi:10.1007/s10346-010-0206-z
- 34 Ridley, J., J. M. Gregory, P. Huybrechts, and J. Lowe, 2010: Thresholds for irreversible decline of the Greenland ice
35 sheet. *Climate Dynamics*, **35**, 1065-1073. doi:10.1007/s00382-009-0646-0
- 36 Rignot, E., 2008: Changes in West Antarctic ice stream dynamics observed with ALOS PALSAR data. *Geophys. Res.*
37 *Lett.*, **35**. doi:L12505, 10.1029/2008gl033365
- 38 Rignot, E., and R. H. Thomas, 2002: Mass balance of polar ice sheets. *Science*, **297**, 1502-1506.
39 doi:10.1126/science.1073888
- 40 Rignot, E., and S. S. Jacobs, 2002: Rapid bottom melting widespread near Antarctic ice sheet grounding lines. *Science*,
41 **296**, 2020-2023. doi:10.1126/science.1070942
- 42 Rignot, E., and P. Kanagaratnam, 2006: Changes in the velocity structure of the Greenland ice sheet. *Science*, **311**, 986-
43 990. doi:10.1126/science.1121381
- 44 Rignot, E., and J. Mouginot, 2012: Ice flow in Greenland for the International Polar Year 2008-2009. *Geophys. Res.*
45 *Lett.*, **39**. doi:10.1029/2012gl051634
- 46 Rignot, E., A. Rivera, and G. Casassa, 2003: Contribution of the Patagonia Icefields of South America to sea level rise.
47 *Science*, **302**, 434-437. doi:10.1126/science.1087393
- 48 Rignot, E., M. Koppes, and I. Velicogna, 2010: Rapid submarine melting of the calving faces of West Greenland
49 glaciers. *Nature Geoscience*, **3**, 187-191. doi:10.1038/ngeo765
- 50 Rignot, E., J. Mouginot, and B. Scheuchl, 2011a: Ice Flow of the Antarctic Ice Sheet. *Science*, **333**, 1427-1430.
51 doi:10.1126/science.1208336
- 52 —, 2011b: Antarctic grounding line mapping from differential satellite radar interferometry. *Geophys. Res. Lett.*, **38**.
53 doi:10.1029/2011gl047109
- 54 Rignot, E., J. E. Box, E. Burgess, and E. Hanna, 2008a: Mass balance of the Greenland ice sheet from 1958 to 2007.
55 *Geophys. Res. Lett.*, **35**. doi:L20502, 10.1029/2008gl035417
- 56 Rignot, E., I. Velicogna, M. R. van den Broeke, A. Monaghan, and J. Lenaerts, 2011c: Acceleration of the contribution
57 of the Greenland and Antarctic ice sheets to sea level rise. *Geophys. Res. Lett.*, **38**, 5.
58 doi:L05503,10.1029/2011gl046583
- 59 Rignot, E., G. Casassa, P. Gogineni, W. Krabill, A. Rivera, and R. Thomas, 2004: Accelerated ice discharge from the
60 Antarctic Peninsula following the collapse of Larsen B ice shelf. *Geophys. Res. Lett.*, **31**, 4. doi:L18401,
61 10.1029/2004gl020697

- 1 Rignot, E., J. L. Bamber, M. R. Van Den Broeke, C. Davis, Y. H. Li, W. J. Van De Berg, and E. Van Meijgaard, 2008b:
2 Recent Antarctic ice mass loss from radar interferometry and regional climate modelling. *Nature Geoscience*, **1**,
3 106-110. doi:10.1038/ngeo102
- 4 Riseborough, D. W., 1990: Soil latent heat as a filter of the climate signal in permafrost. *Proceedings of the Fifth*
5 *Canadian Permafrost Conference, Collection Nordicana*, Universite Laval, Quebec, 199-205.
- 6 Riva, R. E. M., et al., 2009: Glacial Isostatic Adjustment over Antarctica from combined ICESat and GRACE satellite
7 data. *Earth and Planetary Science Letters*, **288**, 516-523. doi:10.1016/j.epsl.2009.10.013
- 8 Rivera, A., G. Casassa, J. Bamber, and A. Kaab, 2005: Ice-elevation changes of Glaciar Chico, southern Patagonia,
9 using ASTER DEMs, aerial photographs and GPS data. *Journal of Glaciology*, **51**, 105-112.
10 doi:10.3189/172756505781829557
- 11 Rivera, A., F. Bown, D. Carrion, and P. Zenteno, 2012: Glacier responses to recent volcanic activity in Southern Chile.
12 *Environmental Research Letters*, **7**. doi:10.1088/1748-9326/7/1/014036
- 13 Rivera, A., T. Benham, G. Casassa, J. Bamber, and J. A. Dowdeswell, 2007: Ice elevation and areal changes of glaciers
14 from the Northern Patagonia Icefield, Chile. *Glob. Planet. Change*, **59**, 126-137.
15 doi:10.1016/j.gloplacha.2006.11.037
- 16 Robinson, D. A., K. F. Dewey, and R. R. Heim, 1993: GLOBAL SNOW COVER MONITORING - AN UPDATE.
17 *Bulletin of the American Meteorological Society*, **74**, 1689-1696
- 18 Roer, I., W. Haeberli, M. Avian, V. Kaufmann, R. Delaloye, C. Lambiel, and A. Käab, 2008: Observations and
19 considerations on destabilizing active rock glaciers in the European Alps. *Ninth International Conference on*
20 *Permafrost*, Institute of Northern Engineering, University of Alaska, Fairbanks, 1505-1510.
- 21 Rogers, J. C., and M. P. Hung, 2008: The Odden ice feature of the Greenland Sea and its association with atmospheric
22 pressure, wind, and surface flux variability from reanalyses. *Geophys. Res. Lett.*, **35**, 5. doi:L08504,
23 10.1029/2007gl032938
- 24 Romanovskii, N. N., H. W. Hubberten, A. Gavrilov, V. E. Tumskey, and A. L. Kholodov, 2004: Permafrost of the east
25 Siberian Arctic shelf and coastal lowlands. *Quat. Sci. Rev.*, **23**, 1359-1369. doi:10.1016/j.quascirev.2003.12.014
- 26 Romanovskii, N. N., H. W. Hubberten, A. V. Gavrilov, A. A. Eliseeva, and G. S. Tipenko, 2005: Offshore permafrost
27 and gas hydrate stability zone on the shelf of East Siberian Seas. *Geo-Mar. Lett.*, **25**, 167-182.
28 doi:10.1007/s00367-004-0198-6
- 29 Romanovsky, V. E., S. L. Smith, and H. H. Christiansen, 2010a: Permafrost Thermal State in the Polar Northern
30 Hemisphere during the International Polar Year 2007-2009: a Synthesis. *Permafrost and Periglacial Processes*,
31 **21**, 106-116. doi:10.1002/ppp.689
- 32 Romanovsky, V. E., et al., 2010b: Thermal State of Permafrost in Russia. *Permafrost and Periglacial Processes*, **21**,
33 136-155. doi:10.1002/ppp.683
- 34 Ross, N., et al., 2012: Steep reverse bed slope at the grounding line of the Weddell Sea sector in West Antarctica.
35 *Nature Geosci*, **5**, 393-396
- 36 Rothrock, D. A., Y. Yu, and G. A. Maykut, 1999: Thinning of the Arctic sea-ice cover. *Geophys. Res. Lett.*, **26**, 3469-
37 3472
- 38 Rothrock, D. A., D. B. Percival, and M. Wensnahan, 2008: The decline in arctic sea-ice thickness: Separating the
39 spatial, annual, and interannual variability in a quarter century of submarine data. *Journal of Geophysical*
40 *Research-Oceans*, **113**. doi:10.1029/2007jc004252
- 41 Rott, H., F. Muller, T. Nagler, and D. Floricioiu, 2011: The imbalance of glaciers after disintegration of Larsen-B ice
42 shelf, Antarctic Peninsula. *Cryosphere*, **5**, 125-134. doi:10.5194/tc-5-125-2011
- 43 Sannel, A. B. K., and P. Kuhry, 2011: Warming-induced destabilization of peat plateau/thermokarst lake complexes.
44 *Journal of Geophysical Research-Biogeosciences*, **116**, 16. doi:G03035, 10.1029/2010jg001635
- 45 Sasgen, I., H. Konrad, E. Ivins, M. R. van den Broeke, J. L. Bamber, Z. Martinec, and V. Klemann, 2012a: Antarctic
46 ice-mass balance 2002 to 2011: regional re-analysis of GRACE satellite gravimetry measurements with
47 improved estimate of glacial-isostatic adjustment. *The Cryosphere*
- 48 Sasgen, I., et al., 2012b: Timing and origin of recent regional ice-mass loss in Greenland. *Earth and Planetary Science*
49 *Letters*. doi:EPSL-D-11-00870
- 50 Sato, T., et al., 2011: Reevaluation of the viscoelastic and elastic responses to the past and present-day ice changes in
51 Southeast Alaska. *Tectonophysics*, **511**. doi:10.1016/j.tecto.2010.05.009
- 52 Scambos, T. A., C. Hulbe, M. Fahnestock, and J. Bohlander, 2000: The link between climate warming and break-up of
53 ice shelves in the Antarctic Peninsula. *Journal of Glaciology*, **46**, 516-530. doi:10.3189/172756500781833043
- 54 Scambos, T. A., J. A. Bohlander, C. A. Shuman, and P. Skvarca, 2004: Glacier acceleration and thinning after ice shelf
55 collapse in the Larsen B embayment, Antarctica. *Geophys. Res. Lett.*, **31**, 4. doi:L18402, 10.1029/2004gl020670
- 56 Schaefer, K., T. J. Zhang, L. Bruhwiler, and A. P. Barrett, 2011: Amount and timing of permafrost carbon release in
57 response to climate warming. *Tellus Ser. B-Chem. Phys. Meteorol.*, **63**, 165-180. doi:10.1111/j.1600-
58 0889.2011.00527.x
- 59 Scherler, D., B. Bookhagen, and M. R. Strecker, 2011: Spatially variable response of Himalayan glaciers to climate
60 change affected by debris cover. *Nature Geoscience*. doi:DOI: 10.1038/NCEO1068
- 61 Schiefer, E., B. Menounos, and R. Wheate, 2007: Recent volume loss of British Columbian glaciers, Canada. *Geophys.*
62 *Res. Lett.*, **34**, 6. doi:L16503, 10.1029/2007gl030780

- 1 Schmidt, S., and M. Nusser, 2012: Changes of High Altitude Glaciers from 1969 to 2010 in the Trans-Himalayan Kang
2 Yatzé Massif, Ladakh, Northwest India. *Arct. Antarct. Alp. Res.*, **44**, 107-121. doi:10.1657/1938-4246-44.1.107
- 3 Schneider, C., M. Schnirch, C. Acuna, G. Casassa, and R. Kilian, 2007: Glacier inventory of the Gran Campo Nevado
4 Ice Cap in the Southern Andes and glacier changes observed during recent decades. *Glob. Planet. Change*, **59**,
5 87-100. doi:10.1016/j.gloplacha.2006.11.023
- 6 Schoeneich, P., X. Bodin, J. Krysiacki, P. Deline, and L. Ravel, 2010: Permafrost in France, 1st report. Grenoble:
7 Institut de Géographie Alpine, Université Joseph Fourier, 68 pp.
- 8 Schoof, C., 2007: Ice sheet grounding line dynamics: Steady states, stability, and hysteresis. *J. Geophys. Res.-Earth*
9 *Surf.*, **112**. doi:10.1029/2006jf000664
- 10 ———, 2010: Ice-sheet acceleration driven by melt supply variability. *Nature*, **468**, 803-806. doi:10.1038/nature09618
- 11 Schrama, E. J. O., and B. Wouters, 2011: Revisiting Greenland ice sheet mass loss observed by GRACE. *Journal of*
12 *Geophysical Research-Solid Earth*, **116**. doi:B02407, 10.1029/2009jb006847
- 13 Schuur, E. A. G., J. G. Vogel, K. G. Crummer, H. Lee, J. O. Sickman, and T. E. Osterkamp, 2009: The effect of
14 permafrost thaw on old carbon release and net carbon exchange from tundra. *Nature*, **459**, 556-559.
15 doi:10.1038/nature08031
- 16 Schweiger, A., R. Lindsay, J. L. Zhang, M. Steele, H. Stern, and R. Kwok, 2011: Uncertainty in modeled Arctic sea ice
17 volume. *Journal of Geophysical Research-Oceans*, **116**. doi:10.1029/2011jc007084
- 18 Selmes, N., T. Murray, and T. D. James, 2011: Fast draining lakes on the Greenland Ice Sheet. *Geophys. Res. Lett.*, **38**,
19 5. doi:L15501, 10.1029/2011gl047872
- 20 Shahgedanova, M., G. Nosenko, T. Khromova, and A. Muraveyev, 2010: Glacier shrinkage and climatic change in the
21 Russian Altai from the mid-20th century: An assessment using remote sensing and PRECIS regional climate
22 model. *Journal of Geophysical Research-Atmospheres*, **115**, 12. doi:D16107, 10.1029/2009jd012976
- 23 Shahgedanova, M., G. Nosenko, I. Bushueva, and M. Ivano, 2012: Changes in area and geodetic mass balance of small
24 glaciers, Polar Urals, Russia, 1950–2008. *Journal of Glaciology*, **58**, 953-964
- 25 Shakhova, N., I. Semiletov, I. Leifer, A. Salyuk, P. Rekant, and D. Kosmach, 2010a: Geochemical and geophysical
26 evidence of methane release over the East Siberian Arctic Shelf. *Journal of Geophysical Research-Oceans*, **115**,
27 14. doi:C08007, 10.1029/2009jc005602
- 28 Shakhova, N., I. Semiletov, A. Salyuk, V. Yusupov, D. Kosmach, and O. Gustafsson, 2010b: Extensive Methane
29 Venting to the Atmosphere from Sediments of the East Siberian Arctic Shelf. *Science*, **327**, 1246-1250.
30 doi:10.1126/science.1182221
- 31 Shannon, S., et al., 2012: Enhanced basal lubrication and the contribution of the Greenland ice sheet to future sea level
32 rise *Proceedings of the National Academy of Sciences*
- 33 Sharkhuu, A., et al., 2007: Permafrost monitoring in the Hovsgol mountain region, Mongolia. *J. Geophys. Res.-Earth*
34 *Surf.*, **112**, 11. doi:F02s06, 10.1029/2006jf000543
- 35 Sharp, M., D. O. Burgess, J. G. Cogley, M. Ecclestone, C. Labine, and G. J. Wolken, 2011: Extreme melt on Canada's
36 Arctic ice caps in the 21st century. *Geophys. Res. Lett.*, **38**. doi:10.1029/2011gl047381
- 37 Shepherd, A., and others, 2012: A Reconciled Estimate of Ice Sheet Mass Balance. *Science*, **submitted**
- 38 Shepherd, A., D. Wingham, T. Payne, and P. Skvarca, 2003: Larsen ice shelf has progressively thinned. *Science*, **302**,
39 856-859
- 40 Shepherd, A., A. Hubbard, P. Nienow, M. King, M. McMillan, and I. Joughin, 2009: Greenland ice sheet motion
41 coupled with daily melting in late summer. *Geophys. Res. Lett.*, **36**. doi:L01501, 10.1029/2008gl035758
- 42 Shepherd, A., D. Wingham, D. Wallis, K. Giles, S. Laxon, and A. V. Sundal, 2010: Recent loss of floating ice and the
43 consequent sea level contribution. *Geophys. Res. Lett.*, **37**, 5. doi:L13503, 10.1029/2010gl042496
- 44 Shi, H. L., Y. Lu, Z. L. Du, L. L. Jia, Z. Z. Zhang, and C. X. Zhou, 2011: Mass change detection in Antarctic ice sheet
45 using ICESat block analysis techniques from 2003 similar to 2008. *Chinese Journal of Geophysics-Chinese*
46 *Edition*, **54**, 958-965. doi:10.3969/j.issn.0001-5733.2011.04.010
- 47 Shiklomanov, N. I., et al., 2010: Decadal variations of active-layer thickness in moisture-controlled landscapes, Barrow,
48 Alaska. *Journal of Geophysical Research-Biogeosciences*, **115**. doi:10.1029/2009jg001248
- 49 Shuman, C. A., E. Berthier, and T. A. Scambos, 2011: 2001–2009 elevation and mass losses in the Larsen A and B
50 embayments, Antarctic Peninsula. *Journal of Glaciology*, **57**, 737-754
- 51 Sicart, J. E., R. Hock, and D. Six, 2008: Glacier melt, air temperature, and energy balance in different climates: The
52 Bolivian Tropics, the French Alps, and northern Sweden. *Journal of Geophysical Research-Atmospheres*, **113**.
53 doi:10.1029/2008jd010406
- 54 Siegfried, M. R., R. L. Hawley, and J. F. Burkhart, 2011: High-Resolution Ground-Based GPS Measurements Show
55 Intercampaign Bias in ICESat Elevation Data Near Summit, Greenland. *Ieee Transactions on Geoscience and*
56 *Remote Sensing*, **49**, 3393-3400. doi:10.1109/tgrs.2011.2127483
- 57 Silverio, W., and J. M. Jaquet, 2005: Glacial cover mapping (1987-1996) of the Cordillera Blanca (Peru) using satellite
58 imagery. *Remote Sensing of Environment*, **95**, 342-350. doi:10.1016/j.rse.2004.12.012
- 59 Silverio, W., and J.-M. Jaquet, 2012: Multi-temporal and multi-source cartography of the glacial cover of Nevado
60 Coropuna (Arequipa, Peru) between 1955 and 2003. *International Journal of Remote Sensing*, **33**.
61 doi:10.1080/01431161.2012.676742
- 62 Simmonds, I., C. Burke, and K. Keay, 2008: Arctic Climate Change as Manifest in Cyclone Behavior. *J.Clim.*, **21**,
63 5777-5796. doi:10.1175/2008jcli2366.1

- 1 Slobbe, D. C., P. Ditmar, and R. C. Lindenberg, 2009: Estimating the rates of mass change, ice volume change and
2 snow volume change in Greenland from ICESat and GRACE data. *Geophysical Journal International*, **176**, 95-
3 106. doi:10.1111/j.1365-246X.2008.03978.x
- 4 Smith, D. M., 1998: Recent increase in the length of the melt season of perennial Arctic sea ice. *Geophys. Res. Lett.*, **25**,
5 655-658. doi:10.1029/98gl00251
- 6 Smith, S. L., S. A. Wolfe, D. W. Riseborough, and F. M. Nixon, 2009: Active-Layer Characteristics and Summer
7 Climatic Indices, Mackenzie Valley, Northwest Territories, Canada. *Permafrost and Periglacial Processes*, **20**,
8 201-220. doi:10.1002/ppp.651
- 9 Smith, S. L., et al., 2010: Thermal State of Permafrost in North America: A Contribution to the International Polar
10 Year. *Permafrost and Periglacial Processes*, **21**, 117-135. doi:10.1002/ppp.690
- 11 Sole, A. J., D. W. F. Mair, P. W. Nienow, I. D. Bartholomew, M. A. King, M. J. Burke, and I. Joughin, 2011: Seasonal
12 speedup of a Greenland marine-terminating outlet glacier forced by surface melt-induced changes in subglacial
13 hydrology. *J. Geophys. Res.-Earth Surf.*, **116**, 11. doi:F03014, 10.1029/2010jg001948
- 14 Solomon, S. M., A. E. Taylor, and C. W. Stevens, 2008: Nearshore Ground Temperatures, Seasonal Ice Bonding and
15 Permafrost Formation Within the Bottom-Fast Ice Zone, Mackenzie Delta, NWT. *Proceedings of the Ninth
16 International Conference of Permafrost*, D. L. Kane, and K. M. Hinkel, Eds., Institute of Northern Engineering,
17 University of Alaska, Fairbanks, 1675-1680.
- 18 Sorensen, L. S., et al., 2011: Mass balance of the Greenland ice sheet (2003-2008) from ICESat data - the impact of
19 interpolation, sampling and firn density. *Cryosphere*, **5**, 173-186. doi:10.5194/tc-5-173-2011
- 20 Sorg, A., T. Bolch, M. Stoffel, O. Solomina, and M. Beniston, 2012: Climate change impacts on glaciers and runoff in
21 Tien Shan (Central Asia). *Nature Climate Change*. doi:doi:10.1038/nclimate1592
- 22 Spreen, G., R. Kwok, and D. Menemenlis, 2011: Trends in Arctic sea ice drift and role of wind forcing: 1992-2009.
23 *Geophys. Res. Lett.*, **38**, 6. doi:L19501, 10.1029/2011gl048970
- 24 Spreen, G., S. Kern, D. Stammer, and E. Hansen, 2009: Fram Strait sea ice volume export estimated between 2003 and
25 2008 from satellite data. *Geophys. Res. Lett.*, **36**. doi:10.1029/2009gl039591
- 26 Stammerjohn, S., R. Massom, D. Rind, and D. Martinson, 2012: Regions of rapid sea ice change: An inter-hemispheric
27 seasonal comparison. *Geophys. Res. Lett.*, **39**. doi:10.1029/2012gl050874
- 28 Stammerjohn, S. E., D. G. Martinson, R. C. Smith, X. Yuan, and D. Rind, 2008: Trends in Antarctic annual sea ice
29 retreat and advance and their relation to El Nino-Southern Oscillation and Southern Annular Mode variability.
30 *Journal of Geophysical Research-Oceans*, **113**. doi:10.1029/2007jc004269
- 31 Steele, M., W. Ermold, and J. L. Zhang, 2008: Arctic Ocean surface warming trends over the past 100 years. *Geophys.
32 Res. Lett.*, **35**. doi:10.1029/2007gl031651
- 33 Straneo, F., R. G. Curry, D. A. Sutherland, G. S. Hamilton, C. Cenedese, K. Vage, and L. A. Stearns, 2011: Impact of
34 fjord dynamics and glacial runoff on the circulation near Helheim Glacier. *Nature Geoscience*, **4**, 322-327.
35 doi:10.1038/ngeo1109
- 36 Straneo, F., et al., 2010: Rapid circulation of warm subtropical waters in a major glacial fjord in East Greenland. *Nature
37 Geoscience*, **3**, 182-186. doi:10.1038/ngeo764
- 38 Straneo, F., et al., Submitted: Characteristics of ocean waters reaching Greenland's glaciers. *Annals of Glaciology*
- 39 Streletskiy, D. A., N. I. Shiklomanov, F. E. Nelson, and A. E. Klene, 2008: 13 Years of Observations at Alaskan CALM
40 Sites: Long-term Active Layer and Ground Surface Temperature Trends. *9th International Conference on
41 Permafrost*, Institute of Northern Engineering, University of Alaska, Fairbanks, 1727-1732.
- 42 Stroeve, J., M. M. Holland, W. Meier, T. Scambos, and M. Serreze, 2007: Arctic sea ice decline: Faster than forecast.
43 *Geophys. Res. Lett.*, **34**. doi:10.1029/2007gl029703
- 44 Sundal, A. V., A. Shepherd, P. Nienow, E. Hanna, S. Palmer, and P. Huybrechts, 2011: Melt-induced speed-up of
45 Greenland ice sheet offset by efficient subglacial drainage. *Nature*, **469**, 522-U583. doi:10.1038/nature09740
- 46 Surazakov, A. B., V. B. Aizen, E. M. Aizen, and S. A. Nikitin, 2007: Glacier changes in the Siberian Altai Mountains,
47 Ob river basin, (1952-2006) estimated with high resolution imagery. *Environmental Research Letters*, **2**, 7.
48 doi:045017, 10.1088/1748-9326/2/4/045017
- 49 Takala, M., J. Pulliainen, S. J. Metsamaki, and J. T. Koskinen, 2009: Detection of Snowmelt Using Spaceborne
50 Microwave Radiometer Data in Eurasia From 1979 to 2007. *Ieee Transactions on Geoscience and Remote
51 Sensing*, **47**, 2996-3007. doi:10.1109/tgrs.2009.2018442
- 52 Tamura, T., and K. I. Ohshima, 2011: Mapping of sea ice production in the Arctic coastal polynyas. *Journal of
53 Geophysical Research-Oceans*, **116**, 20. doi:C07030, 10.1029/2010jc006586
- 54 Tarnocai, C., J. G. Canadell, E. A. G. Schuur, P. Kuhry, G. Mazhitova, and S. Zimov, 2009: Soil organic carbon pools
55 in the northern circumpolar permafrost region. *Glob. Biogeochem. Cycle*, **23**, 11. doi:Gb2023,
56 10.1029/2008gb003327
- 57 Tedesco, M., M. Brodzik, R. Armstrong, M. Savoie, and J. Ramage, 2009: Pan arctic terrestrial snowmelt trends (1979-
58 2008) from spaceborne passive microwave data and correlation with the Arctic Oscillation. *Geophys. Res. Lett.*,
59 **36**. doi:L21402, 10.1029/2009gl039672
- 60 Tedesco, M., et al., 2011: The role of albedo and accumulation in the 2010 melting record in Greenland. *Environmental
61 Research Letters*, **6**, 6. doi:014005, 10.1088/1748-9326/6/1/014005
- 62 Thomas, E. R., G. J. Marshall, and J. R. McConnell, 2008a: A doubling in snow accumulation in the western Antarctic
63 Peninsula since 1850. *Geophys. Res. Lett.*, **35**, 5. doi:L01706, 10.1029/2007gl032529

- 1 Thomas, I. D., et al., 2011: Widespread low rates of Antarctic glacial isostatic adjustment revealed by GPS
2 observations. *Geophys. Res. Lett.*, **38**, doi:10.1029/2011GL049277
- 3 Thomas, R., E. Frederick, W. Krabill, S. Manizade, and C. Martin, 2006: Progressive increase in ice loss from
4 Greenland. *Geophys. Res. Lett.*, **33**, 4. doi:L10503, 10.1029/2006gl026075
- 5 ———, 2009: Recent changes on Greenland outlet glaciers. *Journal of Glaciology*, **55**, 147-162
- 6 Thomas, R., C. Davis, E. Frederick, W. Krabill, Y. H. Li, S. Manizade, and C. Martin, 2008b: A comparison of
7 Greenland ice-sheet volume changes derived from altimetry measurements. *Journal of Glaciology*, **54**, 203-212
- 8 Thomas, R. H., and C. R. Bentley, 1978: A model for Holocene retreat of the West Antarctic Ice Sheet, **10**, 150-170
- 9 Thompson, D., G. Tootle, G. Kerr, R. Sivanpillai, and L. Pochop, 2011a: Glacier Variability in the Wind River Range,
10 Wyoming. *Journal of Hydrologic Engineering*, **16**. doi:10.1061/(asce)he.1943-5584.0000384
- 11 Thompson, D. W. J., and J. M. Wallace, 1998: The Arctic Oscillation signature in the wintertime geopotential height
12 and temperature fields. *Geophys. Res. Lett.*, **25**, 1297-1300. doi:10.1029/98gl00950
- 13 ———, 2000: Annular modes in the extratropical circulation. Part I: Month-to-month variability. *J. Clim.*, **13**, 1000-1016.
14 doi:10.1175/1520-0442(2000)013<1000:amitec>2.0.co;2
- 15 Thompson, D. W. J., Solomon, S., P. J. Kushner, M. H. England, K. M. Grise, and D. J. Karoly, 2011b: Signatures of
16 the Antarctic ozone hole in Southern Hemisphere climate change. *Nature Geoscience*, doi: 10.1038/NNGEO1296
- 17 Thompson, L. G., E. Mosley-Thompson, M. E. Davis, and H. H. Brecher, 2011c: Tropical glaciers, recorders and
18 indicators of climate change, are disappearing globally. *Annals of Glaciology*, **52**, 23-34
- 19 Thomson, L. I., G. R. Osinski, and C. S. L. Ommanney, 2011: Glacier change on Axel Heiberg Island, Nunavut,
20 Canada. *Journal of Glaciology*, **57**
- 21 Thost, D. E., and M. Truffer, 2008: Glacier recession on Heard Island, southern Indian Ocean. *Arct. Antarct. Alp. Res.*,
22 **40**, 199-214. doi:10.1657/1523-0430(06-084)[thost]2.0.co;2
- 23 Turner, J., et al., 2005: Antarctic climate change during the last 50 years. *International Journal of Climatology*, **25**, 279-
24 294
- 25 UNEP, 2007: *Global outlook for snow and ice*. UNEP/GRID.
- 26 van de Wal, R. S. W., W. Boot, M. R. van den Broeke, C. Smeets, C. H. Reijmer, J. J. A. Donker, and J. Oerlemans,
27 2008: Large and rapid melt-induced velocity changes in the ablation zone of the Greenland Ice Sheet. *Science*,
28 **321**, 111-113. doi:10.1126/science.1158540
- 29 van den Broeke, M., W. J. van de Berg, and E. van Meijgaard, 2006: Snowfall in coastal West Antarctica much greater
30 than previously assumed. *Geophys. Res. Lett.*, **33**. doi:L02505, 10.1029/2005gl025239
- 31 van den Broeke, M., et al., 2009: Partitioning Recent Greenland Mass Loss. *Science*, **326**, 984-986.
32 doi:10.1126/science.1178176
- 33 van Everdingen, R., Ed., 1998a: *Multi-language glossary of permafrost and related ground-ice terms*. National Snow
34 and Ice Data Center / World Data Center for Glaciology.
- 35 ———, 1998b: *Multi-language glossary of permafrost and related ground-ice terms*. National Snow and Ice Data Center
36 / World Data Center for Glaciology.
- 37 van Huissteden, J., C. Berrittella, F. J. W. Parmentier, Y. Mi, T. C. Maximov, and A. J. Dolman, 2011: Methane
38 emissions from permafrost thaw lakes limited by lake drainage. *Nature Climate Change*, **1**, 119-123.
39 doi:10.1038/nclimate1101
- 40 Vasiliev, A. A., M. O. Leibman, and N. G. Moskalenko, 2008: Active Layer Monitoring in West Siberia under the
41 CALM II Program. *9th International Conference on Permafrost*, Institute of Northern Engineering, University of
42 Alaska, Fairbanks, 1815-1821.
- 43 Vaughan, D. G., et al., 2003: Recent rapid regional climate warming on the Antarctic Peninsula. *Climatic Change*, **60**,
44 243-274
- 45 Velicogna, I., 2009: Increasing rates of ice mass loss from the Greenland and Antarctic ice sheets revealed by GRACE.
46 *Geophys. Res. Lett.*, **36**. doi:L19503, 10.1029/2009gl040222
- 47 Velicogna, I., and J. Wahr, 2006a: Acceleration of Greenland ice mass loss in spring 2004. *Nature*, **443**, 329-331.
48 doi:10.1038/nature05168
- 49 ———, 2006b: Measurements of time-variable gravity show mass loss in Antarctica. *Science*, **311**, 1754-1756.
50 doi:10.1126/science.1123785
- 51 Vieira, G., et al., 2010: Thermal State of Permafrost and Active-layer Monitoring in the Antarctic: Advances During the
52 International Polar Year 2007-2009. *Permafrost and Periglacial Processes*, **21**, 182-197. doi:10.1002/ppp.685
- 53 Vigdorichik, M. E., 1980: *Arctic Pleistocene history and the development of submarine permafrost*. Westview Press.
- 54 Vincent, C., E. Le Meur, D. Six, P. Possenti, E. Lefebvre, and M. Funk, 2007: Climate warming revealed by englacial
55 temperatures at Col du Dome (4250 m, Mont Blanc area). *Geophys. Res. Lett.*, **34**. doi:10.1029/2007gl029933
- 56 Vinje, T., 2001: Fram strait ice fluxes and atmospheric circulation: 1950-2000. *J. Clim.*, **14**, 3508-3517.
57 doi:10.1175/1520-0442(2001)014<3508:fsifaa>2.0.co;2
- 58 Vuille, M., G. Kaser, and I. Juen, 2008: Glacier mass balance variability in the Cordillera Blanca, Peru and its
59 relationship with climate and the large-scale circulation. *Glob. Planet. Change*, **62**, 14-28.
60 doi:10.1016/j.gloplacha.2007.11.003
- 61 Wadhams, P., 1990: Evidence for thinning of the Arctic ice cover north of Greenland. *Nature*, **345**, 795-797
- 62 Wadhams, P., and N. R. Davis, 2000: Further evidence of ice thinning in the Arctic Ocean. *Geophys. Res. Lett.*, **27**,
63 3973-3975

- 1 Wadhams, P., N. Hughes, and J. Rodrigues, 2011: Arctic sea ice thickness characteristics in winter 2004 and 2007 from
2 submarine sonar transects. *Journal of Geophysical Research-Oceans*, **116**. doi:10.1029/2011jc006982
- 3 Walsh, J. E., and W. L. Chapman, 2001: 20th-century sea-ice variations from observational data. *Annals of Glaciology*,
4 *Vol 33*, **33**, 444-448. doi:10.3189/172756401781818671
- 5 Wang, Y. T., S. G. Hou, and Y. P. Liu, 2009: Glacier changes in the Karlik Shan, eastern Tien Shan, during 1971/72-
6 2001/02. *Annals of Glaciology*, **50**, 39-45
- 7 Weeks, W. F., and A. J. Gow, 1978: PREFERRED CRYSTAL ORIENTATIONS IN FAST ICE ALONG MARGINS
8 OF ARCTIC OCEAN. *Journal of Geophysical Research-Oceans and Atmospheres*, **83**, 5105-5121.
9 doi:10.1029/JC083iC10p05105
- 10 Weertman, J., 1974: Stability of the junction of an ice sheet and an ice shelf. *Journal of Glaciology*, **13**, 3-11
- 11 Weidick, A., C. E. Boggild, and N. T. Knudsen, 1992: Glacier inventory of west Greenland - Digital media. National
12 Snow and Ice Data Center.
- 13 Wendt, J., A. Rivera, A. Wendt, F. Bown, R. Zamora, G. Casassa, and C. Bravo, 2010: Recent ice-surface-elevation
14 changes of Fleming Glacier in response to the removal of the Wordie Ice Shelf, Antarctic Peninsula. *Annals of
15 Glaciology*, **51**, 97-102
- 16 Weyhenmeyer, G. A., D. M. Livingstone, M. Meili, O. Jensen, B. Benson, and J. J. Magnuson, 2011: Large
17 geographical differences in the sensitivity of ice-covered lakes and rivers in the Northern Hemisphere to
18 temperature changes. *Global Change Biology*, **17**, 268-275. doi:10.1111/j.1365-2486.2010.02249.x
- 19 WGMS, 1989: World glacier inventory - Status 1988. IAHS (ICSU) / UNEP / UNESCO, World Glacier Monitoring
20 Service, 458 pp.
- 21 ———, 2008: *Global Glacier Changes: facts and figures*. UNEP.
- 22 ———, 2009: Glacier Mass Balance Bulletin No. 10 (2006-2007). *World Glacier Monitoring Service*. .
- 23 White, D., et al., 2007: The arctic freshwater system: Changes and impacts. *Journal of Geophysical Research-
24 Biogeosciences*, **112**. doi:G04s54, 10.1029/2006jg000353
- 25 White, W. B., and R. G. Peterson, 1996: An Antarctic circumpolar wave in surface pressure, wind, temperature and sea-
26 ice extent. *Nature*, **380**, 699-702. doi:10.1038/380699a0
- 27 Whitehouse, P. L., M. J. Bentley, and A. M. Le Brocq, 2012: Lower Antarctic sea level contribution from GRACE
28 using a new glacial isostatic adjustment model. *Quat. Sci. Rev.*, **32**, 1-24
- 29 Willis, M. J., A. K. Melkonian, M. E. Pritchard, and A. Rivera, 2012: Ice Loss from the Southern Patagonian Icefield.
30 *Geophys. Res. Lett.*, **accepted**
- 31 Wingham, D. J., D. W. Wallis, and A. Shepherd, 2009: Spatial and temporal evolution of Pine Island Glacier thinning,
32 1995-2006. *Geophys. Res. Lett.*, **36**. doi:L17501, 10.1029/2009gl039126
- 33 Wingham, D. J., A. Shepherd, A. Muir, and G. J. Marshall, 2006a: Mass balance of the Antarctic ice sheet.
34 *Philosophical Transactions of the Royal Society a-Mathematical Physical and Engineering Sciences*, **364**, 1627-
35 1635. doi:10.1098/rsta.2006.1792
- 36 Wingham, D. J., A. J. Ridout, R. Scharroo, R. J. Arthern, and C. K. Shum, 1998: Antarctic elevation change from 1992
37 to 1996. *Science*, **282**, 456-458. doi:10.1126/science.282.5388.456
- 38 Wingham, D. J., et al., 2006b: CryoSat: A mission to determine the fluctuations in Earth's land and marine ice fields.
39 *Natural Hazards and Oceanographic Processes from Satellite Data*, **37**, 841-871. doi:10.1016/j.asr.2005.07.027
- 40 Worby, A. P., C. A. Geiger, M. J. Paget, M. L. Van Woert, S. F. Ackley, and T. L. DeLiberty, 2008: Thickness
41 distribution of Antarctic sea ice. *Journal of Geophysical Research-Oceans*, **113**. doi:10.1029/2007jc004254
- 42 Wouters, B., D. Chambers, and E. J. O. Schrama, 2008: GRACE observes small-scale mass loss in Greenland.
43 *Geophys. Res. Lett.*, **35**. doi:L20501, 10.1029/2008gl034816
- 44 Wu, Q., T. Zhang, and Y. Liu, 2012: Thermal state of the active layer and permafrost along the Qinghai-Xizang (Tibet)
45 Railway from 2006 to 2010. *Cryosphere*, **6**, 607-612. doi:10.5194/tc-6-607-2012
- 46 Wu, Q. B., and T. J. Zhang, 2008: Recent permafrost warming on the Qinghai-Tibetan plateau. *Journal of Geophysical
47 Research-Atmospheres*, **113**. doi:D13108, 10.1029/2007jd009539
- 48 ———, 2010: Changes in active layer thickness over the Qinghai-Tibetan Plateau from 1995 to 2007. *Journal of
49 Geophysical Research-Atmospheres*, **115**. doi:D09107, 10.1029/2009jd012974
- 50 Wu, X. P., et al., 2010: Simultaneous estimation of global present-day water transport and glacial isostatic adjustment.
51 *Nature Geoscience*, **3**, 642-646. doi:10.1038/ngeo938
- 52 Xie, H., et al., 2011: Sea-ice thickness distribution of the Bellingshausen Sea from surface measurements and ICESat
53 altimetry. *Deep-Sea Research Part II-Topical Studies in Oceanography*, **58**, 1039-1051.
54 doi:10.1016/j.dsr2.2010.10.038
- 55 Xu, Y., E. Rignot, D. Menemenlis, and M. Koppes, 2012: Numerical experiments on subaqueous melting of Greenland
56 tidewater glaciers in response to ocean warming and enhanced subglacial discharge. *Annals of Glaciology*, **53**.
57 doi:10.3189/2012/AoG60A139
- 58 Yao, T., et al., 2012: Different glacier status with atmospheric circulations in Tibetan Plateau and surroundings. *Nature
59 Climate Change*. doi:doi:10.1038/nclimate1580
- 60 Yde, Y. C., and O. Pasche, 2010: Reconstructing climate change: Not all glaciers suitable. *EOS*, **91**, 189-190
- 61 Ye, Q. H., S. C. Kang, F. Chen, and J. H. Wang, 2006a: Monitoring glacier variations on Geladandong mountain,
62 central Tibetan Plateau, from 1969 to 2002 using remote-sensing and GIS technologies. *Journal of Glaciology*,
63 **52**, 537-545. doi:10.3189/172756506781828359

- 1 Ye, Q. H., T. D. Yao, S. C. Kang, F. Chen, and J. H. Wang, 2006b: Glacier variations in the Naimona'nyi region,
2 western Himalaya, in the last three decades. *Annals of Glaciology*, Vol 43, 2006, **43**, 385-389.
3 doi:10.3189/172756406781812032
- 4 Young, D. A., et al., 2011a: A dynamic early East Antarctic Ice Sheet suggested by ice-covered fjord landscapes.
5 *Nature*, **474**, 72-75. doi:10.1038/nature10114
- 6 Young, N. E., J. P. Briner, Y. Axford, B. Csatho, G. S. Babonis, D. H. Rood, and R. C. Finkel, 2011b: Response of a
7 marine-terminating Greenland outlet glacier to abrupt cooling 8200 and 9300 years ago. *Geophys. Res. Lett.*, **38**.
8 doi:10.1029/2011gl049639
- 9 Zamolodchikov, D., 2008: Recent climate and active layer changes in Northeast Russia: regional output of Circumpolar
10 Active Layer Monitoring (CALM). *9th International Conference on Permafrost*, Institute of Northern
11 Engineering, University of Alaska, Fairbanks, 2021-2027.
- 12 Zdanowicz, C., A. Smetny-Sowa, D. Fisher, N. Schaffer, L. Copland, J. Eley, and F. Dupont, 2012: Summer melt rates
13 on Penny Ice Cap, Baffin Island: Past and recent trends and implications for regional climate. *Journal of*
14 *Geophysical Research*, **117**. doi:10.1029/2011JF002248
- 15 Zeeberg, J., and S. L. Forman, 2001: Changes in glacier extent on north Novaya Zemlya in the twentieth century. *The*
16 *Holocene*, **11**, 161-175
- 17 Zemp, M., H. J. Zumbhul, S. U. Nussbaumer, M. H. Masiokas, L. E. Espizua, and P. Pitte, 2011: Extending glacier
18 monitoring into the Little Ice Age and beyond. 67-69.
- 19 Zhang, M., S. Wang, Z. Li, and F. Wang, 2012: Glacier area shrinkage in China and its climatic background during the
20 past half century. *Journal of Geographical Sciences*, **22**. doi:10.1007/s11442-012-0908-3
- 21 Zhang, T., R. G. Barry, K. Knowles, F. Ling, and R. L. Armstrong, 2003: Distribution of seasonally and perennially
22 frozen ground in the Northern Hemisphere. *Proceedings of the 8th International Conference on Permafrost, 21-*
23 *25 July 2003*, Zurich, Switserland, A.A. Balkema, Lisse, the Netherlands, 1289–1294.
- 24 Zhang, T. J., 2005: Influence of the seasonal snow cover on the ground thermal regime: An overview. *Reviews of*
25 *Geophysics*, **43**. doi:Rg4002 10.1029/2004rg000157
- 26 Zhang, T. J., et al., 2005: Spatial and temporal variability in active layer thickness over the Russian Arctic drainage
27 basin. *Journal of Geophysical Research-Atmospheres*, **110**. doi:D16101 10.1029/2004jd005642
- 28 Zhao, L., Q. B. Wu, S. S. Marchenko, and N. Sharkhuu, 2010: Thermal State of Permafrost and Active Layer in Central
29 Asia during the International Polar Year. *Permafrost and Periglacial Processes*, **21**, 198-207.
30 doi:10.1002/ppp.688
- 31 Zheltykhina, N., 2005: GLIMS Glacier Database - Digital Media. National Snow and Ice Data Center/World Data Center
32 for Glaciology.
- 33 Zhou, C. P., W. B. Yang, L. A. Wu, and S. Y. Liu, 2009: Glacier changes from a new inventory, Nianchu river basin,
34 Tibetan Plateau. *Annals of Glaciology*, **50**, 87-92
- 35 Zhou, Y., D. Guo, G. Qiu, G. Cheng, and S. Li, 2000: *Geocryology in China*. Science Press, 450 pp.
- 36 Zimov, S. A., E. A. G. Schuur, and F. S. Chapin, 2006: Permafrost and the global carbon budget. *Science*, **312**, 1612-
37 1613. doi:10.1126/science.1128908
- 38 Zuo, Z., and J. Oerlemans, 1997: Contribution of glacier melt to sea level rise since AD 1865: a regionally
39 differentiated calculation. *Climate Dynamics*, **13**, 835-845
- 40 Zwally, H. J., and P. Gloersen, 2008: Arctic sea ice surviving the summer melt: interannual variability and decreasing
41 trend. *Journal of Glaciology*, **54**, 279-296. doi:10.3189/002214308784886108
- 42 Zwally, H. J., and M. B. Giovinetto, 2011: Overview and Assessment of Antarctic Ice-Sheet Mass Balance Estimates:
43 1992-2009. *Surveys in Geophysics*, **32**, 351-376. doi:10.1007/s10712-011-9123-5
- 44 Zwally, H. J., D. H. Yi, R. Kwok, and Y. H. Zhao, 2008: ICESat measurements of sea ice freeboard and estimates of
45 sea ice thickness in the Weddell Sea. *Journal of Geophysical Research-Oceans*, **113**. doi:C02s15,
46 10.1029/2007jc004284
- 47 Zwally, H. J., J. C. Comiso, C. L. Parkinson, D. J. Cavalieri, and P. Gloersen, 2002: Variability of Antarctic sea ice
48 1979-1998. *Journal of Geophysical Research-Oceans*, **107**, 21. doi:10.1029/2000jc000733
- 49 Zwally, H. J., et al., 2005: Mass changes of the Greenland and Antarctic ice sheets and shelves and contributions to sea
50 level rise: 1992-2002. *Journal of Glaciology*, **51**, 509-527. doi:10.3189/172756505781829007
- 51 Zwally, H. J., et al., 2011: Greenland ice sheet mass balance: distribution of increased mass loss with climate warming;
52 2003-07 versus 1992-2002. *Journal of Glaciology*, **57**, 88-102.
- 53
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Appendix 4.A: Details of Available and Selected Ice Sheet Mass Balance Estimates 1992 to 2009

All comprehensive mass balance estimates available for Greenland, and the sub-set of those selected for this assessment (Section 4.4.2) are listed in Tables 4.A.1 and 4.A.2. Those available for Antarctica are shown in Tables 4.A.3 and 4.A.4. These studies include estimates made from satellite gravimetry, satellite altimetry and the mass balance method. The studies selected are the latest made by 12 different research groups, for each of Greenland and Antarctica. The tables also note whether smaller glaciers peripheral to the ice sheet are included, or excluded, in the estimate, and explain why some studies were not selected (e.g., earlier estimates from the same researchers have been updated by more recent analyses using extended data).

Table 4.A.1: Sources used for calculation of ice loss from Greenland.

Source	Method	Start	End	Gt yr ⁻¹	Cited Uncertainty	Peripheral Glaciers	Comment
(Wu et al., 2010)	GRACE +GPS	2002.375	2008.958	-104	23	incl.	Global inversion technique affected due to paucity of GPS data around Greenland
(Sorensen et al., 2011)	ICESAT	2003.875	2008.292	-221	38	Not known	Density assumption for snow vs ice is listed
(Sasgen et al., 2012b)	GRACE	2002	2009	-236	4	incl.	
(Schrama and Wouters, 2011)	GRACE	2003.3	2010.2	-201	18	incl.	
(Cazenave et al., 2009)	GRACE	2003	2008	-136	18	incl.	CNES fields are truncated to lower harmonics than other fields
Zwally (Zwally et al., 2011)	ERS1,2 /ICESAT	1992	2002	-7	3	Not known	SRALT not effective in SE Greenland where most losses are located
Zwally (Zwally et al., 2011)	ERS1,2 /ICESAT	2003	2007	-171	4	Not known	Density assumption for snow vs ice is not clear
(Velicogna, 2009)	GRACE	2002	2009	-224	33	incl.	Yearly estimates given and used in compilation
(Pritchard et al., 2010)	GRACE	2003.9	2009.875	-195	22	incl.	
(Baur et al., 2009)	GRACE	2002.9	2008.583	-177	12	incl.	
(Slobbe et al., 2009)	GRACE	2002.5	2007.5	-214	78	incl.	
(Slobbe et al., 2009)	ICESAT	2003.2	2007.417	-139	68	Not known	Density assumption for snow vs ice
(Rignot et al., 2011c)	Flux	1992.00	2009.92	-154	51	excl.	Yearly estimates given and used in compilation
(Chen et al., 2011)	GRACE	2002.3	2005.25	-157	38	incl.	
(Chen et al., 2011)	GRACE	2005.3	2009.917	-248	38	incl.	
(Barletta et al., 2012)	GRACE	2003.0	2011.9	-234	20	incl.	
(Harig and Simons, Submitted)	GRACE	2003.0	2011.0	-199	6	incl.	PGR uncertainty is not included in error estimate

Table 4.A.2: Sources NOT used for calculation of ice loss from Greenland.

Source	Method	Start	End	Gt yr ⁻¹	Cited Uncertainty	Peripheral Glaciers	Comment
(Krabill et al., 1999)	Airborne	1993	1999	-47		excl.	Includes only half the ice sheet and fills in the rest with melt model
(Rignot and Kanagaratnam, 2006)	Flux	1996	1996	-83	28	excl.	Superseded (Rignot et al., 2011c)

(Rignot and Kanagaratnam, 2006)	Flux	2000	2000	-127	28	excl.	Superseded (Rignot et al., 2011c)
(Rignot and Kanagaratnam, 2006)	Flux	2002	2005	-75	38	excl.	Superseded (Rignot et al., 2011c)
(Rignot et al., 2008a)	Flux	1996	1996	-97	47	excl.	Superseded (Rignot et al., 2011c)
(Rignot et al., 2008a)	Flux	2000	2000	-156	44	excl.	Superseded (Rignot et al., 2011c)
(Rignot et al., 2008a)	Flux	2004	2007	-264	39	excl.	Superseded (Rignot et al., 2011c)
(Thomas et al., 2006)	Altimetry	1994	1999	-27	23	excl.	Includes only half the ice sheet and fills in the rest with melt model
(Zwally et al., 2005)	ERS1,2	1992.375	2002.875	-12	3	???	Superseded (Zwally et al., 2011)
(Thomas et al., 2006)	Altimetry	1999	2004	-81	24	excl.	Includes only half the ice sheet and fills in the rest with melt model
(Velicogna and Wahr, 2006a)	GRACE	2002	2004	-75	21	incl.	Superseded (Velicogna, 2009)
(Ramillien et al., 2006)	GRACE	2002	2005	-169	66	incl.	Superseded (Cazenave et al., 2009)
(Chen et al., 2006)	GRACE	2002.375	2005	-219	21	incl.	Superseded (Chen et al., 2011)
(Luthcke et al., 2006)	GRACE	2003	2005	-101	16	incl.	Superseded (Pritchard et al., 2010)
(Wouters et al., 2008)	GRACE	2003.2	2008.1	-179	25	incl.	Superseded (Schrama and Wouters, 2011)
(van den Broeke, 2009)	Flux	2003	2008	-237	20	excl.	Superseded (Rignot et al., 2011c)

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3**Table 4.A.3:** Sources used for calculation of ice loss from Antarctica.

Source	Method	Start	End	Gt yr ⁻¹	Cited Uncertainty	Reliability	Comment
(Wu et al., 2010)	GRACE +GPS	2002.375	2008.958	-87	43	incl.	Paucity of GPS data around Antarctica to constrain the inversion
(Wingham et al., 2006a)	ERS-1/2	1992.83	2003.17	27	69	???	No data in Peninsula (uncertainty increased to compensate); series truncated within 100 km of coast
(Velicogna, 2009)	GRACE	2002	2009	-144	73	incl.	Yearly estimates given and used in compilation
(Chen et al., 2009)	GRACE	2002.33	2006	-144	58	incl.	
(Chen et al., 2009)	GRACE	2006	2009	-220	89	incl.	
(Rignot et al., 2011c)	Flux	1992.00	2009.92	-83	91	excl.	Yearly estimates given and used in compilation
(Horwath and Dietrich, 2009)	GRACE	2002.33	2008	-109	48	incl.	Not clear why value is so much lower than other estimates with same data
(Moore and King, 2008)	GRACE	2002.33	2006	-164	80	incl.	
(Cazenave et al., 2009)	GRACE	2003	2008	-198	22	incl.	Fields truncated to lower number of harmonics than

							other estimates
(Dong-Chen et al., 2009)	GRACE	2002.583	2007.75	-78	37	incl.	Methodology unclear and incompletely described; quantification of errors not explained
(Shi et al., 2011)	ICESAT	2003.167	2008.25	-78	5	Not known	Methodology unclear and incompletely described; quantification of errors not explained
(Zwally et al., 2005)	Altimetry	1992.29	2001.29	-30	52	Excluded	No data in Peninsula (uncertainty increased to compensate); firm compaction not validated, unclear performance at coast
(Ivins et al., 2005)	GRACE +GPS	2003	2009.25	-42	9	Not known	Uses Antarctic Peninsula from GRACE, INSAR, GPS to correct altimetry estimates; GPS network and data quality insures a higher quality of this local solution versus global solutions of Wu et al., 2010
(King et al., Submitted)	GRACE	2002.6	2011.0	-71	12	incl.	
(Barletta et al., 2012)	GRACE	2003.0	2011.9	-61	34	incl.	

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3**Table 4.A.4:** Sources NOT used for calculation of ice loss from Antarctica.

Source	Method	Start	End	Gt yr ⁻¹	Cited Uncertainty	Reliability	Comment
(Rignot et al., 2008b)	Flux	1996	1996	-112	91	excl.	Superseded (Rignot et al., 2011c)
(Rignot et al., 2008b)	Flux	2006	2006	-196	92	excl.	Superseded (Rignot et al., 2011c)
(Wingham et al., 1998)	Altimetry	1992.29	1997	-60	76	Not known	Superseded (Wingham et al., 2006a)
(Rignot and Thomas, 2002)	Flux	1995	2000	-26	37	excl.	Not an ice sheet wide estimate
(Davis et al., 2005)	Altimetry	1992	2002	43	23	Not known	Superseded (Helsen et al., 2008)
(Velicogna and Wahr, 2006b)	GRACE	2002	2005	-139	73	incl.	Superseded (Velicogna, 2009)
(Ramillien et al., 2006)	GRACE	2002	2005	-40	32	incl.	Superseded (Cazenave et al., 2009)
(Jia et al., 2009)	GRACE	?	?	-82	29	incl.	Methodology to obtain this number is unclear
(Gunter et al., 2009)	Altimetry	2003.17	2007.17	-100	?	Not known	No error bar and no final estimate

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Chapter 4: Observations: Cryosphere

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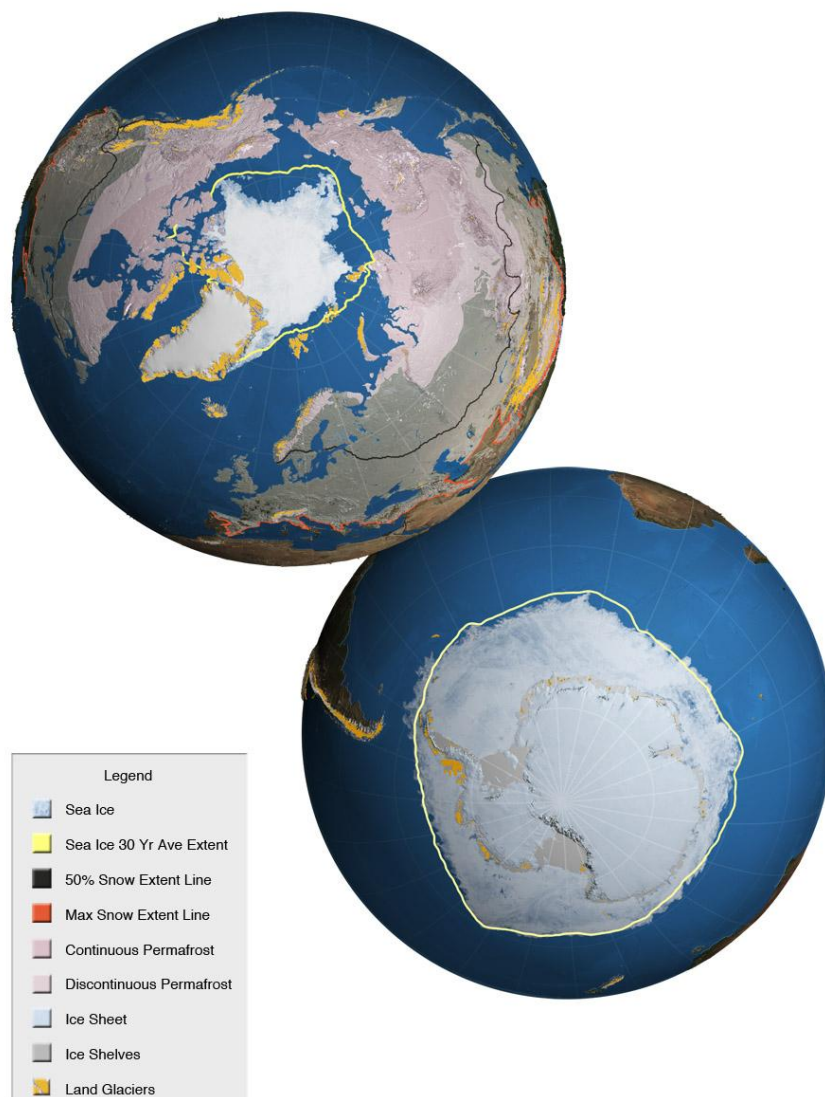
Review Editors: Jonathan Bamber (UK), Philippe Huybrechts (Belgium), Peter Lemke (Germany)

Date of Draft: 5 October 2012

Notes: TSU Compiled Version

1 **Figures**

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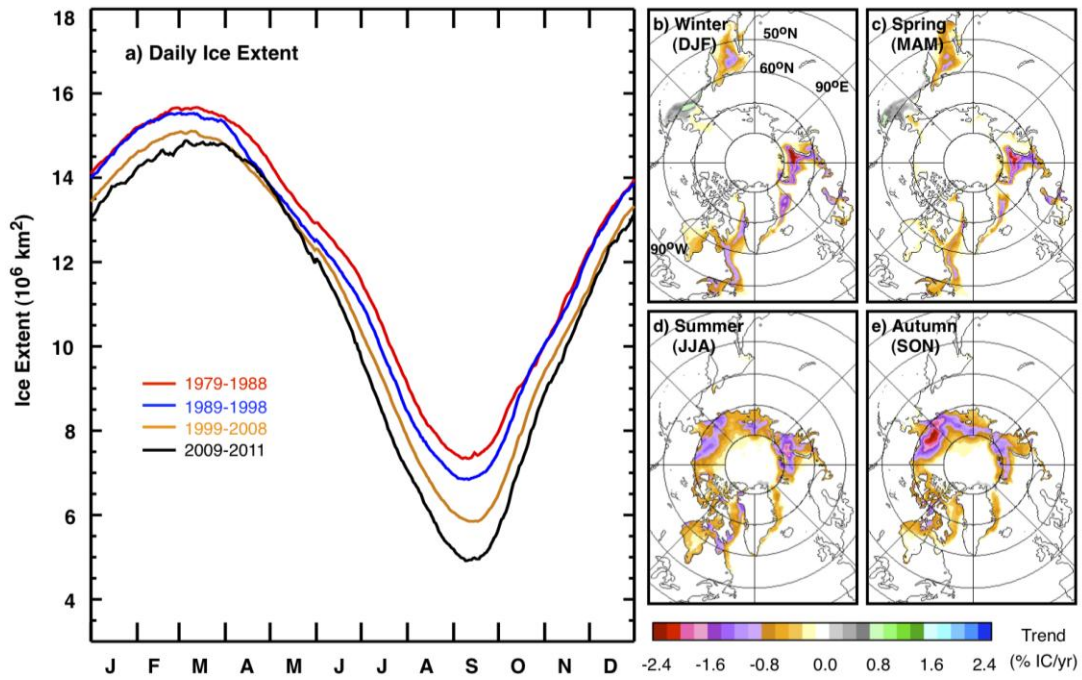
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5 **Figure 4.1:** The cryosphere in the Northern and Southern Hemispheres in polar projection. The map of the Northern
6 Hemisphere shows the sea ice cover during minimum extent (9th September 2011). The yellow line is the average
7 location of the ice edge (15% ice concentration) for the yearly minima for 1979 to 2011. Areas of continuous
8 permafrost are shown in darker pink, discontinuous permafrost in lighter pink. The shaded area over land and
9 permafrost shows snow cover as derived from MODIS data (July 2009 to March 2010) with the greatest extent during
10 that period represented by the black line. The Greenland ice sheet (blue/grey) and locations of glaciers (yellow) are also
11 shown, but the glaciers within the ice sheet are shown as part of the ice sheet. The map of the Southern Hemisphere
12 shows approximately the maximum sea ice cover during an austral winter (9th September 2011). The yellow line shows
13 the average ice edge (15% ice concentration) during maximum extents of the sea ice cover for 1979 to 2011. Some of
14 the elements (e.g., some glaciers and snow) located at low latitudes are not visible in this projection (see Figure 4.8).
15 The source of the data for sea ice, permafrost, snow and ice sheet are datasets held at the National Snow and Ice Data
16 Center (NSIDC), University of Colorado, on behalf of the North American Atlas, Instituto Nacional de Estadística,
17 Geografía e Informática (Mexico), Natural Resources Canada, U.S. Geological Survey, Government of Canada, Canada
18 Centre for Remote Sensing and The Atlas of Canada. Glacier outlines were derived from multiple datasets (Weidick et
19 al., 1992; Zheltyhina, 2005). Figure courtesy of the NASA Visualization Group.

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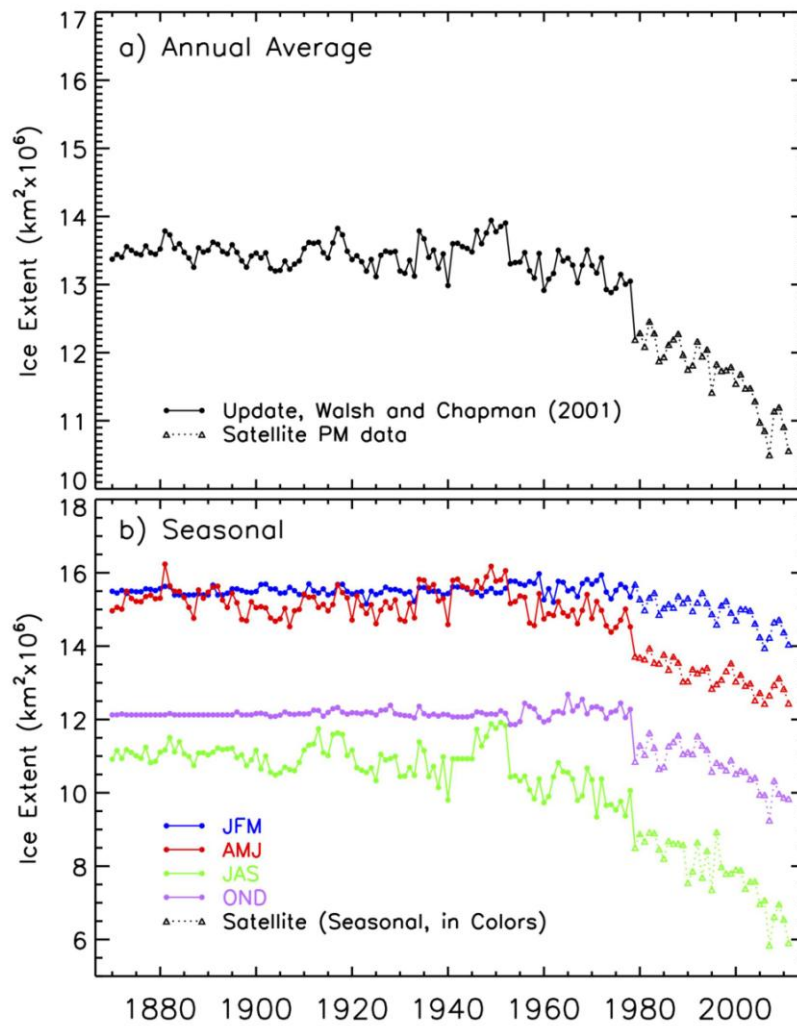
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Figure 4.2: (a) Plots of decadal averages of daily sea ice extent in the Arctic (1979 to 1988 in red, 1989 to 1998 in blue, 1999 to 2008 in gold) and a five-year average daily ice extent from 2009 to 2011; ice concentration trends (1979–2011) in (b) winter, (c) spring, (d) summer and (e) autumn (Comiso and Nishio, 2008).

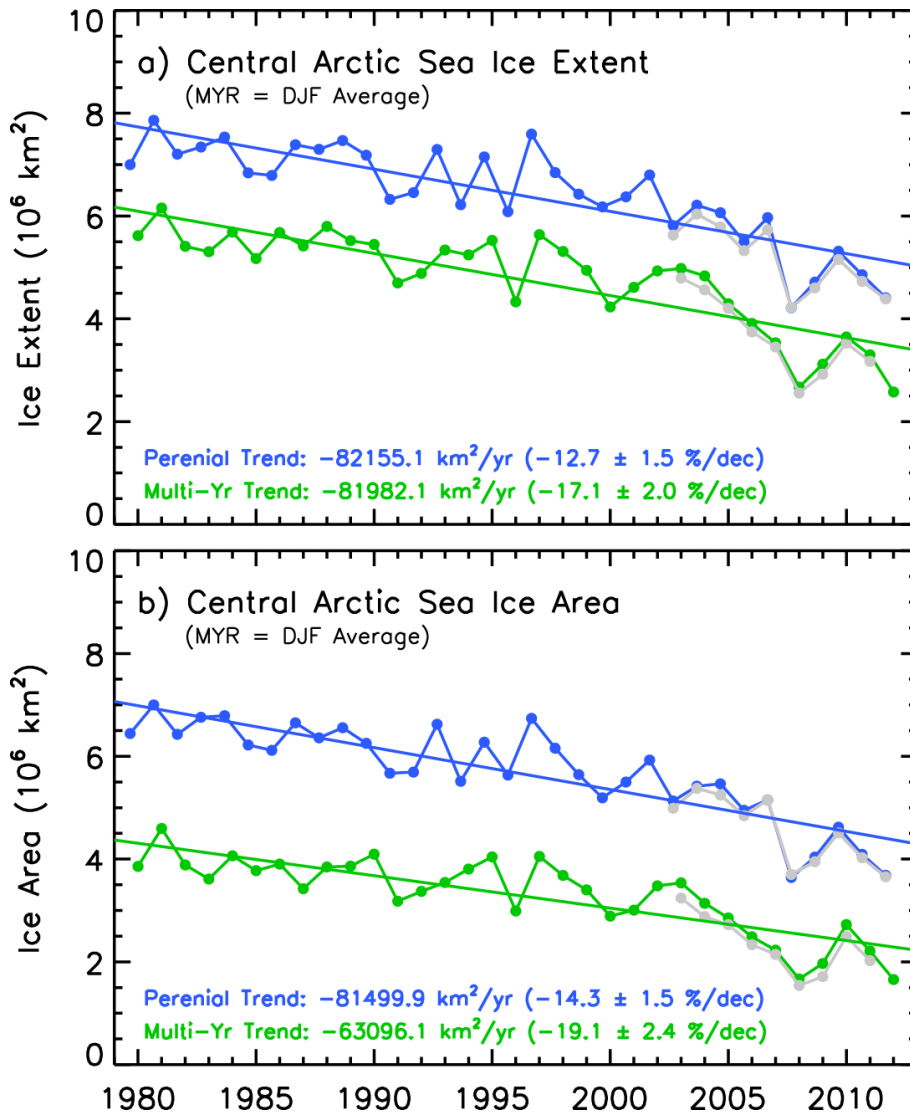
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Figure 4.3: (a) Yearly and (b) seasonal ice extent in the Arctic using averages of mid-month values derived from in situ and satellite data from 1870 to 2011 (updated from, Walsh and Chapman, 2001). The yearly and seasonal averages for the period from 1979 to 2011 are from satellite passive microwave data (triangle fonts) which have also been updated.

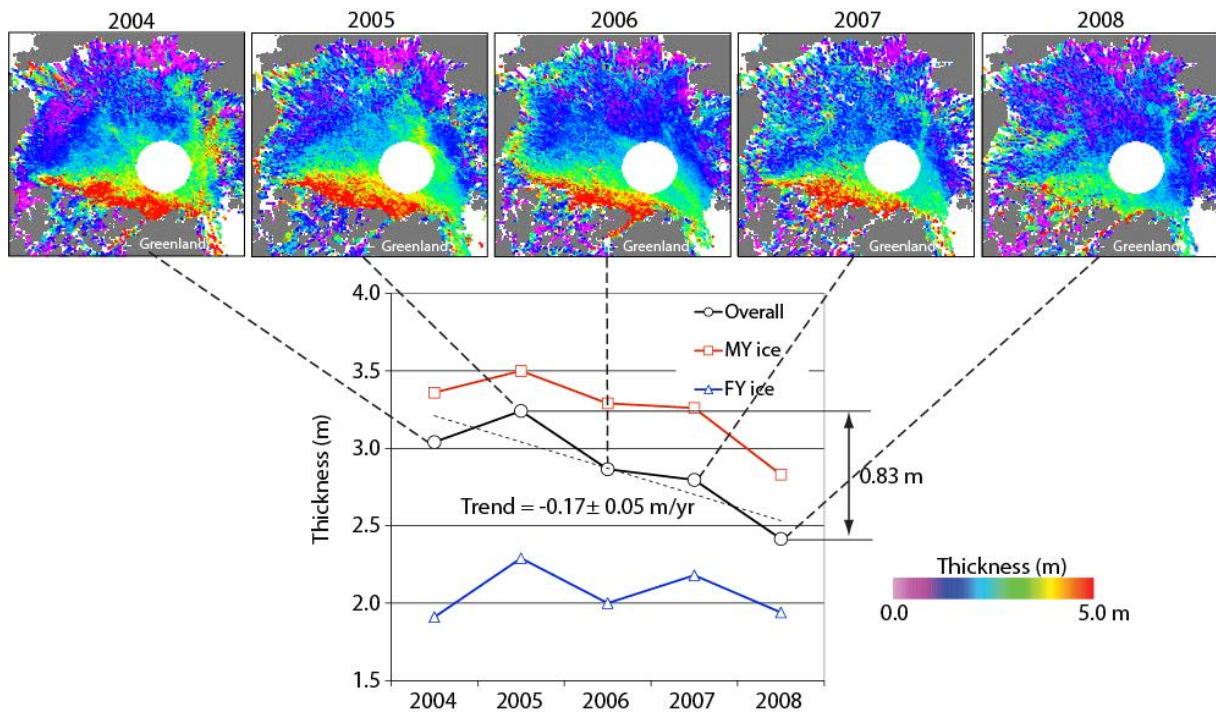
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Figure 4.4: Yearly perennial (blue) and multiyear (green) ice extent (a) and ice area (b) in the Central Arctic for each year from 1979 to 2011 as derived from satellite passive microwave data. Perennial ice values are derived from summer ice minimum values, while the multiyear ice values are averages of December, January and February data. The gray lines (after 2002) are derived from AMSR-E data (Comiso, 2011b).

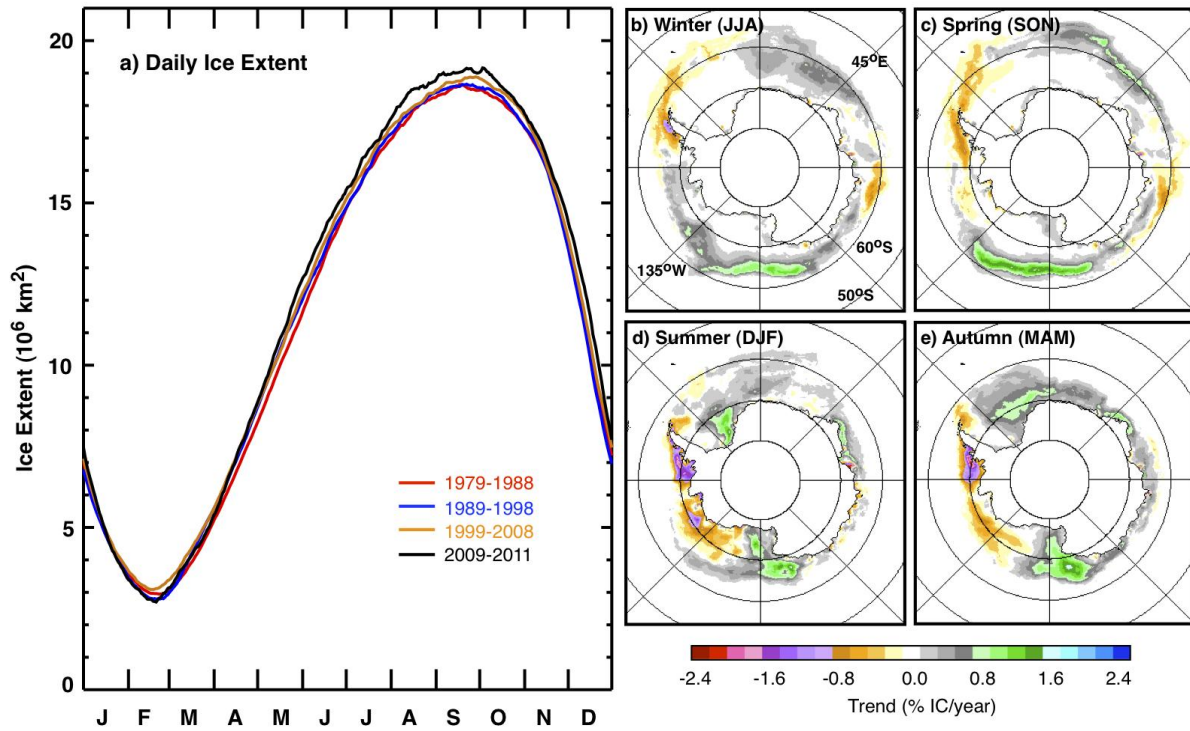
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Figure 4.5: The distribution of winter sea ice thickness in the Arctic and the trends in average, first-year (FY) ice, and multiyear (MY) ice thickness derived from ICESat records, 2004–2008 (Kwok, 2009).

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Figure 4.6: (a) Plots of decadal averages of daily sea ice extent in the Antarctic (1979 to 1988 in red, 1989 to 1998 in blue, 1999 to 2008 in gold) and average values of daily ice extents in 2009 to 2011; ice concentration trends (1979–2011) in (b) winter, (c) spring, (d) summer and (e) autumn.

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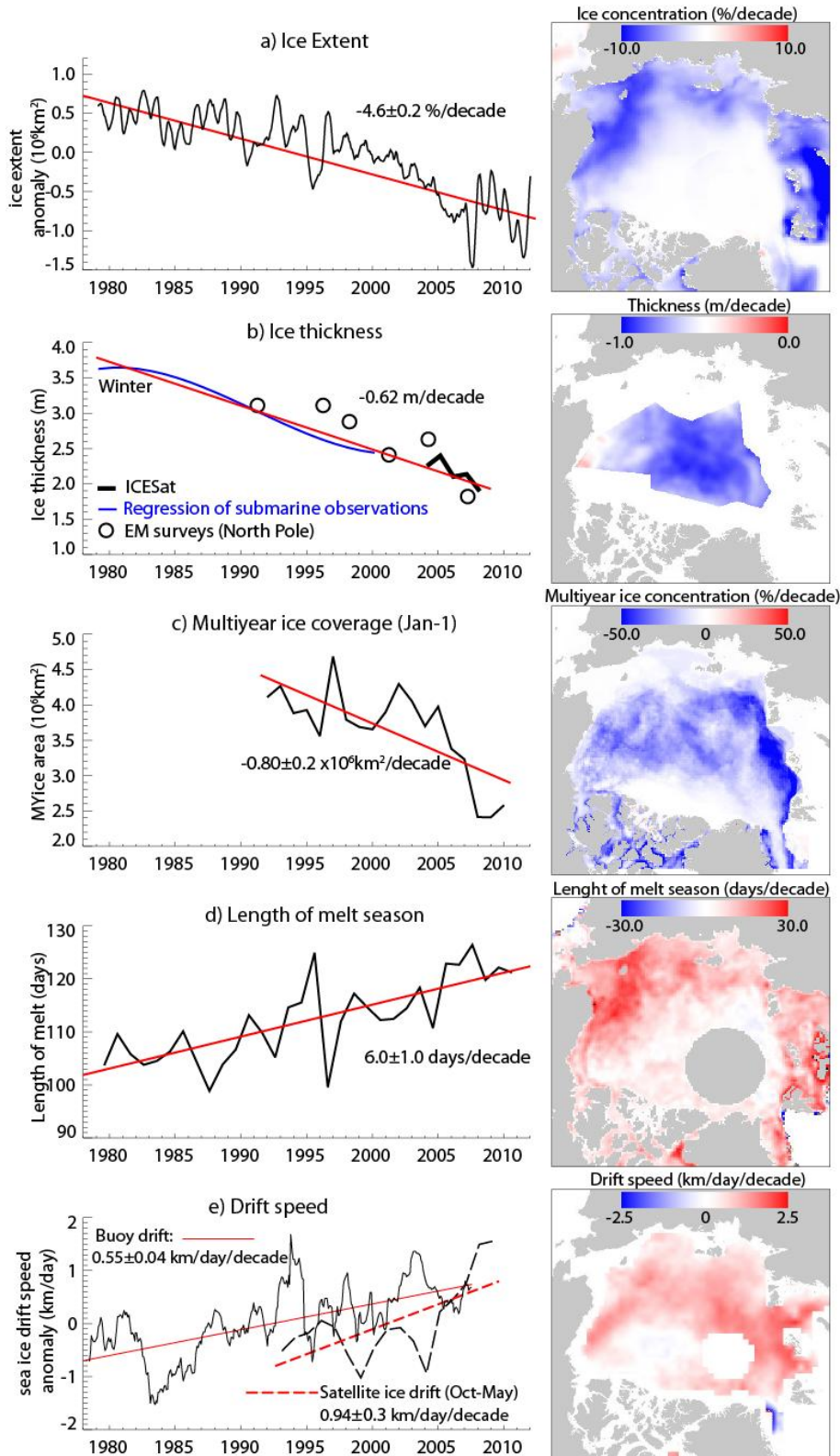
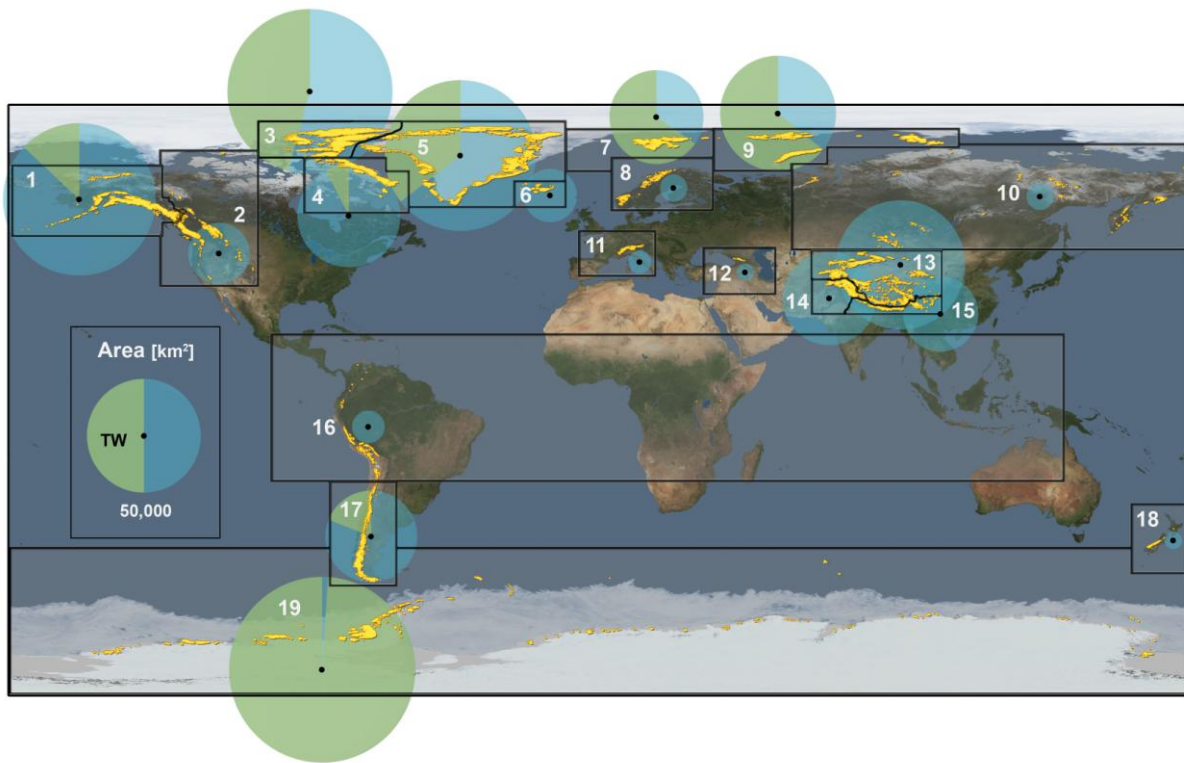


Figure 4.7: Summary of linear decadal trends (red) in: (a) Arctic ice extent and concentration from satellite passive microwave observations (Comiso and Nishio, 2008) ; (b) sea ice thickness from submarine (blue), satellites (black), and in-situ/EM surveys (circles); trend in submarine ice thickness is from multiple regression of available observations to separate the interannual changes, the annual cycle, and the spatial field of thickness within the data release area (Haas et al., 2008; Kwok and Rothrock, 2009); (c) multiyear sea ice coverage from analysis of QuikSCAT (Kwok, 2009); (d) length of melt season (Markus et al., 2009); and (e) satellite-derived drift speed (Spreen et al., 2011).

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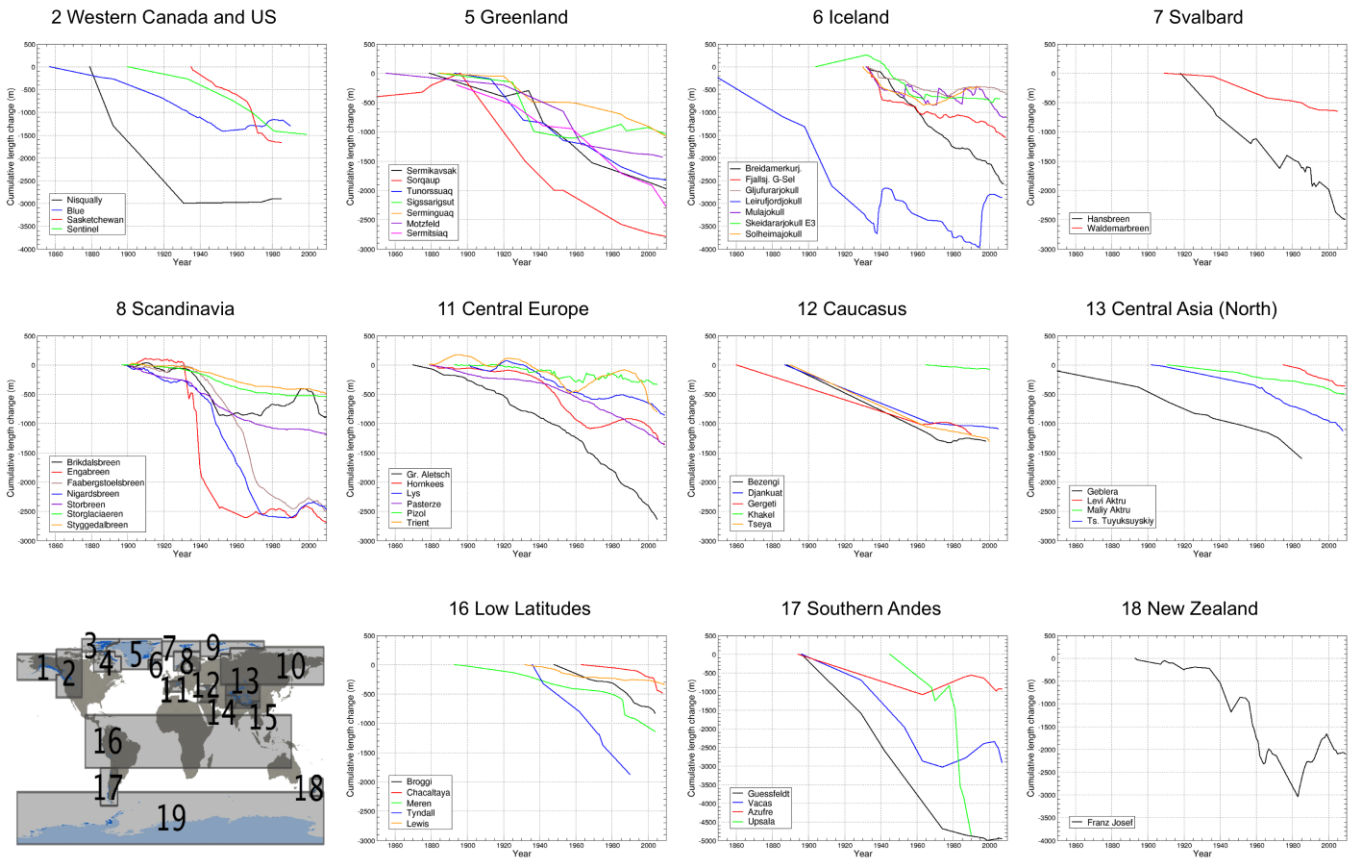
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Figure 4.8: Global distribution of glaciers (yellow) and area covered (size of the circle), sub-divided into the 19 RGI regions (white number) referenced in Table 4.3. The area percentage covered by tidewater glaciers in each region is shown in green. Data from Arendt et al. (2012) and Gardner et al. (Submitted).

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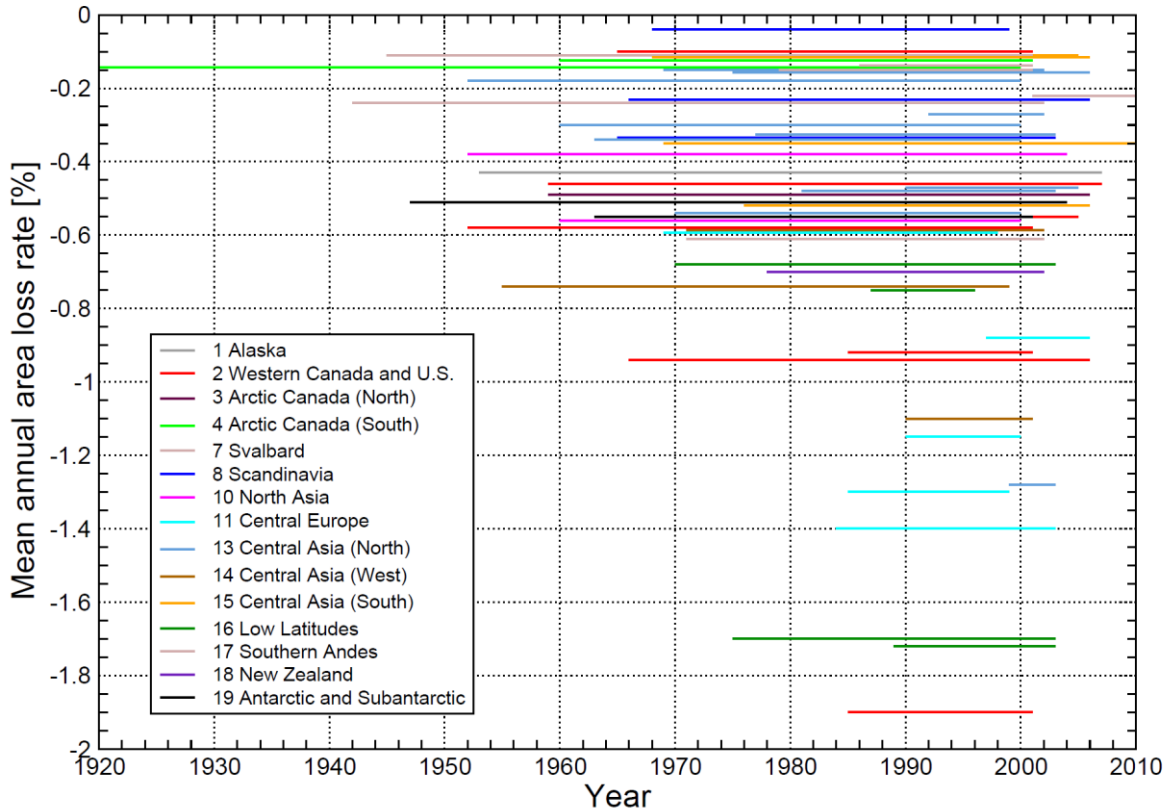
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4 **Figure 4.9:** Selection of cumulative glacier length changes as compiled from long-term in situ measurements in 11
 5 regions. Two of the glaciers in Iceland (Leirufjordjökull and Mulajökull) show surge-type behaviour. Data are from
 6 WGMS (2008) and Leclercq et al. (Submitted) for region 5 (Greenland).

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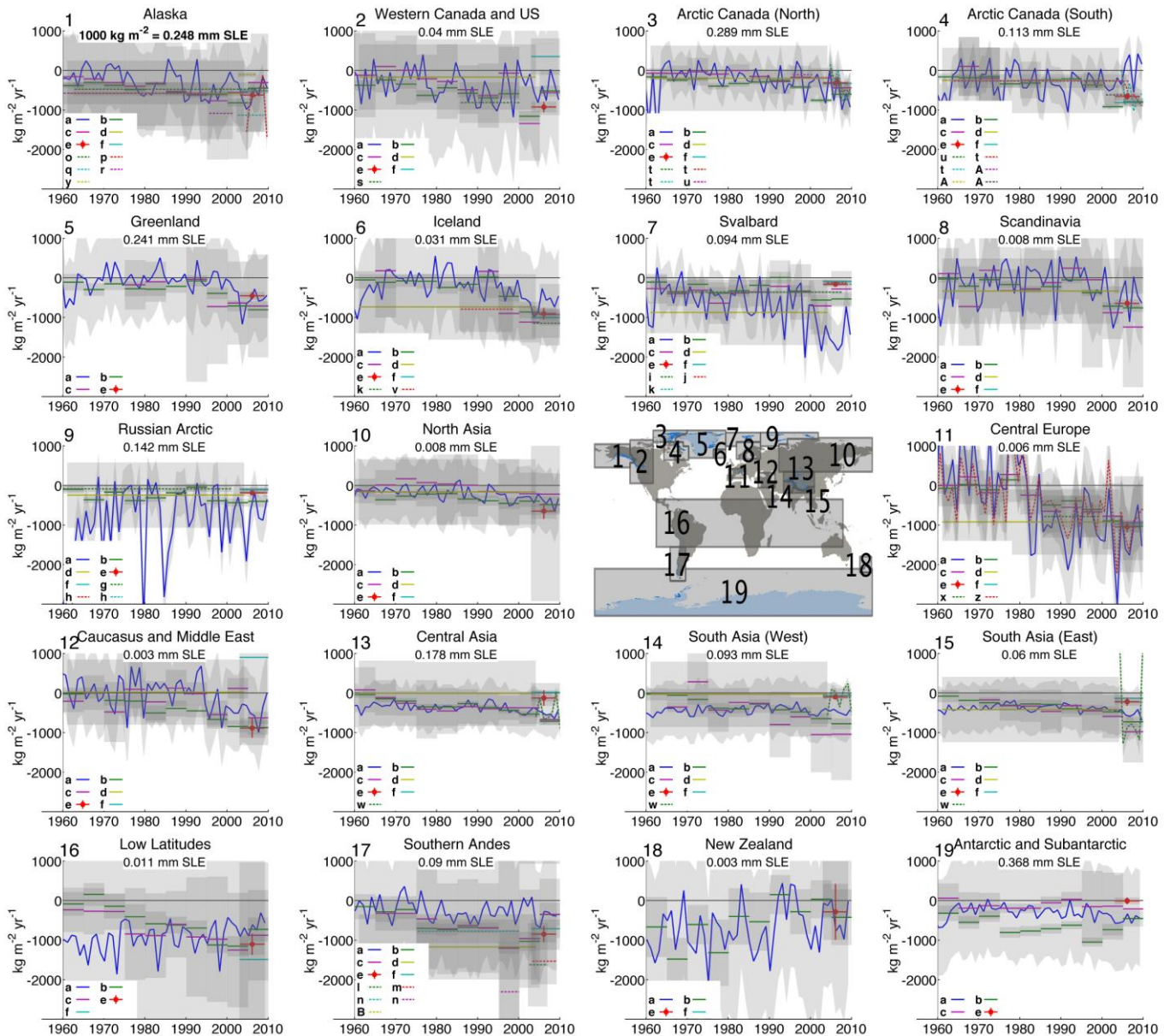
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Figure 4.10: Mean annual area-loss rates for 14 out of the 19 regions depicted in Figure 4.8. Each line refers to the observed relative area loss from a specific publication and its length is related to the period used for averaging. The publications considered for each subregion (in brackets) are: (1) (Le Bris et al., 2011); (2) (Barrand and Sharp, 2010; Bolch et al., 2010; Debeer and Sharp, 2007; Jiskoot et al., 2009; Thompson et al., 2011a) (3) (Huss et al., 2008); (4) (Dowdeswell et al., 2007; Paul and Kaab, 2005); (7) (Kaab, 2008); (8) (Andreassen et al., 2008; Paul and Andreassen, 2009; Paul et al., 2011a); (10) (Shahgedanova et al., 2010; Shahgedanova et al., 2012); (11) (Abermann et al., 2009; Lambrecht and Kuhn, 2007; Paul et al., 2011b; Paul et al., 2004); (13) (Aizen et al., 2007; Bolch et al., 2010; Cia et al., 2005); (Kutuzov and Shahgedanova, 2009; Li et al., 2006; Narama et al., 2006; Narozhniy and Zemtsov, 2011; Surazakov et al., 2007; Wang et al., 2009; Ye et al., 2006a; Ye et al., 2006b; Zhang et al., 2012; Zhou et al., 2009); (14) (Bolch, 2007; Khromova et al., 2006; Narama et al., 2006); (15) (Bhambri et al., 2011; Bolch et al., 2008; Kulkarni et al., 2007; Nie et al., 2010; Schmidt and Nusser, 2012); (16) (Cullen et al., 2006; Klein and Kincaid, 2006; Peduzzi et al., 2010; Racoviteanu et al., 2008; Silverio and Jaquet, 2005; Silverio and Jaquet, 2012); (17) (Davies and Glasser, Submitted; Rivera et al., 2005; Rivera et al., 2007; Schneider et al., 2007); (18) (Gjermundsen et al., 2011); (19) (Berthier et al., 2009; Thost and Truffer, 2008). Data compilation by M. Mahrer, University of Zurich.

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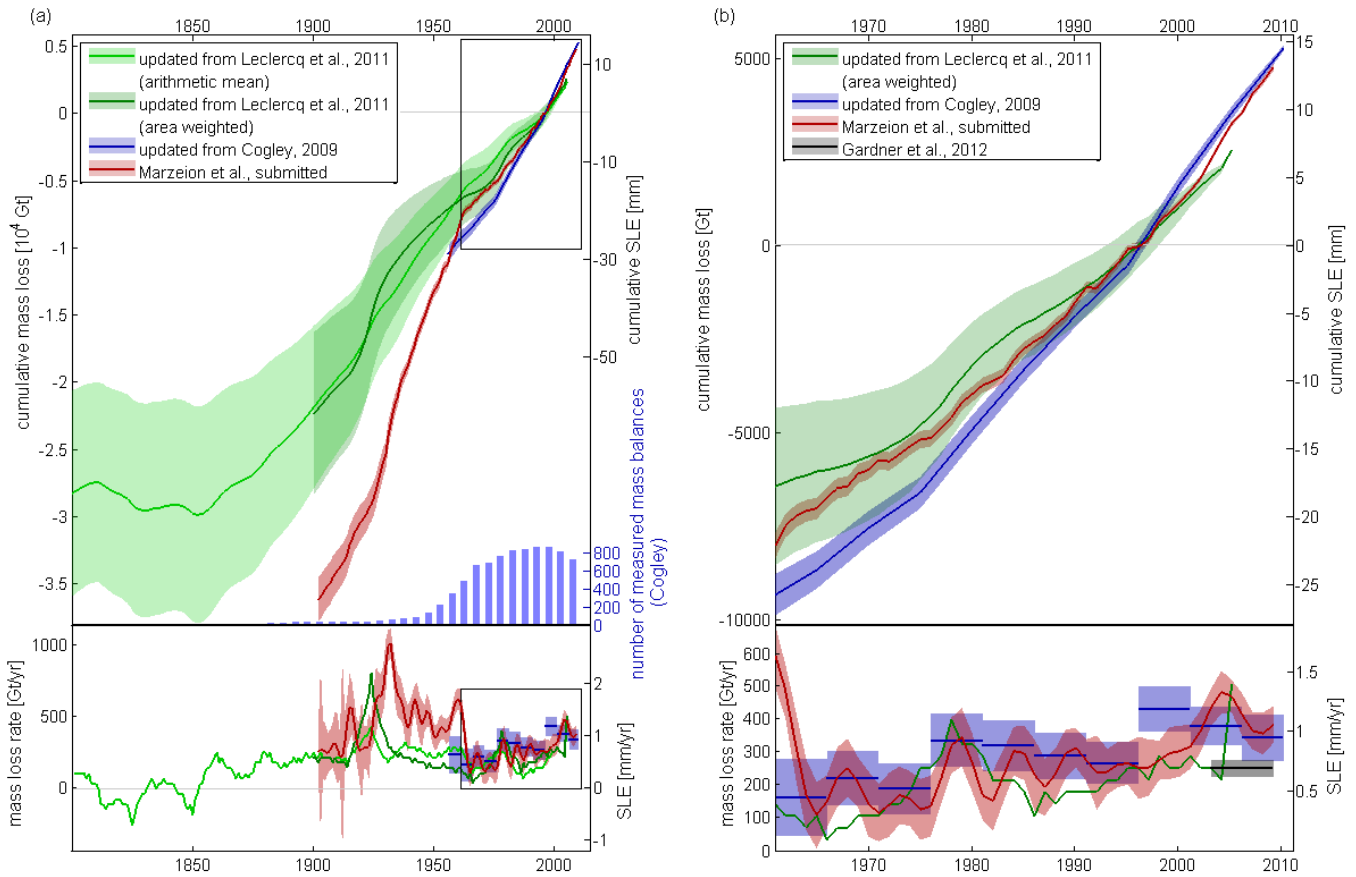
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Figure 4.11: Glacier mass change rates in $[\text{kg m}^{-2} \text{yr}^{-1}]$ for the 19 regions from Figure 4.8 and Table 4.2. Regional values are either from airborne and/or satellite repeat topographic mapping (**u**: (Abdalati et al., 2004); **r**: (Arendt et al., 2002); **o**: (Berthier et al., 2010); **v**: (Magnusson et al., 2005); **h**: (Moholdt et al., 2012); **j**: (Moholdt et al., 2010); **i**: (Nuth et al., 2010); **x**: (Paul and Haeberli, 2008); **n**: (Rignot et al., 2003); **s**: Schiefer (Schiefer et al., 2007); **B**: (Willis et al., 2012), from repeated gravity field measurements (GRACE) (**l**: (Chen et al., 2007); **m**: (Ivins et al., 2011); **f**: (Jacob et al., 2012); **p**: (Luthcke et al., 2008); **w**: (Matsuo and Heki, 2010); **y**: (Peltier, 2009); **k**: (Schrama and Wouters, 2011); **q**: (Wu et al., 2010), volume area scaling (**g**: (Glazovsky and Macheret, 2006)); or from extrapolation from single glacier measurements (**b** [spatial interpolation] and **c** [arithmetic mean]: (Cogley, 2009b); (Huss, 2012)), and modelling with atmospheric input variables (**a**: (Marzeion et al., submitted), **d**: (Hock et al., 2009) . In addition multiple methods are used by: (**A**: (Gardner et al., 2011), **t**: (Gardner et al., Submitted) and **e**: (Gardner et al., Submitted)). Incomplete regional measurements are up-scaled. Uncertainties bounds show the 90% confidence envelope. Uncertainties not provided by the authors, are assigned a random error of 500 kg m^{-2} for non-elevation difference studies and as a cumulative error of 5 m for elevation change studies. Conversions from $(\text{kg m}^{-2} \text{yr}^{-1})$ to mm SLE are given for each region below the region names. GRACE estimates of glacier mass change are often not accompanied by estimates of glaciers area that are required for conversion from Gt yr^{-1} to $\text{kg m}^{-2} \text{yr}^{-1}$. In such cases Gt yr^{-1} values were divided by the total RGI regional glacier area with the exception of **l** and **m** that were divided by the area of the Patagonian Icefields.

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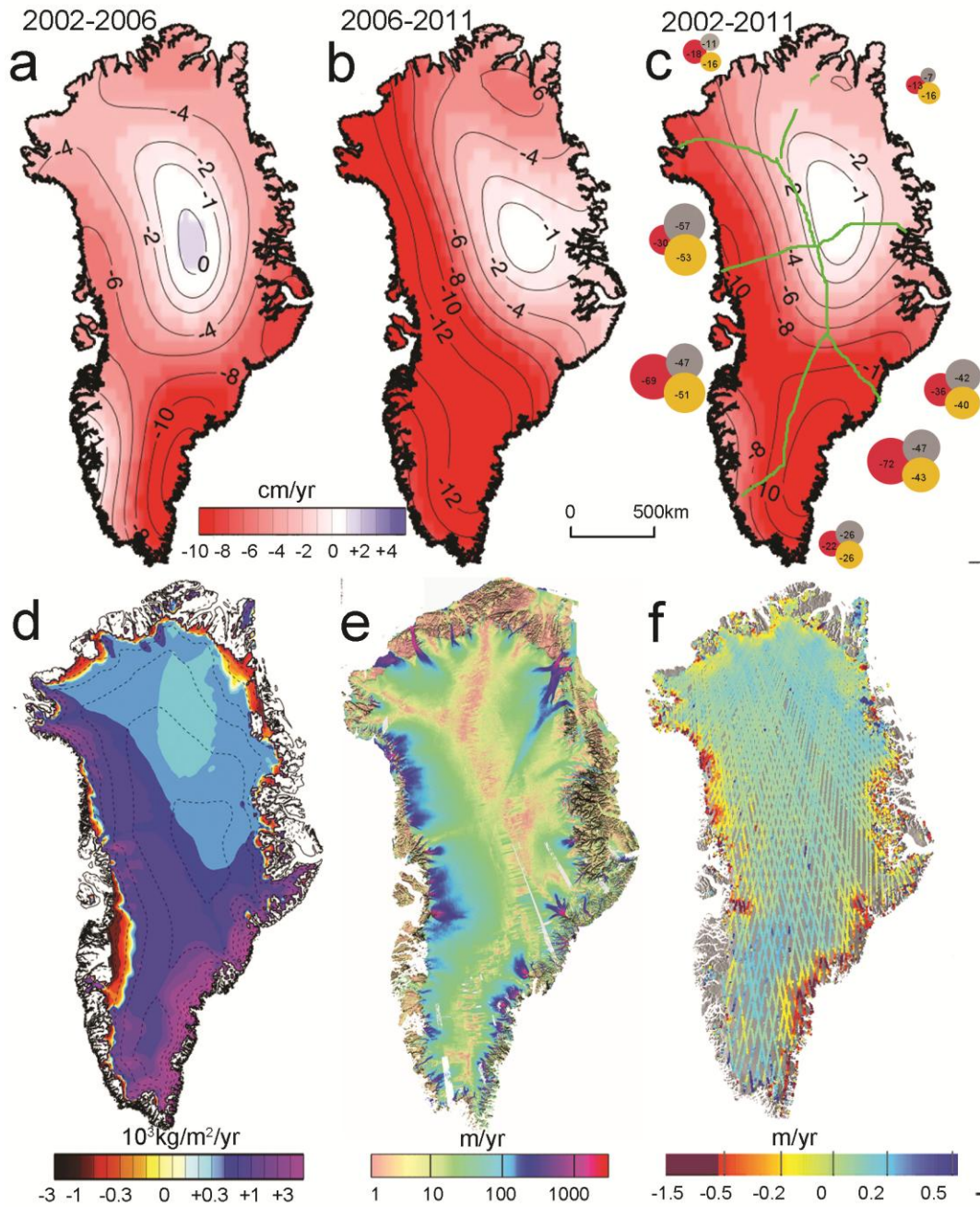
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4 **Figure 4.12:** Global cumulative (top graphs) and annual (lower graphs) glacier mass change 1800–2010 and 1960–2010
 5 in panel (a) and (b) respectively. The 1986–2005 averages of the different cumulative estimates are all set to zero mm
 6 SLE. Estimates are from glacier length variations (updated from, Leclercq et al., 2011), from area-weighted
 7 extrapolations of individual directly and geodetically measured glacier mass budgets (updated from, Cogley, 2009b),
 8 and from modelling with atmospheric variables as input (Marzeion et al., submitted). Uncertainties are based on
 9 comprehensive error analyses in Cogley (2009b) and Marzeion et al. (submitted) and on assumptions about the
 10 representativeness of the sampled glaciers in Leclercq et al. (2011). The latter uncertainties are not shown in the lower
 11 panels for better clarity. The blue bars show the number of measured single glacier mass balances used in the updated
 12 Cogley (2009b) time series. The mean 2003–2009 estimate by Gardner et al. (Submitted) is added to the bottom panel
 13 b. Figure drawn by U. Blumthaler, Institute of Meteorology and Geophysics, University of Innsbruck

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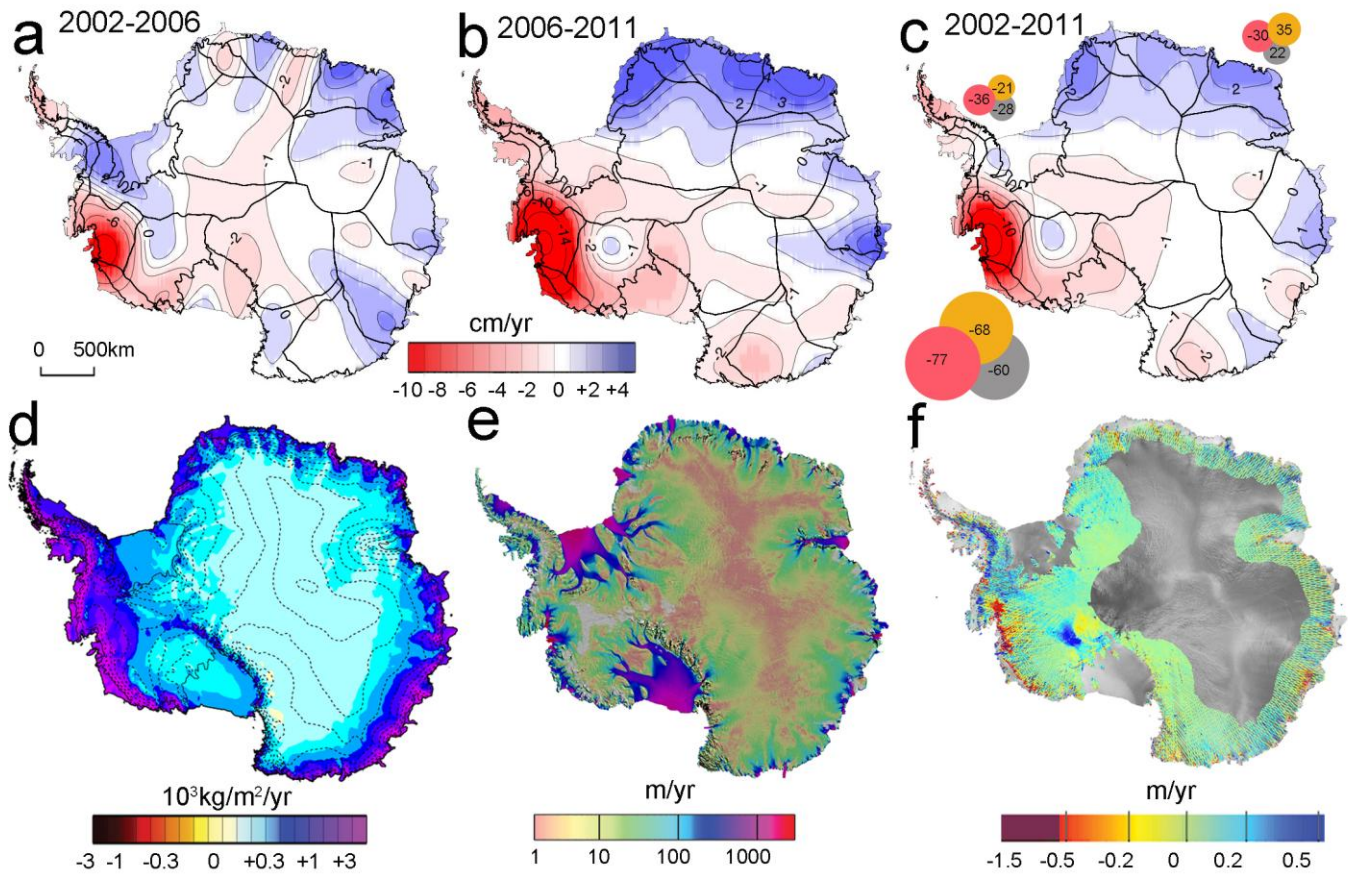
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Figure 4.13: Temporal pattern of ice loss in Greenland from (a-c) GRACE time-variable gravity in centimetres of water per year for the periods (a) 2002 to 2006, (b) 2006 to 2011 and (c) 2002 to 2011, color coded red (loss) to blue (gain) (updated from, Velicogna, 2009). Circles in c) indicate average ice loss (Gt/yr) from the mass budget (red), GRACE (orange) and ICESat (blue) (Sasgen et al., 2012a; Sasgen et al., 2012b)); (d) mean surface mass balance for years 1957–2009 from regional atmospheric climate modelling (Ettema et al., 2009); (e) ice velocity from satellite radar interferometry data for years 2007–2009 showing fastest flow in red, fast flow in blue, and slower flow in green and yellow (Rignot and Mouginot, 2012), and (f) ice-thinning rates from ICESat data for years 2003–2008 with thinning in red to thickening in blue (Pritchard et al., 2009).

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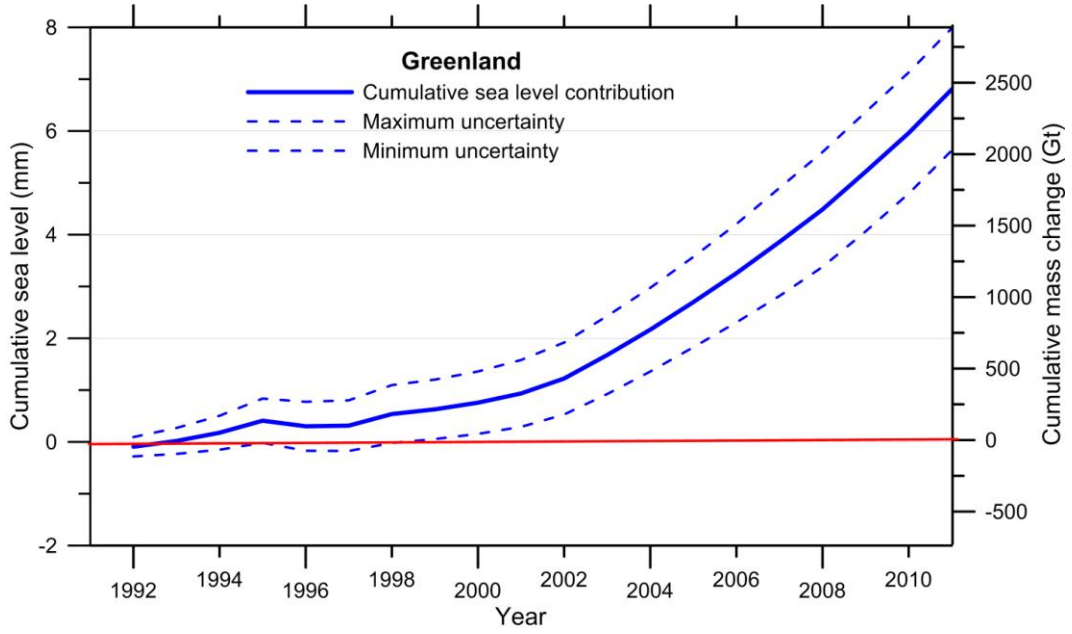
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Figure 4.14: Temporal evolution of ice loss in Antarctica from (a-c) GRACE time-variable gravity in centimetres of water per year for the periods (a) 2002 to 2006, (b) 2006 to 2011 and (c) 2002 to 2011, color coded red (loss) to blue (gain) (adapted from, Velicogna, 2009) Circles in (c) indicate average ice loss (Gt/yr) for years 2003–2008 for the Antarctic Peninsula, West Antarctica and East Antarctica (red = mass budget, orange = GRACE, grey = ICESat) (Shepherd and others, 2012); (d) mean surface mass balance in Antarctica for years 1989–2004 from regional atmospheric climate modelling (van den Broeke et al., 2006); (e) ice sheet velocity for 2007–2009 showing fastest flow in red, fast flow in blue, and slower flow in green and yellow (Rignot et al., 2011a); (f) ice thinning rates from ICESat for years 2003–2008 with thinning in red to thickening in blue (Pritchard et al., 2009).

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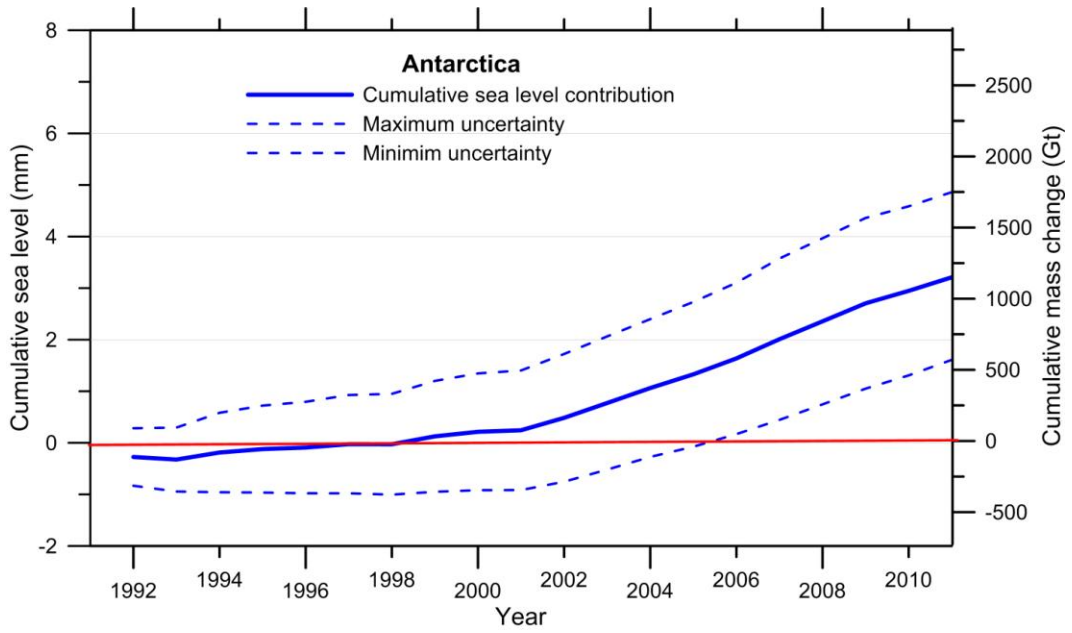
Figure 4.15: Cumulative sea level rise contribution (and ice loss equivalent) from Greenland derived from the unweighted annual averages from 14 recent studies (see main text and Appendix 4.A for details).

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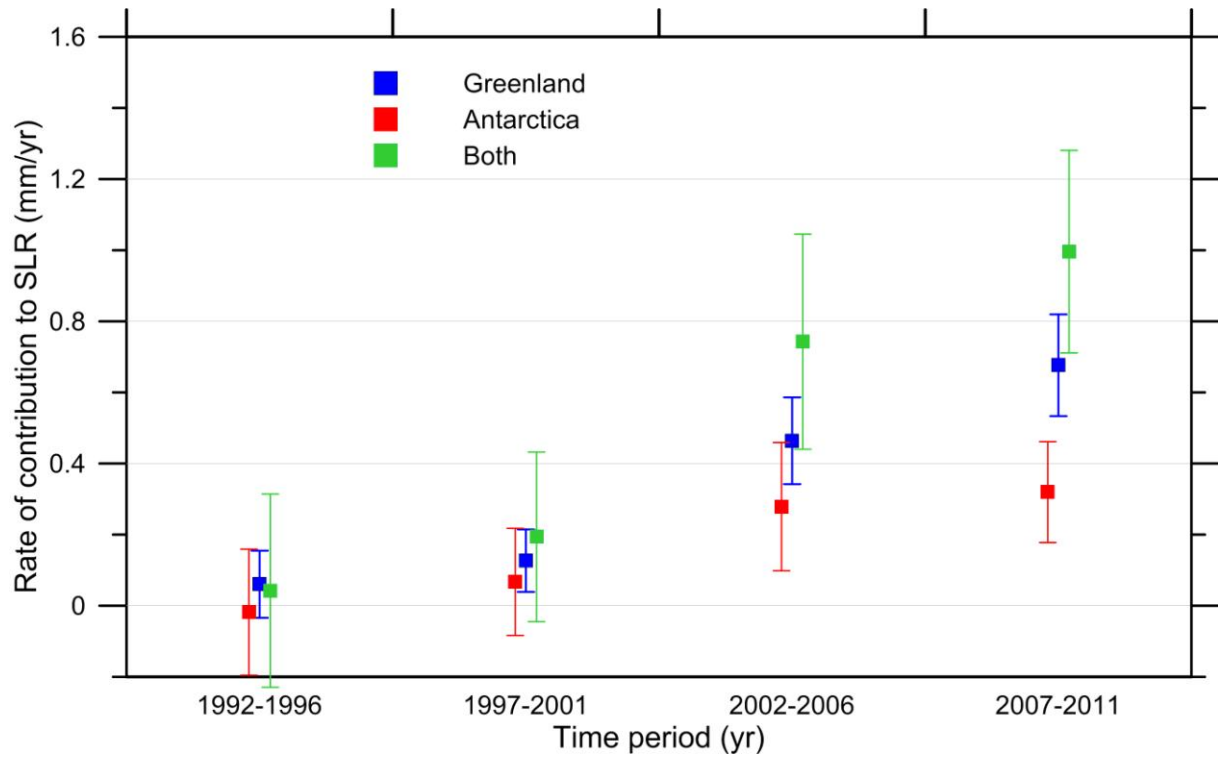
Figure 4.16: Cumulative sea level rise contribution (and ice loss equivalent) from Antarctica derived from the unweighted annual averages from 14 recent studies (see main text and Appendix 4.A for details).

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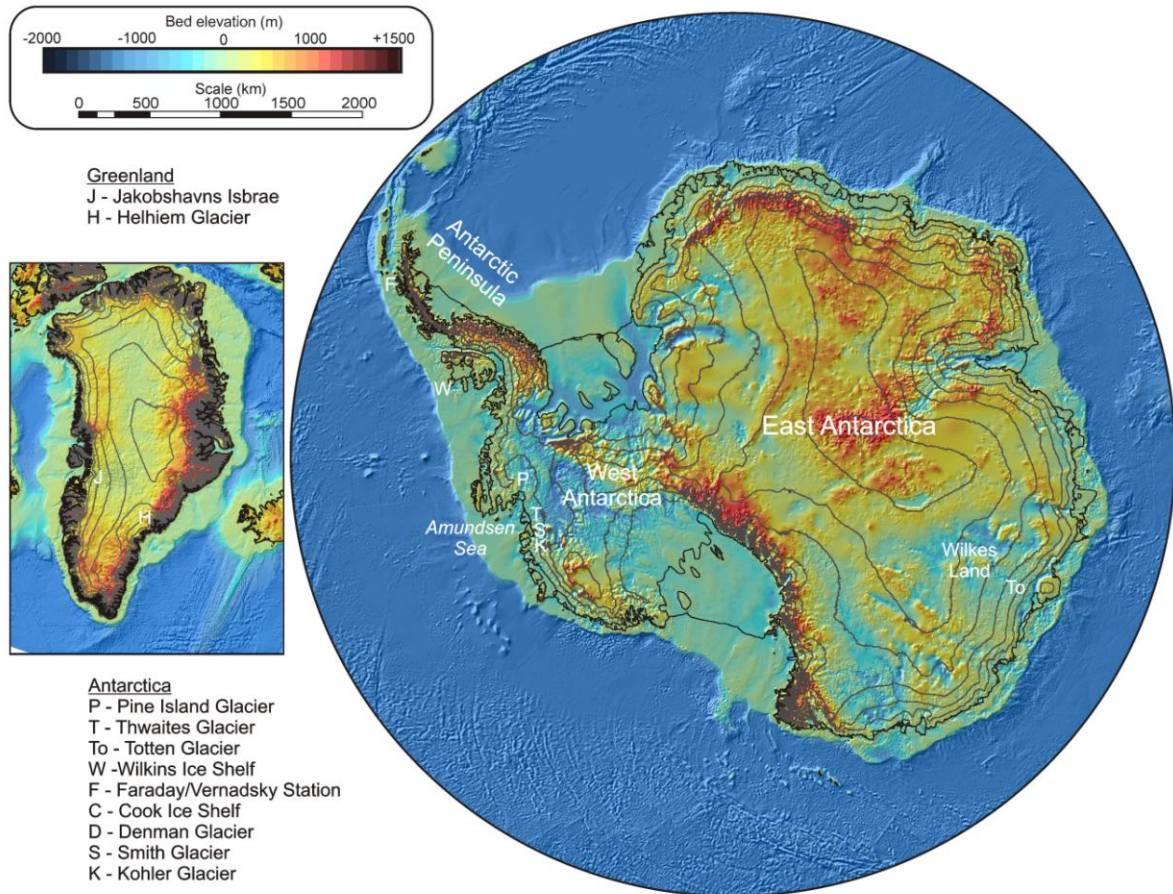
Figure 4.17: Rate of ice sheet contribution to sea level rise averaged over 5 year periods between 1992 and 2011. These estimates are derived from the data in Figure 4.15 and Figure 4.16.

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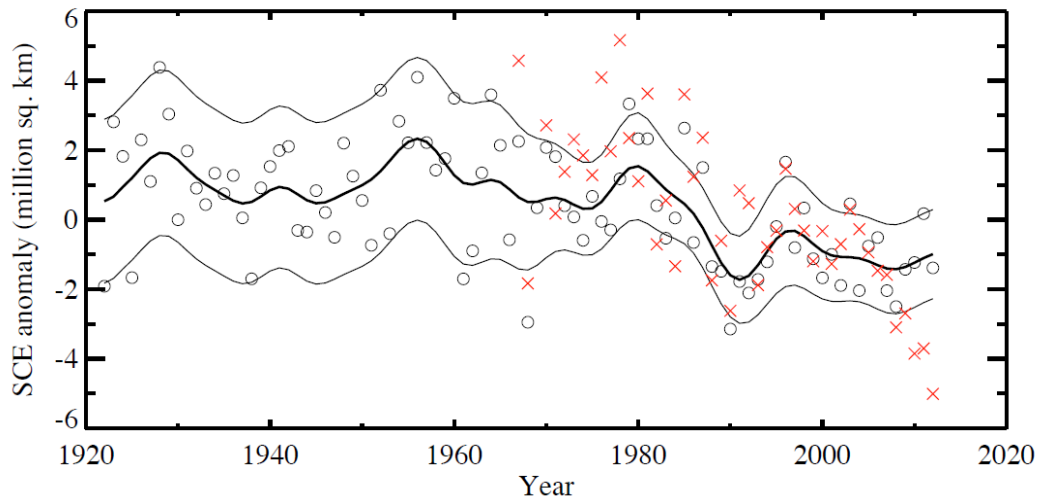
Figure 4.18: Bed topography for Greenland and Antarctica, derived from (Fretwell et al., Submitted; Griggs and Bamber, 2011b) with marine-based parts of the ice sheet highlighted and arrows showing access routes for rapid discharge of marine-based sectors. Figure drawn by P. Fretwell, British Antarctic Survey.

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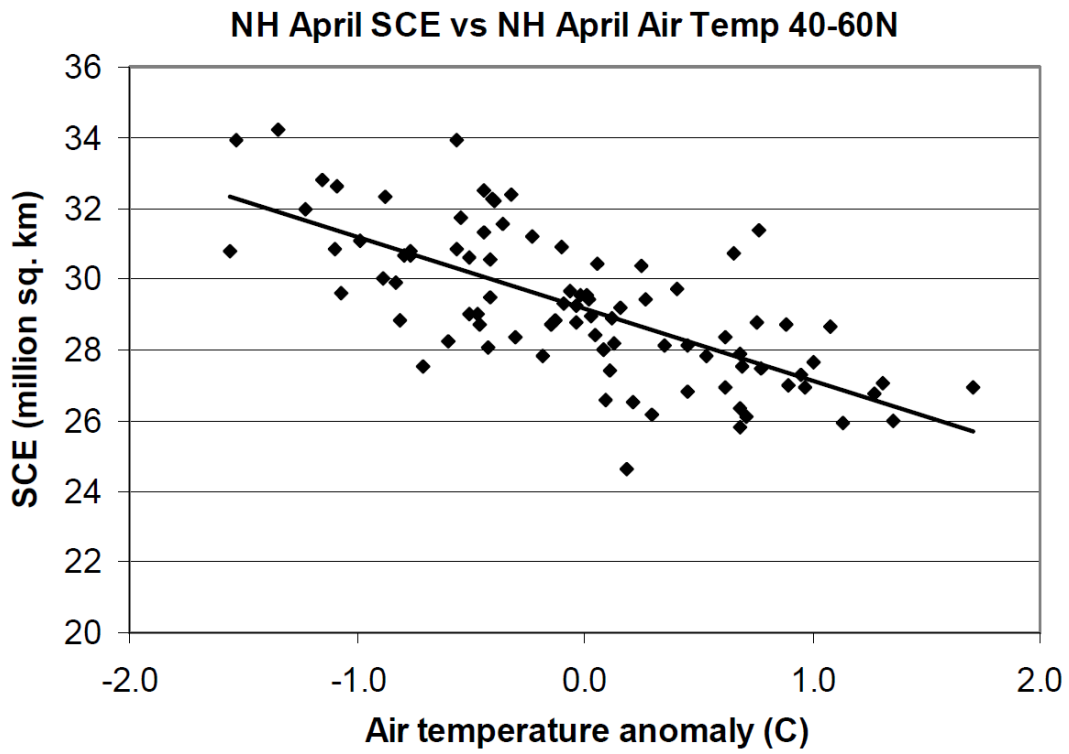
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Figure 4.19: March-April NH snow cover extent (SCE, circles) over the period of available data, shown with the 13-term filtered running mean and 95% confidence interval; and June SCE (x's, from satellite data alone). The width of the smoothed confidence interval is also influenced by the interannual variability in SCE. Updated, from Brown and Robinson (2011). For both time series the anomalies are calculated relative to the 1971–2000 mean.

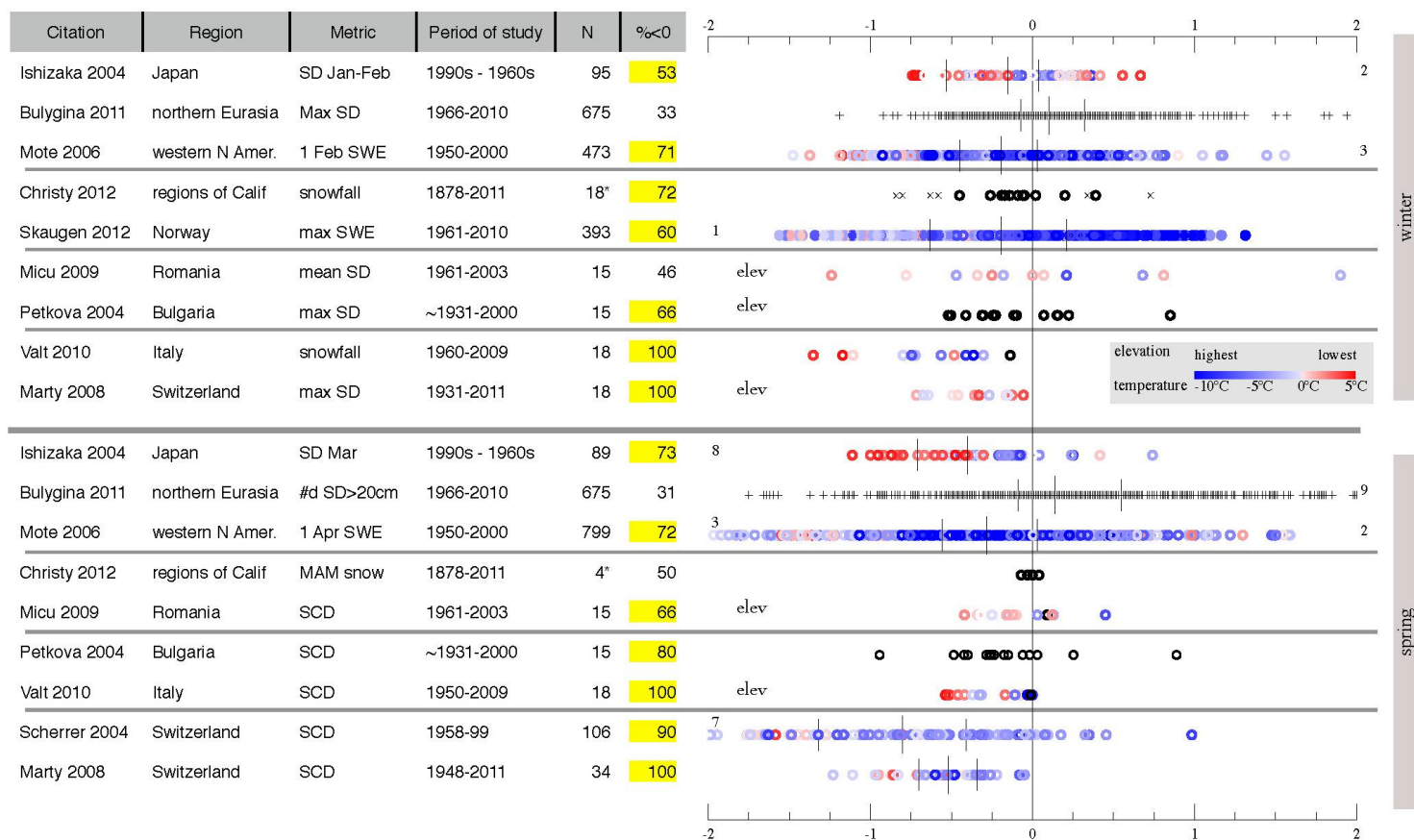
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Figure 4.20: Relationship between NH April SCE and corresponding land area air temperature anomalies over 40°N–60°N from the CRU dataset (Jones et al., 2012). Air temperature explains 48.7% of the variance. From Brown and Robinson (2011).

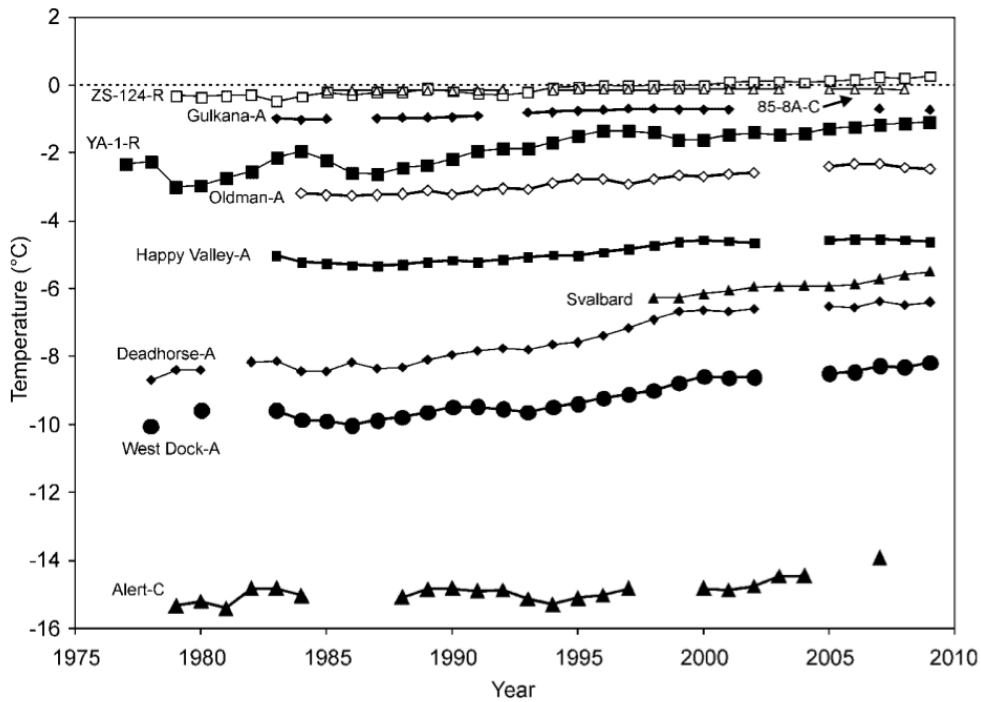
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3 **Figure 4.21:** Summary of station trends in metrics of snow that, based on the work of Brown and Mote (2009), are (top half) more reflective of mid-winter conditions and (bottom
 4 half) more reflective of spring conditions. See text for definitions of abbreviations. Where symbols are circles, the quantity plotted is the percentage change of a linear fit divided by
 5 the number of years of the fit. For the Bulygina study, the quantity plotted is the trend in cm a^{-1} (top) and \# days a^{-1} (bottom). Solid circles in the Skaugen study were statistically
 6 significant. Asterisks in the ‘N’ column emphasize that Christy 2011 combined records from over 500 stations into 18 regions; none of the trends was statistically significant. He
 7 judged time series from some regions to be unsuitable for statistical analysis and these are indicated here by an ‘x’. For studies with more than 50 sites, the median, upper and lower
 8 quartiles are shown with vertical lines. In a few cases, some trends lie beyond the edges of the graph; these are indicated by a numeral at the corresponding edge of the graph, e.g., 2
 9 sites $>2\% \text{ a}^{-1}$ for the Ishizaka study. Colours indicate temperature or, where indicated, elevation using the lowest and highest site to set the colour scale. Note the prevalence of
 10 negative trends at lower/warmer sites, especially in spring.

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Figure 4.22: Time series of mean annual ground temperatures at depths between 10 and 20 m for boreholes throughout the circumpolar northern permafrost regions (Romanovsky et al., 2010a). Data sources for North American, Russian and Nordic sites are Smith et al. (2010), Romanovsky et al. (2010b) and Christiansen et al. (2010), respectively. **C** Canadian site; **A** Alaskan site; **R** Russian site. The Svalbard site is Janssonhaugen (PACE-10) (Isaksen et al., 2007). Measurement depth for Russian boreholes and 85–8A is 10 m, Gulkana, Oldman and Alert are 15 m, and 20 m for all other boreholes. Borehole locations are: ZS-124 – 67.48°N 063.48°E; 85-8A – 61.68°N 121.18°W; Gulkana – 62.28°N 145.58°W; YA-1 – 67.58°N 648°E; Oldman – 66.48°N 150.68°W; Happy Valley – 69.18°N 148.88°W; Svalbard – 78.28°N 016.58°E; Deadhorse –70.28°N 148.58°W; West Dock – 70.48°N 148.58°W; Alert – 82.58°N 062.48°W.

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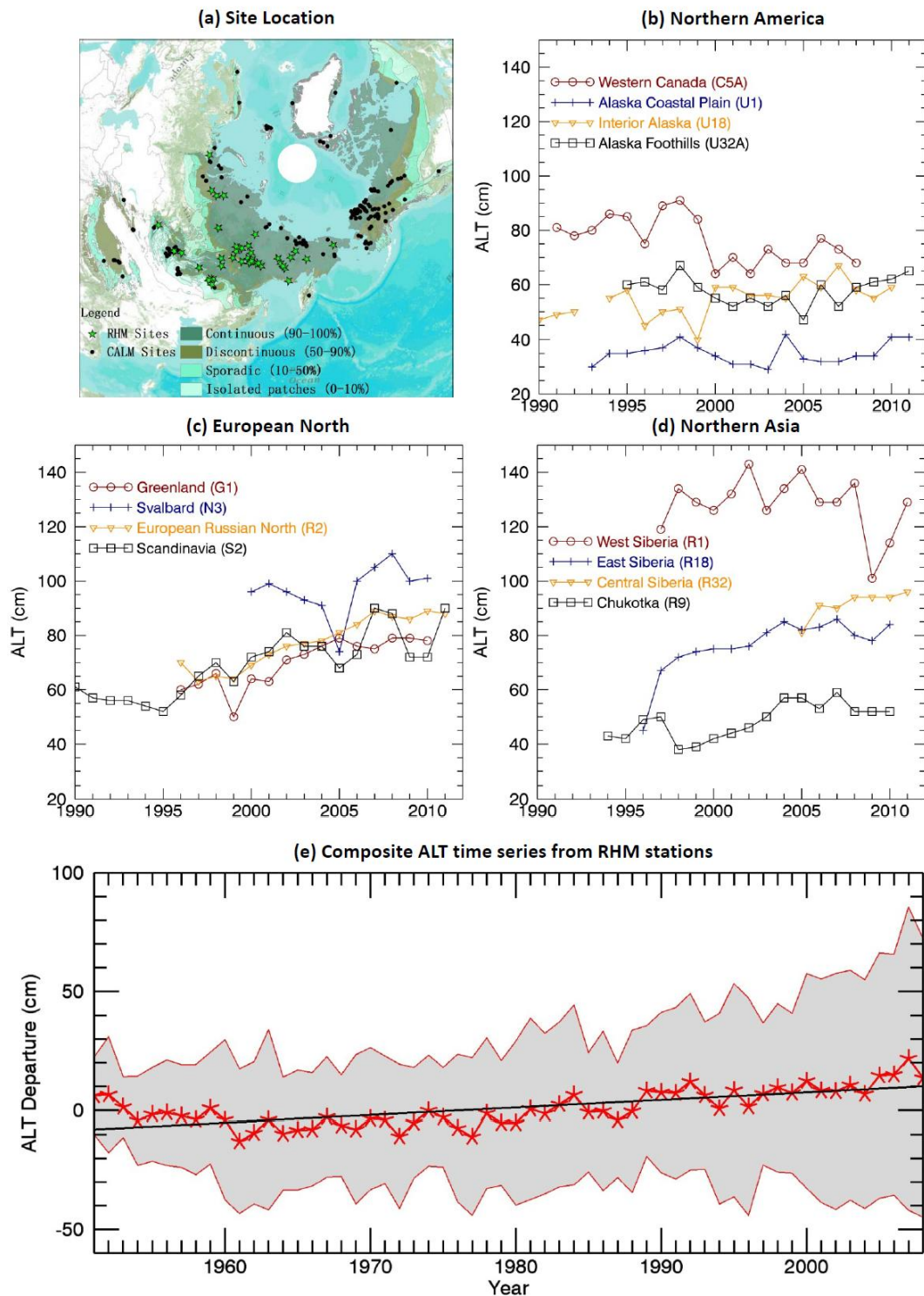
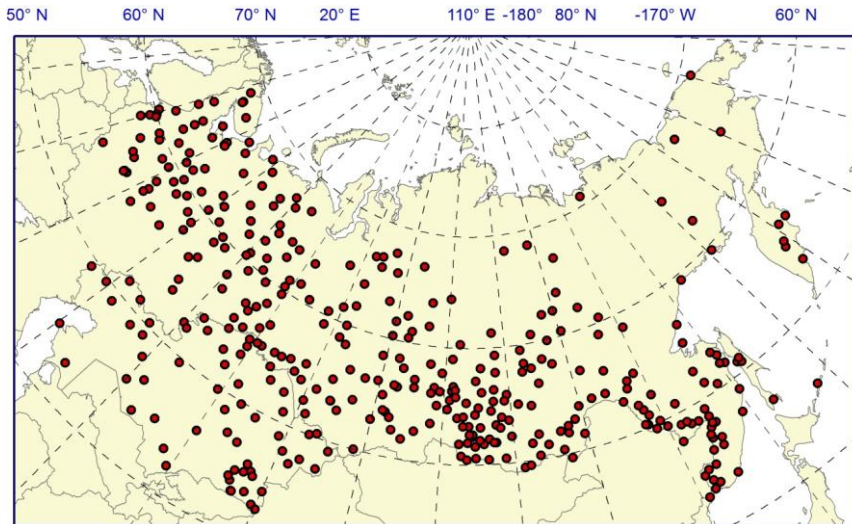


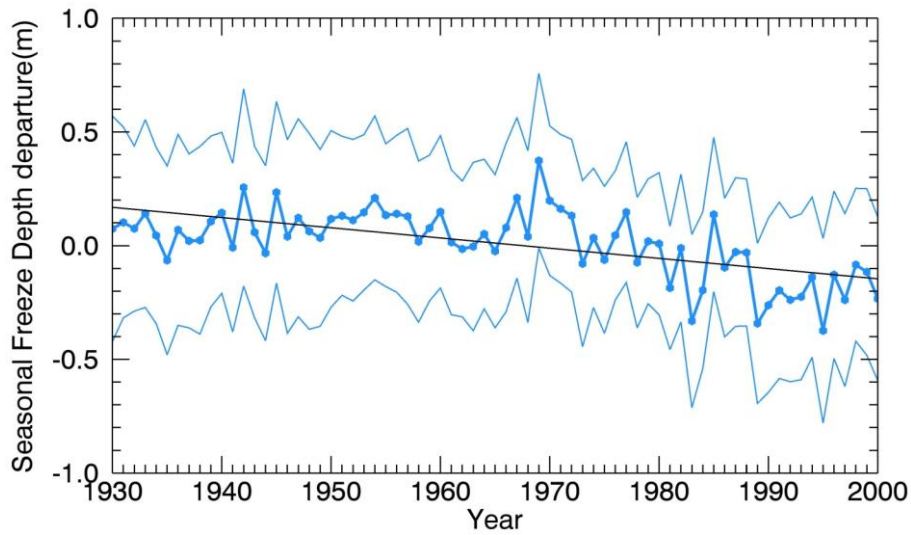
Figure 4.23: Locations for the Circumpolar Active Layer Monitoring (CALM) sites (Brown et al., 2000) and the Russian Hydrometeorological Stations (Frauenfeld et al., 2004; Zhang et al., 2005) (a), and changes in active layer thickness in Northern America (b), Northern Asia (c), European North (d), and Siberia (e). ALT data for Northern America, Northern Asia and European North are obtained from the International Permafrost Association (IPA) CALM website at <http://www.udel.edu/Geography/calm/about/permafrost.html>. Data for Siberia stations are obtained from the Russian Hydrometeorological Stations (RHM). Figure 4. 23 (e) shows a composite of ALT changes extracted from RHM stations soil temperature data. The number of RHM stations has expanded from 31 stations as reported from Frauenfeld et al. (2004) and Zhang et al. (2005) to 44 stations and the time series has extended from 1990 to 2008. The red star represents the mean composite value, the shaded area indicates the standard deviation, and the line is the trend.

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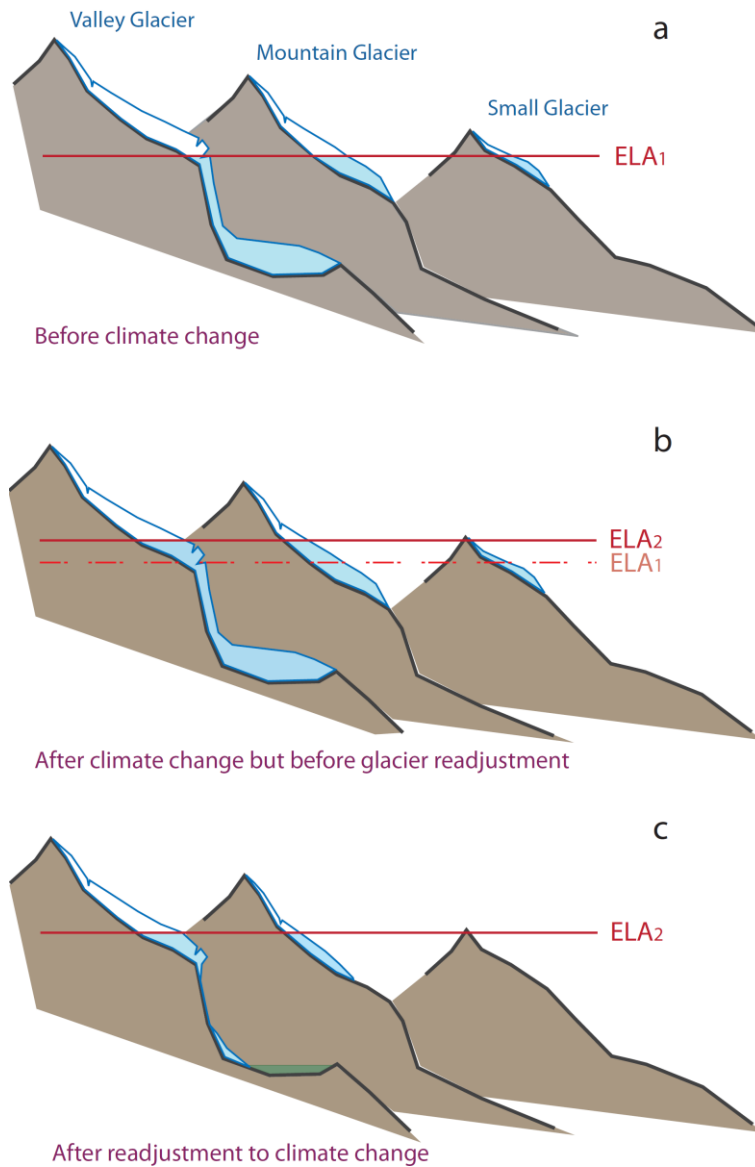
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Figure 4.24: Top: distribution of hydrometeorological stations across Russia; bottom: changes in thickness of seasonally frozen ground (From, Frauenfeld and Zhang, 2011).

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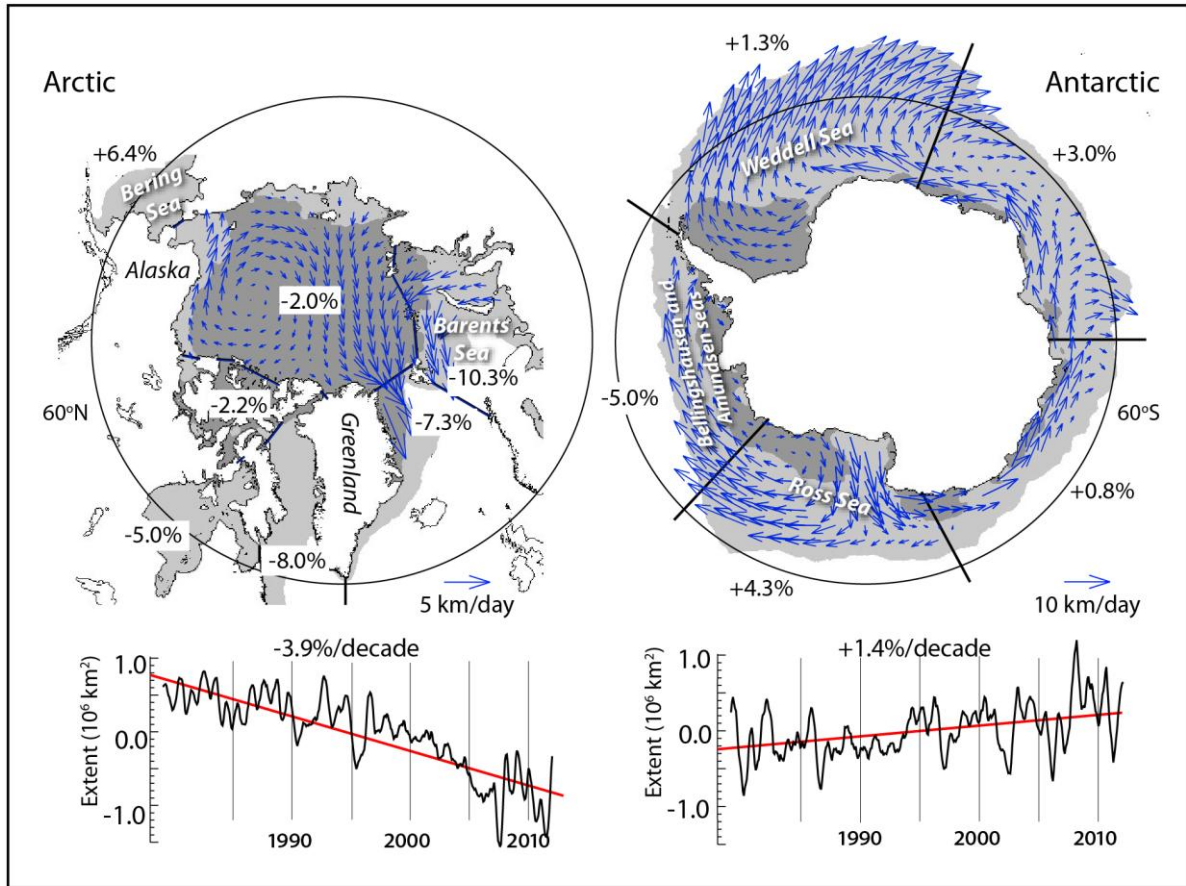
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4 **FAQ 4.1, Figure 1:** Schematic of three types of glaciers located at different elevations, and their response to an upward
 5 shift of the equilibrium line altitude (ELA). a) For a given climate, the ELA has a specific altitude (ELA₁), and all
 6 glaciers have a specific size. b) Due to a temperature increase, the ELA shifts upwards to a new altitude (ELA₂),
 7 initially resulting in reduced accumulation and larger ablation areas for all glaciers. c) After glacier size has adjusted to
 8 the new ELA, the valley glacier (left) has lost its tongue and the small glacier (right) has disappeared entirely.

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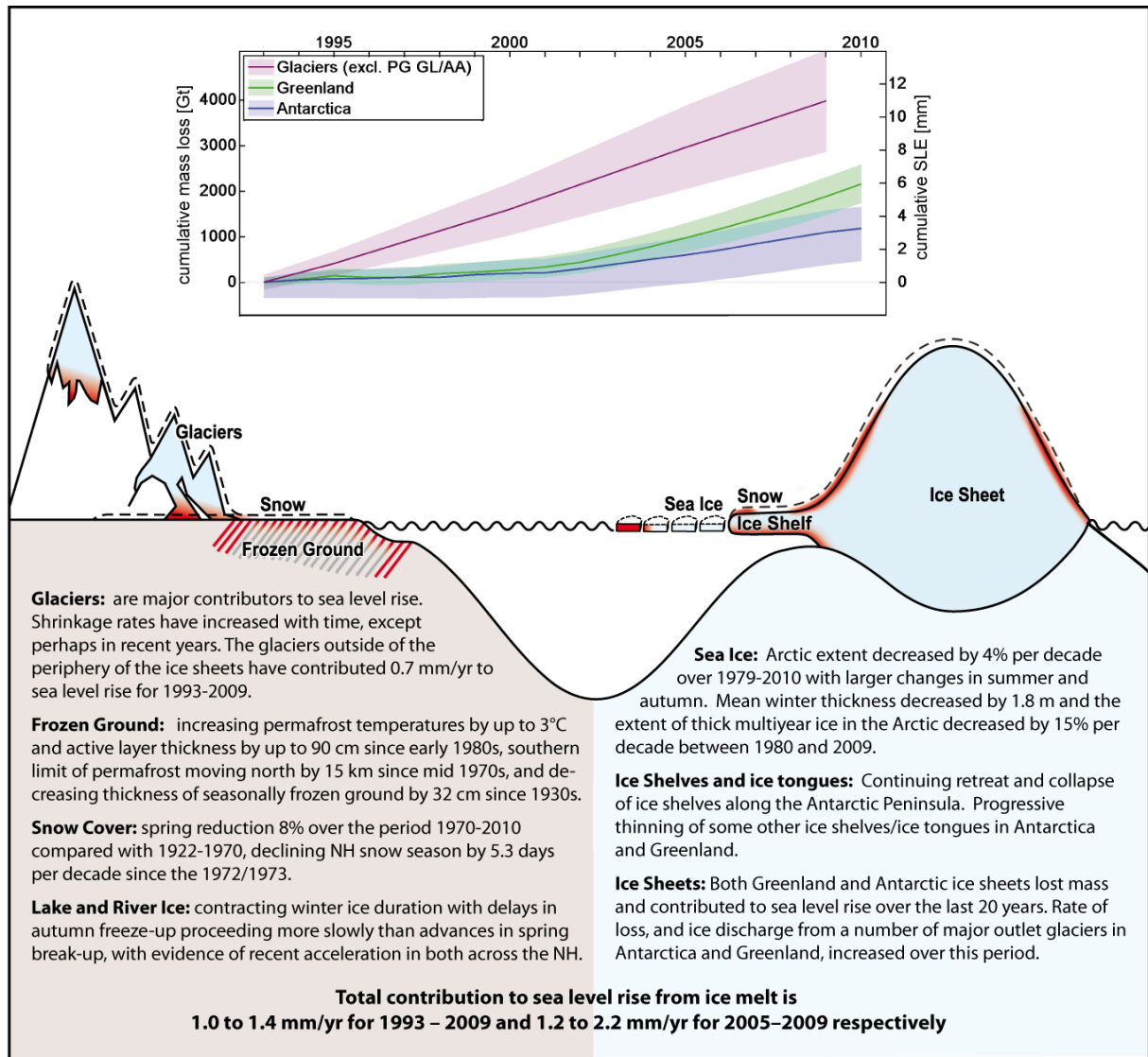
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FAQ 4.2, Figure 1: The mean circulation pattern of sea ice and the decadal trends (%) in annual average ice extent in different sectors of the Arctic and Antarctic. The average sea ice cover for the period 1979 through May 2011, from satellite observations, at maximum (minimum) extent is shown as light (dark) grey shading.

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Figure 4.25: Schematic summary of the observed variations in the cryosphere. The insert figure summarises our assessment of the contribution of ice loss from the ice sheets of Greenland and Antarctica to global sea level rise, together the contribution from all glaciers except those in the periphery of the ice sheets (Section 4.3.4).