Chapter 4: Observations: Changes in Snow, Ice and Frozen Ground

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Table of Contents

Executive Summary

In the climate system the cryosphere (which consists of snow, river and lake ice, sea ice, ice shelves and ice sheets, and frozen ground) is intricately linked to the surface energy budget, the water cycle and sea level change. The cryosphere integrates climate variations over a wide range of time-scales making it a natural sensor of climate variability, and providing a visible expression of climate change. In the past, the cryosphere has undergone significant variations on many time scales, associated with ice ages and with shorter-term variations like the Younger Dryas or the Little Ice Age. Recent decreases in ice mass are correlated with rising surface air temperatures over the past century. This is especially true for the region north of 65°N, where temperatures have increased by about twice the global average from 1965 to 2005.

- Snow cover has decreased in most regions, especially in spring. April Northern Hemisphere (NH) snow cover observed by satellite decreased during 1966–2004 by 0.5×10^6 km^2 per decade, or about 5% in total. NH snow cover decreased in every month but November and December. In the Southern Hemisphere (SH), the small number of long records or proxies mostly shows either decreases or no changes in the past 40+ years. Where snow cover or snow pack decreased, temperature often dominates; where snow increased, precipitation almost always dominates. For example, NH April snow cover extent is significantly correlated (r = –0.68) with 40–60°N April temperature, and declines in the mountains of western North America and in the Swiss Alps are largest at low elevations.

- Freeze-up and break-up dates for river and lake ice exhibit considerable spatial variability (with some regions showing trends of opposite sign). Averaged over available data for the NH, spanning the past 150 years, freeze-up date has occurred later at a rate of 5.8 ± 1.9 days per century, while the break-up date has occurred earlier at a rate of 6.5 ± 1.4 days per century.

- Satellite data indicate a continuation of the roughly 2.7 ± 0.7% per decade decline in annual mean Arctic sea-ice extent since 1978. The decline for summertime extent is larger than for wintertime, with the summer minimum declining at a rate of about 7.4 ± 2.9% per decade. Other data indicate that the summer decline began around 1970. Similar observations in the Antarctic reveal larger inter-annual variability but no consistent trends.

- Submarine-derived data for the central Arctic indicate a reduction in sea-ice thickness of about 1m from 1987 to 1997. Model-based reconstructions suggest an Arctic-wide reduction of 0.6 to 0.9 m over the same period. Large-scale trends prior to 1987 are ambiguous.

- Mass loss of glaciers and ice caps is estimated to be 0.51 ± 0.32 mm in sea level equivalent (SLE) per year between 1961 and 2003, and 0.81 ± 0.43 mm SLE per year between 1993 and 2003. The late 20th century glacier wastage is likely a response to post-1970 warming. Strongest mass losses per unit area are observed in Patagonia, Alaska and NW USA/SW Canada. Because of the corresponding large areas, the biggest contributions to sea level rise come from Alaska, the Arctic, and the Asian high mountains.

- Taken together, the ice sheets in Greenland and Antarctica are shrinking. Thickening in central regions of Greenland is more than offset by increased melting near the coast. Some outlet glaciers, which drain ice from the interior, are accelerating in both Greenland and Antarctica. Assessment of the data and techniques suggests a mass balance of the Greenland Ice Sheet of between +25 and –60 Gt (–0.07 to 0.17 mm SLE) per year from 1961–2003, and –50 to–100 Gt (0.14 to 0.28 mm SLE) per year from 1993–2003, with even larger losses in 2005. Estimates for the overall Antarctic ice-sheet mass balance range from +100Gt to –200 Gt (–0.28 to 0.55 mm SLE) per year for 1961–2003, and from +50 Gt to –200 Gt (–0.14 to 0.55 mm SLE) per year for 1993–2003. Acceleration of mass loss may have occurred, but not so dramatically as in Greenland.

- Permafrost temperature has increased by up to 3°C since the 1980s in the Arctic. The permafrost base is thawing at a rate ranging from 0.02 m/year in Alaska to 0.4 m/year on the Tibetan Plateau. Permafrost degradation is leading to widespread changes in land surface characteristics and drainage systems.

- The maximum extent of seasonally frozen ground has decreased by about 7% in the NH, and its maximum depth has decreased about 0.3 m in Eurasia since the mid-20th century. In addition, maximum

seasonal thaw depth has increased about 0.2m in the Russian Arctic. Onset dates of thaw in spring and freeze in autumn advanced five to seven days in Eurasia from 1988–2002, leading to an earlier growing season but no change in duration.

- Results summarized here indicate that the total cryospheric contribution to sea level change ranges from –0.2 to 1.5 mm per year between 1961–2003, and from 0.4 to 2.1 mm per year between 1993 and 2003, at rates that increased during the period primarily due to increasing losses from mountain glaciers and ice caps. Assuming a midpoint-mean ± uncertainties and Gaussian error summation of estimates for glaciers and ice sheets, a total cryospheric contribution of 1.2 ± 0.6 mm SLE per year is derived for 1993–2003.

4.1 Introduction

The main components of the cryosphere are snow, river and lake ice, sea ice, glaciers and ice caps, ice shelves, ice sheets, and frozen ground (Figure 4.1.1). In terms of the ice mass and its heat capacity, the cryosphere is the second largest component of the climate system (after the ocean). Its relevance for climate variability and change is based on physical properties, such as its high surface reflectivity (albedo) and the latent heat associated with phase changes, which have a strong impact on the surface energy balance. The presence/absence of snow or ice in polar regions is associated with an increased/decreased meridional temperature difference, which impacts on winds and ocean currents. Because of the positive temperature-ice albedo feedback, cyrospheric components act to amplify changes and so represent sensitive indicators of climate variation and change. Elements of the cryosphere are found at all latitudes, enabling a near-global assessment of cryosphere-related climate changes.

The cryosphere stores about 75% of the world's fresh water. The volume of the Greenland and Antarctic ice sheets are equivalent to approximately 7 m and 57 m of sea level rise, respectively. Changes of the ice mass on land have contributed significantly to recent changes of the sea level. On a regional scale, many glaciers and small ice caps play a crucial role in fresh water availability.

Presently, ice permanently covers 10% of the land surface, of which only a tiny fraction lies in ice caps and glaciers outside Antarctica and Greenland (Table 4.1.1). Ice also covers approximately 7% of the oceans in the annual mean. In mid-winter, snow covers approximately 49% of the land surface in the Northern Hemisphere (NH). Frozen ground has the largest area of any component of the cryosphere. Changes in the components of the cryosphere occur at different time-scales, depending on their dynamic and thermodynamic characteristics (Figure 4.1.1). All parts of the cryosphere contribute to short-term climate changes, with frozen ground, ice shelves and ice sheets contributing also to longer term changes including the ice-age cycles.

[INSERT FIGURE 4.1.1 HERE]

Table 4.1.1. Area, volume and sea level equivalent of cryospheric components. Indicated are the annual minimum and maximum for snow, sea ice and seasonally frozen ground, and the annual mean for the other components. The sea ice area is represented by the extent (area enclosed by the sea ice edge). The values for glaciers and ice caps denote the smallest and largest estimates excluding glaciers and ice caps surrounding Greenland and Antarctica.

Cryospheric Components	Area (10^6 km^2)	Ice Volume (10^6 km^3)	Sea Level Equivalent (SLE) (m)[f]
Snow on land (NH)	1.9 ~ 45.2	0.0005 ~ 0.005	0.001 ~ 0.01
Sea ice	19 ~ 27	0.019 ~ 0.025	
Glaciers and ice caps[a, (b)]	0.51 (0.54)	0.05 (0.13)	0.15 (0.37)
Ice shelves[c]	1.5	0.7	
Ice sheets	14.0	27.6	63.9
Greenland[d]	1.7	2.9	7.3
Antarctica[c]	12.3	24.7	56.6
Seasonally frozen ground (NH)[e]	5.9 ~ 48.1	0.006 ~ 0.065	
Permafrost (NH)[e]	22.8	4.5	

Notes:
(a) Ohmura (2004)
(b) Dyurgerov and Meier (2005)
(c) Lythe et al. (2001)
(d) Bamber et al. (2001)
(e) Zhang et al. (2003), excluding permafrost under ocean, ice sheets and glaciers.
(f) Assuming an oceanic area of 3.62×10^8 km^2, an ice density of 917 kgm^{-3}, a sea water density of 1028 kgm^{-3}, and sea water replacing grounded ice below sea level.

Seasonally, the area covered by snow in the NH ranges from a mean maximum in January of 45.2×10^6 km^2 to a mean minimum in August of 1.9×10^6 km^2. Snow covers more than 33% of lands north of the equator from November to April, reaching 49% coverage in January. The role of snow in the climate system includes strong positive feedbacks related to albedo and other, weaker feedbacks related to moisture storage, latent heat, and insulation of the underlying surface (Clark et al., 1999a).

The seasonal ice cover that forms on high-latitude rivers and lakes plays an important role in freshwater ecosystems, winter transportation, bridge and pipeline crossings, etc. Changes in the thickness and duration of these ice covers can therefore have consequences for both the natural environment and human activities. Of particular importance is the use of river and lake ice as a part of the northern road transportation network. In many northern countries river crossings can only be made in winter using 'ice bridges' (sections of ice that have been made thicker by clearing snow or flooding the surface with water). Similarly, ships and barges are often used on rivers and lakes to supply remote settlements and to transport ore, minerals and other resources. This can only be accomplished during the ice-free summer period. Finally, the breakup of river ice is often accompanied by 'ice jams' (blockages formed by accumulation of broken ice); these jams impede the flow of water and may lead to severe flooding.

At maximum extent Arctic sea ice covers more than 15 million km^2, reducing to only 7 million km^2 in summer. Antarctic sea ice is considerably more seasonal, ranging from a winter maximum of over 19 million km^2 to a minimum extent of about 3 million km^2. Sea ice less than one year old is termed *first-year ice* and that which survives more than one year is called *multi-year ice*. Most sea ice is part of the mobile *pack ice*, which circulates in the polar oceans, driven by winds and surface currents. This pack ice is extremely inhomogeneous, with differences in ice thicknesses and age, snow cover, open water distribution, etc. occurring on spatial scales from metres to hundreds of kilometres.

Glaciers and ice caps have a relatively high ratio between mass turnover and mass storage, and so adapt to changing climate conditions much more rapidly than the ice sheets. Changes of glaciers and ice caps reflect climate variations, in many cases in remote areas where no direct climate records are available such in high latitudes or on the high mountains that penetrate high into the mid troposphere. They contribute to sea level changes and affect the fresh water availability in many mountains and surrounding regions. Formation of large and hazardous lakes is occurring as glacier termini retreat from prominent Little Ice Age moraines, especially in the steep Himalaya and the Andes.

The ice sheets of Greenland and Antarctica are the main reservoirs capable of affecting sea level. Ice formed from snowfall, spreads under gravity towards the coast, where it melts or calves into the ocean to form icebergs. Until recently (including the TAR) it was assumed that the spreading velocity would not change rapidly, so that impacts of climate change could be estimated primarily from expected changes in snowfall and surface melting. Observations of rapid ice-flow changes since the TAR have complicated this picture, with strong indications that floating ice shelves "regulate" the motion of tributary glaciers, which can accelerate manifold following ice-shelf breakup.

Frozen ground includes seasonally frozen ground and permafrost. The permafrost region occupies approximately 22.79×10^6 km^2 or 23.9% of the land area in the Northern Hemisphere. On average, the long-term maximum area extent of the seasonally frozen ground, including the active layer over permafrost, is about 48.12×10^6 km^2 or 50.5% of the land area in the Northern Hemisphere. In terms of the area extent, frozen ground is the single largest component, hence the most vulnerable part to climate change, of the cryosphere. Permafrost records air temperature changes and other proxy information about environmental changes. Frozen ground can translate climatic change to other environmental components and facilitate further climate change through the impacts on greenhouse gas exchange between the atmosphere and the land surface.

4.2 Changes in Snow Cover

4.2.1 Background

The high albedo of snow (0.8–0.9 for fresh snow) has an important influence on the surface energy budget and on Earth's radiative balance (e.g., Groisman et al., 1994). Snow albedo, and hence the strength of the feedback, depends on a number of factors such as the depth and age of a snow cover, vegetation height, the amount of incoming solar radiation, and cloud cover. The albedo of snow may be decreasing because of anthropogenic increases in the sootiness of snow (Hansen and Nazarenko, 2004); see Section 2.5.4 for details.

In addition to the direct snow-albedo feedback, snow may influence climate through indirect feedbacks (i.e., those in which there are more than two causal steps) on summer soil moisture, which is typically enhanced in years with more snow or later snowmelt, and on atmospheric circulation. Indirect feedbacks on atmospheric circulation may involve two types of circulation, monsoonal (e.g., Lo and Clark, 2001) and annular (e.g., Saito and Cohen, 2003), though these connections are statistically tenuous and controversial (Robock et al., 2003; Bamzai, 2003).

In this section, observations of snow cover extent are updated from the TAR. In addition, several new topics are covered: Changes in snow depth and snow water equivalent; relationships of snow to temperature and precipitation; and observations and estimates of changes in snow in the southern hemisphere. Changes in the fraction of precipitation falling as snow or other frozen forms are covered in Chapter 3, Section 3.3.2.3. This section covers only snow on land; snow on various forms of ice is covered in subsequent sections.

4.2.2 Observations of Snow Cover, Snow Duration, and Snow Quantity

4.2.2.1 Sources of snow data

Daily observations of the depth of snow and of new snowfall have been made by various methods in many countries. The number of stations reporting snow depths climbed from just a few in the early 1900s to a maximum during the 1970s before declining in the 1990s. In the mountains of western North America, routine measurements of snow water equivalent (SWE) at roughly monthly intervals became widespread by 1950. In situ snow data suffer from changes in station location, observing time, and land cover, which must be considered in evaluating long-term trends, and are not uniformly distributed.

The premier dataset used to evaluate large-scale snow covered area (SCA) is the weekly visible-wavelength satellite maps of Northern Hemisphere snow cover produced by the U.S. National Oceanic and Atmospheric Administration's (NOAA) National Environmental Data and Information Service (NESDIS) (Robinson et al., 1993). Trained meteorologists produced the weekly NESDIS snow product from visual analyses of visible satellite imagery. These maps constitute the longest satellite-derived environmental dataset available, dating to late 1966, and are well-validated against surface observations. Changes in mapping procedures in 1999 have affected the continuity of data series at a small number of mountain and coastal gridpoints, but the areas in question are very small and estimates of hemispheric snow cover are unaffected. For the southern hemisphere, mapping of SCA began only in 2000 with the advent of MODIS.

Spaceborne remote sensing of microwave emissions provides the potential for global monitoring of snow cover with several advantages over the visible imagery: retrievals are unimpeded by cloud cover and winter darkness, and permit estimates of depth and SWE, which cannot be derived from visible images. Microwave brightness temperature data are available from 1978 for estimating snow cover extent, snow depth and snow water equivalent, although differences in sensor calibration in the switch between SMMR and SSM/I in 1987 must be resolved in order to generate homogeneous depth or SWE data series (Derksen et al., 2003). Estimates of SCA from microwave compare moderately well with visible data except in autumn (when microwave estimates are too low) and over the Tibetan plateau (microwave too high) (Armstrong and Brodzik, 2001). Work is ongoing to develop reliable depth and SWE retrievals from passive microwave for areas with heavy forest or deep snowpacks, and the relatively coarse spatial resolution (~10-25 km) still limits applications over mountainous regions.

4.2.2.2 Variability and trends in Northern Hemisphere snow cover

In this subsection, following the hemispheric view provided by the large-scale analyses by Brown (2000) and Robinson et al. (1993), regional and national-scale studies are discussed. The mean annual Northern Hemisphere SCA is 23.9×10^6 km^2, not including the Greenland ice sheet.

Interannual variability of SCA is largest not in winter, when mean SCA is greatest, but in autumn (in absolute terms) or summer (in relative terms). Monthly standard deviations range from 1.0×10^6 km^2 in August and September to 2.7×10^6 km^2 in October, and are generally just below 2×10^6 km^2 in non-summer months. There remains some uncertainty as to whether the microwave satellite data show similar interannual variability and trends except in autumn (see Section 4.2.2.1).

Since the early 1920s, and especially since the late 1970s, SCA has declined in spring (Figure 4.2.1) and summer but not substantially in winter (Table 4.2.1) despite winter warming (see Chapter 3, Section 3.2.2). Recent declines in SCA in the months of February through August have resulted in (1) a shift in the month of maximum SCA from February to January; (2) a statistically significant decline in annual mean SCA; and (3) a shift toward earlier spring melt by almost 2 weeks in the 1972–2000 period (Dye, 2002). Early in the satellite era, between 1967 and 1987, mean annual SCA was 24.4×10^6 km^2. An abrupt transition occurred between 1986 and 1988, and since 1988 the mean annual extent has been 23.1×10^6 km^2, a statistically significant (T test, $p < 0.01$) reduction of approximately 5% (Robinson and Frei, 2000). April SCA also declined significantly over the 1922–2004 period, by $2.7 \pm 0.7 \times 10^6$ km^2 or about 8% (updated from Brown, 2000).

[INSERT FIGURE 4.2.1 HERE]

Table 4.2.1. Trend (10^6 km^2/decade) in monthly NH SCA from satellite data (Rutgers corrected, D. Robinson) over the 1966–2004 period and for three months covering the 1922–2004 period based on the NH SCA reconstruction of Brown (2000).

Years	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
1966–2004	–0.15	–0.52	–0.72[a]	–0.50[a]	–0.60[a]	–0.98[a]	–1.10[a]	–0.55[a]	–0.00	–0.39	0.15	0.28
1922–2004	n/a	n/a	–0.21[a]	–0.32[a]	n/a	n/a	n/a	n/a	n/a	0.22[a]	n/a	n/a

Notes:
(a) Statistically significant (0.05) trends.
n/a: not available.

Warming plays a significant role in variability and trends of NH SCA, especially in March and April (Figure 4.2.2). Two related pieces of evidence support this conclusion. First, April NH SCA and April air temperature (40–60°N) over the 1922–2004 period are highly correlated on an interannual timescale ($r = -0.68$) (updated from Brown, 2000), reflecting the strength of the snow-albedo feedback, which also helps determine the longer-term trends (for temperature see Section 3.2.2). Second, the swath of largest declines in snow cover in March and April over middle latitudes of North America and Eurasia corresponds to the areas where snow cover and temperature are strongly correlated (Clark et al., 1999a). Snow-albedo feedbacks are likely contributing to this elevated spring response as demonstrated by Groisman et al. (1994).

The following paragraphs discuss regional details including information not available or missing from the satellite data and from Brown’s (2000) hemispheric reconstruction.

[INSERT FIGURE 4.2.2 HERE]

4.2.2.2.1 North America

From 1915 to 2004, North American SCA increased in November, December and January owing to increases in precipitation (Chapter 3, Section 3.3.2; Groisman et al., 2004). Over the same period of record, trends in other months are not significant; they become significant only when trend analysis begins after mid-century. Declines are most pronounced over western North America (Groisman et al., 2004). Shifts toward earlier melt were also observed in northern Alaska by about 8 days since the mid-1960s (Stone et al., 2002).

Another dimension of change in snow is provided by the annual measurements of mountain SWE near April 1 in western North America, which indicate declines since 1950 at about 75% of locations monitored (Mote et al., 2005). The date of maximum mountain SWE appears to have shifted earlier by about two weeks since 1950, as inferred from streamflow measurements (Stewart et al., 2005). That these reductions are predominantly due to warming is demonstrated by regression analysis (e.g., Stewart et al., 2005), and by the dependence of trends in SWE (Mote et al., 2005) on elevation or equivalently mean winter temperature (Figure 4.2.3a), with largest percentage changes near snowline.

[INSERT FIGURE 4.2.3 HERE]

4.2.2.2.2 Europe and Eurasia

The reconstruction from instrumental records by Brown (2000) did not include data over western and central Europe. Snow cover trends in mountain regions of Europe are characterized by large regional and altitudinal variations. Recent declines in snow cover have been documented in the mountains of Switzerland (e.g., Scherrer et al., 2004) and Slovakia (Vojtek et al., 2003), but no change was observed in Bulgaria over the 1931–2000 period (Petkova et al., 2004). Each of the studies showing declines noted that the declines were largest at lower elevations, and Scherrer et al. (2004) statistically attributed the declines in the Swiss Alps to warming as is clear when trends are plotted against winter temperature (Figure 4.2.3b).

Lowland areas of central Europe are characterized by recent reductions in annual snow cover duration by ~1 day/yr (e.g., Falarz, 2002). Trends toward greater maximum snow depth but shorter snow season have been noted in Finland (Hyvärinen, 2003), the former Soviet Union 1936–1995 (Ye and Ellison, 2003), and in the Tibetan (Zhang et al., 2004) and Qinghai-Xizang (Chen and Wu, 2000) Plateaus since the late 1970s. Qin et al. (2006) reported no trends in snow depth or snow cover in western China, even though correlation with temperature was about –0.5 (correlation with winter precipitation was about +0.6).

4.2.2.3 Southern Hemisphere

Outside of Antarctica, very little land area in the southern hemisphere experiences snow cover annually; snow cover over Antarctica is covered in Section 4.6. Long-term records of snow cover, snowfall, snow depth, or SWE are scarce. In some cases, proxies for snowline can be used, but the quality of data is much lower than for most northern hemisphere areas.

4.2.2.3.1 South America

A long term increasing trend in the number of snow days was found in the eastern side of the central Andes region (33ºS) from 1885 to 1996, derived from newspaper reports of Mendoza city (Prieto et al., 2001). Estimates from microwave satellite observations for mid-latitude alpine regions of South America for the period of record 1979 to 2002 show substantial interannual variability with little or no long-term trend.

Other approaches suggest some response of snowline to warming in South America. The 0°C isotherm altitude (ZIA), an indication of snowline, has been derived from the daily temperature profile obtained from radiosonde data located at Quintero (32°47’S, 71°33’W, 8 m above sea level) (Carrasco et al., 2005), which represents the snowline behaviour in western Andes from about 30°S to 36°S. Over the 1975–2001 period of record, the linear change in winter ZIA was 121.9 ± 7.7 m. However, no significant change was observed for ZIA on the days when there was precipitation in central Chile, implying that the observed winter ZIA trend is caused by enhanced snow melt on dry days.

4.2.2.3.2 Australia and New Zealand

For the mountainous southeastern area of Australia, studies of late-winter (August-September) snow depth have shown some significant declines, as much as 40%. Trends in maximum snow depth were more modest. The stronger declines in late winter are attributed to spring season warming, while maximum snow depth is largely determined by winter precipitation, which has declined only slightly (Nicholls, 2004; Hennessy et al., 2003).

In New Zealand, annual observations of end-of-summer snowline on 47 glaciers have been made by airplane since 1977, and reveal large interannual variability primarily associated with atmospheric circulation anomalies (Clare et al., 2002); it is noteworthy, however, that the four years with highest snowline occurred in the 1990s. The only study of seasonal snow cover in the Southern Alps covered only the 1930–1985 period and has not been updated.

4.3 Changes in River and Lake Ice

4.3.1 Background

Because of the importance to many human activities, freeze-up and break-up dates of river and lake ice have been recorded for a long time at many locations. These records provide useful climate information, although they must be interpreted with care. In the case of rivers, both freeze-up and break-up at a given location can be strongly affected by conditions far upstream (for example, heavy rains or snow-melt in a distant portion of the watershed). In the case of lakes, the historical observations have typically been made at coastal locations (often protected bays and harbours) and so may not be representative of the lake as a whole, or comparable to more recent satellite-based observations. Nevertheless, these observations represent some of the longest records of cryospheric change available.

Observations of ice thickness are considerably sparser and are generally made using direct drilling methods. Long-term records are available at a few locations; however it should be noted that, just as for sea-ice, the variations and trends in lake and river ice thickness are a consequence of changes in snow-fall and redistribution along with changes in temperature and radiative forcing.

4.3.2 Changes in Freeze-up and Break-up Dates

Freeze-up is defined conceptually as the time at which a continuous and immobile ice cover forms; however, operational definitions range from local observations of the presence/absence of ice, to inferences drawn from river discharge measurements. Break-up is typically the time at which the ice cover begins to move downstream in a river or at which open water becomes extensive at the measurement location for lakes. Here again, there is some ambiguity in the specific date, and in the extent to which local observations reflect conditions elsewhere on a large lake or in a large river basin.

Selected time series from a recent compilation of river and lake freeze-up and break-up records by Magnuson et al. (2000) are shown in Figure 4.3.1. They limited consideration to records spanning at least 150 years. 9 out of 15 records showed significant trends toward later freeze-up and 16 out of 25 records showed significant trends toward earlier break-up (at the 5% confidence level). When averaged together, the freeze-up date has become later at a rate of 5.8 days per century, while the break-up date has occurred earlier at a rate of 6.5 days per century.

[INSERT FIGURE 4.3.1 HERE]

A larger sample of Canadian rivers spanning the last 30 to 50 years was analyzed by Zhang et al. (2001). These freeze-up and break-up estimates (based on inferences from streamflow data) exhibit considerable variability, with a trend toward earlier freeze-up and break-up over much of the country of the country. The earlier break-up dominates however, leading to a significant increase in open water duration at many locations as shown in Figure 4.3.2. A recent analysis of Russian river data by Smith (2000) revealed a trend toward earlier freeze-up of western Russian rivers and later freeze-up in rivers of eastern Siberia over the last 50 to 70 years. Break-up dates did not exhibit statistically significant trends.

[INSERT FIGURE 4.3.2 HERE]

A comparable analysis of freeze-up and break-up dates for Canadian lakes has recently been completed by Duguay et al. (2005, in press). These results (shown in Figure 4.3.3) indicate a fairly general trend toward earlier break-up (particularly in western Canada), while freeze-up exhibited a mix of early and later dates.

[INSERT FIGURE 4.3.3 HERE]

There are insufficient published data on river and lake ice thickness to allow assessment of trends. Modelling studies (e.g., Duguay et al., 2003) indicate that, as with the landfast sea-ice case, much of the variability in maximum ice thickness and break-up date is driven by variations in snowfall.

4.4 Changes in Sea Ice

4.4.1 Background

Sea ice is formed by freezing of sea water in the polar oceans. It is an important, interactive component of the global climate system because: a) it is central to the powerful 'ice-albedo' feedback mechanism that enhances climate response at high latitudes; b) it modifies the exchange of heat, gases and momentum between the atmosphere and polar oceans, and c) it redistributes freshwater via the transport and subsequent melt of relatively fresh sea ice, and hence alters ocean buoyancy forcing.

The thickness of sea ice is a consequence of past growth, melt and deformation, and so is an important indicator of climatic conditions. Ice thickness is also closely connected to ice strength, and so changes in thickness are important to navigability by ships, to the stability of the ice as a platform for use by humans and marine mammals, to light transmission through the ice cover, etc. Sea ice increases in thickness as bottom freezing balances heat conduction through the ice to the surface (heat conduction is strongly influenced by the insulating thickness of the ice itself and the snow on it). Most of the inhomogeneity in the pack results from deformation of the ice due to differential movement of individual pieces of ice (called 'floes'). Open water areas created within the ice pack under divergence or shear (called 'leads') are a major contributor to ocean-atmosphere heat exchange (turbulent heat loss from the ocean in winter and shortwave heating in the summer). In some locations, due either to persistent ice divergence or to persistent upwelling of oceanic heat, open water areas within an otherwise ice-covered region can be sustained over much of the winter. These are called 'polynyas' and are important feeding areas for marine mammals and birds.

Under convergence, thin ice sheets may 'raft' on top of each other, doubling the ice thickness, and under strong convergence (for example, when wind drives sea ice against a coast), the ice buckles and crushes to form sinuous 'ridges' of thick ice. In the Arctic, ridges can be tens of meters thick, account for nearly half of the total ice volume, and constitute a major impediment to transportation on, through, or under the ice. Although ridging is generally less severe in the Antarctic, ice deformation is still an important process in thickening the ice cover.

Near shore, in bays and fjords, and amongst islands like those of the Canadian Arctic Archipelago, sea ice can be attached to land and therefore be immobile. This is termed 'landfast' ice. In the Arctic such ice (and in particular its freeze-up and break-up) is of special importance to local residents as it is used as a platform for hunting and fishing, and is an impediment to shipping.

Some climatically important characteristics of sea ice include its concentration (that fraction of the ocean covered by ice); its extent (the area enclosed by the ice edge – operationally defined as the 15% concentration contour); the total area of ice within its extent (i.e., extent weighted by concentration), the area of multi-year ice within the total extent, its thickness (and the thickness of the snow cover on it); and its velocity; its growth and melt rates (and hence salt or freshwater flux into the ocean). Ice extent, or ice edge position, is the only sea ice variable for which observations are available for more than a few decades. The position of the ice edge, particularly in winter, reflects the location where the ice supplied by advection is balanced by melt, which is in turn determined largely by transport of heat in the atmosphere and ocean. Expansion or retreat of the ice edge may be amplified by the ice albedo feedback.

4.4.2 Sea Ice Extent and Concentration

4.4.2.1 Data sources and time periods covered

The most complete record of sea ice extent is provided by passive microwave satellite data available since the early 1970s. Prior to that, aircraft, ship and coastal observations are available at certain times and in certain locations. Portions of the north Atlantic are unique in having ship observations extending well back into the 19th century. Far fewer historic data exist from the Southern Hemisphere.

Estimation of sea-ice properties from passive microwave emission requires an algorithm to convert observed radiance into ice concentration (and type). Several such algorithms are available (e.g., Steffen et al., 1992) and their accuracy has been evaluated using high-resolution satellite and aircraft imagery (e.g., Cavalieri, 1992; Kwok, 2002) and operational ice charts (e.g., Agnew and Howell, 2003). The accuracy of satellite-derived ice concentration is usually 5% or better, although errors of 10–20% can occur during the melt season. The accuracy of the ice edge (relevant to estimating ice extent) is largely determined by the spatial resolution of the satellite radiometer, and is on the order of 25 km (recently-launched instruments provide improved resolution of about 12.5 km). Summertime concentration errors do lead to a bias in estimated ice-covered area in the warm seasons of both northern and southern hemisphere (Agnew and Howell, 2003; Worby and Comiso, 2004). This is an important consideration when comparing the satellite period with older proxy records of ice extent.

Distinguishing between first-year and multi-year ice from passive microwave data is more difficult. Comparisons of passive and active microwave estimates of multi-year ice fraction indicate large differences (e.g., Kwok et al., 1996) and so the derived multi-year ice concentration is probably not a reliable climate indicator. However the summer minimum ice extent, which is by definition the multi-year ice extent at that time of year, is not as prone to algorithm errors (e.g., Comiso, 2002).

4.4.2.2 Hemispheric, regional and seasonal time series from passive microwave

Most analyses of variability and trend in ice extent using the satellite record have focussed on the period after 1978 when the satellite sensors have been relatively constant. A notable result is the asymmetry between Arctic and Antarctic changes. An updated version of the analysis done by Comiso (2003), spanning the period from November 1978 through December 2005 is shown in Figure 4.4.1. The annual mean ice extent anomalies are shown. There is a significant decreasing trend in Arctic sea ice extent of -33 ± 8.8 x 10^3 km^2 per year (equivalent to approximately -2.7 ± 0.7% per decade), whereas the Antarctic results show a small positive trend of 5.6 ± 11 x 10^3 km^2 per year (0.47 ± 0.9%)which is not statistically significant. The uncertainties represent the 95% confidence interval around the trend estimate and the percentages are based on the 1978–2005 mean. In both hemispheres the trends are larger in summer and smaller in winter. In addition, there is considerable variation in the magnitude, and even the sign, of the trend from region to region within each hemisphere.

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[INSERT FIGURE 4.4.1 HERE]

The most remarkable change observed in the Arctic ice cover has been the decrease in ice that survives the summer, shown in Figure 4.4.2. Trend in the minimum Arctic sea ice extent, between 1979 and 2005, was -60 ± 24 x 10^3 km^2 per year (–7.4 ± 2.9% per decade). These trends are superimposed on substantial interannual to decadal variability which is associated with variability in atmospheric circulation (Belchansky et al., 2005).

[INSERT FIGURE 4.4.2 HERE]

4.4.2.3 Longer records of hemispheric extent

The lack of comprehensive sea ice data prior to the satellite era hampers estimates of hemispheric-scale trends over longer time scales. Rayner et al. (2003) compiled a data set of sea ice extent for the 20th century from available sources and account for the inhomogeneity between them (Figure 4.4.3). There is a clear indication of sustained decline in Arctic ice extent since about the early 1970s, particularly in summer. On a regional basis, portions of the North Atlantic have sufficient historical data, based largely on ship reports and coastal observations, to permit trend assessments over periods exceeding 100 years. Vinje (2001) compiled information from ship reports in the Nordic Seas to estimate April sea-ice extent in this region for the period since about 1860. This time series is also shown in Figure 4.4.3 and indicates a generally continuous decline from the start of the record to the end. Ice extent data from Russian sources have recently been published (Polyakov et al., 2003), and cover essentially the entire 20th century for the Russian coastal seas (Kara, Laptev, East Siberian and Chukchi). These data, which exhibit large interdecadal variability, show a declining trend since the 1960s until a reversal in the late 1990s. It is particularly notable that the Russian data indicate anomalously little ice during the 1940s and 1950s, whereas the Nordic Sea data indicates anomalously large extent at this time. Omstedt and Chen (2001) obtained a proxy record of the annual maximum extent of sea ice in the region of the Baltic Sea over the period 1720–1997. This record showed a substantial decline in sea ice that occurred around 1877, and that there was greater variability in sea ice extent in the colder 1720–1877 period than in the warmer 1878–1997 period. Hill et al., 2002 have examined sea ice information for the Canadian maritime region and deduced that sea ice incursions occurred during the 1800s in the Grand Banks and surrounding areas that are now ice-free. Although there are problems with homogeneity of all these data (with quality declining further back in history), and with the disparity in spatial scales represented by each, they are all consistent in terms of the declining ice extent during the latter decades of the 20th century, with the decline beginning prior to the satellite era. Those data that extend far enough back in time imply, with high confidence, that sea-ice was more extensive in the North Atlantic during the 19th century.

[INSERT FIGURE 4.4.3 HERE]

Continuous long-term data records for the Antarctic are lacking, as systematic information on the entire Southern Ocean ice cover became available only with the advent of routine microwave satellite reconnaissance in the early 1970s. Parkinson (1990) examined ice edge observations from four late-18th to early 19th century exploration voyages. Her analysis suggested that the summer Antarctic sea ice was more extensive in the eastern Weddell Sea in 1772 and in the Amundsen Sea in 1839 than the present day range from satellite observations. But many of the early observations are within the present range for the same time of year. An analysis of whaling records by de la Mare (1997) suggested a step decline of Antarctic sea ice coverage by 25% (a 2.8° poleward shift in average ice edge latitude) between the mid-1950s and the early 1970s. A re-analysis by Ackley et al. (2003), which accounted for offsets between satellite-derived ice edge and whaling ship locations, challenged evidence of significant change in ice edge location. Curran et al. (2003) made use of a correlation between methanesulphonic acid (MSA) concentration (a by-product of marine phytoplankton) in a near coastal Antarctic ice core and the regional sea ice extent in the sector from 80E to 140E to infer a quasi decadal pattern of interannual variability in the ice extent in this region, along with a roughly 20% decline (approximately 2 degrees of latitude) since the 1950s.

In summary, the Antarctic data provides evidence of a decline in sea-ice extent in some regions, but there is insufficient data to draw conclusions about hemispheric changes prior to the satellite era.

4.4.3 Sea Ice Thickness

4.4.3.1 Sea ice thickness data sources and time periods covered

Until recently there have been no satellite remote sensing techniques capable of mapping sea ice thickness, and this parameter has primarily been determined by drilling or by under-ice sonar measurement of draft (the submerged portion of sea ice).

Sub-sea sonar from submarines or moored instruments can be used to measure ice draft over a footprint of 1–10 m diameter. Draft is converted to thickness assuming an average density for the ice-snow in the measured floe. The principal challenges to accurate observation with sonars are uncertainties in sound speed and atmospheric pressure, and the identification of spurious targets. Upward-looking sonar has been on submarines operating beneath Arctic pack ice since 1958). US and UK naval data are now being released for science, and some dedicated Arctic submarine missions were made for science during 1993–1999. Ice-draft measurement by moored ice-profiling sonar, which are best suited to studies of ice transport or change at fixed sites, began in the Arctic in the late 1980s. Instruments have operated since 1990 in the Beaufort and Greenland Seas and for shorter intervals in other areas, but few records span more than 10 years. In the Southern Hemisphere there are no data from submarines and only short time series from moored sonar.

Other techniques such as electromagnetic-induction sounders deployed on the ice surface, ships or aircraft, or air-borne laser altimetry to measure freeboard, have limited applicability to wide-scale climate analysis of sea ice thickness. Indirect estimates, based on measurement of surface gravity waves, are available in some regions for the 1970s and 80s (Nagurnyi et al., 1999 as reported in Johannessen et al., 2004), but the accuracy of these estimates is difficult to quantify.

Quantitative data on the thickness of Antarctic pack ice only started to become available in the 1980s from sparsely scattered drilling programs covering only small areas and primarily for use in validating other techniques. Visual observations of ice characteristics from ships (Worby and Ackley, 2000) are not adequate for climate monitoring, but are providing one of the first broad pictures of Antarctic sea ice thickness.

4.4.3.2 Evidence of changes in Arctic pack ice thickness from submarine sonar

Estimates of thickness change over limited regions are possible when submarine transects are repeated (e.g. Wadhams, 1992). The North Pole is a common way point in many submarine cruises and this allowed McLaren et al. (1994) to analyze data from twelve submarine cruises near the Pole between 1958 and 1992. They found considerable interannual variability, but no significant trend. Shy and Walsh (1996) examined the same data in relation to ice drift and found that much of the thickness variability was due to the source location and path followed by the ice prior to arrival at the Pole.

Rothrock et al. (1999) provided the first 'basin-scale' analysis and found that ice draft in the mid 1990s was less than that measured between 1958 and 1977 at every available location (including the North Pole). The change was least (–0.9 m) in the southern Canada Basin, greatest (–1.7 m) in the Eurasian Basin [with an estimated overall error of less than 0.3 m]. The decline averaged about 42% of the average 1958–1977 thickness. Their study included very few data within the seasonal sea ice zone and none within 200 miles of Canada or Greenland.

Subsequent studies indicate that the reduction in ice thickness was not gradual, but occurred abruptly before 1991. Winsor (2001) found no evidence of thinning along 150°W from six springtime cruises during 1991–1996, but Tucker et al. (2001), using springtime observations from 1976 to 1994 along the same meridian, noted a decrease in ice draft sometime between the mid 1980s and early 1990s, with little subsequent change. The observed change in mean draft resulted from a decrease in the fraction of thick ice (more than 3.5-m draft) and an increase in the fraction of thin ice, which was probably due to reduced storage of multi-year ice in a smaller Beaufort gyre and the export of "surplus" via Fram Strait. Yu et al. (2004) presented evidence of a similar change in ice thickness over a wider area. However, ice thickness varies considerably from year to year at a given location and so the rather sparse temporal sampling provided by submarine data makes inferences regarding long-term change difficult.

4.4.3.3 Other evidence of sea ice thickness change in the Arctic and Antarctic

Haas (2004, and references therein) use ground-based electromagnetic induction measurements to show a decrease of approximately 0.5 m between 1991 and 2001 in the modal thickness (i.e., the most commonly observed thickness) of ice floes in the Arctic Trans-Polar Drift. Their survey of 120 km of ice on 146 floes during four cruises is biased by an absence of ice-free and thin-ice fractions, and underestimation of ridged ice, but the data are descriptive of floes that are safe to traverse in summer, and the observed changes are most likely due to thermodynamic forcing.

There are no available data on change in the thickness of Antarctic sea ice, much of which is considerably thinner and less ridged than ice in the Arctic Basin.

4.4.3.4 Model-based estimates of change

Physically-based sea ice models, forced with winds and temperatures from atmospheric reanalyses and sometimes constrained by observed ice concentration fields, can provide continuous time series of sea ice extent and thickness which can be compared to the sparse observations, and used to interprete the observational record. Models such as those described by Rothrock et al. (2003) and references therein are able to reproduce the observed interannual variations in ice thickness, at least when averaged over fairly large regions. In particular, model studies can elucidate some of the forcing agents responsible for observed changes in ice thickness.

A comparison of various model simulations of historical Arctic ice thickness or volume is shown in Figure 4.4.4 (based on figures in Rothrock et al., 2003 and Koeberle and Gerdes, 2003). All the models indicate a marked reduction in ice thickness of 0.6 to 0.9 m starting in the late 1980s, but disagree somewhat with respect to trends and/or variations earlier in the century. Most models indicate a maximum in ice thickness in the mid 1960s, with local maxima around 1980 and 1990 as well. There is an emerging suggestion from both models and observations that much of the decrease in thickness occurred between the late 1980s and late 1990s.

It is not possible to attribute the abrupt decrease in thickness inferred from submarine observations entirely to the (rather slow) observed warming in the Arctic, and some of the dramatic decrease may be a consequence of spatial redistribution of ice volume over time (e.g., Holloway and Sou, 2002). Low-frequency, large-scale modes of atmospheric variability (such as interannual changes in circulation connected to the Northern Annular Mode) affect both wind-driving of sea ice and heat transport in the atmosphere, and therefore contribute to interannual variations in ice formation, growth and melt (e.g., Rigor et al., 2002; Dumas et al., 2003).

[INSERT FIGURE 4.4.4 HERE]

For the Antarctic, Fichefet et al. (2003) conducted one of the few long-term simulations of ice thickness using observationally-based atmospheric forcing covering the period 1958 to 1999. They noted pronounced decadal variability, with area-average ice thickness varying by ±0.1m (over a mean thickness of roughly 0.9m), but no long-term trend.

4.4.3.5 Landfast ice changes

Inter-annual variations in landfast ice thickness for selected stations in northern Canada were analysed by Brown and Coté (1992). At each of the four sites studied, where ice typically thickens to about 2 m at the end of winter, they detected both positive and negative trends in ice thickness, but no spatially coherent pattern. Inter-annual variation in ice thickness at the end-of season was determined principally by variation in the amount and timing of snow accumulation, not variation in air temperature. An analysis of several half-century records in Siberian seas has provided evidence that trends in landfast ice thickness over the past century in this area have been small, diverse and generally not statistically significant (Polyakov et al., 2003). Some variability is correlated with a low-frequency atmospheric oscillation of multi-decadal period.

In the Antarctic, fast ice thickness measurements have been intermittently made at the coastal sites of Mawson and Davis for about the last 50 years. Although there is no long term trend in maximum ice thickness. At both sites there is a trend for the date of maximum thickness to become later at a rate of about 4 days per decade (Heil and Allison, 2002).

4.4.3.6 Snow on sea ice

Warren et al. (1999) analysed 37 years (1954–1991) of snow depth and density measurements made at Soviet drifting stations on multiyear Arctic sea ice. They found a weak negative trend for all months, with the largest trend, a decrease of 8 cm (23%) over 37 years in May, the month of maximum snow depth.

There are few data on snow cover and distribution in the Antarctic, and none adequate for detecting any trend in snow cover. Massom et al. (2001) collated available ship observations (between 1981 and 1987) to show that average Antarctic snow thickness is typically 0.15–0.20 m, and varies widely both seasonally and regionally. An important process in the Antarctic sea ice zone is the formation of snow-ice, which occurs when a snow loading depresses thin sea ice below sea level, causing sea water flooding of the near surface snow and subsequent rapid freezing.

4.4.3.7 Assessement of changes to sea ice thickness

Sea ice thickness is one of the most difficult geophysical parameters to measure on large-scales and, because of the large variability inherent in the sea-ice-climate system, evaluation of ice thickness trends from the available observational data is difficult. Nevertheless, on the basis of submarine sonar data and interpolation of the Arctic basin average sea ice thickness from a variety of physically-based sea ice models, it is very likely that the average sea ice thickness in the central Arctic has decreased by up to 1 metre since the late 1980s, and that most of this decrease occurred between the late 1980s and 1990s. The steady decrease in the area of the summer minimum Arctic sea ice cover since the 1980s, resulting in less thick multi-year ice at the start of the next growth season, is consistent with this. This recent decrease however occurs within the context of longer term decadal variability, with stong maxima in Arctic ice thickness in the mid-1960s and around 1980 and 1990, due to both dynamic and thermodynamic forcing of the ice by circulation changes associated with low-frequency modes of atmospheric variability.

There are insufficient data to draw any conclusions about trends in the thickness of Antarctic sea ice.

An emerging new technique, using satellite radar or laser altimetry to estimate ice freeboard (that part of the ice above water surface) from the measured ranges to the ice and sea surface in open leads (and assuming an average floe density and snow depth) offers promise for future monitoring of large-scale sea ice thickness. Laxon et al. (2003) estimated average Arctic sea ice thickness over the cold months (October-March) for 1993–2001 from satellite-borne radar altimeter measurements. Their data reveal a realistic geographic variation of thickness (increasing from about 2 m near Siberia to 4.5 m off the coasts of Canada and Greenland) and a significant (9%) inter-annual variability in winter ice thickness, but no indication of trend over this time.

4.4.4 *Pack Ice Motion*

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

4.4.4.1 Data sources and time periods covered

Sea ice motion data are primarily derived from the drift of ships, manned stations, and buoys set on or in the pack ice. Although some individual drift trajectories date back to the late 19th century in the Arctic and the early 20th century in the Antarctic, a coordinated observing program did not begin until the International Arctic Buoy Programme (IABP) in the late 1970s. The IABP currently maintains an array of about 25 buoys at any given time and produces gridded fields of ice motion from these using objective analysis (Rigor et al., 2002 and references therein).

Sea ice motion may also be derived from satellite data by estimating the displacement of sea ice features found in two consecutive images from a variety of satellite instruments (e.g., Agnew et al., 1997; Kwok, 2000). The passive microwave sensors provide the longest period of coverage (1979-present) but their spatial resolution limits the precision of motion estimates. The optimal interpolation of satellite and buoy data, (e.g., Kwok et al., 1998) seems to be the most consistent data set to assess interannual variability of sea ice motion.

In the Antarctic, buoy deployments have only been reasonably frequent since the late 1980s. Since 1995, buoy operations have been organised within the WCRP International Programme for Antarctic Buoys (IPAB), although spatial and temporal coverage remain poor. A digital atlas of Antarctic sea ice has been compiled from two decades of combined passive microwave and IPAB buoy data (Schmitt et al., 2004).

4.4.4.2 Changes in patterns of sea ice motion and modes of climate variability that affect sea ice motion

Gudkovich (1961) hypothesized the existence of two regimes of Arctic ice motion driven by large scale variations in atmospheric circulation. Using a coupled atmosphere-ocean-ice model, Proshutinsky and Johnson (1997) showed that the regimes proposed by Gudkovich (1961) alternated on 5–7 year intervals. Similarly, Rigor, et al. (2002) showed that the changes in the patterns of sea-ice motion from the 1980's to the 1990's are related to the Northern Annular Mode (NAM). There is, however, no indication of a long-term trend in ice motion.

In the Antarctic, ice motion undergoes an annual cycle caused by stronger winds in winter. Interannual oscillations are found in all regions, most regularly in the Ross, Amundsen, and Bellingshausen Seas with periods of about 3–6 years (Venegas et al., 2001). As for the Arctic, no trend in ice motion is apparent based on the limited data available.

4.4.4.3 Ice export and advection

The sea ice outflow through Fram Strait is a major component of the ice mass balance of the Arctic Ocean. Approximately 14% of the sea ice mass is exported each year through Fram Strait. Vinje (2001) constructed a time series of ice export during 1950–2000 using available moored ice-profiling sonar observations and a parameterization based on geostrophic wind. He found substantial inter-decadal variability in export but no trend.

Kwok and Rothrock (1999) assembled an 18 year time series of ice area and volume flux through Fram Strait based on satellite-derived ice motion and concentration estimates. They found an annual mean area flux of 919,000 km^2/yr, (nearly 10% of the Arctic Ocean area) with large interannual variability that is correlated in part with the NAM or NAO index. Using the thickness data of Vinje et al. (1998), they estimate a mean volume flux of 2366 km^3. Subsequent modelling by Hilmer and Jung (2000) indicated that the correlation between NAO (or nearly equivalently, the NAM) and Fram Strait ice outflow is somewhat transient, with significant correlation during the period 1978–1997, but no correlation during 1958–1977 (Figure 4.4.5). This was a consequence of rather subtle shifts in the spatial pattern of surface pressure (and hence wind) anomalies associated with the NAO. A recent update of this record (Kwok et al., 2004) to 24 years shows only minor variations in the mean volume and area flux and the correlation with NAO persists.

[INSERT FIGURE 4.4.5 HERE]

Overall, while there is considerable low frequency variability in the pattern of sea ice motion, there is no evidence of a trend in either hemisphere.

4.5 Changes in Glaciers and Ice Caps

4.5.1 Background

Those glaciers and ice caps not immediately adjacent to the large ice sheets of Greenland and Antarctica cover an area between 512 and 540 x 10^3 km^2 according to inventories from different authors (Table 4.5.1); volume estimates differ considerably from 51 to 133 x 10^3 km^3, with respective sea level rise equivalents (SLE) between 0.15 and 0.37 m. Including glaciers and ice caps surrounding the Greenland ice sheet and West Antarctica, but excluding those on the Antarctic peninsula and those surrounding East Antarctica, yields 0.71 ± 0.2 m SLE. These estimates are about 40% higher than those given in the IPCC Third Assessment Report (2001), but area inventories are still incomplete and volume measurements more so, despite increasing efforts.

Table 4.5.1. Extents of glaciers and ice caps as given by different authors. Area, A, volume, V, and respective sea level rise equivalent, SLE.

	R&B 05[a]	O 04[a]	D&M 05[a]	D&M 05[b]	IPCC 01[b]
A [10^3 km^2]	522 ± 42	512	540 ± 30	785 ± 100	680
V [10^3 km^3]	87 ± 10	51	133 ± 20		180 ± 40
SLE [m]	0.24 ± 0.03	0.15	0.37 ± 0.06	0.71 ± 0.2	0.50 ± 0.1

Notes:
(a) glaciers and ice caps surrounding Greenland and Antarctic ice sheets are excluded
(b) glaciers surrounding Greenland and West Antarctic ice sheets are included.
R&B 05 (Raper and Braithwaite, 2005): volume derived from hypsometry and volume/area scaling within 1° × 1° grid cells
O 04 (Ohmura, 2004): volume derived from a statistical relationship between glacier volume and area, calibrated with 61 glacier volumes derived from radio-echo-sounding measurements
D&M 05 (Dyurgerov and Meier, 2005): volume derived from a statistical relationship between glacier volume and area, calibrated with 144 glacier volumes derived from radio-echo-sounding measurements.

Glaciers and ice caps provide among the most visible indications of the effects of climate change. Their surface mass balance - the gain or loss of snow and ice over a hydrological year or season - is determined by the climate and the departure of the glacier from its equilibrium extent. In high and mid latitudes, mass balance seasons are determined by the annual cycle of air temperature, with accumulation dominating in winter and ablation in summer. In the low latitudes, ablation occurs year-round and accumulation is controlled by precipitation seasonality. In wide parts of the Himalaya most accumulation and ablation occur during summer (Fujita and Ageta, 2000). Small and steep glaciers are controlled primarily by their vertical mass balance profile, whereas the horizontal mass balance profiles are more important on larger and flatter glaciers. Mass balance gradients along these profiles and the associated glaciers' sensitivity to temperature change are strong for wet and warm conditions (maritime - large mass turnover) and weak under cold and dry conditions (continental - small mass turnover). The latter are most sensitive to changes of moisture-related conditions (Kaser, 2001). If climate changes the intensity and duration of the respective mass balance seasons and/or the mass balance gradients, a glacier will change its extent toward a size that allows the mass balance to become zero again. With few exceptions, mass balance of glaciers and ice caps always tends toward zero, although climate variability and the time lag of glacial response prevent a static equilibrium. Changes in glacier extent lag behind climate changes by only a few years on the short, steep and shallow glaciers of the tropical mountains with year-round ablation, but by up to several centuries on the largest glaciers and ice caps with small slopes and cold ice. Calving of icebergs into tidewater is not immediately and straightforwardly linked to climate, but general relations to climate can often be discerned.

4.5.2. Large and Global Scale Analyses

Records of glacier length changes go far back in time (written reports as far back as 1600 in a few cases) and are directly related to low-frequency climate change. From 169 glacier-length records, Oerlemans (2005) has compiled mean length variations of glacier tongues for large scale regions from 1700 to 2000 (Figure 4.5.1). Although much local to regional and high-frequency variability is superimposed, the smoothed series give an apparently homogeneous signal. General retreat of glacier termini started after 1800, with considerable mean retreat rates in all regions after 1850 lasting throughout the 20th century. A slowdown of retreats between about 1970 and 1990 is more evident in the raw data. Retreats were again generally rapid in the 1990s; the Atlantic and the Southern Hemisphere curves reflect precipitation driven advances of glaciers in Western Scandinavia and New Zealand (Chinn et al., 2005).

[INSERT FIGURE 4.5.1 HERE]

Records of directly measured glacier mass balances are few and stretch back only to the mid 20th century. Because of the very intensive fieldwork required, these records are biased toward logistically and morphologically "easy" glaciers. Uncertainty of directly obtained annual surface mass balance is typically ±200 kg m^{-2} y^{-1} due to measurement and analysis errors (Cogley, 2005). Data are originally collected and distributed by the World Glacier Monitoring Service (WGMS (ICSI-IAHS), various years). From these and from several other new and historical sources, annual mass balance time series for about 300 individual glaciers have been constructed, quality-checked, analyzed and presented in three databases (Cogley, 2003; Dyurgerov and Meier, 2005; Ohmura, 2004). Dyurgerov and Meier (2005) also incorporated recent findings from repeat altimetry of glaciers and ice caps in Alaska (Arendt et al., 2002) and Patagonia (Rignot et al., 2003). Only a few individual series stretch over the entire period. From these statistically small samples, global estimates have been obtained by arithmetic averaging (C1 in Figure 4.5.2), area weighting (DM and O in Figure 4.5.2), and by spatial interpolation (C2 in Figure 4.5.2). Although mass balances reported from individual glaciers include the effect of changing glacier area, deficiencies in the inventories do not allow for general consideration of area changes. The effect of this methodological inaccuracy is considered minor. Table 4.5.2 summarizes the data plotted in Figure 4.5.2.

[INSERT FIGURE 4.5.2 HERE]

Table 4.5.2. Area, *A* [10^3 km^2], mean annual rates of specific mass balance, *b* [kg m^{-2} y^{-1}], mass balance, *B* [Gt y^{-1}], and respective sea level rise equivalents, *SLE* [mm y^{-1}], of glaciers and ice caps for different periods. Global$^{\mathbf{excl}}$ excludes glaciers and ice caps around the ice sheets; Global$^{\mathbf{incl}}$ includes those around Greenland and West Antarctic ice sheets. Numbers for Global$^{\mathbf{excl}}$ correspond to MB in Figure 4.5.2, those for Global$^{\mathbf{incl}}$ are modified from Dyurgerov and Meier (2005) by applying pentadal DM/MB ratios. The uncertainty given for SLE applies to all variables. Sources: Cogley (2005), Dyurgerov and Meier (2005), and Ohmura (2004).

		1960/1961–2002/2003			1960/1961–1991/1992			1992/1993–2002/2003		
	A	*b*	*B*	*SLE*	*b*	*B*	*SLE*	*b*	*B*	*SLE*
Global$^{\mathbf{excl}}$	539.8	–270	–146	0.40 ± 0.22	–206	–111	0.31 ± 0.18	–399	–215	0.59 ± 0.30
Global$^{\mathbf{incl}}$	785.0	–233	–183	0.51 ± 0.32	–163	–128	0.35 ± 0.26	–375	–294	0.81 ± 0.43

The histories of global-mean mass balance from different authors have very similar shapes despite some offsets in magnitude. Around 1970 mass balances were close to zero or slightly positive in most regions as well as in the global mean (Figure 4.5.2), indicating near-equilibration with climate after the strong earlier mass loss particularly during the 1940s. This gives confidence that the late 20th century glacier wastage is essentially a response to post-1970 global warming (Greene, 2005). Strong mass losses are indicated for the 1940s but uncertainty is great since the arithmetic mean values (C1 in Figure 4.5.2) are from few glacier sites. Mass loss rates for 1992/1993 to 2002/2003 are roughly double those for 1960/1961–1991/1992 (Table 4.5.2). The most recent "pentade" stretches from 2000/01 to 2003/04 only. There, the arithmetic mean C1 is strongly biased by the strong negative mass balances in the European Alps in 2002/2003 indicating the caution that is appropriate when interpreting the arithmetic average of individual mass balance series.

Over the last half century, both global mean winter accumulation and summer melting have increased steadily (Dyurgerov and Meier, 2005; Greene, 2005; Ohmura, 2004); at least in the northern hemisphere, winter accumulation and summer melting correlate positively with hemispheric air temperature, whereas the net balance correlates negatively with hemispheric air temperature (Greene, 2005). Dyurgerov and Dwyer (2001) analysed time series of 21 Northern Hemisphere glaciers and found a rather uniformly increased mass-turnover rate, qualitatively consistent with moderately increased precipitation and substantially increased low-altitude melting. This general trend is also indicated by reports from Alaska (Arendt et al., 2002), the Canadian Arctic Archipelago (Abdalati et al., 2004) and Patagonia (Rignot et al., 2003).

Regional analyses by Dyurgerov and Meier (2005) show strongest negative specific mean mass balances in Patagonia, NW USA + SW Canada, and Alaska, with losses especially rapid in Patagonia and Alaska after the mid-1990s (Figure 4.5.3, a). A cumulative specific mass balance of -10×10^3 kg m^{-2} corresponds to loss of 10 m of water or about 11 m of ice averaged over the glaciated area; cumulative losses in Patagonia are approximately 40 m of ice thickness averaged over the glaciers. Only Europe showed a mean value close to zero, with strong mass losses in the Alps compensated by mass gains in maritime Scandinavia until the end of the 20th century. High spatial variability of climate and, thus, of glacier variations also exists in other large regions such as in the High Mountains of Asia (Dyurgerov and Meier, 2005; Liu et al., 2003). Values for Patagonia and Alaska are mainly derived from altimetry evaluations made by Arendt et al. (2002) and Rignot et al. (2003), and authors of both papers note that the observed mass losses cannot be explained by surface mass balance only but also by increased ice discharge from enhanced ice velocities. The latter, in turn, has possibly been triggered by previous negative mass balances of glaciers calving into tidewater as well as by increased meltwater production that enhances basal sliding. Some glaciers exhibit quasi-periodic internal instabilities (surging), which can affect data from those glaciers (Arendt et al., 2002; Rignot et al., 2003), but these effects are expected to average very close to zero over large regions and many years or decades. Because of lack of suitable information, the temporal variation of the mass loss of the Patagonian ice fields has been adapted to the course of Alaskan mass balances assuming similar climate regimes (Dyurgerov and Meier, 2005).

[INSERT FIGURE 4.5.3 HERE]

The surface mass balance of snow and ice is determined by a complex interaction of energy fluxes toward and away from the surface, and the occurrence of solid precipitation. Nevertheless, glacier fluctuations show a strong statistical correlation with air temperature at least on a large spatial scale throughout the 20th century (Greene, 2005), and a strong physical basis exists to explain why warming would cause mass loss. Changes in snow accumulation also matter, and may dominate in response to strong circulation changes or when temperature is not changing greatly. For example, analyses of glacier mass balances, volume changes, length variations and homogenized temperature records for the western portion of the European Alps (Vincent et al., 2005) clearly indicate the role of precipitation changes in glacier variations in the 18th and 19th centuries. Similarly, Nesje and Dahl (2003) explained glacier advances in southern Norway in the early 18th century based on increased winter precipitation rather than cold temperatures.

Mass balances are the effect of specific mass balances (climate signal) on the existing glacier area. Consequently, the biggest mass losses and, thus, contribution to sea level rise are from Alaska with 0.1 mm y^{-1} SLE from 1961–1992 and 0.28 mm y^{-1} SLE from 1993–2003, the Arctic (0.12 and 0.22), and the High Mountains of Asia (0.08 and 0.12) (Figure 4.5.3, b).

4.5.3 Special Regional Features

Although reports on individual glaciers or limited glacier areas support the global picture of ongoing strong ice shrinkage in almost all regions, some exceptional results indicate the complexity of both regional to local scale climate and respective glacier regimes.

For glaciers in the dry and cold *Taylor Valley, Antarctica*, Fountain et al. (2004) hypothesized that an increase in average air temperature by 2°C alone can explain the observed glacier advance through ice softening.

Altimetry in *Svalbard* suggested ice-cap growth (Bamber et al., 2004). Careful evaluation of the processes involved shows a sea level contribution of only 0.01 mm a^{-1} for the last 3 decades of the 20th century (Hagen et al., 2003).

Scandinavia: Norwegian coastal glaciers, which advanced in the 1990s due to increased accumulation in response to a positive swing in the North Atlantic Oscillation (NAO) (Nesje et al., 2000), started to shrink around 2000 as a result of almost simultaneous reduced winter accumulation and greater summer melting (Kjøllmoen, 2005). Norwegian glacier termini farther inland have retreated continuously at a more moderate rate. Storglaciären, a poly-thermal glacier in Northern Sweden, lost 8.3 m (22% of the average thickness) of the cold surface layer between 1989 and 2001, primarily from increased wintertime temperatures yielding a longer melt season; summer ablation was normal (Pettersson et al., 2003). As with coastal Scandinavia, glaciers in the *New Zealand Alps* advanced during the 1990s, but have started to shrink since 2000. Increased precipitation may have caused the glacier growth, perhaps associated with more-frequent El Niño events (Chinn et al., 2005).

In the *European Alps*, exceptional mass loss during 2003 removed an average of 2500 kg m^{-2} over 9 measured Alpine glaciers, almost 60% higher than the previous record of 1.6 m we loss in 1996 and four times more than the mean loss from 1980 to 2001 (600 kg m^{-2}) (Zemp et al., in press). This was caused by extraordinarily high air temperatures over a long period, extremely low precipitation, and albedo feedback from Sahara dust depositions and a previous series of negative mass balance years.

Whereas *Himalayan* glaciers have generally shrunk at varying rates (Solomina et al., 2004; Su and Shi, 2002; Wang et al., 2004), several high glaciers in the central *Karakoram* are reported to have advanced and/or thickened at their tongues (Hewitt, 2005), probably due to enhanced transport of moisture to high altitudes.

Tropical Glaciers have shrunk from a mid 19th Century maximum, following the global trend (Figure 4.5.4). Strong shrinkage rates in the 1940s were followed by relatively stable extents that lasted into the 1970s. Since then, shrinkage has become stronger again; as in other mountain ranges, the smallest glaciers are more strongly affected. Tropical glaciers, being in principle very sensitive to both temperature changes and those related to atmospheric moisture, have shrunk mostly in response to regional changes in atmospheric moisture content and related energy and mass balance variables such as solar radiation, precipitation, albedo, and sublimation during the 20th century. Inter-annual variation in moisture seasonality, which is tied to sea surface temperature anomalies and related atmospheric circulation modes, strongly dominates the behaviour of tropical glaciers (Francou et al., 2004; Francou et al., 2003; Kaser, 2001; Kaser and Osmaston, 2002; Mölg and Hardy, 2004; Mölg et al., 2003; Wagnon et al., 2001). Glaciers on Kilimanjaro behave exceptionally (Figure 4.5.4). Even though the thickness of the tabular ice on the summit plateau has not changed dramatically over the 20th century, the ice has shown an incessant retreat of the vertical ice walls at its margins, for which solar radiation is identified as the main driver (Mölg et al., 2003). The mass balance on the horizontal top ice surfaces is governed by precipitation amount and frequency and associated albedo (Mölg and Hardy, 2004), and has sporadically reached positive annual values even in recent years (Thompson et al., 2002). In contrast to the plateau ice, the shrinkage of the glaciers on Kilimanjaro's slopes is constantly decelerating.

[INSERT FIGURE 4.5.4 HERE]

4.5.4 Changing Runoff from Glaciers and Ice Caps

Observations show that glaciers significantly modify stream flow in quantity, variability and timing by temporarily storing water as snow and ice. Annual basin runoff is enhanced or decreased in years of negative or positive mass balances, respectively (Hock et al., 2005). Year-to-year runoff variability is reduced to a minimum at moderate (~10 to 40%) basin ice coverage. Glacier discharge shows pronounced melt-induced diurnal and seasonal cyclicity, the latter beneficial to many areas since glacier meltwater is typically released during periods of otherwise low flow conditions. The effects of glacier wastage on glacier runoff include initial increases in total glacier runoff and peak flows, and considerable amplification of diurnal melt runoff amplitudes, followed by significantly diminished runoff totals and diurnal amplitudes as the glaciers continue to shrink. Figure 4.5.5 shows the example of Vernagtferner, Austrian Alps. Moderate summer discharge totals, peak flows and diurnal amplitudes during the 1970s, when positive mass balances prevailed, were significantly enhanced by the end 1990s due to a two-decade period of continuous glacier mass loss in general and a loss of the firn cover in particular. Effects are particularly strong on water availability in parts of the low latitude Andes (Kaser et al., 2003) and the Himalaya and the semi-arid central Eurasia.

[INSERT FIGURE 4.5.5 HERE]

4.6 Changes and Stability of Ice Sheets and Ice Shelves

4.6.1 Background

The ice sheets of Greenland and Antarctica hold enough ice to raise sea level about 64 m if fully melted (Lythe and Vaughan, 2001). Even a modest change could strongly affect future sea-level and freshwater flux to the oceans, with possible climatic implications. These ice sheets consist of vast central reservoirs of slow-moving ice drained by ice-walled ice streams or rock-walled outlet glaciers flowing rapidly into floating ice shelves or narrower ice tongues, or directly into the ocean; few terminate on land. Ice shelves often form in embayments, or run aground on local bedrock highs to form ice rumples or ice rises, and friction with embayment sides or local grounding points helps restrain the motion of the ice shelves and their tributaries. About half of the ice lost from Greenland is by surface melting and runoff into the sea, but there is little surface melting in Antarctica. Dynamics of the slow-moving ice and of ice shelves are reasonably well understood and can be modeled adequately, but this is not so for ice streams and outlet glaciers. Until recently (including the TAR), it was assumed that velocities of these glaciers cannot change very rapidly, and impacts of climate change were estimated primarily as changes in snowfall and surface melting. Recent observations show that glacier speeds can change rapidly, for reasons that are still under investigation. Consequently, this assessment will not adequately quantify such effects.

4.6.2 Mass Balance of the Ice Sheets and Ice Shelves

The current state of balance of the Greenland and Antarctic ice sheets is discussed here, focusing on the substantial progress made since the Third Assessment Report (TAR) of the IPCC (2001). Possible future changes are considered in Chapter 10, and in Chapter 19 of WGII.

4.6.2.1 Techniques

Several techniques are used to measure the mass balance of large ice masses. The mass-budget approach compares input from snow accumulation with output by ice flow and meltwater runoff. Repeated altimetry assesses volume changes. Time variations in gravity over the ice sheets reveal mass changes. Changes in length of day and in the direction of the Earth's rotation axis also reveal mass redistribution.

(i) **Mass-budget**

Snow accumulation is often estimated from annual layering in ice cores, with interpolation between core sites from satellite microwave measurements or radar sounding (Jacka et al., 2004). Increasingly, atmospheric-modelling techniques are also applied (e.g., Monaghan et al., in press). Ice discharge is calculated from radar or seismic measurements of ice thickness, and from in situ or remote measurements of ice velocity, usually where the ice begins to float and velocity is nearly depth-independent. A major advance since the TAR has been widespread application of interferometric synthetic aperture radar (InSAR) techniques to measure ice velocity over very large areas of the ice sheets (e.g., Rignot et al., 2005). Calculation of mass discharge also requires estimates for runoff of surface meltwater, which is large for low-elevation regions of Greenland and parts of the Antarctic Peninsula but small or zero elsewhere on the ice sheets. Surface-melt amounts usually are estimated from modelling driven by atmospheric reanalyses, global models or climatology, and often calibrated against surface observations where available (e.g., Hanna et al., 2005; Box et al., in press). The typically small mass loss by melting beneath grounded ice is usually estimated from models. Mass loss from melting beneath ice shelves can be large, and is difficult to measure; it is generally inferred from differences between mass input and output.

Ice-sheet mass inputs and outputs are large and difficult to estimate with high accuracy. For example, van de Berg et al. (in press) summarized six estimates of net accumulation on the grounded section of Antarctica published between 1999 and 2006, which ranged from 1811 to 2076 Gt a^{-1} or ±7% about the midpoint. (1 Gt = 1 Pg, and transfer of 360 Gt of grounded—non-floating—ice to the ocean would raise sea level ~1 mm.)

Greenland accumulation rate is probably uncertain by about 5%. Although broad InSAR coverage and progressively improving estimates of grounding-line ice thickness have substantially improved ice-discharge estimates, incomplete data coverage implies uncertainties in discharge estimates of a few percent. 5% uncorrelated errors on input and output would imply mass-budget uncertainties of about 40 Gt a^{-1} for Greenland and 140 Gt a^{-1} for Antarctica. Large interannual variability and trends also complicate interpretation. Box et al. (in press) estimated average accumulation on the Greenland ice sheet of 543 Gt a^{-1} from 1988-2004, but with an annual minimum of 482 Gt a^{-1}, a maximum of 613 Gt a^{-1}, and a best-fit linear trend yielding increase of 68 Gt a^{-1} during the period. Glacier velocities can change substantially, sometimes in months or years, adding to the overall uncertainty of mass-budget calculations.

(ii) Repeated altimetry

Surface-elevation changes reveal ice-sheet mass changes after correction for changes in depth-density profiles and in bedrock elevation, or for hydrostatic equilibrium if the ice is floating. Satellite radar altimetry (SRALT) has been widely used to estimate elevation changes (Shepherd et al., 2002; Davis et al., 2005; Johannessen et al., 2005; Zwally et al., in press), together with laser altimetry from airplanes (Krabill et al., 2004) and from ICESat (Thomas et al., in press). Modeled corrections for isostatic changes in bedrock elevation are small (few mm a^{-1}), but with uncertainties nearly as large as the corrections in some cases (Zwally et al., in press). Corrections for near-surface firn density changes are larger (>10 mm a^{-1}) (Cuffey, 2001) and also uncertain.

Radar altimetry provides long-term and widespread coverage for more than a decade, but with important challenges. The available SRALT data are from altimeters with a beam width of 20 km or more, designed and demonstrated to make accurate measurements over the almost flat, horizontal ocean. Data interpretation is more complex over sloping and undulating ice-sheet surfaces with spatially and temporally varying dielectric properties and thus penetration into near-surface firn. Empirical corrections are applied for some of these effects, and for inter-satellite biases. The correction for the offset between the ERS-1 and ERS-2 altimeters is reported by Zwally et al. (in press) to affect mass-change estimates for the interval 1992–2002 by ~50 Gt a^{-1} for Greenland, and to differ from the corresponding correction of Johannessen et al. (2005) by about 20 Gt a^{-1}, although some of this difference may reflect differences in spatial coverage of the studies combined with spatial dependence of the correction. Changes in surface dielectric properties affect the returned waveform and thus the measured range, so a correction is made for elevation changes correlated to returned-power changes. This effect is small averaged over an ice sheet but often of the same magnitude as the remaining signal at a point, and could remove signal if climate change affected both elevation and surface character, hence returned power.

SRALT tracking algorithms use leading edges of reflected radar waveforms, thus primarily sampling higher-elevation parts of the large footprint. This probably introduces only small errors over most of an ice sheet, where surfaces are nearly flat. But glaciers and ice streams often flow in surface depressions that can be narrower than the radar footprint, so that SRALT-derived elevation changes are weighted toward slower-moving ice at the glacier sides. This is of most concern in Greenland, where other studies show thinning along outlet glaciers just a few kilometres wide (Abdalati et al., 2001). SRALT elevation-change estimates have not been validated against independent data except at higher elevations, where surfaces are nearly flat and horizontal and dielectric properties nearly unchanging (Thomas et al., 2001). SRALT coverage is lacking within 900 km of the poles, and data reliability is limited in steep regions, but coverage has now been achieved for about 90% of the Greenland ice sheet and 80% of the Antarctic ice sheet (Zwally et al., in press).

Laser altimeters reduce some of the difficulties with SRALT by having negligible penetration of near-surface layers and a smaller footprint (about 1 m for airborne laser, and 60 m for ICESat). However, clouds limit data acquisition, and accuracy is affected by atmospheric conditions and particularly by laser-pointing errors. Airborne surveys over Greenland in 1993/1994 and 1998/1989 yielded estimates of elevation change accurate to ±14 mm a^{-1} along survey tracks (Krabill et al., 2002); however, the large gaps between flight lines must be filled, often by simple interpolation in regions of weak variability or by interpolation using physical models in more-complex regions (Krabill et al., 2004).

(iii) Geodetic measurements, including measurement of temporal variations in Earth gravity

Since 2002, the GRACE mission is providing routine measurement of the Earth's gravity field and its temporal variability. After removing the effects of tides, atmospheric loading etc., high-latitude data contain information on temporal changes in the mass distribution of the ice sheets and underlying rock (Velicogna and Wahr, 2005). Estimates of ice-sheet mass balance are sensitive to modeled estimates of bedrock vertical motion. Velicogna and Wahr estimated a correction for Greenland ice-sheet mass balance of 5 ± 17 Gt a^{-1} for the bedrock motion; Antarctica, which is a few times larger, will likely have a few times larger uncertainty in the correction. (Note that stated uncertainties for ice-sheet mass balances reported here and below are those from the published papers. Some papers include error terms that were estimated without formal statistical derivations, and other papers note omission of estimates for certain possible systematic errors.)

Other geodetic data provide constraints on mass changes in the high latitudes. These data include the history of changing length of day from eclipse records, the related ongoing changes in the spherical-harmonic coefficients of the geopotential, and changes in the rotation vector of the planet (true polar wander; Peltier, 1998; Munk, 2002; Mitrovica et al., in press). At present, unique solutions are not possible from these techniques, but hypothesized histories of ice-sheet changes can be tested against the data for consistency, and progress is being made rapidly.

4.6.2.2 Measured balance of the ice sheets and ice shelves

Mass balance of the large ice sheets was summarized by Rignot and Thomas (2002) and Alley et al. (2005).

(i) **Greenland**

Many recent studies have addressed Greenland mass balance. They yield a broad picture (Figure 4.6.1) of slight inland thickening (Thomas et al., 2001; Johannessen et al., 2005; Zwally et al., 2006), and strong near-coastal thinning, primarily in the south along fast-moving outlet glaciers (Abdalati et al., 2001; Rignot and Kanagaratnam, 2006) with overall accelerating shrinkage.

[INSERT FIGURE 4.6.1 HERE]

Analysis of GRACE data showed total losses of 75 ± 26 Gt a^{-1} between April, 2002 and July, 2004 (Velicogna and Wahr, 2005). Because of the low spatial resolution of GRACE, this includes losses from isolated mountain glaciers and ice caps near the coast, whereas the other results discussed next do not.

Mass loss from the ice-sheet surface (net snow accumulation minus meltwater runoff) has increased recently. Box et al. (in press) used calibrated atmospheric modelling and a single approximation for ice-flow discharge to estimate average ice-sheet mass loss of more than 100 Gt a^{-1} during 1988–2004; they also found acceleration of surface mass loss during this interval of 43 Gt a^{-1}. A similar analysis by Hanna et al. (2005) for 1961–2003 found somewhat higher net accumulation but similar trends, with ice-sheet growth of 22 ± 51 Gt a^{-1} from 1961–1990, shifting to shrinkage of 14 ± 55 Gt a^{-1} from 1993–1998, and shrinkage of 36 ± 59 Gt a^{-1} from 1998–2003. Again, ice-flow acceleration was not included in these estimates.

In a study especially using SRALT but incorporating laser-elevation measurements from aircraft and a correction for the effect of changing temperature on near-surface density, Zwally et al. (in press) estimated slight growth of the ice sheet by 11 ± 3 Gt a^{-1} from 1992–2002. However, they noted that mass loss of 18 ± 2 Gt a^{-1} would be indicated if the thickness changes at higher elevations are largely low-density firn rather than high-density ice, as might apply if increasing accumulation rate were also taken into account (Box et al., in press; Hanna et al., 2005). The more spatially limited results of Johannessen et al. (2005) from the same radar dataset indicate slightly less shrinkage or slightly more growth than those of Zwally et al. (2006) in regions of overlap. Krabill et al. (2000) also found slight thickening of central regions (~10 mm a^{-1}) from laser measurements covering 1993/1994 to 1998/1999.

Krabill et al. (2004) used repeat laser altimetry and modelled surface mass balance to estimate mass loss of about 45 Gt a^{-1} from 1993/1994 to 1998/1999, with acceleration to loss of 73 ± 11 Gt a^{-1} during the overlapping interval 1997 to 2003. These values may underestimate total losses, because they probably do not take account of rapid thinning in sparsely-surveyed regions such as the southeast, where mass-budget studies show large losses (Rignot and Kanagaratnam, 2006).

Rignot and Kanagaratnam (2006) combined several data sets, with special focus on the acceleration in velocity of outlet glaciers measured by SAR interferometry. Starting from an estimated excess ice-flow discharge of 51 ± 28 Gt a^{-1} in 1996, these authors estimated that the ice-flow loss increased to 83 ± 27 Gt a^{-1} in 2000 and 150 ± 36 Gt a^{-1} in 2005. Adding surface-mass-balance deviations from the long-term average as calculated by Hanna et al. (2004) yielded mass losses of 82 ± 28 Gt a^{-1} in 1996, 124 ± 28 Gt a^{-1} in 2000, and 202 ± 37 Gt a^{-1} in 2005. The more-pronounced ice-flow accelerations were restricted to regions south of 66°N before 2000 but extended to 70°N by 2005. These estimates of rapid mass loss would be reduced somewhat if ice-surface velocities are higher than depth-averaged velocities, which may apply in some places.

[INSERT FIGURE 4.6.2 HERE]

Greenland ice sheet mass-balance estimates are summarized in Figure 4.6.2. Most results indicate accelerating mass loss from Greenland during the 1990s and to 2005. The different estimates are not fully independent (there is, for example, some commonality in the isostatic corrections used for GRACE and altimetry estimates, and other overlaps can be found), but sufficient independence remains to increase confidence in the result. Different techniques have not fully converged quantitatively, with mismatches larger than formal error estimates suggesting structural uncertainties in the analyses, some of which were discussed above. The SRALT results showing overall balance or slight thickening, in contrast to other estimates, may result from the SRALT limitations over narrow glaciers discussed earlier.

Assessment of the data and techniques suggests mass balance of the Greenland Ice Sheet ranging between growth by 25 Gt a^{-1} and shrinkage by 60 Gt a^{-1} for 1961–2003, shrinkage by 50 to 100 Gt a^{-1} for 1993–2003 and by even higher rates between 2003 and 2005. Formal error bounds are not warranted in light of the differences between estimates from different techniques. Moreover, interannual variability is very large, driven mainly by variability in summer melting, but also by sudden glacier accelerations (Rignot and Kanagaratnam, 2006). Consequently, the short time interval covered by instrumental data is of concern in separating fluctuations from trends.

(ii) Antarctica

Rignot and Thomas (2002) combined several data sets including improved estimates of glacier velocities from InSAR to obtain Antarctic mass-budget estimates. For East Antarctica, growth of 20 ± 21 Gt a^{-1} was indicated, with estimated losses of 44 ± 13 Gt a^{-1} from West Antarctica. The balance of the Antarctic Peninsula was not assessed. Combining the East and West Antarctic numbers yields loss of 24 ± 25 Gt a^{-1} for the region monitored. The time interval covered by these estimates is not tightly constrained, because ice input was estimated from data sets of varying length; output data were determined primarily in the few years before 2002.

Zwally et al. (in press) obtained SRALT coverage of ~80% of the ice sheet, including some portions of the Antarctic Peninsula, and interpolated to the rest of the ice sheet. The resulting balance included West Antarctic loss of 47 ± 4 Gt a^{-1}, East Antarctic gain of 17 ± 11 Gt a^{-1}, and overall loss of 30 ± 12 Gt a^{-1}. If all the ice-thickness changes were low-density firn rather than ice, the loss would be smaller, at 14 ± 5 Gt a^{-1}. Davis et al. (2005) compiled SRALT data for ~70% of the ice sheet, and did not interpolate to the rest. The same pattern of East Antarctic thickening and West Antarctic thinning was observed (Figure 4.6.3). Davis et al. (2005) suggested that the East Antarctic change was primarily from increased snowfall. Assigning all ice-thickness change to low-density firn produces growth of the monitored portions of the ice sheet by 45 ± 8 Gt a^{-1}; if all change were ice, this growth would be 105 ± 20 Gt a^{-1}. Following the suggestion that the East Antarctic changes are from increased snow accumulation and the West Antarctic changes are more likely to be ice-dynamical would yield growth of monitored regions of 33 ± 9 Gt a^{-1}. Notice, however, that Monaghan et al. (in press) did not find the strong increase in snow accumulation suggested by Davis et al. (2005) in arguing for use of low-density firn in East Antarctic changes.

[INSERT FIGURE 4.6.3 HERE]

Using input-output techniques, Rignot et al. (2005) documented discharge 84 ± 30% larger than accumulation rate for the glaciers that fed the former Wordie Ice Shelf on the west coast of the northern Antarctic Peninsula, a region largely absent from the SRALT studies. Consideration of strong imbalances in glaciers feeding the former Larsen B ice shelf across the Peninsula, and extrapolation of the results to undocumented basins, suggested mass loss from the ice of the northern part of the Antarctic Peninsula of 42 ± 7 Gt a^{-1}. Observation of widespread glacier-front retreat in the region (Cook et al., 2005) motivates the extrapolation, although mass loss would be overestimated if snow accumulation has been systematically underestimated (van de Berg et al., in press).

Taking the Rignot and Thomas (2002), Zwally et al. (in press), and Rignot et al. (2005) results as providing the most-complete Antarctic coverage suggests ice-sheet thinning of ~60 Gt a^{-1}, with uncertainty of similar magnitude to the signal. Consideration of acceleration of some near-coastal glaciers, discussed below, and the difficulty of SRALT sampling of such regions, might allow slightly faster mass loss. The time interval considered is not homogeneous; the Rignot et al. (2005) results include changes after the collapse of the Larsen B ice shelf in 2002, younger than data in the other studies, and suggest the possibility of accelerating mass loss. Use of the more spatially restricted Davis et al. (2005) SRALT data rather than the Zwally et al. (in press) results illustrates the persistent uncertainties; depending on the assumed density structure of the changes, Davis et al. (2005) combined with the Rignot et al. (2005) estimate for the Antarctic Peninsula would suggest near-balance or Antarctic growth.

Trends in gravity data measured by the GRACE satellite indicate mass loss from the Antarctic ice sheet, including ice in the Antarctic Peninsula and in small glaciers and ice caps nearby (Velicogna and Wahr, 2006). GRACE data indicate loss of 139 ± 73 Gt a^{-1} between April, 2002 and July, 2005. Near-balance is indicated for East Antarctica, at 0 ± 51 Gt a^{-1}, with mass loss in West Antarctica of 136 ± 21 Gt a^{-1}.

Assessment of the data and techniques suggests overall Antarctic ice-sheet mass balance ranging from growth by 50 Gt a^{-1} to shrinkage by 200 Gt a^{-1} from 1993–2003. Formal error bars are not warranted, and there is no implication that the midpoint of this range provides the best estimate. Lack of older data complicates a similar estimate for the period 1961–2003. Acceleration of mass loss is likely to have occurred, but not so dramatically as in Greenland. Considering the lack of estimated strong trends in accumulation rate, assessment of the possible acceleration and of the slow time scales affecting central regions of the ice sheets, it is reasonable to estimate that the behavior from 1961–2003 falls between ice-sheet growth by 100 Gt a^{-1} and shrinkage by 200 Gt a^{-1}.

Simply summing the 1993–2003 contributions from Greenland and Antarctica produces a range from balance (0 Gt a^{-1}) to shrinkage by 300 Gt a^{-1}, or contribution to sea-level rise of 0 to 0.8 mm a^{-1}. For 1961–2003, the same calculation spans growth by 125 Gt a^{-1} to shrinkage by 260 Gt a^{-1}. Geodetic data on Earth rotation and polar wander provide additional insight (Peltier, 1998). Munk (2002) suggested that the geodetic data did not allow much contribution from ice sheets, but subsequent reassessment of the errors involved in some of the data sets and analyses allows an anomalous late-20th century sea-level rise of up to ~1 mm a^{-1} (360 Gt a^{-1}) from melting of land ice (Mitrovica et al., in press). Estimated mountain-glacier contributions do not supply this, so a contribution from the polar ice sheets is consistent, although little change in polar ice is also consistent.

(iii) Ice shelves

Most ice shelves are in Antarctica, where they cover an area of ~1.5M km^2, or 11% of the entire ice sheet, and where nearly all ice streams and outlet glaciers flow into ice shelves. By contrast, Greenland ice shelves occupy only a few thousand km^2, and many are little more than floating glacier tongues. Mass loss by surface-meltwater runoff is not important for most ice-shelf regions, which lose mass primarily by iceberg calving and basal melting, although basal freeze-on occurs in some regions.

Developments since the TAR include improved velocity and thickness data to estimate fluxes, and interpretation of repeated SRALT surveys over ice shelves to infer thickening/thinning rates. Melting of up to tens of m a^{-1} has been estimated beneath deeper ice near grounding lines (Rignot and Jacobs, 2002; Joughin and Padman, 2003). Significant changes are observed on most ice shelves, with both positive and negative trends, and with faster changes on smaller shelves. Overall, Zwally et al. (in press) estimated mass loss from ice shelves fed by glaciers flowing from West Antarctica of 95 ± 11 Gt a^{-1}, and mass gain to ice shelves fed by glaciers flowing from East Antarctica of 142 ± 10 Gt a^{-1}. Rapid thinning of more than 1 m a^{-1}, and locally more than 5 m a^{-1}, was observed between 1992 and 2001 for many small ice shelves in the Amundsen Sea and along the Antarctic Peninsula. Thinning of ~1 m a^{-1} (Shepherd et al., 2003; Zwally et al., in press) preceded the fragmentation of almost all (3300 km^2) of the Larsen-B ice shelf along the Antarctic Peninsula in fewer than 5 weeks in early 2002 (Scambos et al., 2003).

4.6.3 Causes of Changes

4.6.3.1 Changes in snowfall and surface melting

For Greenland, modelling driven by reanalysis data and calibrated against surface observations indicates recent increases in temperature, precipitation minus evaporation, surface meltwater runoff, and net mass loss from the surface of the ice sheet, as well as areal expansion of melting and reduction in albedo (Hanna et al., 2005; in press; Box et al., in press). High interannual variability means that many of the trends are not highly significant, but the trends are supported by the consistency between the various component data sets and of results from different groups. The best estimates of net snowfall minus meltwater runoff include an increase in the Greenland contribution to sea-level rise of 58 Gt a^{-1} between the 1961–1990 and 1998–2003 intervals (Hanna et al., 2005), or of 43 Gt a^{-1} from 1998-2004 (Box et al., in press).

For Antarctica, the recent summaries by van de Berg et al. (in press), van den Broeke et al (2006) and Monaghan et al. (in press) have updated trends in accumulation rate. Contrary to some earlier work, these new studies find no continent-wide significant trends in accumulation over the interval 1980–2004 (van de Berg et al.; in press; van den Broeke et al., 2006) or 1985–2001 (Monaghan et al., in press) from atmospheric reanalysis products (NCEP, ECMWF, Japanese), or from two mesoscale models driven by ECMWF and one by NCEP reanalyses. Strong interannual variability is found, approaching 5% for the continent, and important regional and seasonal trends that fit into larger climatic patterns, including an upward trend in accumulation in the Antarctic Peninsula. Studies of temperature trends similarly show regional changes (Schneider et al., 2004) including the strong warming in the Antarctic Peninsula region. A recent reanalysis of data poleward of 50°S from 1958 to 2002 shows overall warming for land, ocean, and the whole domain over that interval (Chapman and Walsh, in press). Trends are not especially large, typically less than 0.3°C/decade, both regional warming and cooling are observed, different seasons show different behavior, and the strongest trends are the wintertime warming over the Antarctic Peninsula. Furthermore, the trends are dependent on the intervals considered, with overall trend analyses started between 1966 and 1982 showing cooling but the full interval showing warming. Thus, the results are consistent with the cooling reported by Thompson and Solomon (2002) owing to the different interval of coverage (also see van den Broeke, 2000; Vaughan et al., 2001; Turner et al., 2005; Doran et al., 2002).

4.6.3.2 Ongoing dynamic ice sheet response to past forcing

Because some portions of ice sheets respond only slowly to climate changes (decades to thousands of years or longer), past forcing may be influencing ongoing changes. A comprehensive attempt to discern such long-term trends contributing to recently measured imbalances was made by Huybrechts (2002) and Huybrechts et al. (2004). They found little long-term trend in volume of the Greenland Ice Sheet, but a trend of Antarctic shrinkage of about 90 Gt a^{-1}, primarily because of post-ice-age retreat of the West Antarctic grounding line. This trend is modelled to largely disappear over the next millennium. Most of the sensitivity studies by Huybrechts (2002) produced such a thinning trend, but one produced an opposite trend at present; in addition, simulated trends for today were highly dependent on the poorly known timing of grounding-line retreat in West Antarctica. Moreover, the ice-flow model does not include the full stress solution for ice shelves, ice streams and outlet glaciers, nor full interaction between ice shelves and the ocean because of lack of knowledge of oceanic changes. Strong conclusions about long-term trends are probably not warranted. The long-term sea-level trends discussed in chapter 6, and geodetic data discussed above, probably are consistent with the trend suggested by Huybrechts et al. (2004) but not with a much larger trend. Some geologic data also support recent and perhaps ongoing Antarctic mass loss (e.g., Stone et al., 2003).

4.6.3.3 Dynamic response to recent forcing

Numerous papers since the TAR have documented rapid changes in marginal regions of the ice sheets. Attention has especially focused on ice-flow accelerations of glaciers along the Antarctic Peninsula (Scambos et al., 2004; Rignot et al., 2004; Rignot et al., 2005), the glaciers draining into Pine Island Bay and nearby parts of the Amundsen Sea from West Antarctica (Thomas et al., 2004; Shepherd et al., 2004), and Jakobshavn Glacier (Thomas et al., 2003; Joughin et al., 2004) and other glaciers south of about 70°N in Greenland (Howat et al., 2005; Rignot and Kanagaratnam, 2006). Important accelerations may have occurred in some coastal parts of East Antarctica (Zwally et al., in press), and ice-flow slowdown has been observed on Whillans and Bindschadler Ice Streams on the Siple Coast of West Antarctica (Joughin and Tulaczyk, 2002). Rignot and Kanagaratnam (2006) estimated that ice-discharge increase in Greenland caused mass loss in 2005 to be about 100 Gt a^{-1} larger than in 1996; consideration of the changes in the Amundsen Sea and Antarctic Peninsula regions of West Antarctica (and the minor opposing trend on Whillans and Bindschadler Ice Streams) suggests a similar-magnitude Antarctic signal, although with greater uncertainty and occurring perhaps over a longer interval (Joughin and Tulaczyk, 2002; Thomas et al., 2004; Rignot et al., 2005; van den Broeke et al., 2006).

Most of the other coastal changes appear to have involved inland acceleration following reduction or loss of ice shelves. Very soon after breakup of the Larsen B ice shelf along the Antarctic Peninsula, the speeds of tributary glaciers increased up to 8-fold, but with little change in velocity of adjacent ice still buttressed by remaining ice shelf (Rignot et al., 2004; Scambos et al., 2004). Thinning and breakup of the floating ice tongue of Jakobshavn Glacier were accompanied by approximate doubling of the ice flow velocity (Thomas et al., 2003; Thomas; 2004; Joughin et al., 2004). Ice-shelf thinning has occurred with the speed-up of tributary glaciers entering the Amundsen Sea (Shepherd et al., 2002; 2004; Joughin et al., 2003).

Because of drag between ice shelves and embayment sides or localized regrounding points on sea-bed topographic highs, shortening or thinning of ice shelves is expected to speed ice flow (Mercer, 1978; Thomas, 1979), with even small ice shelves potentially important (Dupont and Alley, in press). Targeted models addressing speed-up of particular glaciers in response to ice-shelf reduction are capable of simulating the observed time scales (notable changes in years or less) and patterns of change (largest thinning and speed-up near the coast, decreasing inland and following ice streams) (Payne et al., 2004; Dupont and Alley, 2005). Comprehensive-model runs for ice-sheet behaviour over the last century, using known forcings and flow processes but omitting full stress coupling with ice shelves and poorly-known details of sub-ice-shelf oceanographic changes, match overall ice-sheet trends rather well (Huybrechts et al., 2004), but fail to show these rapid marginal thinning events. This suggests that the changes are in response to processes (either forcings from ocean-temperature or ocean-circulation changes, or ice-flow processes) not included in the comprehensive modelling.

The acceleration of Helheim Glacier, Greenland may be akin to ice-shelf-linked changes. Enhanced calving removed not-quite-floating ice at Helheim, reducing restraint on the remaining ice and allowing faster flow (Howat et al., 2005).

Other ice-flow changes have occurred that are not linked to ice-shelf reduction. The Siple Coast, Antarctica changes likely reflect inherent flow variability rather than recent forcing (Parizek et al., 2003). Zwally et al. (2002) showed for one site near the equilibrium line on the west coast of Greenland that the velocity of comparatively slow-moving ice increased just after seasonal onset of drainage of surface meltwater into the ice sheet, and that greater meltwater input produced greater ice-flow speed-up. The total speed-up was not large (order of 10%), but the effect is not included in most ice-flow models. Inclusion in one model (Parizek and Alley, 2004) somewhat increased the sensitivity of the ice sheet to future climate change, mostly beyond the year 2100. Much uncertainty remains, especially related to whether fast-moving glaciers and ice streams are similarly affected, and whether access of meltwater to the bed through more than 1 km of cold ice would migrate inland if warming caused surface melting to migrate inland (Alley et al., in press). This could thaw ice that is frozen to the bed, allowing faster flow through onset of basal sliding or subglacial sediment deformation. Data are not available to assess whether effects of increased surface melting in Greenland have been transmitted to the bed and contributed to ice-flow acceleration.

4.6.3.4. Melting and calving of ice shelves

Many of the largest and fastest ice-sheet changes thus appear to be at least in part responses to ice-shelf shrinkage or loss. Although ice-shelf shrinkage does not directly contribute to sea-level change because shelf ice is already floating, the very tight coupling to inland ice means that ice-shelf balance does matter to sea level. The available data suggest that the ice-shelf changes have resulted from environmental warming, with both oceanic and atmospheric temperatures important, although changes in oceanic circulation cannot be ruled out as important contributors.

The southward-progressing loss of ice shelves along the Antarctic Peninsula is consistent with a thermal limit to ice-shelf viability (Morris and Vaughan, 2003). Cook et al. (2005) found that no ice shelves exist on the warmer side of the –5°C mean annual isotherm, whereas no ice shelves on the colder side of the –9°C isotherm have broken up. Before the 2002 breakup of Larsen B ice shelf, local air temperatures had increased by more than 1.5°C over the previous 50 years (Vaughan et al., 2003), increasing summer melting and formation of large melt ponds on the ice shelf. These likely contributed to breakup by draining into and wedging open surface crevasses that linked to bottom crevasses filled with seawater (Scambos et al., 2000). Large ice-flow models do not accurately capture the physical processes involved in such dramatic iceberg calving, or in more common calving behavior.

Despite an increased ice supply from tributary glaciers, thinning of up to several meters per year has been measured for ice shelves on the Amundsen Sea coastline in the absence of large surface-mass-balance changes. This suggests that increased basal ice melting is responsible for the thinning (Shepherd et al., 2003; 2004). Similarly, the 15-km floating ice tongue of Jakobshavn Glacier survived air temperatures during the 1950s similar to or even warmer than those associated with thinning and collapse near the end of the century, implicating oceanic heat transport in the more-recent changes, although air-temperature increase may have contributed.

The basal mass balance of an ice shelf depends on temperature and ocean circulation beneath. Isolation from direct wind forcing means that the main drivers of sub-ice-shelf circulation are tidal and density (thermohaline) forces. Lack of knowledge of sub-ice bathymetry has hampered the use of three-dimensional models to simulate circulation beneath the thinning ice shelves. Both the west side of the Antarctic Peninsula and the Amundsen Sea coast are exposed to warm Circumpolar Deep Water (CDW) (Hellmer et al., 1998), capable of causing rapid ice-shelf basal melting. Increased melting in the Amundsen Sea is consistent with observed recent warming by 0.2°C of ocean waters seaward of the continental shelf break (Jacobs et al., 2002; Robertson et al., 2002). Simple regression analysis of available data including those from the Amundsen Sea indicated that 1°C warming of sub-ice-shelf waters increases basal melt rate by about 10 m a^{-1} (Shepherd et al., 2004).

Box 4.1: Ice Sheet Dynamics and Stability

The ice sheets of Antarctica and Greenland could raise sea level greatly. Central parts of these ice sheets change slowly, but near the coast rapid changes over quite large areas have been observed. In these areas, uncertainties about glacier basal conditions, ice deformation, and interactions with the surrounding ocean seriously limit the ability to make accurate projections.

Ice sheets are thick, broad masses of ice formed mainly from compaction of snow (Paterson, 1994). They spread under their own weight, transferring mass towards their margins where it is lost primarily by runoff of surface meltwater or by calving of icebergs into marginal seas or lakes. Water vapor fluxes (sublimation/condensation), and basal melting/freezing (especially beneath ice shelves) may also be important local processes of mass gain and loss.

Ice sheets flow by internal deformation, basal sliding, or a combination of both. Deformation in ice occurs through solid-state processes analogousto those involved in polycrystalline metals which are relatively close to their melting points. Deformation rates depend on the gravitational stress (which increases with ice thickness and with the slope of the upper surface), temperature, impurities, and size and orientation of the crystals (which in turn depend in part on the prior deformational history of the ice). While these characteristics are not completely known, slow ice flow by deformation can be modeled with reasonable accuracy.

For basal sliding to be an important component of the total motion, meltwater or deformable wet sediment slurries at the base are required for lubrication. While the central regions of ice sheets (typically above 2000 m elevation) seldom experience surface melting, the basal temperature may be raised to the melting point by heat from the earth's interior, delivered by meltwater transport, or from the "friction" of ice motion. Sliding velocities under a given gravitational stress can differ by orders of magnitude, depending on the presence or absence of unconsolidated sediment, the roughness of the substrate, and the supply and distribution of water.

Basal conditions are well-characterized in very few regions, introducing important uncertainties to the modeling of basal sliding.

Ice flow is often channeled into fast-moving ice streams (which flow between slower-moving ice walls) or outlet glaciers (with rock walls). Enhanced flow in ice streams arises either from higher gravitational stress linked to thicker ice in bedrock troughs, or from increased basal lubrication.

Cold ice discharged across the coast often remains attached to the ice sheet to become a floating ice shelf. An ice shelf moves forward, spreading and thinning under its own weight, and fed by snowfall on its surface and ice input from the ice sheet. Friction at ice-shelf sides and over local shoals slows the flow of the ice shelf and thus the discharge from the ice sheet. An ice shelf loses mass by calving icebergs from the front and by basal melting into the ocean cavity beneath. Estimates based on available data suggest a 1°C ocean warming could increase ice-shelf basal melt by 10 m per year, but inadequate knowledge of the bathymetry and circulation in the largely inaccessible ice shelf cavities restricts the accuracy of such estimates.

Ice deformation is nonlinear, increasing approximately proportional to the cube of the applied stress. Moreover, an increase in any of the six independent applied stresses (three stretching stresses and three shears) increases the deformation rate for all other stresses. For computational efficiency, most long simulations with comprehensive ice-flow models use a simplified stress distribution, but recent changes in ice-sheet margins and ice streams cannot be simulated with these models, demonstrating a need for resolving the full stress configuration. Development of such models is still in its infancy, with few results yet available.

Ice-sheets respond to environmental forcing on numerous time scales. A surface warming may take more than 10,000 years to penetrate to the bed and change temperatures there; a meltwater-filled crevasse might penetrate to the bed and affect the temperature locally within minutes. Ice velocity over most of an ice sheet changes slowly in response to changes in the ice sheet shape or surface temperature, but large velocity changes may occur rapidly on ice streams and outlet glaciers in response to changing basal conditions or changes in the ice shelves into which they flow.

The palaeo-record of previous ice ages indicates that ice sheets shrink in response to warming and grow in response to cooling. The data also indicate that shrinkage can be far faster than growth. Understanding of the processes suggests that this arises both because surface melting rates can be much larger than the highest snowfall rates, and because ice discharge may be accelerated by strong positive feedbacks (Paterson, 1994; Clark et al. 1999b). Thawing of the bed, loss of restraint from ice shelves, or changes in meltwater supply and transmission can increase flow speed greatly. The faster flow may then generate additional lubrication from frictional heating and from erosion to produce wet sediment slurries. Surface lowering as the faster flow thins the ice will enhance surface melting, and will reduce basal friction where the thinner ice becomes afloat. Despite competition from stabilizing feedbacks, warming-induced changes have led to rapid shrinkage and loss of ice sheets in the past, and remain of concern in considering the future.

4.7 Changes in Frozen Ground

4.7.1 Background

Frozen ground, in a broad sense, includes near-surface soil affected by short-term freeze/thaw cycles, seasonally frozen ground, and permafrost. In terms of the area extent, frozen ground is the single largest component, hence the most vulnerable part, of the cryosphere. The presence of frozen ground depends on the ground temperature that is controlled by the surface energy balance. While the climate is an important factor determining the distribution of frozen ground, local factors are also important such as vegetation conditions, snow cover, physical and thermal properties of soils, and soil moisture conditions. The permafrost temperature regime is a sensitive indicator of the decade-to-century climatic variability (Lachenbruch and Marshall, 1986; Osterkamp, 2005). Thawing of permafrost can lead to subsidence of the ground surface as masses of ground ice melt and to the formation of uneven topography know as thermokarst, generating dramatic changes in ecosystems, landscape, and infrastructure performance (Nelson et al., 2001; Walsh et al., 2005). Surface soil freezing and thawing processes play a significant role in the land-surface energy and moisture balance, hence in climate and hydrologic systems. Changes in permafrost and soil seasonal freezing/thawing processes have dramatic impacts on spatial patterns, seasonal to inter-annual variability and long-term trends in terrestrial carbon budgets and surface-atmosphere trace gas exchange, directly through biophysical controls on both photosynthesis and respiration, and indirectly through controls on soil nutrient availability.

4.7.2 Changes in Permafrost

4.7.2.1 Data sources

Measurements of permafrost temperature can go back as early as 1829 in Siberia. Systematic permafrost temperature monitoring started in the 1950s from both the standard hydrometeorological stations up to 3.2 m from the standard Russian Hydrometerological Stations (Zhang et al., 2001) and deep boreholes up to >100 m (Pavlov, 1996). The U. S. Geological Survey has measured permafrost temperatures from deep boreholes in northern Alaska since the 1940s (Lachenbruch and Marshall, 1986) and from shallow boreholes (generally <80 m) since the mid 1980s (Osterkamp, 2005). Deep permafrost temperature measurements on the Tibetan Plateau were conducted in the early 1960s, while continuous permafrost monitoring only started in the late 1980s (Zhao et al., 2003). Monitoring of deep permafrost temperatures started in the early 1980s in northern Canada (Smith S. et al., 2005) and in the 1990s in Europe through the Permafrost and Climate in Europe (PACE) program (Harris et al., 2003).

4.7.2.2 Changes in permafrost temperature

Permafrost in the Northern Hemisphere has experienced temperature increases in recent decades (Table 4.7.1). Permafrost surface temperature has in general increased about 2 to 4°C from the turn of the 20th century to the early 1980s on the North Slope of Alaska (Lachenbruch and Marshall, 1986) although at some sites there was little warming or even a cooling trend. Permafrost temperature has increased additional 2 to 3°C since the early 1980s (Osterkamp, 2005). Measurements (Osterkamp, 2003) and modelling results (see Hinzman et al., 2005; Walsh et al., 2005) indicate that permafrost temperature has increased up to 2°C in the Interior of Alaska since the 1980s. Changes in air temperature alone over the same period cannot account for the permafrost temperature increase, while increased snow cover may be responsible for a significant proportion of the permafrost surface temperature increase (Zhang, 2005).

Data from the Northern Mackenzie Valley in the continuous permafrost zone show that permafrost temperature between depths of 20 to 30 m has increased about 1°C in the 1990s (Smith S. et al., 2005). The magnitude of the temperature increase reduced significantly in the Central Mackenzie Valley and no significant trend of permafrost temperature change is observed in the Southern Mackenzie valley, where permafrost is thin (less than 10 to 15 m thick) and warmer than –0.3°C (Smith S. et al. 2005, Couture et al., 2003). The absence of a trend is likely due to the absorption of latent heat required for phase change. Similar results are reported for warm permafrost in the southern Yukon Territory (Haeberli and Burn, 2002). Cooling of permafrost was observed from the late 1980s to the early 1990s at a depth of 5 m at Iqaluit in the eastern Canadian Arctic. This cooling however, was followed by warming of 0.4°C per year between 1993 and 2000 (Smith S. et al., 2005). This trend is similar to that observed in Northern Quebec, where cooling of permafrost was observed between the mid 1980s and mid 1990s at a depth of 10 m (Allard et al., 1995) which was followed by warming beginning in 1996 (Brown et al., 2000).

Evidence of permafrost warming was also observed in the Russian Arctic. Permafrost temperature increased approximately 1°C at depths between 1.6 m to 3.2 m from the 1960s to the 1990s in East Siberia, about 0.3 to 0.7°C at depth of 10 m in northern West Siberia (Pavlov, 1996), and about 1.2 to 2.8°C at depth of 6 m from 1973 through 1992 in northern European Russia (Oberman and Mazhitova, 2001). Fedorov and Konstantinov (2003) reported that permafrost temperatures from three central Siberian stations did not increase between 1991 and 2000. Mean annual temperature in Central Mongolia in the recent 30 years at depth from 10 to 90 m increased 0.05 to 0.15^{o}C/decade (Sharkhuu, 2003).

Results from six years continuous ground temperature monitoring in the 100 m deep permafrost borehole on Janssonhaugen, Svalbard, indicate that the permafrost has warmed significantly, the mean annual ground surface temperature currently increasing at a rate of about 0.4°C/decade in the past 60 to 80 years (Isaksen et al., 2000). Results from five years of continuous ground temperature monitoring in Juvvasshøe, Southern Norway, indicate that the permafrost is currently also strongly warming. From 1999 to 2004, ground temperatures have increased by ~0.3°C at 15 m depth. Because at both these sites wind action prevents snow accumulation in winter, a close relationship is observed between air, ground surface, and ground subsurface temperatures, which makes the geothermal records from Janssonhaugen and Juvvasshøe powerful indicators of climate change. At the Murtèl-Corvatsch borehole, permafrost temperatures in 2001 and 2003 at a depth of 11.5m in ice-rich coarse frozen debris, were only slightly below –1°C, and were the highest since readings began in 1987 (Vonder Mühll et al., 2004). Analysis of the long-term thermal record from this site has shown that in addition to summer air temperatures, the depth and duration of snow cover, particularly in early winter, have a major influence on permafrost temperatures (Harris et al., 2003).

Permafrost temperature increased about 0.2 to 0.5°C from the 1970s to 1990s over the hinterland of the Tibetan Plateau (Zhao et al., 2003), up to 0.5°C along the Qinghai-Xizang Highway over a period from 1995 to 2002 (Wu and Liu, 2003; Zhao et al., 2004), and about 0.2 to 0.4°C from 1973 to 2002 in 16 to 20 m depths in Tianshan Mountain regions (Qiu et al., 2000; Zhao et al., 2004). Over the Da Hinggan Mountains in northeastern China, permafrost surface temperature increased about 0.7 to 1.5°C over a period from 1978 through 1991 from the valley bottom to the north-facing slopes (Zhou et al., 1996). Permafrost temperature at the depth of the zero annual temperature variation increased about 2.1°C on the valley bottom, 0.7°C on the north-facing slopes, and 0.8°C on south-facing slopes. In areas of the south-facing slopes where no permafrost exists, soil temperature at the bottom of the seasonally frozen ground increased about 2.4°C (Zhou et al., 1996).

Table 4.7.1. Recent Trends in Permafrost Temperature

Region	Depth (m)	Period of Record	Permafrost Temperature Change (°C)	Reference
United States				
Northern Alaska	~1	1910's–1980's	2-4	Lachenbruch and Marshall, 1986
Northern Alaska	20	1983–2003	2–3	Osterkamp, 2005
Interior of Alaska	20	1983–2003	0.5–1.5	Osterkamp, 2005
Canada				
Alert, Nunavut	15–30	1995–2000	0.9	Smith et al., 2003
Northern Mackenzie Valley	10–20	Mid-1980s–2003	1.1	Smith S. et al., 2005
Central Mackenzie Valley	10–20	Mid-1980s–2003	0.5	Smith S. et al., 2005
Southern Mackenzie Valley & Southern Yukon Territory	~20	Mid-1980s–2003	0	Smith S. et al., 2005
Northern Quebec	10	Late 1980s-mid-1990s	<–1	Allard et al., 1995
Northern Quebec	10	1996–2001	1.0	Allard et al., 1995
Lake Hazen	2.5	1994–2000	1.0	Broll et al., 2003
Iqaluit, Eastern Canadian Arctic	5	1993–2000	2.0	Smith S. et al., 2005
Russia				
East Siberia	1.6–3.2	1960–1002	~1.3	Walsh, 2005
Northern West Siberia	10	1980–1990	0.3–0.7	Pavlov, 1996
European north of Russia, continuous permafrost zone	6	1973–1992	1.6–2.8	Pavlov, 1996
Northern European Russia	6	1970–1995	1.2–2.8	Oberman and Mazhitova, 2001
Europe				
Juvvasshoe, Southern Norway	~5	Past 30–40 years	0.5 to 1.0	Isaksen et al., 2000
Janssonhaugen, Svalbard	~5	Past 60–80 years	1 to 2	Isaksen et al., 2000
Murtel-Corvatsch	11.5	1987–2001	1.0	Vonder Muhll et al., 2004
China				
Tibetan Plateau	~10	1970's–1990's	0.2-0.5	Zhao et al., 2004
Qinghai-Xizang Highway	3–5	1995–2002	Up to 0.5	Wu and Liu, 2003; Zhao et al., 2004
Tianshan Mountains	16-20	1973–2002	0.2–0.4	Qiu et al., 2000; Zhao et al., 2004
Da Hinggan Mountians, Northeastern China	~2	1978–1991	0.7–1.5	Zhou et al., 1996

4.7.2.3 Permafrost degradation

Permafrost degradation refers to a naturally or artificially caused decrease in the thickness and/or areal extent of permafrost. Evidence of change in the southern boundary of discontinuous permafrost zone in the past decades has been reported. In North America, the southern boundary has migrated northward in response to warming since the Little Ice Age, and continues to do so today (Halsey et al., 1995). In recent years, widespread permafrost warming and thawing have occurred on the Tibetan Plateau, China. Based on data from ground penetration radar and in-situ measurements, the lower limit of permafrost has moved up about 25 m from 1975 through 2002 on the north-facing slopes of the Kunlun Mountains (Nan et al., 2003). From Amdo to Liangdehe along the Qinghai-Xizang Highway, areal extent of permafrost islands decreased approximately 36% over the past three decades (Wang, 2002). Areal extent of taliks expanded about 1.2 km on both sides of the Tongtian River (Wang, 2002). Overall, the northern limit of permafrost retreated about 0.5 to 1.0 km southwards and the southern limit moved northwards about 1.0 to 2.0 km (Wu and Liu , 2003; Wang and Zhao, 1997).

When the warming at the permafrost surface eventually penetrates to the base of permafrost and the new surface temperature remains stable, thawing at the base of the ice-bearing permafrost occurs, especially for the thin discontinuous permafrost. At Gulkana, Alaska, basal thawing of permafrost is at an average rate of 0.04 m per year since 1992 (Osterkamp, 2003). Over the Tibetan Plateau, the basal thawing rate of about 0.01 to 0.02 m per year was observed since the 1960s (Zhao et al., 2003). It is expected that the basal thawing rate will accelerate over the Tibetan Plateau when current permafrost surface warming continues.

When the ice-rich permafrost thaws, the ground surface subsides; this downward displacement of the ground surface is called thaw settlement. Typically, thaw settlement does not occur uniformly over space, yielding a chaotic surface with small hills and wet depressions known as thermokarst terrain; this is particular common in areas underlain by ice wedges. On slopes, particularly in mountainous regions, thawing of ice-rich, near-surface permafrost layers can create mechanical discontinuities in the substrate, leading to active-layer detachment slides (Lewkowicz, 1992), which have a capacity for damage to structures similar to other types of rapid mass movements. Climate-induced thermokarst may have detrimental impacts on infrastructure built upon permafrost. Thermokarst processes pose a serious threat to arctic biota through either over-saturation or drying (Hinzman et al., 2005; Walsh et al., 2005). Extensive thermokarst development has been discovered near Council, Alaska (Yoshikawa and Hinzman, 2003) and in central Yakutia (Gavrilov and Efremov, 2003). Significant expansion and deepening of thermokarst lakes were observed near Yakutsk (Fedorov and Konstantinov, 2003) between 1992 and 2001. The largest subsidence rates of 17 to 24 cm/yr were observed in depressions holding young thermokarst lakes. Satellite data reveal that in the continuous permafrost zone of Siberia, total lake area increased by about 12% and lake number rose by 4% during the past three decades (Smith L. et al., 2005). Over the discontinuous permafrost zone, total area and lake number decreased by up to 9% and 13%, respectively, probably due to the lake water drainage through taliks.

The most sensitive regions of permafrost degradation are coasts with ice-bearing permafrost that are exposed to thermo-abrasion around the Arctic Ocean. The mean annual erosion rate varies from 2.5–3.0 m/yr for the ice-rich coasts to 1.0 m/yr for the ice-poor permafrost coast along the Russian Arctic Coast (Rachold et al., 2003). Over the Alaskan Beaufort Sea coast, the mean annual erosion rate ranges from 0.7 to 3.2 m/yr with maximum rate up to 16.7 m/yr (Jorgenson and Brown, 2005). The current circum-arctic coastal erosion results in a sediment flux of 430.8×10^6 t/yr and a total carbon flux of 6.69×10^6 t/yr into the Arctic Ocean. Lowering in permafrost stability and intensification of coastal erosion due to global warming would definitely increase sediment and carbon input to the Arctic Ocean, potentially causing considerable transformation of the Arctic coastal currents and circulation.

4.7.2.4. Subsea permafrost

Subsea permafrost (or offshore) permafrost refers to permafrost occurring beneath the seabed. It exists in continental shelves in the Polar Regions. At present, the thermal regime of subsea permafrost is primarily controlled by seawater temperature. Subsea permafrost has formed either in response to the negative mean annual sea-bottom temperature or as the result of inundation of terrestrial permafrost (Walsh et al., 2005). Subsea permafrost derives its importance from current interests in the development of offshore petroleum and other natural resources in the continental shelves of Polar Regions. Design, construction and operation of coastal facilities; structures founded on the seabed; subsea pipelines, and wells drilled for exploration and production must consider the presence and characteristics of subsea permafrost. Subsea permafrost contains or overlies large volumes of CH_4 in the form of gas hydrates at depths of up to several hundred meters. As subsea permafrost warms and thaws, destabilization of the gas hydrates could increase the flux of CH_4 to the atmosphere (Walsh et al., 2005).

4.7.3 Changes in Seasonally Frozen Ground

4.7.3.1 Changes in the active layer

The active layer is that the portion of the soil above permafrost that thaws and freezes seasonally and plays an important role in cold regions because most ecological, hydrological, biogeochemical, and pedogenic activity takes place within it (Kane et al., 1991). Changes in active layer thickness are influenced by many factors, including surface temperature, physical and thermal properties of the surface cover and substrate, vegetation, soil moisture, and duration and thickness of snow cover (Brown et al., 2000; Frauenfeld et al., 2004; Zhang et al., 2005). The inter-annual and spatial variations in thaw depth at point locations can be large, an artifact of year-to-year and microtopographic variations in both surface temperature and soil moisture. It is therefore important to obtain robust, replicate samples of active-layer thickness for monitoring purposes. When the other conditions remain constant, changes in active layer thickness could be expected to increase in response to the warming of climate, especially summer air temperature.

Long-term monitoring of the active layer has been conducted over the past several decades in Russia. By the early 1990s, there were about 25 stations, each containing 8–10 plots and 20–30 boreholes to depth 10–15 m for measuring ground temperatures (Pavlov, 1996). Measurements of soil temperature in permafrost have been carried out in the former Soviet Union from 31 stations, most of them started in the 1950s but a few as early as in the 1930s (Figure 4.7.1). Over the period 1956–1990, the active layer exhibited a statistically significant deepening by about 20 cm. Changes in air temperature and snow depth are responsible for the increase in active layer thickness.

[INSERT FIGURE 4.7.1 HERE]

The Circumpolar Active Layer Monitoring (CALM) program, now as part of the GTN-P program, was developed in the 1990s and currently incorporates more than 125 sites at locations including the Arctic, the Antarctic, and several midlatitude mountain ranges (Brown et al., 2000; Nelson, 2004; Figure 4.7.2). CALM is designed to observe the response of the active layer and near-surface permafrost to climate change. The results from northern high-latitude sites demonstrate substantial inter-annual and inter-decadal fluctuations in active layer thickness. The active layer responds consistently to forcing by air temperature on an inter-annual basis. During the mid- to late-1990s in Alaska and northwestern Canada, maximum and minimum thaw depth was observed in 1998 and in 2000, corresponding to the warmest and coolest summers, respectively. Evidence of increase in active layer thickness and thermokarst development are observed, indicating degradation of warmer permafrost (Brown et al., 2000). Evidence from the Permafrost and Climate in Europe (PACE) program indicates that active layer thickness has been the greatest in the summers of 2002 and 2003, approximately 20% greater than the previous years (Harris et al., 2003). Active layer thickness has increased by up to 1.0 m along the Qinghai-Xizang Highway over the Tibetan Plateau since the early 1980s (Zhao et al., 2004).

[INSERT FIGURE 4.7.2 HERE]

4.7.3.2 Seasonally frozen ground in non-permafrost area

Seasonally frozen ground refers to the top layer, which freezes and thaws annually regardless of whether there is underlain permafrost. Significant changes in seasonally frozen ground have been observed worldwide. The thickness of seasonally frozen ground has decreased by more than 0.30 m from 1956 through 1990 in Russia (Figure 4.7.1), primarily controlled by the increase in winter air temperature and snow depth (Frauenfeld et al., 2004).

Over the Tibetan Plateau, the thickness of seasonally frozen ground has decreased over a range from 0.05 to 0.22 m from 1967 through 1997 (Zhao et al., 2004). The driving force for the decrease in thickness of the seasonally frozen ground is the significant warming in cold seasons, while changes in snow cover depth plays a minor role. The duration of seasonally frozen ground shortened by more than 20 days from 1967 through 1997 over the Tibetan Plateau, mainly due to the earlier onset of thaw in spring rather than the late onset of freeze in autumn (Zhao et al., 2004).

Over the 20th century, there was less area where soil experienced seasonal freezing and thawing, especially in the late 20th century. The maximum extent of seasonally frozen ground has decreased by about 7% in the Northern Hemisphere since the mid-20th century, while in spring, the decrease in areal extent ranges up to 15% (Figure 4.7.3; Zhang et al., 2003). There was little change in the area extent of seasonally frozen ground during the early and mid winters.

[INSERT FIGURE 4.7.3 HERE]

4.7.3.3 Near-surface soil freeze-thaw cycle

Satellite remote sensing data have been used to detect the near-surface soil freeze/thaw cycle at regional and hemispheric scales. Evidence from the satellite passive microwave remote sensing record indicates that the onset dates of thaw in spring and freeze in autumn advanced five to seven days in Eurasia over the period 1988–2002, leading to a forward shift of the growing season but no change in its length (Smith et al., 2004). In North America, there was a trend toward late freeze dates in autumn by about five days that led, in part, to a lengthening of the growing season by eight days. Overall, the timing of the seasonal thawing and subsequent initiation of the growing season in early spring has advanced by approximately eight days from 1988 to 2001 for the pan-Arctic basin and Alaska (McDonald et al., 2004).

4.7.4 Consequences

Changes in the thickness and areal extent of frozen ground have considerable influence on local and regional environments and potential for disturbing human activities.

The primary control on local hydrological processes in northern regions is dictated by the presence or absence of permafrost, but is also influenced by the thickness of the active layer and the total thickness of the underlying permafrost. As permafrost becomes thinner or decreases in areal extent, the interaction of surface and sub-permafrost ground water processes becomes more important (Woo, 1986; Hinzman et al., 2003). The inability of soil moisture to infiltrate to deeper groundwater zones due to ice rich permafrost maintains very wet soils in arctic regions. However, in the slightly warmer regions of the subarctic, the permafrost is thinner or discontinuous. In permafrost-free areas, surface soils can be quite dry as infiltration is not restricted, impacting ecosystem dynamics, fire frequency and latent and sensible heat fluxes. Thickening of the active layer and melting of ice-rich permafrost in the Russian Arctic drainage basin may have already contributed, in part, to the increased river runoff (Zhang et al., 2005).

Thickening of the active layer directly results in thawing the decomposed plant materials frozen in the upper permafrost and exposing the carbon to microbial decomposition, which can release carbon dioxide and methane to the atmosphere. In seasonally frozen environments, the growing season is determined primarily by the length of the unfrozen period. Variations in both the timing of spring thaw and the resulting growing season length have been found to have a major impact on terrestrial carbon exchange and atmospheric CO_2 source/sink strength in boreal regions (Randerson et al., 1999). The timing of spring thaw, in particular, can influence boreal carbon uptake dramatically through temperature and moisture controls to net photosynthesis and respiration processes.

Deepening of the active layer has substantial effect on slope instability and rock falls within the steep mountain terrain. The summer of 2003 was the warmest on record in much of the Alps, and the depth of seasonal thaw penetration increased significantly, particularly in bedrock. Associated with these rapid increases in bedrock active-layer thickness, rock-fall activity in the Alps during 2003 was exceptional, with the majority of source areas lying within perennially frozen rock walls (Gruber et al., 2004).

4.8 Synthesis

Meteorological and cryospheric observations show a consistent picture of surface warming and ice declining (Question 4.1, Figure 1), except for latitudes south of 65S, where the sea ice exhibits a small positive but insignificant trend since 1978, which is in accordance with the evolution of the surface temperature. In all other regions of the globe shrinking ice masses are in accordance with warmer temperatures (Figure 4.8.1).

[INSERT FIGURE 4.8.1 HERE]

Since publication of the TAR the cryosphere has undergone significant changes, such as the strong retreat of the Arctic sea ice, especially in summer; the continued shrinking of mountain glaciers; the decrease of the extent of snow cover and seasonally frozen ground particularly in spring and the earlier break-up of river and lake ice; and wide-spread thinning of ice shelves along the Amundsen Sea coast, indicating increased basal melting due to increased ocean heat fluxes in the cavities below the ice shelves.

An additional new feature is the increasingly visible fast dynamic response of ice shelves, e.g. the dramatic break-up of the Larsen B ice shelf in 2002, and the acceleration of tributary glaciers and ice streams, with possible consequences for the adjacent part of the ice sheets.

One difficulty with using cryospheric quantities as indicators of climate change is the sparse historical data base. Although 'extent' of ice (sea-ice and glacier margins for example) have been observed for a long time at a few locations, the 'amount' of ice (thickness or depth) is difficult to measure. Therefore, reconstructions of past mass balance are often not possible.

The most important cryospheric contributions to sea level variations arise from changes of the ice on land, e.g., glaciers, ice caps, and ice sheets. In the TAR recent ice contribution was estimated as 0.2–0.4 mm/yr (of 1–2 mm/yr total sea level rise). New results presented here indicate that all glaciers contributed about 0.51±0.32 mm/yr during 1963–2003, increasing to 0.81 ± 0.43 from 1993–2003 (Table 4.5.2). Estimates of both ice sheets combined provided a contribution ranging from –0.35 to +0.72 for 1961–2003 and 0 to 0.8 mm/year through 1993–2003, increasing during this period. A conservative error estimate in terms of summing ranges is given in Table 4.8.1. Assuming a midpoint-mean, interpreting the range as uncertainty, and using Gaussian error summation of estimates for glaciers and ice sheets, suggests that the total ice contribution to sea level rise during 1993 to 2003 was approximately 1.2 ± 0.6 mm/yr.

Table 4.8.1. Estimates of cryospheric contribution to sea level change

Cryospheric component	Sea Level Equivalent (mm/yr)	
	1961–2003	1993–2003
Glaciers and Ice Caps	+0.19 to +0.83	+0.38 to +1.24
Greenland	−0.07 to +0.17	+0.14 to +0.28
Antarctica	−0.28 to +0.55	−0.14 to +0.55
Total	−0.16 to +1.55	+0.38 to +2.07

The large uncertainties reflect the difficulties to estimate the global ice mass and its variability, because global monitoring of ice thickness is impossible (even the total area of glaciers is not exactly known), and extrapolation from local measurements is therefore necessary. A regional extension of the monitored ice masses and an improvement of measurement and extrapolation techniques are urgently required.

In spite of the large uncertainties, the data that are available portray a rather consistent picture of a cryosphere in decline over the 20th century, increasingly so during 1993 to 2003.

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Question 4.1: Is the Amount of Snow and Ice on the Earth Decreasing?

Yes. Despite growth in some places and little change in others, the majority of observations show melting over many years, especially since 1980 and increasing during the past decade (Question 4.1, Figure 1). Most mountain glaciers are getting smaller. Snow cover is retreating earlier in the springtime. Sea ice in the Arctic is shrinking, especially in summer. Reductions are reported in seasonally frozen ground, river and lake ice, together with warming of permafrost. And important coastal regions of the ice sheets on Greenland, West Antarctica, and the Antarctic Peninsula are thinning to a degree that, taken together, the ice sheets are contributing to sea-level rise. The total contribution of glaciers and ice sheets to sea level rise is estimated as 1.2±0.6 mm per year.

Continuous satellite measurements since 1966 capture most of the Earth's seasonal snow cover on land, despite limitation of such measurements to the northern hemisphere. Snow cover has decreased about 5% since 1966. The decrease has been especially prominent in late winter and spring, with little change in fall or early winter, and occurred in many places despite increases in precipitation.

Satellite data do not yet allow similarly reliable measurement of ice conditions in lakes and rivers, or in seasonally or permanently frozen ground. Numerous local to regional reports have been published, however, and generally seem to indicate warming of permafrost and increase in the thickness of the summertime thawed layer, decrease in wintertime freeze depth in seasonally frozen areas, and decrease in duration of seasonal river and lake ice.

Since 1978, satellite data have provided continuous coverage of sea-ice extent in both polar regions. From 1978–2005 total sea-ice extent in the Arctic decreased by 2.7 ± 0.7% per decade. In the Antarctic, sea ice shows a slight positive but insignificant trend of 0.5 ± 0.9% per decade. Trends in both hemispheres were seasonal with the largest changes observed in summer. In the Arctic, the summer sea ice extent exhibits a reduction of 7.4 ± 2.9% per decade. Thickness data, especially from submarines, are available but restricted to the Central Arctic, where they indicate thinning of more than 40% between the 1958–1977 period and the 1990s.

[INSERT QUESTION 4.1, FIGURE 1 HERE]

Most mountain glaciers have been shrinking, with such retreat probably having started at about 1850. Glacial melt has contributed 0.51 ± 0.32 mm per year to sea level rise between 1961 and 2003. Many northern hemisphere glaciers had a few years of near-balance around 1970, followed by enhanced shrinkage, with sea level contributions of 0.81 ± 0.43 mm per year between 1993 and 2003.

The large ice sheets of Greenland and Antarctica are likely shrinking overall contributing about 0.4 ± 0.4 mm/yr to sea-level rise during 1993–2003, with accelerated loss likely through 2005. Thickening of high-altitude, cold regions of Greenland and East Antarctica, perhaps from increased snowfall, is more than offset by thinning in coastal regions of Greenland and West Antarctica in response to increased flow velocity and increased Greenland surface melting. Coastal flow velocities increased through some combination of reduced restraint from shrinkage of floating extensions called ice shelves, increased stress from coastal steepening, and increased lubrication from additional meltwater, all in response to increased melting.

Ice interacts with local energy balance and transport processes in complex ways, so unique explanations of changes are difficult. Nonetheless, when the local temperature is too high, ice melts. Reduction in snow cover and in mountain glaciers has occurred despite increased snowfall in many cases, implicating increased local air temperatures. Similarly, although snow-cover changes affect frozen ground, lake and river ice, they do not seem sufficient to explain the observed changes, suggesting that increased local air temperatures have been important. Arctic sea-ice reductions can be simulated fairly well in models driven by historical circulation and temperature patterns; temperatures show an upward trend but without a significant trend in ice export from the Arctic, implicating warming. The observed increases of ice-sheet snowfall in some cold central regions, surface melting in warmer coastal regions, and sub-ice-shelf melting along many coasts are all consistent with local environmental warming. The geographically widespread nature of these cryospheric changes suggests that overall the Earth is losing ice because of warming.