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

Coordinating Lead Authors: Peter Lemke, Jiawen Ren

Lead Authors: Richard Alley, Ian Allison, Jorge Carrasco, Gregory Flato, Yoshiyuki Fujii, Georg Kaser, Philip Mote, Robert Thomas, Tingjun Zhang

Contributing Authors: J. Box, D. Bromwich, R. Brown, G. Cogley, J. Comiso, M. Dyurgerov, B. Fitzharris, O. W. Frauenfeld, H. Fricker**,** H. Gudmundsson, C. Haas, J. O. Hagen, C. Harris, L. Hinzman, R. Hock, M. Hoelzle, P. Huybrechts, K. Isaksen, P. Jansson, A. Jenkins, I. Joughin, C. Kottmeier, R. Kwok, S. Laxon, D. MacAyeal, K. C. McDonald, H. Melling, A. Ohmura, A. Payne, T. Prowse, B. Raup, C. Raymond, E. Rignot, I. Rigor, D. Robinson, D. Rothrock, L. Shiyin, S. Smith, O. Solomina, D. Vaughan, J. Walsh, A. Worby, T. Yamada, L. Zhao

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Table of Contents

Executive Summary ... 3
4.1 Introduction ... 5
Box 4.1: The Ice Albedo Feedback ... 6
Question 4.1: Is the Amount of Snow and Ice on the Earth Decreasing? ... 6
4.2 Changes in Snow Cover and Albedo ... 7
4.2.1 Background ... 7
4.2.2 Snow – Albedo Feedback ... 8
4.2.3 Other Feedbacks of Snow on Climate ... 8
4.2.4 Observations of Snow Cover, Snow Duration, and Snow Quantity ... 8
4.3 Changes in River and Lake Ice ... 11
4.3.1 Introduction ... 11
4.3.2 Changes in Freeze-up and Break-up Dates ... 11
4.4 Changes in Sea Ice ... 12
4.4.1 Background ... 12
4.4.2 Sea Ice Extent and Concentration ... 13
4.4.3 Sea Ice Thickness ... 15
4.4.4 Pack Ice Motion ... 18
4.5 Changes in Glaciers and Ice Caps ... 19
4.5.1 Background ... 19
4.5.2. Large and Global Scale Analyses and Interpretation of Measured Glacier Changes ... 20
4.5.3 Special Regional Features ... 22
4.5.4 Little Ice Age and Medieval Warm Period ... 23
4.5.5 Changing Runoff from G&IC and Glacier Related Hazards ... 24
4.6 Changes and Stability of Ice Sheets and Ice Shelves ... 24
4.6.1 Background ... 24
4.6.2 Mass Balance of the Ice Sheets and Ice Shelves ... 24
4.6.3 Causes of Changes ... 27
4.6.4 Other Issues in Projection of Balance Changes ... 31
4.7 Changes in Frozen Ground ... 32
4.7.1 Background ... 32
4.7.2 Changes in Permafrost ... 33
4.7.3 Changes in Seasonally Frozen Ground ... 36
4.7.4 Consequences ... 37
4.8 Synthesis ... 38
4.8.1 Observed changes of the cryosphere ... 38
4.8.2 Cryospheric Contributions to Sea Level Change ... 39

References 40

Executive Summary

In the past, the cryosphere has undergone significant variations associated with the ice ages and shorter-term climate variations like the Younger Dryas or the Little Ice Age. Recent changes over the past century are in accordance with the rise of the global surface air temperature. This is especially true for the polar regions of the Northern Hemisphere (NH), where the surface air temperature north of 70N has increased over the past 50 years by about three times the global average.

- Since the early 1920s, and especially since the late 1970s, snow cover in the NH has declined in spring though not substantially in winter. Mountain snowpack in western North America has also declined in spring at 75% of locations monitored since 1950. Though snow is poorly monitored in the Southern Hemisphere (SH), most records or proxies show either declines or no changes in the past 40+ years. In Australia, as in the NH, declines have been larger in spring than in winter. Where there are declines in snow cover or snowpack, temperature often plays a dominant role; where there are increases, precipitation almost always dominates. For example, NH April snow cover extent is significantly correlated ($r = -0.68$) with 40–60°N April temperature.

- Both the freeze-up and break-up estimates of river and lake ice exhibit considerable variability, with some evidence of opposing trends in the western and eastern parts of the northern continents. On average, the general trend over the past 150 years indicates that the freeze-up date has become later at a rate of 5.8 days per century, while the break-up date has advanced at a rate of 6.5 days per century.

- Satellite data indicate a continuation of the roughly 3% per decade decline in annual mean Arctic ice extent since 1978. The decline for wintertime extent is smaller than for summertime, with the summer minimum extent declining at a rate of 7.3 % per decade. The actual area covered by ice at summer minimum has declined by 9.2% per decade, reflecting the amount of multi-year ice that survives summer melt. Similar observations in the Antarctic reveal larger interannual variability but no significant trends. Longer term records are available for some NH locations and reveal large decadal and longer scale variability, superimposed on a declining trend extending well prior to the satellite record.

- Submarine-derived data for the Arctic indicate a reduction in pack ice thickness particularly since the late 1980s. These sparse data are augmented by model simulations that are able to reproduce this behaviour and imply a combination of thermodynamic forcing and driving by large-scale atmospheric variability. Landfast ice stations from both the Canadian and Russian Arctic indicate both increasing and decreasing thickness trends over the last 50–100 years, depending on the location. Ice thickness information in the Antarctic is too sparse and too recent to make inferences regarding variability and trends.

- Sea ice motion is forced predominantly by the wind and so is closely connected to large-scale atmospheric variability. There are no apparent trends in sea-ice circulation, or in ice area outflow from the Arctic to the North Atlantic.

- During the 20th century, glaciers and ice caps (G&IC) have generally experienced considerable mass losses with strongest retreats in the 1930s and 1940s and after 1990. Late 20th century glacier wastage is likely a response to post-1970 global warming. Mass loss of G&IC, excluding those around the two ice sheets, is estimated 0.36 mm in sea level equivalent per year between 1967/1968 and 1996/1997, with about twice as high rates from 1992 to 2003. Most mass is discharged from retreating G&IC in Alaska, the Arctic, and the Asian High Mountains. Tropical glacier changes are synchronous with global ones. Kilimanjaro is a special case with the solar driven incessant retreat of the vertical walls of the tabular plateau ice.

- Glacier length variations allow for reconstructing temperature variations on a global and regional scale. Before 1900 precipitation anomalies are also important.

- Extreme mass losses of glaciers in the European Alps in 2003 were caused by extraordinarily high air temperatures over a long period and extremely low precipitation amounts, as well as albedo feedback from a previous series of negative mass balance years.

- As an intermediate effect of glacier retreat, increases in total glacier runoff and peak flows, and considerable amplification of diurnal melt runoff amplitudes are observed. Glacier retreat causes the development of hazardous lakes particularly in the Andes and the Himalaya.

- The ice sheets appear to be near-balance or thickening slightly in central regions, but thinning around the margins of Greenland and important parts of West Antarctica, broadly consistent with expectations for a warming world. Greenland contributed 0.1–0.2 mm per year to sea-level rise between 1993 and 2003, with ice-sheet losses increasing during this time, whereas the sign of the contribution in Antarctica remains uncertain. Important regions remain undersampled, knowledge of accumulation-rate trends is uncertain, and ground observations are almost wholly lacking to check satellite altimetry and assess the role of density change in elevation change. For Greenland and Antarctica combined, a small contribution to sea-level rise of 0.1 mm per year for the past decade with an increase to 0.2 mm per year during the past five years appears most consistent with published results, with uncertainties of similar magnitude.

- Since the TAR, attention has especially focused on notable acceleration of tributary glaciers contributing to sea-level rise following thinning or loss of ice shelves in some near-coastal regions of Greenland and Antarctica. Prognostic models currently configured for long integrations remain most reliable in their treatment of surface accumulation and ablation, as for the TAR, but do not include full treatments of the sub-ice-shelf melting, ice-sheet/ice-stream/ice-shelf coupling, and related issues contributing to these rapid marginal changes; thus, projections from such models may underestimate ice-flow contributions to sea-level rise by poorly constrained amounts.

- Permafrost and seasonally frozen ground have experienced significant changes in the past decades. Permafrost surface temperature has increased by 2 to 4°C from the turn of the 20th century to the early 1980s in the Alaskan Arctic. The warming is accelerating since the 1980s with additional increase of 2 to 3°C. Permafrost warming is also observed with variable magnitude but a consistent trend in Canadian Arctic, Siberia, Tibetan Plateau, and Europe. Permafrost boundaries have moved northwards in Canada and upwards on the Tibetan Plateau. Observed evidence indicates that the base thawing of permafrost at an average rate of 0.02 to 0.4 m/yr in northern Alaska and the Tibetan Plateau. Thickness of the active layer over permafrost has increased about 20 cm from the mid-1950s to 1990 in the Russian Arctic although recent measurements show no obvious trends but with large inter-annual variability since the 1990s in the Arctic as whole. Maximum depth of seasonally frozen ground has decreased by about 30 cm from the mid-1950s to 1990 in Russia and about 20 cm on the Tibetan Plateau, China. Maximum area extent of seasonally frozen ground has decreased by 10 to 15% in spring over the 20th century in the Northern Hemisphere. Evidence from 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 from 1988–2002, leading to a forward shift of the growing season but no change in its length.

- The total cryospheric contribution to sea level change has been estimated in the TAR to be 0.2–0.4 mm/year. Recent data analyses indicate a significant increase to 1 mm/year during the past five to ten years.

4.1 Introduction

The cryosphere represents an important part of the climate system. Concerning its mass and 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 very special physical properties, such as high reflectivity (albedo), low heat conductivity and high value of latent heat associated with phase changes, all of which have a large impact on the surface energy balance with important consequences for the thermal and dynamical structure of both the atmosphere and the ocean.

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). 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. Ice also covers approximately 6.5% of the oceans in the annual mean (Table 4.1.1). In mid-winter, snow covers approximately 49% of the land surface in the Northern Hemisphere (NH). Frozen ground has the largest areal extent of any component of the cryosphere. The components of the cryosphere exhibit 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 and volume of cryospheric components. Indicated are the seasonal minimum and maximum for snow, sea ice and seasonally frozen ground, and the annual mean for the other components.

Cryospheric Components	Area (10^6 km^2)	Volume (10^6 km^3)
Snow on land (NH)	3.6 ~ 46.9	0.0005 ~ 0.005
Sea ice	18.0 ~ 25.0	0.019 ~ 0.025
Glaciers and ice caps[a, b]	0.51 (0.54)	0.05 (0.13)
Ice shelves[c]	1.54	0.73
Ice sheets	13.80	32.33
Seasonally frozen ground (NH)[d]	5.85 ~ 64.9	0.006 ~ 0.065
Permafrost (NH)[d]	22.8	4.5

Notes:
(a) Ohmura (2005)
(b) Dyurgerov and Meier (2005)
(c) Drewry (1983)
(d) Zhang et al. (2005)

An important property of snow and ice is its high surface albedo. Up to 90% of the incident solar radiation is reflected by snow and ice surfaces. Only a small part of this reflected energy is absorbed in the atmosphere, most of it is lost to space. Over the open ocean and over forested land, on the other hand, 90% of the solar radiation is absorbed, and heats the climate system. In addition, snow and ice are effective insulators. Therefore, snow and ice surfaces act as energy sinks, where cold air is produced.

Because of the spherical shape of the Earth and the high albedo of snow and ice, the polar regions absorb significantly less solar radiation than the tropics. The resulting temperature differences induce winds and currents, which are influenced by many interactions within the climate system. The cryosphere plays a major role in these interactions, because the tracks of low pressure systems are influenced by the large temperature gradients at snow and ice margins, and the interactions with sea ice and ice shelves affect the oceanic deep water formation, and, therefore, the thermohaline circulation in the ocean. The climate system is regulated by a large variety of positive (destabilizing) and negative (stabilizing) feedbacks. The cryosphere plays an important role in this climatic feedback system. Because of the positive, destabilizing temperature – ice albedo – feedback the cryospheric components, especially those with short response times, represent very effective indicators of climate variations (Box 4.1). An advantage of this is that cryospheric components are found over all latitudes.

In addition, 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 55 m of sea level rise, respectively. Changes of the ice mass on land contributed significantly to recent changes of the sea level (see Section 4.8). On a regional scale, many glaciers and small ice caps play a crucial role in fresh water availability.

[START OF BOX 4.1]

Box 4.1: The Ice Albedo Feedback

Almost all of the Earth's energy comes from the sun. About 30% of the arriving solar energy is reflected back to space, and the other 70% warms the planet, especially its surface. Since the atmosphere is nearly transparent, 50% of the solar energy reaches the surface. The fraction of solar energy reflected by a surface is termed the 'albedo'. Snow and ice are especially reflective, with typical albedos of 0.6 to 0.9 (although values above and below this range are possible).

If the climate warms for any reason the area covered by snow and sea-ice will decrease and the underlying land and ocean will be revealed. Since these underlying surfaces have lower albedos, less sunlight will be reflected and more absorbed hence warming the system. Similarly, if the climate cools, the area covered by snow and sea-ice will increase, more solar radiation will be reflected and less absorbed and the system will cool. This "ice/albedo" behaviour constitutes a positive feedback mechanism, one that acts to amplify both warm and cold perturbations to the climate system.

The climate system involves many feedback mechanisms, some of which may be locally positive (amplifying) or negative (damping). On average, negative feedbacks must be predominant or the system would be unstable. Feedback mechanisms nevertheless vary from place to place on the globe and ice-albedo feedback is an example of this since it operates at high latitudes but not in the tropics (e.g., Boer and Yu, 2002; 2003). This explains, at least in part, why climate warming is amplified at high latitudes even though the radiative effect causing it is more or less uniform, as is the case for an increase in CO2.

Climate model experiments in which the ice albedo feedback is suppressed in some way (e.g., Rind et al., 1995; Hall, 2004) indicate that roughly 1/3 of the global temperature response to increasing greenhouse gas concentrations is a consequence of the ice albedo feedback and, interestingly, that its effect is 'felt' (via transport and other feedbacks) even in the Tropics (Box 4.1, Figure 1).

[INSERT BOX 4.1, FIGURE 1 HERE]

[END OF BOX 4.1]

[START OF QUESTION 4.1]

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 (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, so that the ice sheets are contributing to sea-level rise.

Consistent satellite measurements since 1966 capture most of the Earth's seasonal snow cover on land, despite limitation 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, contributing to the lack of consistent global observations.

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 passive-microwave data have provided consistent coverage of sea-ice extent in both polar regions. From 1978–2004, total sea-ice extent in the Arctic decreased by 2.7 ± 0.2% per decade. In the Antarctic, sea ice shows a slight positive but insignificant trend of 0.7 ± 0.2% 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 7.3 ± 1.7% 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 retreating, probably having started about 1850, with a contribution of 0.36 mm a^{-1} to sea level rise between 1968 and 1997. Many glaciers had a few years of near-balance around 1970, followed by enhanced retreat with sea level contributions about twice as high as between 1968 and 1997 through the past decade.

The large ice sheets of Greenland and Antarctica are changing in many ways. Net shrinkage probably has been occurring overall, contributing slightly (~0.1 mm/yr, with large uncertainties) to sea-level rise during the 1990s, probably with an accelerating trend. The high-altitude, cold regions of Greenland and East Antarctica are thickening or are near balance, but coastal thinning likely offsets this. Increased surface melting and ice-flow velocity have contributed to thinning in coastal Greenland, and increased flow velocity has contributed to thinning in the Amundsen Sea embayment region of West Antarctica and along parts of the Antarctic Peninsula. Much of this increased flow has resulted from reduction or loss of floating extensions called ice shelves, in response to increased basal melting in sea water, and in one case to increased surface melting. Dynamic changes include a slowdown of an ice stream that has switched the Siple Coast region of West Antarctica from thinning to thickening over the last decade, but large speedups and slowdowns have occurred over the last millennium or longer in this region, and the thickening does not offset the thinning elsewhere in West Antarctica.

Many processes control the extent of ice, and simple explanations of the observed changes are not always available. However, the effects of warming are probably more important than anything else in these observations. Snow cover has shrunk in the springtime, often despite increased precipitation, indicating warming. Reduced snow cover does affect frozen ground, river and lake ice, but is unlikely to be sufficient to explain the changes observed, with warming implicated. Sea ice thinning and areal shrinkage in the Arctic are strongly linked to the upward trend in the surface air temperature and to the atmospheric circulation pattern. Mountain glaciers are sensitive to many things, but most are more sensitive to temperature than to anything else, and the wastage of mountain glaciers indicates warming; indeed, estimates of warming from calibrated glaciers agree with independent, thermometer-based estimates. The increased snowfall in interior regions of the ice sheets, as well as the strong coastal changes, are consistent with warming, although the coastal changes are spatially restricted and are especially indicative only of water temperatures under ice shelves. Overall, the Earth appears to be losing ice because of warming.

[END OF QUESTION 4.1]

4.2 Changes in Snow Cover and Albedo

4.2.1 Background

The role of snow in the climate system includes strong positive feedbacks related to albedo (Box 4.1) and other indirect feedbacks related to moisture storage, latent heat, and insulation of the underlying surface (Clark et al., 1999). Snow cover also helps determine the ice growth rate. Chemical reactions in snow also affect the concentration of bromine and other atmospheric trace constituents.

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, precipitation and atmospheric circulation; 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 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 Snow – Albedo Feedback

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) although the size 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.

Hansen and Nazarenko (2004) speculated that the albedo of snow may be changing under human influence, separately from changes driven by the accumulation of greenhouse gases. Black carbon, or soot, in snow can reduce its albedo by as much as 0.14, and the sootiness of snow has evidently been increasing in many parts of the northern hemisphere, possibly contributing to recent Arctic warming.

4.2.3 Other Feedbacks of Snow on Climate

The indirect feedbacks may involve two types of circulation, monsoonal and annular, though these connections are statistically tenuous and controversial. Asian snow cover in spring, especially in southern China, appears to influence the strength of the Asian monsoon in the summer (e.g., Bamzai and Shukla, 1999; see Section 3.7.1 for more about the Asian monsoon). In low-snow years, drier soils allow the surface to warm more, and intensify land-ocean heat contrasts, hence strengthening the monsoonal circulation. Similar results have been suggested for the North American monsoon (Lo and Clark, 2001). Other studies, however, have found no evidence for a relationship between soil moisture and monsoon strength (e.g., Robock et al., 2003).

Second, statistical analysis suggests a lag relationship between Eurasian snow cover in autumn and the strength of the Arctic Oscillation (AO; see Section 3.6.4) that winter (Saito and Cohen, 2003). Although the existence and strength of the AO in a GCM appears to be independent of snow cover variations, only with interannual snow cover variations does the AO extend into the stratosphere as observed (Gong et al., 2002). The mechanism for snow cover influence on AO appears to be an effect on large-scale wave activity fluxes over Siberia, weakening the polar vortex in high-snow years (Gong et al., 2003). However, Bamzai (2003) found that the AO leads snow cover on timescales from weeks to months, suggesting that the dominant causal pathway is for AO to influence snow, not the other way around.

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

4.2.4.1 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.

The premier dataset used to evaluate large-scale snow cover extent (SCE) is the weekly visible-wavelength satellite maps of Northern Hemisphere snow cover produced by the 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, and there is recent evidence that the NOAA dataset overestimates spring snow cover over a portion of northern

Canada (Wang et al., 2005). However, these issues are unlikely to have affected the homogeneity of continental-scale estimates of snow cover extent derived from the NOAA dataset. For the southern hemisphere, mapping of SCE began only in 2000 with the advent of MODIS.

Microwave remote sensing provides the potential for global monitoring of snow cover unimpeded by cloud cover and winter darkness. 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 SCE 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.

4.2.4.2 Variability and trends in snow cover: Northern Hemisphere

The mean annual Northern Hemisphere SCE is 25.6×10^6 km^2. This includes snow over the Greenland ice sheet, which is discussed in section 4.6. Seasonally, the area covered by snow ranges from a mean maximum in January of 46.9×10^6 km^2 to a mean minimum in August of 3.6×10^6 km^2. Snow covers more than 33% of lands north of the equator from November to April, reaching 49% coverage in January.

Interannual variability of SCE is largest not in winter, when mean SCE 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. The range in October is 13×10^6 km^2 about a mean of 19×10^6 km^2. There remains some uncertainty as to whether the microwave satellite data show similar interannual variability and trends except in autumn (see 4.2.4.1).

Since the early 1920s, and especially since the late 1970s, SCE has declined in spring (Figure 4.2.1) but not substantially in winter (Table 4.2.1) despite winter warming (Jones and Moberg, 2003; section 3.2.2). Recent declines in SCE in the months of February through April have resulted in (1) a shift in the month of maximum SCE from February to January; and (2) a statistically significant decline in annual mean SCE. Early in the satellite era, between 1967 and 1987, mean annual SCE was 26.1×10^6 km^2. An abrupt transition occurred between 1986 and 1988, and since 1988 the mean annual extent has been 24.8×10^6 km^2, a statistically significant (T test, $p < 0.01$) reduction of approximately 5% (Robinson and Frei, 2000) April SCE also declined significantly over the 1922–2004 period, by $-0.032\ (\pm 0.008) \times 10^6$ km^2/yr for an aggregate loss of 2.7×10^6 km^2 or about ~1% per decade (updated from Brown, 2000).

[INSERT FIGURE 4.2.1 HERE]

Table 4.2.1. Trend (10^6 km^2/decade) in monthly NH SCE 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 SCE 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

Warming plays a significant role in variability and trends of NH SCE, especially in March (Figure 4.2.2) and April. Two related pieces of evidence support this conclusion. First, April NH SCE 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., 1999). Snow-albedo feedbacks are likely contributing to this elevated spring response as demonstrated by Groisman et al. (1994).

[INSERT FIGURE 4.2.2 HERE]

4.2.4.2.1 North America

From 1915 to 2004, North American SCE increased in November, December and January owing to increases in precipitation (Section 3.3.2; Groisman et al., 2004; Zhang et al., 2000). Over the same period of record, trends in other months are not significant; they become significant only after mid-century. Declines are most pronounced over western North America (Groisman et al., 2004) including northern Alaska, where the date of snowmelt has advanced about 8 days since the mid-1960s (Stone et al., 2002). In New England (northeastern U.S.), a reduction in spring snow is implied by the 1–2 week advance in spring snowmelt runoff that has occurred mostly since 1970 (Hodgkins et al. (2003).

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) and hydrologic modeling (Hamlet et al., 2005). That these reductions are predominantly due to temperature increases is demonstrated by modeling (Hamlet et al., 2005), regression analysis (Stewart et al., 2005), and in the dependence of trends in peak flow (Regonda et al. 2005) and in SWE (Mote et al., 2005) on elevation or equivalently mean winter temperature (Figure 4.2.3a), with largest changes near snowline.

[INSERT FIGURE 4.2.3 HERE]

4.2.4.2.2 Europe and Eurasia

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 (Laternser and Schneebeli, 2003; 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) and snow cover duration increased over mountainous areas of Poland since about 1950 (Falarz, 2004). In the Swiss Alps, statistically significant declines since about 1980 in the mean snow depth, the duration of continuous snow cover and the number of snowfall days followed increases over the 1931–1980 period (Laternser and Schneebeli, 2003). 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 (Figure 4.2.3b).

Lowland areas of central Europe are characterized by recent reductions in annual snow cover duration by ~1 day/yr (Falarz, 2002; Tooming and Kadaja, 2000) and an increase in the interannual variability of spring snow cover (Falarz, 2004). Trends toward greater maximum snow depth but shorter snow season have been noted in Finland (Hyvärinen, 2003), the former Soviet Union 1936–1995 (Kitaev et al., 2002; Ye and Ellison, 2003, corroborated by earlier snowmelt runoff (Ye et al., 2003)), and in the Tibetan (Zhang et al., 2004) and Qinghai-Xizang (Chen and Wu, 2000) Plateaus since the late 1970s. The most significant decreases in snow cover duration over northern Eurasia (over the 1956–2000 period) have occurred in Siberia (Groisman et al., 2005).

4.2.4.3 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.

4.2.4.3.1 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 (32º47'S, 71º33'W, 8 m above sea level) (Carrasco et al., 2005). Over the 1975–2001 period of record, the linear trend in winter ZIA was 121.9 ± 7.7 m (Figure 4.2.4). However, no significant change was observed for ZIA on the days when there was precipitation in central Chile: the observed warming has occurred mainly during days with no precipitation. The ZIA trend during the period suggests enhanced snow melt on dry days.

[INSERT FIGURE 4.2.4 HERE]

4.2.4.3.2 Australia and New Zealand

For the mountainous southeastern area of Australia, studies of late-winter (August-September) snow depth have shown some significant declines; trends in maximum snow depth are more modest. Hennessy et al. (2003) examined four sites and found declines at three of them over the 1957–2002 period. At Spencers Creek, maximum snow depth declined somewhat since 1962 but spring snow depth declined by about 40% (Nicholls, 2004). The stronger declines in late winter are attributed to the dominant role of temperature and the large temperature trends, while the more modest declines in maximum snow depth are attributed to the greater importance of 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. Over the 1930–1985 period, there was no clear trend in the amount of seasonal snow in the Southern Alps (Fitzharris and Garr, 1995), but this study has not been updated.

4.3 Changes in River and Lake Ice

4.3.1 Introduction

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 and possible 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 lead to severe flooding.

Because of the importance to many human activities, the 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 give 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 data and 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), Figure 4.3.2. These freeze-up and break-up estimates (based on inferences from streamflow data) exhibit considerable variability, but show a propensity toward later freeze-up across Canada while break-up tends to be earlier in western Canada and later in the easternmost part of the country. A recent analysis of Russian river data by Smith (2001) 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 analyisis 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 (see Box 4.1); 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.

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 18 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.

The thickness of sea ice is a consequence of past growth, melt and deformation, and so is an important indicator of climatic conditions and of the ability of the ice pack to store and transport fresh water. 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); its thickness (and the thickness of the snow cover on it); its velocity; its growth and melt rates (and hence salt or freshwater flux into the ocean), and the area of multi-year ice within the total extent. 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 be 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. Similar but not as comprehensive data exist in the Southern Hemisphere since the middle of the 19th Century.

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% or more 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 extent 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 area, which is by definition the multi-year ice area 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 October 2004 is shown in Figure 4.4.1 and illustrates a significant trend in Arctic sea ice extent of –2.7 ± 0.2% per decade.The Antarctic results show a slight but insignificant positive trend of 0.7 ± 0.2% per decade. 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.

[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. Trends in the minimum Arctic sea ice extent, between 1979 and 2004, were –7.3 ± 1.7% per decade in the minimum ice extent and –9.2 ± 1.5% per decade for actual ice area (updated from Comiso, 2002). 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. Several compilations of available data spanning the 20th century are compared in Rayner et al. (2003), with the most recent shown in Figure 4.4.3. There is a clear indication of sustained decline in Arctic ice extent since about the 1960s, 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 decline in recent decades, along with much more extensive ice in the late 19th and early 20th century. 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 also show a declining trend since the 1960s, but exhibit large interdecadal variability. 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. On a more local level, the Icelandic sea ice index (the ‘Koch Index’), recently updated by Ogilvie and Jonsson (2001), is a combination of residence time and length of Icelandic coastline experiencing ice in a given year (larger numbers therefore imply more ‘severe’ ice years). This index is shown by the symbols in Figure 4.4.3, Ruffman et al. (2003) 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. 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 significant decline in sea ice 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. 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 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. The earliest observations of Antarctic sea ice are from Cook’s 1772–1774 expedition, and the first maps of mean sea ice extent for each month were compiled by Mackintosh and

Herdman (1940) from observations from the Discovery scientific expeditions and from whaling voyages. Parkinson (1990) examined ice edge observations from four late-18th to early 19th century exploration voyages to suggest 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 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. However 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 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.

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 ice thickenss).

Direct drilling is best suited to level landfast ice that annually forms in coastal areas. Weekly measurements of fast ice became routine at Arctic weather stations in Siberia in the late 1930s and in northern Canada in the late 1940s. Unfortunately, few of the Arctic observations continued uninterrupted into the 21st century. Similar observations in the Antarctic commenced in the 1950s and have continued at some coastal sites, albeit intermittently, until the present.

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 from both submarine and moored sonars are uncertainties in sound speed and atmospheric pressure, and the correct identification of spurious targets. Upward-looking sonar has been on submarines operating beneath Arctic pack ice since 1958 and the first published map of ice thickness in the central Arctic was prepared from classified data by Bourke and Garrett (1987). US and UK naval data are now being released for science (with position and time of observation slightly “blurred” for security), and some dedicated Arctic submarine missions were made for science during 1993–1999. Ice-draft measurement by moored ice-profiling sonar (IPS), 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. In the Southern Hemisphere there are no data from submarines and only short time series from moored sonar.

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.

An emerging new technique uses satellite radar altimetery to measure range to the ice and, when leads are present, range to the sea surface. The ice freeboard (that fraction of the ice above the water surface) is the difference between the two ranges and the thickness can be estimated assuming an average floe density (which varies with snow loading). An approach to processing radar altimetry for sea ice has been described by Laxon et al. (2003): this is presently limited to the cold months (October to April in the central Arctic) and to ice thicker than about 0.5 m. While it has not been operational long enough yet to provide data for climate change studies, this is a promising technique for future ice thickness monitoring.

Electromagnetic-induction sounders deployed on the ice surface, ships or aircraft, can measure the thickness of cold, level floes to an accuracy of 0.1 m, but may underestimate the thickness of deformed ice if conductive seawater layers are present within the floe structure. Applicability to climate analysis is hampered by the localized nature of these measurements. Aircraft-mounted laser systems have also been used to measure sea ice freeboard, but accurate estimates of ice thickness demand meticulous analysis and the technique is better suited to determiming ridging statistics.

Finally, physically-based sea-ice models, driven by observationally-based atmospheric and oceanic forcing, provide continuous time series of ice extent and thickness which can be compared to the sparse observations, and used to interpolate the observational record. In particular, model studies can elucidate some of the forcing agents responsible for observed changes in ice thickness.

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

Wadhams (1992), using data from two surveys conducted in 1976 and 1987, documented a 15% decrease in ice draft (5.34 to 4.55 m) over a wide area north of Greenland and a progressive thinning of ice within the East Greenland Current. McLaren et al. (1994) analyzed data from twelve submarine cruises near the Pole between 1958 and 1992 and found no significant trend in ice draft. 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) showed from 50-km mean values of draft sampled along the same track in different years, that drafts in the mid 1990s were less than those measured between 1958 and 1977 at every common point (including the North Pole). The change was least (–0.9 m) in the southern Canada Basin, greatest (–1.7 m) in the Eurasian Basin, and averaged about 42%. Their study included very few data within the seasonal sea ice zone and none within 200 miles of Canada or Greenland. Additional data from 1976 and 1996 in the area between Fram Strait and the Pole revealed a comparable 43% reduction in average ice draft (Wadhams and Davis, 2000).

It has been suggested 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 from1976 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. Yu et al. (2004) presented evidence of a similar change in ice thickness distribution over a wider area. Tucker et al. (2001) argue that the likely cause of thinner ice along 150°W in the 1990s is reduced storage of multi-year ice in a smaller Beaufort gyre and the export of “surplus” via Fram Strait.

It is apparent that 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

There are very few moored ice-profiling sonar time series for the Arctic spanning more than 10 years, and none for the Antarctic, so these data have generally been used to determine seasonal and inter-annual variability and physical processes, not trends (e.g., Melling and Riedel, 1996; Strass and Fahrbach, 1998).

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 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.

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 of ice freeboard. They corrected for snow burden using a climatological estimate of snow amount. 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. The thickness variability is highly correlated to the length of the summer melt season, indicating that changes in ice mass are likely to be thermal in origin and that the submarine observed thinning (Rothrock et al., 1999) may be due, in part, to an increase in melt season length. There is no significant trend in thickness over the short period of their time-series.

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 contrained by observed ice concentration fields, are useful for interpreting or augmenting thickness observations. 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. 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 starting in the late 1980s, but disagree somewhat with respect to trends and/or variations earlier in the century. However, 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 indication from both models and observations that much of the change in thickness occurred between the late 1980s and late 1990s. Although some of the dramatic change inferred from submarine observations may be a consequence of spatial redistribution of ice volume over time (e.g., Holloway and Sou, 2002), thermodynamic changes are also believed to be important. It is not possible to attribute the abrupt change in thickness entirely to the (rather slow) observed warming in the Arctic. Indeed, low-frequency atmospheric variability (such as interannual changes in circulation connected to the Arctic Oscillation) appears to be important in flushing ice out of the Arctic Basin, thus increasing the amount of summer open water and enhancing thermodynamic thinning through the ice-albedo feedback (e.g., Lindsay and Zhang, 2005). Large-scale modes of variability affect both wind-driving 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]

Fichefet et al. (2003) conducted one of the few long-term simulations of Antarctic 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 variation in landfast ice thickness for selected stations in northern Canada was 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. The principal determinant of inter-annual variation in end-of season ice thickness was not variation in air temperature, but variation in the amount and timing of snow accumulation. 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.

Fast ice thickness measurements have been intermittently made at the coastal Antarctic sites of Mawson and Davis for about the last 50 years. Although there is no long term trend in maximum ice thickness at these sites, interannual variability increased significantly in the 1980s at Mawson and in the 1990s at Davis, and 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. The computed interannual variability of snow depth in May is 6 cm, with a weak negative trend for all months. The largest trend, a decrease of 8 cm over 37 years, occurs in May, the month of maximum snow depth, and appears to be due to a reduction in accumulation-season snowfall. The snow cover on sea ice observed in the Beaufort Sea during 1997–1998 (Sturm et al., 2002) was very similar to the 37-year average, even though the ice in that year was unusually thin.

For the Antarctic there are fewer data on snow cover and distribution, and no data adequate for detecting any trend in snow cover. Massom et al., (2001) collated all available ship observations to show that average snow thickness is typically 0.15–0.20 m, and varies widely both seasonally and regionally due to differences in precipitation regimes and the age of the underlying ice. An important process in the Antarctic sea ice zone is the formation of snow-ice. This 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. Different regional estimates from ice cores suggest that the fraction of snow-ice in the Antarctic pack varies from 8 to 38% and is seasonally-dependent, with a lower percentage observed early in the growth season compared to spring (Massom et al., 2001). Due to the widespread occurrence of snow-ice, snow thickness observations underestimate the total snow accumulation because some of the snow cover has been entrained in the floes as snow ice.

4.4.4 Pack Ice Motion

4.4.4.1 Data sources and time periods covered

Our observational understanding of sea ice motion is based primarily on 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. Buoy-derived observations of sea ice motion have velocity estimation errors of less than 0.02 cm/s (Rigor et al., 2002 and references therein). Gridded fields of ice motion are then produced using objective analysis.

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.

Sea ice motion may also be derived from satellite data by estimating the displacement of sea ice features found in two consecutive images (e.g., Agnew et al., 1997; Kwok, 2000). This can be accomplished using imagery from a variety of satellite instruments. 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.

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). Comparisons to the drifting-buoy data show an rms error of the satellite derived drift speed of 4.9 cm/s with a mean bias of 2.4 cm/s. The digital atlas focuses on Antarctic sea ice variability over periods ranging from sub-monthly to interannual.

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

The drift of sea-ice is primarily forced by the winds and ocean currents. On time scales of days to years, the winds explain most of the variance in sea-ice motion (e.g., over 70% of daily ice motion variance is explained by wind forcing -- Thorndike and Colony, 1982). On longer time scales, the patterns of ice motion also follow the evolving patterns of wind forcing. Gudkovich (1961) hypothesized the existence of two regimes of Arctic ice motion driven by large scale variations in atmospheric circulation. Using a coupled ocean-ice model, Proshutinsky and Johnson (1997) showed that the regimes proposed by Gudkovich (1961) alternated on 5–7 year intervals between anti-cyclonic and cyclonic circulation patterns. Similarly, Rigor, et al. (2002) showed that the changes in the patterns of sea-ice motion from the 1980’s to the 1990’s is related to the Arctic Oscillation (AO). For example, in Figure 4.4.5, the mean field of ice motion for 1979 shows some typical features during low AO conditions, i.e., a large Beaufort Gyre and a transpolar drift stream sweeping along the Eurasian coast. The sea ice motion field for 1994 exhibits a smaller Beaufort Gyre, and a transpolar drift steam which sweeps across a broad area of the central Arctic. This comparison illustrates the dominant role of the AO in driving interannual changes in Arctic ice motion, which in turn have a profound affect on the production of sea ice during winter, through the increased advection of sea ice away from the Eurasian Coast, increased Ekman divergence, and transport of ice out of the Arctic Basin.

[INSERT FIGURE 4.4.5 HERE]

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., 2002). No mean trend of ice motion has been detected from the limited data available for the past 25 years.

4.4.4.3 Ice export and advection; freshwater fluxes

The sea ice outflow through Fram Strait is a major component of the mass balance of the Arctic Ocean. Approximately 14% of the sea ice mass is exported each year through the Fram Strait. Vinje (2001) constructed a time series of ice export during 19502000 using available observations and a parameterization based on geostrophic wind, and finds 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 AO or NAO index. Using the thickness data of Vinje et al. (1998), they estimate a mean volume flux of 2366 km^3. Subsequent modelling work by Hilmer and Jung (2000) indicated that the correlation between NAO (or nearly equivalently, the AO) and Fram Strait ice outflow is somewhat transient, with significant correlation during the period 1978–1997, but no correlation during the period 1958–1977 (Figure 4.4.6). 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 4.4.6 HERE]

4.5 Changes in Glaciers and Ice Caps

4.5.1 Background

Glaciers and Ice Caps (G&IC) have a high ratio between mass turnover and mass storage, and so adapt to changing climate conditions more rapidly than do the ice sheets (Figure 4.1.1). G&IC changes reflect changing climate, affect fresh water availability in many mountains and surrounding regions, and have direct input to oceans. Those G&IC not immediately adjacent to the large ice sheets of Greenland and Antarctica cover an area between 512 and 540H10^3 km^2 according to inventories from different authors; volume estimates differ considerably from 51 to 133H10^3 km^3, with respective sea level equivalents, SLE, between 0.15 and 0.37 m (Table 4.5.1). The sole estimate for the G&IC that surround the ice sheets is 0.34 ± 0.06 m SLE. Area inventories are still incomplete, and volume measurements more so, despite increasing efforts.

The surface mass balance of a glacier—the gain or loss of mass 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 that leads to accumulation dominating in winter, and ablation in summer. In the low latitudes, ablation occurs year-round and accumulation periods are determined by precipitation seasons. Small and steep glaciers are controlled primarily by vertical gradients in mass balance, whereas horizontal gradients are more important on larger and flatter glaciers. Atmospheric temperature lapse rates, precipitation gradients and the balance between melting and sublimation (which consumes much more energy than melting) are the primary controls for the mass balance gradients. Accordingly, gradients and associated 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. 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 which allows the mass balance to become zero again. Mass balance always tends toward zero, although climate variability and the time-lag of glacial response prevent a static equilibrium. Changes in geometry lag 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 G&IC with small slopes and cold ice.

Table 4.5.1: Extents of G&IC as compiled by Dyurgerov and Meier (2005) from different sources. Area (10^3 km^2), volume (10^3 km^3), and respective sea level rise equivalent, SLE (m) are given.

Source	Area		Volume / SLE		G&IC + AA + GL	
	G&IC[a]	AA+GL[b]	G&IC[a]	AA+GL[b]	Area	Volume/SLE
MB-96	540	140			680	180 0.50[c]
RB-05	522 ± 42		87 ± 10 0.24 ± 0.03			
O-04	512		51 0.15			
DM-05	540 ± 30	245 ± 100	133 ± 20 0.37 ± 0.06[c]	125 ± 60 0.34 ± 0.17[c]	785 ± 100	258 ± 65 0.71 ± 0.2[c]

Notes:
MB-96: Meier and Bahr (1996): volume derived from a statistical relationship between glacier volume and area, calibrated with 144 glacier volumes derived from radio-echo-sounding measurements
RB-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.
DM-05: Dyurgerov and Meier (2005): volume estimates as in MB-96.
(a) G&IC excluding those in Antarctica and Greenland outside the ice sheets
(b) Ice caps in Antarctica and Greenland outside the ice sheets
(c) In cases where SLE is not given by the authors it is calculated by dividing the ice volume by the ocean area of 362 10^6 km^2.

4.5.2. Large and Global Scale Analyses and Interpretation of Measured Glacier Changes

Records of directly measured glacier mass balances are few and stretch back only to the mid 20th century. Because of the very intensive fieldwork, they are biased toward logistically "easy" glaciers. Uncertainty of directly obtained annual surface mass balance is typically ±0.2 m water equivalent 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). Only a few individual series stretch over the entire period. From these statistically small samples, global estimates have been obtained by area weighting (Dyurgerov and Meier, 2005; Ohmura, 2004) and by spatial interpolation (Cogley, 2005) (Table 4.5.2). Applied for the common period 1967/1968 to 1996/1997 and for G&IC excluding those around the ice sheets, the different approaches estimate sea-level equivalent mass loss between 0.28 and 0.44 mm a^{-1}, the mean being 0.36 mm a^{-1}. In addition to directly measure mass balances, Dyurgerov and Meier (2005) have also incorporated recent findings from altimetry evaluations of G&IC in Alaska (Arendt et al., 2002) and Patagonia (Rignot et al., 2003) in their data base..

Table 4.5.2. Global mean annual specific mass balance, b^*, the respective glacier area, A, the resulting total mean annual mass balance, B^*, and the respective rate of sea level rise equivalent, SLE_{B^*}, for G&IC excluding those surrounding the ice sheets and for the common period 1967/1968–1996/1997.

Sources	(Ohmura, 2004)	(Cogley, 2005)	(Dyurgerov and Meier, 2005)
b^* (mm we)	–265	–185[a]	204[a]
A (10^3 km^2)	512	539[a]	540[a]
B^* (km^3 we)	–136	–100[a]	–160[a]
SLE_{B^*} (mm a^{-1})	0.37	0.28[a]	0.44[a]

Notes:
(a) calculated from the numbers given in the references.

For those G&IC outside of Greenland and Antarctica, mass loss for 1992/1993–2002/2003 was twice as large as for 1960/19611997/1998 (Table 4.5.3 and Figure 4.5.1). The specific balance, b^* (the average thickness gained across the surface of a glacier over a year) during the earlier period was especially negative in Patagonia, followed by Alaska + Coast Mountains and the NW USA + SW Canada (Table 4.5.3, Figure 4.5.2). Only Europe showed a mean value close to zero, reflecting the strong mass losses in the Alps being compensated by mass gains in maritime Scandinavia until the end of the 20th century. Also over the most recent decade, Patagonia and Alaska + Coast Mountains had by far most negative b* followed by NW USA + SW Canada and Europe, where Norwegian glaciers have changed sign and added to the mass losses of the Alps. In terms of absolute mass losses, the extended Alaska + Coast Mountains G&IC have contributed most, followed by those in the Arctic and on Asian High Mountains both on a long term and during the last decade. Uncertainty remains due to the fact that measured mass balances generally do not consider iceberg calving, which is only measured and estimated in few cases (Hagen et al., 2003; Rott et al., 1998) but not extrapolated to the global scale. For Svalbard calving is estimated being about16% of the volume lost by melting; in the colder conditions of Severnaya Zemlya, iceberg calving accounts for 35 to 40% of total mass loss from the 5500 km^2 Academy of Sciences Ice Cap (Dowdeswell et al., 2002).

[INSERT FIGURE 4.5.1 HERE]

[INSERT FIGURE 4.5.2 HERE]

Table 4.5.3. Area, A (10^3 km^2), mass losses, *B* (km^3 we), of different G&IC regions worldwide and respective mean annual specific mass balances, b^* (mm we), and mean annual rates of sea level rise equivalents, SLE_{B*} (mm a^{-1}) for different periods (Dyurgerov and Meier, 2005).

		1960/1961–2002/2003			1960/1961–1997/1998			1992/1993–2002/2003		
	A	*B*	b^*	SLE_{B*}	*B*	b^*	SLE_{B*}	*B*	b^*	SLE_{B*}
Arctic	315.0	–1910	–141	0.12	–1302	–109	0.10	–883	–255	0.22
High Mtns. Asia	116.2	–1322	–265	0.09	–1128	–256	0.08	–464	–363	0.12
Alaska + Coast Mtns.	90.0	–2312	–600	0.15	–1892	–550	0.14	–1102	–1110	0.28
NW-USA + SW-CAN	39.2	–848	–503	0.05	–771	–518	0.06	–193	–447	0.05
Europe	17.3	–32	–43	0.00	23	35	–0.00	–72	–378	0.02
Patagonian Ice Fields	19.9	–726	–850	0.05	–588	–780	0.04	–351	–1680	0.09
S.America – PIF	4.7	–21	–103	0.00	–12	–68	0.00	–17	–333	0.00
Antarctica[a]	176.0	–737	–97	0.05	–469	–70	0.03	–408	–211	0.10
Total[b]	778.3	–7953	–240	0.510	–6191	–210	0.43	–3495	–400	0.88
Total[b] SLE_B (mm)		*22*			*17.1*			*9.6*		

Notes:
(a) Includes the Subantarctic islands (7×10^3 km^2)
(b) Not included are glaciers in New Zealand, Kamchatka, Siberia and the tropics for which no respective data or reliable extrapolation are available. Their total contribution to sea level rise over the last half century is considered minor.

The histories of G&IC global-mean mass balance from different authors have very similar shapes despite some offsets in magnitude, (Cogley, 2005; Dyurgerov and Meier, 2005; Green, in review; Ohmura, 2004). Around 1970 mass balances were close to zero or slightly positive in most regions as well as in the global mean (Figure 4.5.3), indicating near-equilibration with climate after the strong earlier retreats particularly during the 1940s. This suggests confidence that the late 20th century glacier wastage is essentially a response to post-1970 global warming (Green, in review).

[INSERT FIGURE 4.5.3 HERE]

Over the last half century, both global mean winter accumulation and summer melting have increased steadily (Dyurgerov and Meier, 2005; Green, in review; Ohmura, 2004), and at least in the northern hemisphere, winter accumulation and summer melting correlate positively with hemispheric air temperature (Green, in review); the negative correlation of net balance with temperature indicates the primary role of temperature in forcing the respective glacier fluctuations. Dyurgerov and Dwyer (2001) have analysed time series of 21 Northern Hemisphere glaciers and have found a rather uniform moderate increase mass turnover, qualitatively consistent with increased precipitation and low-altitude melting with warming. This general trend is indirectly also indicated by reports from Alaska (Arendt et al., 2002), the Canadian Arctic Archipelago (Abdalati et al., 2004) and Patagonia (Rignot et al., 2003), where substantial thinning of ablation areas and moderate thickening of accumulation areas were measured.

Records of glacier length changes are more common and go further back in time (written reports as far back as 1600 in a few cases) than mass balance studies or volume measurements. Furthermore, they are more directly related to low-frequency climate change. Oerlemans (2005, Figure 4.5.4) has constructed a temperature history for different parts of the world from 169 glacier-length records using simplified glacier dynamics that accounts for a response time and climate sensitivity estimated for each glacier. This shows that moderate global warming started in the middle of the 19th century, with about 0.6 K warming by the middle of the 20th century. Following a 25-year cooling period, temperatures rose again after 1970. Much local-regional and high-frequency variability is superimposed on this apparently homogeneous signal. The dataset suggests that the Little Ice Age was at its maximum around 1850 rather than at the end of the 19th century. The model applied, however, does not allow for changing glacier sensitivity over time, which may limit the information before 1900. In fact, analyses of glacier mass balances, volume changes, length variations and homogenized temperature records of the western portion of the European Alps (Vincent et al., 2005) indicate that a 25% positive departure from 20th century average precipitation with little summer temperature change forced 1760 to 1830 glacier growth, reduced winter precipitation caused glacier retreat after 1830 and summer warming only started to be efficient at the beginning of the 20th century. In southern Norway early 18th century glacier advances are also explained by increased winter precipitation rather than cold temperatures (Nesje and Dahl, 2003).

[INSERT FIGURE 4.5.4 HERE]

The continuation and expansion of directly measured mass balance series is a prime demand for monitoring mass loss and sea level contribution from G&IC. Improvement in extrapolation from directly measured mass balances to global estimates via larger scale volume change measurements is also needed. Altimetry methods have recently become increasingly productive because of the use of airborne laser altimetry not least in remote and large areas which cannot be covered by in situ measurements. The expected derivation of climate signals from glacier changes demands enhanced development and application of inverse glacier – climate modelling.

4.5.3 Special Regional Features

Reports on individual glaciers or limited glacier areas indicate ongoing retreat in almost all regions. Important results are summarized below.

For *Taylor Valley, Antarctica*, Fountain et al. (in press) hypothesize 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 including calving 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), started to retreat around 2000 as an almost simultaneous result of reduced winter accumulation and greater summer melting (Kjøllmoen, 2005). Norwegian glaciers further 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 for coastal Scandinavia, glaciers in the *New Zealand Alps* advanced until about 2000 but have started to retreat since then. Increased precipitation may have caused the glacier growth, perhaps associated with more-frequent El Niño events (Chinn et al., 2005).

In the *European Alps*, glaciers lost on average 18% of their area between 1985 and 1999, shrinkage seven times faster than between 1850 and 1973 (Paul et al., 2004). Exceptional mass loss during 2003 removed an average of 2.5 m water equivalent (we) over 9 measured Alpine glaciers, almost 60% higher than the previous record of 1.6 m we in 1996 and four times more negative than the mean loss from 1980 to 2001 (0.6 m we) (Frauenfelder et al., in press). This was caused by extraordinarily high air temperatures over a long period, extremely low precipitation, and albedo feedback from a previous series of negative mass balance years. Strong increases in *Western Alpine* summer ablation since 1982 are attributed to increased summer melting rates and to prolongation of the ablation period into September and October (Vincent et al., 2004).

Most *Himalaya* glaciers have retreated strongly (Solomina et al., 2004; Su and Shi, 2002; Wang et al., 2004). However, several high glaciers in the *Karakoram* are reported to have advanced and/or to thickened at their tongues (Hewitt, 2004) due to enhanced accumulation.

Tropical Glaciers have retreated from a mid 19th Century maximum, following the global trend (Figure 4.5.5). Strong retreat rates in the 1940s were followed by relative stable extents that lasted into the 1970s. Since then, retreat has become stronger again. Small glaciers naturally have the strongest retreat rates, as in all other mountain regions. Tropical glaciers, being in principle very sensitive to both temperature changes and those related to atmospheric moisture, have retreated mostly in response to 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 hygric 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 4.5.5). 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 may have sporadically reached positive annual values even in recent years (Thompson et al., 2002). In contrast to the plateau ice, the glaciers on Kilimanjaro's slopes show a decreasing retreat rate.

[INSERT FIGURE 4.5.5 HERE]

4.5.4 Little Ice Age and Medieval Warm Period

Mountain glaciers in the different parts of the world have shown similar histories of change during the last millennium. Comparing the well-dated records from Alaska, British Columbia, Canadian Rockies, and Patagonia Luckman and Villalba (2001) demonstrated a broad synchronism in the initiation and timing of main glacial events over the last millennium (like the Little Ice Age, LIA, 13th-19th Century) in the extra-tropical North and South America. This pattern is common for most mountain regions of Europe, Asia, and New Zealand, where such records are available (Grove, 2004). In most parts of the World the mountain glaciers reached their maximum and fluctuated around these advanced positions from 17th to 19th centuries, then experienced a retreat, which continues to present. Since the maximum of the LIA the ELA has risen by approximately one hundred meters in the temperate and sub-polar regions and by about two hundred meters in the subtropical areas.

The information on the glacier variations during the beginning of the second millennium A.D. (the Medieval Warm Period, MWP, 10th -13th Century) is rather contradictory. In most cases we do not know exactly how far and for how long the glaciers had retreated between the advances recorded by the frontal moraines. In general glaciers in the Alps, Scandinavia and Alaska and Patagonia (soil horizons and trees buried in the moraines, pro-glacial lake sediments evidences) seem to experience a century-long (or longer) retreat before the beginning of LIA in 13th century. On the other hand glacier advances are recorded between 1050 and 1150 AD (Grove, 2004) in the Alps, Alaska, North and South Patagonia, British Columbia, New Zealand, Franz Josef Land, SE Tibet. Advances in the Alps, Franz Josef Land, N Patagonia, and SE Tibet were very prominent. Despite the low accuracy of some of these dates, the event(s) seems to be of global importance, but do not have yet a plausible explanation.

4.5.5 Changing Runoff from G&IC and Glacier Related Hazards

4.5.5.1 Changing runoff

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 (Figure 4.5.6), followed by significantly diminished runoff totals and diurnal amplitudes as the glaciers continue to shrink. Effects are particularly strong on water availability in parts of the low latitude Andes (Kaser et al., 2003) and the Himalaya and their semi-arid surroundings.

[INSERT FIGURE 4.5.6 HERE]

4.5.5.2 Glacier related hazards

Formation of large lakes is occurring as glaciers retreat from prominent Little Ice Age moraines, especially in the steep Himalaya (Mool et al., 2001a; Mool et al., 2001b; Yamada, 1998) and the Andes (Ames, 1998; Kaser and Osmaston, 2002). At the same time, thawing of buried ice is destabilizing these moraines. These lakes thus have a high potential for Glacier Lake Outburst Floods (GLOFs). Governmental institutions in the respective countries (Nepal, Bhutan, Peru) have provided extensive safety work and many of the lakes are either solidly dammed or drained. Due to the remoteness of some valleys, it is difficult to assess the actual situation. It is estimated that 20 potentially dangerous glacial lakes still exist in Nepal and 24 in Bhutan (Yamada, 1998), several in the Cordillera Blanca and other Peruvian Cordilleras, recommending vigilance.

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 over 60 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 is discussed here, with consideration of possible future changes deferred until Chapter 10. Balance assessment remains difficult, but much progress has been made since the TAR.

4.6.2.1 Techniques

Mass balance is typically determined by measuring the difference between input from snow accumulation and output by ice flow and meltwater runoff, or by measuring volume change of the ice through repeated surface altimetry. “Weighing” the ice sheets using satellite gravity measurements may become important in the future. Issues related to these techniques were recently summarized by the ISMASS Committee (Jacka et al., 2004). Snow accumulation is the primary input, mainly by precipitation but also by vapor deposition and drifting snow. Accumulation is calculated from the thickness and density of snow deposited over some time interval determined by counting of annual layers in ice cores (McConnell et al., 2001), monitoring burial of poles placed in the snow surface (Mosley-Thompson et al., 1999), or other techniques (Jacka et al., 2004). Remote-sensing may be useful in estimating accumulation, especially for interpolation between measurement sites (e.g., Winebrenner et al., 2001; Vaughan et al., 1999). Increasingly, atmospheric-modelling techniques are proving valuable, based on operational analyses (Cullather et al., 1998), reanalysis products, general circulation models (Genthon and Krinner, 2001) or regional models (van Lipzig et al., 2002; Bromwich et al., 2004).

Mass discharge by ice flow is calculated from ice thickness and velocity distribution with depth, normally where the ice begins to float and velocity is nearly depth-independent. Ice thickness is measured by radar, or by seismic techniques. Surface velocity is measured by repeated conventional or GPS surveys, or by interferometric synthetic-aperture radar (InSAR), which is becoming increasingly important (e.g., Joughin et al., 2002). Advances since the TAR have been the combination of surface-elevation mapping (Liu et al., 1999) with improved ice-thickness measurements to obtain a better estimate of the volume of the Antarctic ice sheet (Lythe and Vaughan, 2001), and widespread application of InSAR and new thickness measurements to coastal regions of both ice sheets. Calculation of mass discharge also requires estimates for runoff of surface meltwater, which is large for low-elevation regions of Greenland but small for Antarctica except for the Antarctic Peninsula, and for basal melting of grounded ice, which is generally assumed to be small but can be quite large on fast glaciers. Surface-melt estimates usually are from modelling driven by weather analyses or climatology. This may involve full energy-balance treatments, but more commonly uses a positive degree-day approach (Box et al., 2004). Mass loss from melting beneath ice shelves can be very large (Rignot and Jacobs, 2002).

Surface-elevation changes reveal changes in ice-sheet mass after correction for changes in ice-sheet density and bedrock elevation, or for hydrostatic equilibrium if the ice is floating. Important results have come from ERS-1 and ERS-2 satellite radar altimetry (e.g., Shepherd et al., 2002; Davis et al., 2005), laser altimetry from airplanes (Krabill et al., 2004), and the new ICESat laser altimeter (Thomas et al., in press). Gravity data can contribute to correction for isostatic changes in bedrock elevation, and field data and models can contribute to correction for density changes (Cuffey, 2001). Vaughan et al. (1999) noted that the balance technique requires knowledge of mass input and mass output to accuracies of order 1% to help attribute ongoing sea-level changes. Arthern and Hindmarsh (2003) argued that a combination of the elevation-change technique and the input-output technique yields better estimates than either taken separately.

4.6.2.2 Measured balance of the ice sheets

Mass balance of the large ice sheets was recently summarized by Rignot and Thomas (2002) and Alley et al. (2005a, in press). Between 1993–1994 and 1998–1999, Greenland lost at least 50 Gt of ice a^{-1} (>0.13 mm a^{-1} sea-level rise). Repeat satellite and airborne altimetry and the mass-budget technique showed the ice sheet to be near balance above ~2000 m elevation. In contrast, repeat airborne altimetry showed widespread near-coastal thinning, especially along fast ice streams and outlet glaciers. Moreover, mass loss between 1997 and 2003 increased by about 50%, both from increased surface melting and from increased glacier discharge (Figure 4.6.1, Krabill et al., 2004). An independent analysis by Box and coworkers (Box and Rinke, 2003; Box et al., 2004; Box, in press; Box et al., in review) reached a similar conclusion. They used a calibrated version of the Polar MM5 mesoscale atmospheric model forced by ECMWF reanalysis data from 1991–2003 to estimate all surface mass-balance terms for the ice sheet. Average accumulation (514 Gt a^{-1}) agreed well with a mean of 509 Gt a^{-1} for 8 other recent studies (van der Veen, 2002). Comparing this to estimated ice-flow output of 271 Gt a^{-1} from Reeh et al. (1999), and to meltwater runoff estimated from MM5 (337 Gt a^{-1}) indicated average ice-sheet balance of –94.1 Gt a^{-1}, (0.24 mm a^{-1} sea-level rise); interannual variability was large, with individual-year balances ranging from +52 to –271 Gt a^{-1}. Moreover, estimated output did not include recent dynamic changes, such as doubling of the speed of Jakobshavn Isbrae between 1997 and 2003 (Thomas et al., 2003; Joughin et al., 2004).

[INSERT FIGURE 4.6.1 HERE]

Hanna et al. (2005) reconstructed the surface mass balance of the Greenland ice sheet on a 5 × 5 km grid for the period 1958–2003. Meteorological models forced by ERA-40 reanalysis data for 1958–2001 and ECMWF operational analyses for 2002–2003 were used to retrieve annual precipitation-minus-evaporation and monthly surface temperature to drive the runoff/retention degree-day model of Janssens and Huybrechts (2000). The surface mass balance shows a high year-to-year variability. There are distinct signals (low peaks) in runoff following the major volcanic eruptions of Agung (1964), El Chicon (1983) and Pinatubo (1992) (Figure 4.6.2). Runoff losses from the ice sheet were 264 ± 26 Gt a^{-1} for 1961–1990 and 372 ± 37 Gt a^{-1} for 1998–2003. Significantly rising runoff since the 1990s has been partly offset by increased precipitation. Using the TAR Table 11.5 average numbers for iceberg calving and bottom melting, the best estimate of overall mass balance declined from +22 ± 51 Gt a^{-1} for 1961–1990 to –36 ± 59 Gt a^{-1} for 1998–2003, which is not statistically significant. These results exclude dynamical effects from accelerating glaciers. Three highest runoff years are 1998, 2003, 2002; the three lowest runoff years 1992, 1964, 1983; three highest surface mass balance years: 1972, 1996, 1983; three lowest mass balance years: 1998, 1968, 1971; none of the years between 1958 and 2003 had a negative surface mass balance.

For Antarctica, assessments are less confident, and widespread agreement has not been reached on the sign of any change. Important regions remain undersampled, accumulation-rate and temperature trend retrievals are less reliable, and the role of changing near-surface density in elevation change is more difficult to assess. Mass-balance estimates rely heavily on satellite altimetry that includes important corrections linked to changes in power returned to the sensor (e.g., Davis et al., 2005), but these are not widely “ground-truthed”.

[INSERT FIGURE 4.6.2 HERE]

Vaughan et al. (1999) estimated average snow accumulation of 1811 Gt a^{-1} for grounded ice, and 2288 Gt a^{-1} including ice shelves and ice rises, in broad agreement with other estimates. Errors were estimated at about 5%, but van der Veen (2002) suggested that 15% may be more appropriate. Jacobs et al. (1992) estimated total loss from the ice sheet and ice shelves to be 2613 ± 530 Gt a^{-1}, yielding a balance of –325 ± 594 Gt a^{-1}. This is not significantly different from zero, but encompasses anywhere between 2.5 mm a^{-1} sea-level rise and 0.7 mm a^{-1} sea-level fall if all came from non-floating ice. Rignot and Thomas (2002) constrained this more closely using improved estimates of glacier velocities from InSAR. For East Antarctica, growth was most probable at 22 ± 23 Gt a^{-1}. In contrast, West Antarctica’s balance of –48 ± 14 Gt a^{-1} was significantly negative. More recently, Rignot et al. (2005) concluded that the Antarctic Peninsula is also losing mass. Satellite radar-altimetry coverage extends only to within about 900 km of the poles and cannot resolve changes in steep marginal regions; the interior parts mainly in East Antarctica well-monitored by ERS-1 and ERS-2 thickened during the 1990s, equivalent to growth of about 45 Gt a^{-1} given certain assumptions about evolution of the near-surface density structure (Davis et al., 2005; Vaughan, 2005; Figure 4.6.3). (A region with constant elevation over time may be in balance, may be losing ice but gaining snow--expected during times of rising snowfall--or may be losing snow but gaining ice, introducing possibly important errors that cannot be quantified accurately without better calculations of the transformation of snow to ice driven by accurate accumulation-rate and temperature histories; bedrock motions introduce additional uncertainty.) Reanalysis of ice input and output of the drainage basins feeding the Filchner-Ronne ice shelf from portions of East and West Antarctica indicates slight thickening (39 ± 26 Gt a^{-1}) consistent with the altimetry results (Joughin and Bamber, in press), in a region where Rignot and Thomas (2002) earlier had found little change, possibly reflecting recent accumulation-rate increase.

[INSERT FIGURE 4.6.3 HERE]

The short intervals over which balance estimates are available are clearly of concern. Jakobshavn Isbrae slowed and thickened slightly between 1985 and 1992 before rapidly accelerating after 1997 (Joughin et al., 2004). The Siple Coast of West Antarctica was thinning when assessed in 1987, but switched to thickening because of a slowdown in Whillans ice stream (Joughin et al., 2002). Glacier acceleration in the Amundsen Sea sector of West Antarctica seems to be recent (Joughin et al., 2003) with thinning progressively increasing (Thomas et al., 2004). Longer periods of observation will be required to improve confidence in separation of long-term trends from natural variability and in identification of causes.

4.6.2.3 Measured balance of the ice shelves

Most ice shelves are in Antarctica (Figure 4.6.4), 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. The largest ice shelves (Ross and Filchner-Ronne) are each as big as Spain, with thickness increasing from about 300 m at the seaward ice front to as much as 2 km at the grounding line where ice first becomes afloat. They fill large embayments in the Antarctic coast, and are each fed by numerous ice streams or outlet glaciers. Most ice shelves are far smaller, some occupying bays fed by individual glaciers, others fringing the coast and protected from breakup by ice rises or ice rumples. Pushed by tributary glaciers and spreading under their own weight, ice shelves move seaward at speeds ranging from a few tens to thousands of m/yr. Typical ice shelves are restricted to regions so cold that, even at sea level, there is little summer melting, although some surface melting is observed on a few small shelves along the Antarctic Peninsula and in Greenland. Instead, basal melting and ice-shelf spreading tend to balance thickening from snowfall (or, occasionally, sublimation in strong katabatic winds) or localized basal freezing. Basal freezing typically is slow where it occurs but frozen-on ice is important toward the fronts of a few ice shelves; basal melting is more typical, and melt rates can reach tens of m/yr. Iceberg calving, by poorly quantified mechanisms, occurs at the seaward edges of ice shelves. Terminology for ice shelves versus ice tongues is imprecise with some overlap, but ice shelves are usually wider or have more tributary glaciers.

[INSERT FIGURE 4.6.4 HERE]

Until recently, it was extremely difficult to determine whether ice shelves were thickening or thinning, and most studies assumed ice shelves to be in steady state with thickness profiles unchanged over time. Then, using volume-continuity requirements, basal-melting rates were inferred from measurements of ice thickness, snow accumulation rates, and ice-shelf velocity (Figure 4.6.5). This showed basal melting to predominate, averaging about 0.4 m a^{-1} in Antarctica (Jacobs et al., 1996), but including basal freezing under large parts of the Ronne Ice Shelf, and melting of order 1 m a^{-1} near seaward ice fronts and by up to tens of m a^{-1} beneath deeper ice near inland grounding lines (Rignot and Jacobs, 2002; Joughin and Padman, 2003).

[INSERT FIGURE 4.6.5 HERE]

Occasional calving of very large icebergs from the Ross and Filchner-Ronne ice shelves is approximately balanced by seaward motion if averaged over sufficiently long times, and there is no indication of long-term thickening or thinning, By contrast, progressive break up of ice shelves has occurred along the Antarctic Peninsula, beginning in the late 1980s. In early 2002, almost all of Larsen-B ice shelf, about 3300 km^2 in area, broke into small fragments in less than 5 weeks (Scambos et al., 2003). Very soon after breakup, the speeds of glaciers that had previously drained into Larsen-B ice shelf 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).

During the 1990s, precise satellite-altimeter measurements gave time-series from which the rate of change of ice-shelf surface elevation can be inferred; after correction for tidal rise and fall, these give estimates of ice-thickness change (Shepherd et al., 2003; 2004). Results show thinning of the remaining part of Larsen Ice Shelf, and of most ice shelves along the Amundsen Sea coast.

Until recently, Greenland's fastest glacier – Jakobshavn Isbrae – flowed at ≈7 km a^{-1} into a floating ice tongue or narrow ice shelf about 6 km wide and 15 km long, wedged between the walls of a long fjord connected to the Davis Strait. Near-annual aircraft surveys of glacier surface elevations showed little change during the early 1990s, followed by a rapid surface lowering of both the floating tongue and the lower reaches of the glacier (Thomas et al., 2003). Buoyancy of the floating tongue implies that associated thinning of more than 300 m occurred in 4 years, and was followed by breakup of much of the floating tongue (Joughin et al., 2004). Thinning and retreat of other ice shelves/tongues in Greenland has been documented as well, coupled with faster flow of adjacent ice streams/outlet glaciers (Abdalati et al., 2001). Moreover, ice-shelf changes are unprecedented over several millennia (Pudsey and Evans, 2001; Brachfield et al., 2003) for the Antarctic Peninsula ice shelves, and for more than 40 years for the Jakobshavn ice tongue (Sohn et al., 1998).

4.6.3 Causes of Changes

4.6.3.1 Changes in snowfall and surface melting

For Greenland, Box (2002) compared long instrumental records primarily from coastal stations, to simulations with the Polar-MM5 mesoscale model forced by ECMWF reanalysis data over the years 1991–

2003. The data reveal a complex picture, with seasonal and spatial patterns and strong dependence of some records on trends in the North Atlantic Oscillation. General warming over the 1990s, especially in west Greenland and perhaps at the Summit, is consistent with the trend to greater snowfall and melting in the Polar-MM5 integrations (Box and Rinke, 2003; Box et al., 2004; Box, in press; Box et al., in review; Hanna et al., 2005), but warming generally did not exceed values observed during the 1930s. The Polar MM5 simulations show much variability, with significant trends (>95% confidence) to higher accumulation rate (increasing by ~10 Gt a^{-1}) and to higher meltwater runoff (increasing by ~16 Gt a^{-1}), but with the resulting small negative trend in total mass balance not highly significant. The trends generally continued past the 1990s, with 2001–2003 the three highest years for accumulation, and the first, second and fourth highest years for melting, in the thirteen-year time series. These results are consistent with expectations (summarized in the TAR) that warming increases low-altitude melting and high-altitude precipitation in Greenland, with sufficient warming leading to dominance of the melt increase.

Studies based on reanalysis products and other available data (Bromwich et al., 2004; Davis et al., 2005) indicate a trend to higher accumulation rate in Antarctica over recent decades, although uncertainty remains about the magnitude (also see Mosley-Thompson et al., 1999). Higher accumulation rate is expected from warming, which is indicated especially in coastal regions. Turner et al., 2002 found a spatially weighted warming trend of 0.18 C/decade from 1958–2002, similar to Vaughan et al. 2001 (also see Thompson and Solomon 2002). Van den Broeke (2000) gives a background Antarctic warming trend of +0.13 ± 0.38°C/century, representative of the period 1957–1995, after correcting available temperature records for decadal circulation variability in the Southern Hemisphere. Turner et al. (2005) investigated Antarctic temperature trends over the last 50 years for 19 stations with long records. Eleven of these had warming trends and seven had cooling trends in their annual data, indicating the spatial complexity of change that has occurred across the Antarctic in recent decades.

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 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. Hence, based purely on modelling or available long-term data, it is unclear whether there remains a slow response in ice-sheet volume. This greatly complicates attribution of century-scale trends in sea level.

4.6.3.3 Dynamic response to recent forcing

Numerous recent papers 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 in Greenland (Thomas et al., 2003; Joughin et al., 2004), although the slowdown of Whillans and Bindschadler Ice Streams on the Siple Coast of West Antarctica is also of interest (Joughin and Tulaczyk, 2002). The combined effect of these cases – ice-sheet mass loss of roughly 120 Gt a^{-1} – is notable. All of these changes except those on the Siple Coast appear to share rapid inland response resulting from warming-induced reduction or loss of ice shelves (Thomas, 2004; Thomas et al., in press; Shepherd et al., 2004; Dupont and Alley, 2005, Payne et al., 2004; Rignot et al., 2005); the Siple Coast changes likely do not reflect recent forcing (Parizek et al., 2003).

The ongoing changes support those theoretical and modelling papers showing that ice shelves confined in embayments or with ice rises act to restrain motion of tributary glaciers (e.g., Thomas, 1979), so that thinning, removal, or marginal weakening of an ice shelf will speed the flow of inland ice. The speed-up is accomplished both by the direct effect of reduced longitudinal stresses (Thomas, 2004; Thomas et al., in press; Payne et al., 2004; Dupont and Alley, 2005), and by the rapid advective-diffusive ice-stream response to the thinning resulting from the reduced longitudinal stresses (Payne et al., 2004). In the cases considered here, the time scales (years or less for response) and response style (largest at the coast, decreasing inland, following ice streams) are those expected for loss of ice-shelf buttressing. Importantly, careful model runs for ice-sheet behaviour over the last century, using known forcings and flow processes but omitting the buttressing effects of ice shelves through transmission of longitudinal stresses, match overall ice-sheet trends rather well but fail to show these rapid marginal thinning events, increasing the likelihood that the changes are in response to the ice-shelf-buttressing and longitudinal-stress processes not included in the models (Huybrechts et al., 2004). Notably, even short ice shelves (kilometres or tens of kilometres rather than hundreds of kilometres in length) are shown to be quite important in the flow of the ice feeding them (Thomas, 2004; Thomas et al., in press).

4.6.3.4. Melting of ice shelves

The largest and fastest ice-sheet changes thus appear to be at least in part response to ice-shelf shrinkage or loss, focusing attention on the balance of ice shelves. 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 indicate that the ice-shelf changes have resulted from environmental warming, with both oceanic and atmospheric temperatures important.

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 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 a (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 sea water (Scambos et al., 2000); the speed of breakup may reflect falling-domino-style transmission of forces from capsizing icebergs (MacAyeal et al., 2003). Very soon after breakup, the speeds of glaciers that had previously drained into Larsen-B ice shelf 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), indicating that glaciers do respond to ice-shelf changes.

Observed thinning of Amundsen Sea coast ice shelves despite increased ice supply from tributary glaciers and in the absence of notable surface melting implicates increased basal melting in the thinning (Shepherd et al., 2003; 2004). 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 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). Similarly, oceanic processes seem to have contributed to the thinning and near-loss of the ice shelf in front of Jakobshavn Isbrae in Greenland, although some contribution from surface melting and perhaps even from meltwater lubrication (see below) may have occurred. The ice shelf in front of Jakobshavn survived temperatures warmer than recently during the 1950s (Thomas et al., 2003), implicating oceanic heat transport in the changes of the last decade.

The basal mass balance of an ice shelf thus clearly depends on ocean circulation beneath the ice shelf. Isolation from direct wind forcing means that the main drivers of sub-ice-shelf circulation are tidal and density (thermohaline) forces, with tides primarily a source of energy for vertical mixing (MacAyeal, 1984). Thus, thermohaline circulation has been the subject of most modeling studies. Density variations are generated in the water column beyond the ice front, most notably by wintertime increases in salinity driven by sea-ice growth, and beneath the ice shelf, as a result of melting and freezing. Basal melting cools and freshens the upper part of the water column, producing a stable stratification. The cold water must be removed and replaced by upwelling of warmer, saltier water if melting is to be sustained. Because the ice-shelf base slopes upwards from grounding line to ice front, the vertical stratification induced by melting drives meltwater towards the ice front and draws warm, salty water towards the grounding line. This motion causes mixing and upwelling that sustain melting and maintain forcing on the overturning circulation. Early models (MacAyeal, 1984; Hellmer and Olbers, 1989) focussed mainly on this process, which was found to respond sensitively to changes in temperature of the inflowing salty water. Higher temperatures give rise to more melting, which increases thermohaline forcing, leading to a stronger circulation and further increase in the heat delivered to the ice-shelf base. However, the models assumed an essentially infinite supply of warm inflowing water with defined properties.

In Antarctica, warm inflows are derived from two sources (Figure 4.6.6). Around much of the coastline, the continental shelf seas are dominated by High Salinity Shelf Water (HSSW), formed at the surface melting temperature by salt rejection beneath growing sea ice. Compression during sinking lowers the freezing point below the water temperature, so that HSSW is capable of melting shelf ice, and in most regions HSSW is the warmest water on the continental shelf. Much warmer (3°C above the surface freezing point) Circumpolar Deep Water (CDW) replaces HSSW as the densest water on the continental shelves of the Amundsen and Bellingshausen seas. CDW drives much higher melt rates, and temperatures in the outflow from beneath ice shelves remain high enough to prevent refreezing. Sometimes this outflow reaches the sea surface, where it can melt the sea-ice cover. CDW is the principal mid-depth water mass of the Southern Ocean, but only flows onto the continental shelf in large quantities in the Amundsen and Bellingshausen seas.

[INSERT FIGURE 4.6.6 HERE]

The application of three-dimensional ocean general-circulation models to the study of the waters beneath and in front of ice shelves (e.g., Beckmann et al., 1999; Holland and Jenkins, 2001) has enabled the source of the inflows to be addressed. Most studies have focussed on the large ice shelves of the Ross and Weddell seas, where HSSW provides the inflow. Forcing on the open ocean has been derived by restoring the surface to prescribed values of temperature and salinity (Jenkins et al., 2004), or using a sea-ice model forced by idealised (Grosfeld and Gerdes, 1998) or realistic winds (Timmermann et al., 2002). Results suggest a seasonal supply of inflowing HSSW generated during winter, with the cavity cooling as the supply of HSSW drops in the summer (Nicholls, 1997; Jenkins et al., 2004). Inter-annual changes in sea-ice distribution, resulting from calving of giant icebergs from the ice front or variability in the wind fields, were found to alter the circulation pattern and the resulting net melt rate (Nøst and Østerhus, 1998; Grosfeld et al., 2001; Timmermann et al., 2002). Nicholls (1997) suggested that moderate climatic warming leading to decreased sea-ice production could reduce the supply of HSSW and hence reduce net melting at the base of these large ice shelves. Grosfeld and Gerdes (1998) concluded that this was the most likely response of the Filchner Ice Shelf system to moderate climate warming. In contrast, Williams et al. (2002) used a model with prescribed inflow properties to simulate the response of Amery Ice Shelf to oceanic warming, and concluded that the increase in mass loss for each degree of temperature increase was approximately twice the total loss under current conditions. Such warming could occur only through a gradual replacement of HSSW as the dominant dense water mass on the continental shelf.

Lack of knowledge of the sub-ice bathymetry has hampered the use of three-dimensional models to simulate circulation beneath the thinning ice shelves of the north-western Weddell, the Bellingshausen and the Amundsen seas. Thus, the role of basal melting in driving the changes observed in these regions remains speculative. The latter two regions are dominated by CDW (Jacobs et al., 1996), so changes in melting would be driven by changes in its temperature or flow. The earlier model of Hellmer and Olbers (1989) has been applied to Pine Island Glacier (Hellmer et al., 1998) and used to explore the sensitivity of melting to changes in water temperature. A warming of 1°C was found to double the computed net melt rate. (Simple regression analysis of available data including those from near Pine Island Glacier indicated that 1°C warming of sub-ice-shelf waters increases basal melt rate by about 10 m a^{-1}; Shepherd et al., 2004).

The north-western Weddell Sea represents something of a transition between the two regimes discussed above. True HSSW is absent but the on-shelf version of CDW is so modified by wintertime convection that its temperature is very close to the freezing point (Nicholls et al., 2004). It is possible that reduced sea-ice formation in this area could lead to an increase in the heat carried beneath the ice shelf, in contrast to the reduction postulated for the southern Weddell Sea. However, until better knowledge of seabed topography is available in all these regions, and more sophisticated models are applied to the problems, a detailed understanding will be lacking of how the spatial and temporal variability of the basal mass balance could force rapid ice-shelf changes. The coupled model approach by Grosfeld and Sandhäger (2004) of a dynamic ice shelf and an ocean model for idealized geometries similar to the Filchner Ronne ice shelf system showed that the coupled ice shelf-ocean system is sensitive especially to ocean warming. Increased basal melt rates of 100% for a 0.5°C ocean warming scenario yield an asymmetric development of the ice shelf thickness, suggesting a high vulnerability of ice shelf regions to changed oceanic conditions.

4.6.3.5 Iceberg calving and ice shelf collapse

Iceberg calving is poorly understood, but is currently receiving increasing attention. Understanding the initiation and propagation of fractures in ice shelves is crucial for modeling both the advance and retreat of ice fronts and the catastrophic disintegration of ice shelves. Doake et al. (1998) analyzed modeled stress fields for the Larsen Ice Shelf and pointed out the existence of a "compressive arch", inland of which the ice was laterally confined by the embayment. Seaward of this line the ice was free to spread in all directions. Retreat of the calving front weakens or even breaks this compressive arch, potentially precipitating irreversible ice-shelf collapse. Based on this criterion, Doake et al. (1998) predicted the subsequent collapse of Larsen B Ice Shelf. A similar analysis of the computed stress field was used by Grosfeld and Sandhäger (2004) in their coupled ice shelf-ocean model with cyclical advance and calving of an idealized ice shelf. They identified a line of maximum extension, seaward of the compressive arch, as the site of fracture.

Recent progress towards a more mechanistic theory of crack initiation and growth has been based on application of linear elastic fracture mechanics (van der Veen, 1998). Rist et al. (1999) used a model of the stress distribution in Filchner-Ronne Ice Shelf and a fracture criterion to predict the location of crevassing, while Scambos et al. (2003) applied similar principles to estimate the depth of penetration of surface crevasses on Larsen Ice Shelf and calculate where water-filled crevasses could penetrate the full depth of the ice shelf. More recently, Larour et al. (2004) have applied the principles of fracture mechanics to model the temporal evolution of existing rifts near Filchner-Ronne Ice Front.

Icebergs carry large amounts of fresh water across the Southern Ocean with significant consequences for the oceanic stratification. Only little information is available on the distribution and on the pathways of the icebergs and the associated fresh water release (Schodlock et al., 2005).

4.6.4 Other Issues in Projection of Balance Changes

Projection of balance changes is deferred to Chapter 10. However, some issues related to understanding of the ice sheets and their likely changes are covered briefly here.

4.6.4.1 Data gaps for ice-flow modelling

Ice-flow is often slow, slowly varying in time and space, and can be modelled with considerable accuracy using modern models. However, some regions (ice streams) exhibit much faster flow despite smaller gravitational driving stress, show rapid space- and time-changes in flow, and pose major difficulties for prognostic modelling (see papers in Alley and Bindschadler, 2001). Crossing of the freeze-thaw boundary at the bed of an ice sheet can have order(s)-of-magnitude effects on the ice-flow velocity, with order(s)-of-magnitude uncertainty related to the character of the glacier bed (unconsolidated sediment or rough bedrock). Techniques of basal characterization are well-known, but data on geothermal fluxes for temperature calculations, till versus bedrock distribution (e.g., Anandakrishnan et al., 1998), and bed roughness are restricted to a very small fraction of the ice sheets, precluding accurate prognostic modelling in the face of large perturbations.

4.6.4.2 Fluctuations/Noise in dynamic behaviour

Consideration of ongoing changes in the large ice sheets is complicated by lack of knowledge of the "noise" in ice-sheet behavior. Because some ice-sheet processes have short response times, observed changes may or may not be linked clearly to forcing or to ongoing and predictable changes. As summarized in the papers in Alley and Bindschadler (2001), the Siple Coast/Ross Embayment region of West Antarctica has exhibited much flow variability over the last millennium or longer. The lower reaches of one formerly fast-moving ice stream nearly stopped a century or two ago, a second fast-moving ice stream is slowing in its lower reaches (Joughin and Tulaczyk, 2002), and other changes are occurring or have occurred. However, the changes include speed-up or widening of ice streams as well as slow-down or narrowing (Fahnestock et al., 2000). It remains unclear whether these changes are part of an ongoing trend, which on average would seem to favor slow-down and thus sea-level fall (Joughin and Tulaczyk, 2002), or whether this is "noise" that must be averaged out to observe long-term trends (Parizek et al., 2003).

4.6.4.3 Insights from paleoclimate (model-data mismatches)

Projection of ice-sheet changes is difficult, involving flow within ice, deformation of subglacial materials, sliding over subglacial materials, geothermal fluxes, isostatic deformation of materials beneath and around ice, subglacial lake formation and drainage, ocean-ice and atmosphere-ice coupling involving a great range of processes linked to precipitation, freezing and melting, runoff, etc. Over short times, warming can cause ice growth through enhanced snowfall linked to thermodynamic (higher saturation-vapor pressure) or dynamic processes. However, over longer times, warmer temperatures and reduced ice volume occur together. The warm, high-carbon-dioxide world of the dinosaurs lacked permanent land ice, and recent modeling indicates that growth of ice was caused more by reduction in atmospheric carbon-dioxide concentration than by continental motions (DeConto and Pollard, 2003). Over millennial time scales, ice volume has increased following cooling and decreased following warming of the ice-age cycles, with reduction in ice volume caused primarily by the warming (Huybrechts et al., 2004). Some evidence indicates that the Greenland ice sheet was notably reduced during the previous interglaciation about 125,000 years ago, when temperatures on the ice sheet probably were somewhat warmer than recently (Cuffey and Marshall, 2000). Observations such as these, plus the large volume of water remaining in the ice sheets, motivate additional studies to determine the longer-term (beyond the year 2100, and especially well beyond the year 2100) behavior of the ice sheets.

4.6.4.4 Basal-Lubrication changes from surface meltwater

Zwally et al. (2002) showed for one site near the equilibrium line on the west coast of Greenland that ice-flow velocity 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 the question of 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. This is potentially a very powerful mechanism for thawing ice that is frozen to the bed, allowing onset of perhaps rapid basal sliding or subglacial sediment deformation and possibly causing changes much larger than the 10% observed. Preliminary physical modelling suggests that meltwater access to the bed can migrate inland with warming (Alley et al., 2005b in press).

4.6.4.5 Possible stabilizin feedbacks

It is worth remembering that, despite warming and sea-level rise from the last ice age that caused notable shrinkage of the Greenland and Antarctic ice sheets, they still persist. Hence, stabilizing feedbacks must also exist; the onset of retreat has not (yet) caused runaway collapse. Attention is especially focused on ice-shelf changes propagating into ice sheets. Most modern ice shelves exist in embayments or where anchored by pinning points such as islands. Embayments and islands serve to stabilize the ice shelves both through drag along sides, and through restriction of oceanic circulation that can deliver heat to the sub-ice-shelf cavity to cause melting from below. Further understanding of these processes is essential; ongoing changes may re-form ice shelves deeper in embayments without crossing thresholds leading to greater ice-sheet shrinkage or collapse.

4.6.4.6 Full-Stress-Tensor modelling

Two main traditions in ice-flow modelling have been developed: the inland- or thin-ice approximation, in which the gravitational stress driving ice flow is balanced on the bed immediately beneath the ice, and the ice-shelf approximation, in which stress imbalances are transmitted long distances laterally. As one might expect, the inland-ice approximation is reasonably accurate inland, and the ice-shelf approximation is reasonably accurate on ice shelves. However, the interaction of ice shelves and inland ice, especially through fast-moving ice streams, requires consideration of the full stress state. Because a change in any of the three independent longitudinal-deviatoric stresses and three independent shear stresses affects all six independent deformation components, treatment of the full stress tensor is quite difficult. However, progress is being made towards this goal (e.g., Payne et al., 2004).

4.7 Changes in Frozen Ground

4.7.1 Background

Frozen ground, in a broad sense, includes the near-surface soil freeze/thaw cycle, 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 (Brown et al., 1997; Zhang et al., 1999). 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 mass in the Northern Hemisphere (Zhang et al., 2003), and the maximum area extent of the near-surface soil freeze/thaw cycle is even greater. In terms of the area extent, frozen ground is the single largest component of the cryosphere.

Frozen ground is a product of heat exchange between the ground surface and the atmosphere and is primarily controlled by climate conditions. The permafrost temperature regime is a sensitive indicator of the decade-to-century climatic variability (Lachenbruch and Mashall, 1986; Osterkamp, 2003). The thawing of permafrost can generate dramatic changes in ecosystems, landscape, and infrastructure performance (Nelson et al., 2002). 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 (Solovyev, 2000). Systematic permafrost temperature monitoring started in the 1950s from both the standard hydrometeorological stations up to 3.2 m (Frauenfeld et al., 2004) 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, 2003). Deep permafrost temperature measurements on the Tibetan Plateau, China, were conducted in the early 1960s, while the 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 et al., 2005) and in the 1990s in the Europe through the Permafrost and Climate in Europe (PACE) program (Harris et al., 2003).

Based on the above permafrost monitoring programs from various organizations worldwide, the Global Terrestrial Network for Permafrost (GTN-P), initiated and organized by the International Permafrost Association (IPA), was developed in the 1990s with the long-term goal of obtaining a comprehensive view of the spatial trends and variability of changes in permafrost temperature (Brown et al., 2000; Burgess et al., 2000). The program's two components are: (a) long-term monitoring of the thermal state of permafrost in an extensive borehole network; and (b) monitoring of active-layer thickness and processes at representative locations. Recently, IPA also enhanced the Thermal State of Permafrost program (Romanovsky et al., 2002) as part of IPA's contribution to the International Polar Year. These worldwide permafrost-monitoring programs provide evidence of climate-induced changes.

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 over the last 50–100 years on the North Slope of Alaska (Lachenbruch and Marshall, 1986) although at some sites there was little warming or even cooling trend. Recent measurements conducted by Clow and Urban (see Nelson, 2003) in the same Alaskan norehole network indicated further warming of about 3°C since the mid-1980s. Measurements on the North Slope of Alaska by Osterkamp (2003) show that permafrost temperature has increased about 2 to 3°C (Figure 4.7.1) since the early 1980s. Measurements (Osterkamp, 2003) and modelling results (see Walsh, 2005) indicate that permafrost temperature has increased up to 2°C in the Interior of Alaska. Further analysis indicates that changes in air temperature alone cannot account for the permafrost temperature increase, while increased snow cover may be responsible for a significant proportion of the temperature increase near the surface (Stieglitz et al., 2003).

[INSERT FIGURE 4.7.1 HERE]

Data from the Northern Mackenzie Valley of the continuous permafrost zone show that permafrost temperature between depths of 10 to 20 m has increased about 1.1°C in the past decade (Smith et al., 2003).

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 et al. in press, 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 (Burn 1998; 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 Arctic. This cooling however, was followed by warming of 0.4°C per year between 1993 and 2000 (Smith et al., in press). 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 (Allard et al., 2002; 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 (Romanovsky et al., 2002), about 0.3 to 0.7°C at depth of 10 m in northern West Siberia (Pavlov, 1994), 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). The permafrost temperature increase seems to be connected not only to the increase in air temperature but also an increasing snow thickness preventing stronger cooling of the upper permafrost zone (Pavlov, 1994). Fedorov and Konstantinov (2003) reported that permafrost temperatures from three central Siberian stations did not increase between 1991 and 2000.

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 degrees/decade (Isaksen et al., 2000). Results from five years of continuous ground temperature monitoring in Juvvasshøe indicate that the permafrost is currently also strongly warming. Since 1999 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). Permafrost temperature increased 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. Permafrost temperatures increased 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). 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 in the Da Hinggan Mountains in northeastern China (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	Lanchenbruch and Marshall, 1986
Northern Alaska	20	1983–2003	2–3	Osterkamp, 2003
Interior of Alaska	20	1983–2003	0.5–1.5	Osterkamp, 2003
Canada				
Alert, Nunavut	15–30	1995–2000	0.9	Smith et al., 2003
Northern Mackenzie Valley	10–20	Mid-1980s–2003	1.1	Smith et al., 2005
Central Mackenzie Valley	10–20	Mid-1980s–2003	0.5	Smith et al., 2005

Southern Mackenzie Valley & Southern Yukon Territory	~20	Mid-1980s–2003	0	Couture et al., 2003; Haeberli and Burn, 2002
Northern Quebec	10	Late 1980s-mid-1990s	< -1	Allard et al., 1995
Northern Quebec	10	1996–2001		Allard et al., 2002
Lake Hazen	2.5	1994–2000	1.0	Broll et al., 2003
Igaluit, Eastern Canadian Arctic	5	1993–2000	2.0	Smith 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, 1994
European north of Russia, continuous permafrost zone	6	1973–1992	1.6–2.8	Pavlov, 1994
Northern European Russia	6	1970–1995	1.2–2.8	Oberman and Mazhitova, 2001
Europe				
Juvvasshoe, southern Norway	~5	Past 60–80 years	0.5 to 1.0	Isaksen et al., 2001
Janssonhaugen, Svalbard	~5	Past 60–80 years	1 to 2	Isaksen et al., 2001
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., 2000
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 (Everdingen, 1998). 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 (Thie, 1974; Vitt et al., 1994; Halsey et al., 1995; Laberge and Payette, 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 extend 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 et al., 2003). 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 et al., 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.

Thermokarst topography forms as ice-rich permafrost thaws, either naturally or anthropogenically, and the ground surface subsides into the resulting voids. Extensive thermokarsting 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 Konstantin, 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 et al., 2005). Over the discontunous 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-of-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.

4.7.3 Changes in Seasonally Frozen Ground

4.7.3.1 Changes in the active layer

The active layer is that portion of the soil above permafrost that seasonally experiences thawing and freezing and plays an important role in cold regions because of 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, soil moisture, and duration and thickness of snow cover (Brown et al., 2000; Frauenfeld et al., 2004; Zhang et al., 2005). The inter-annual variation of thaw depth at a site is quite large and consequently, utilizing depth of thaw as an indicator of climatic change may be quite difficult as one would be looking for the response to a subtle change amidst large annual variations. 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 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 more than 30 stations, most of them started in the 1950s but a few was as early as in the 1930s. Over the period 1956–1990, the active layer exhibited a statistically significant deepening by about 20 cm (Figure 4.7.2; Frauenfeld et al., 2004). Changes in air temperature, thawing index, and snow depth are responsible for the increase in active layer thickness.

[INSERT FIGURE 4.7.2 HERE]

The Circumpolar Active Layer Monitoring (CALM) program was developed in the 1990s and currently incorporates more than 100 sites worldwide (Brown et al., 2000). 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, thaw subsidence, and development of thermokarst 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).

4.7.3.2 Seasonally frozen ground in non-permafrost area

Seasonally frozen ground refers to the top layer, which freezes and thaws annually in areas where there is no 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 4.7.3), primarily controlled by the increase in winter air temperature and snow depth (Frauenfeld et al., 2004).

[INSERT FIGURE 4.7.3 HERE]

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. Maximum areal extent of seasonally frozen ground has decreased by 10 to 15% in spring in the 20th century in the Northern Hemisphere (Figure 4.7.4; Zhang et al., 2005). There was little change in the area extent of seasonally frozen ground during the early and mid winters.

[INSERT FIGURE 4.7.4 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). 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. Other hydrologic processes impacted by degrading permafrost include increased winter stream flows, decreased summer peak flows, changes in stream water chemistry, and other fluvial geomorphologic processes (McNamara et al., 1999). Hydrologic changes witnessed among study sites include drying of thermokarst ponds, increased active layer thickness, increasing importance of groundwater in the local water balance and differences in the surface energy balance. By far, the most significant changes occur in response to changing permafrost extent or thickness. As permafrost becomes thinner, the sub-permafrost groundwater becomes more important, either by contributing groundwater to streamflow, or allowing surface water to drain. Thickening of the active layer and melting of ice-rich permafrost in the Russian Arctic drainage basin may have already contributing, in part, to the increased river runoff (Zhang et al., 2005).

The important implications are that in regions over thin permafrost (~<20 m), surface ponds may shrink and surface soils may become drier as the permafrost degrades. In areas over thick permafrost, degradation of massive ice wedges could thaw and catastrophically drain an entire village's water supply. These processes depend upon many complicating factors, such as the regional hydrologic gradients (i.e., whether the region is a groundwater upwelling or downwelling zone). The same mechanisms that allow drying of the ponds may also cause soil drying with significant impacts to latent and sensible heat fluxes. In-depth study and collaboration with villages is needed to project current capacities, future needs and future threats. Village residents need to be trained in appropriate water use, both for current and future water utilization. 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 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 (Frolking et al., 1996; 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 (Jarvis and Linder, 2000; Tanja et al., 2003). With boreal evergreen forests accumulating approximately 1% of annual net primary productivity (NPP) each day immediately following seasonal thawing, variability in the timing of spring thaw can trigger total interannual variability in carbon uptake on the order of 30% (Frolking et al., 1996; Kimball et al., 2004).

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.

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. At Schilthorn the active layer thickness in 2003 was around twice the average of the previous years, and at Stockhorn it increased by around 30%, indicating strong heat conduction coupled with convective heat transfer by water. In the ice-rich frozen debris at Murtèl-Corvatsch, the active layer was deeper than had previously been recorded. 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., 2004b; Schär et al., 2004; Noetzli et al., 2003). It is likely that longer-term permafrost warming observed in the last few decades has also been a factor in more deep-seated destabilisation of rock slopes, such as the rock fall experienced in 2004 near Bormio, Italy.

4.8 Synthesis

4.8.1 Observed changes of the cryosphere

Variations of the cryospheric components are a result of changes in their mass balance, which are determined by dynamic and thermodynamic processes. Dynamic processes determine the motion, i.e. the transport (redistribution) of ice, which is important for sea ice, glaciers and ice caps, and ice shelves and ice sheets; for snow, river and lake ice, and frozen ground it is of minor importance. Thermodynamic processes determine the change of ice mass through melting, freezing and accumulation, which are influenced by energy (radiation and heat) fluxes and precipitation. Areas of net increase (accumulation/freezing) and of net loss (ablation/melting) are generally separate, and, at equilibrium, both are balanced by the motion (redistribution) of ice.

The most important variable for recent changes of the cryosphere is the near-surface air temperature (and the ocean temperature for the marine ice). Meteorological and cryospheric observations show a consistent picture of surface warming (Figure 4.8.1), except for the high southern latitudes, south of 65S, where the sea ice exhibits only a small positive but insignificant trend since late 1970s, which is in accordance with the evolution of the surface temperature. In all other regions of the globe warmer temperatures are in accordance with shrinking ice masses.

The significant decrease of the extent of snow cover and seasonally frozen ground particularly in spring and the earlier break-up of river and lake ice are consistent with the greater temperature increase in spring as opposed to other seasons. Both extent and thickness decline of Arctic sea ice are consistent with the increase in NH temperatures north of 65N. The reduction of sea ice thickness in the Central Arctic is in accordance with strong retreat of sea ice extent in summer. 20th century retreat of mountain glaciers is also consistent with increasing temperatures. Tropical glaciers, retreating synchronously with all others, are mostly driven by changes of seasonal hygric conditions, which are also influenced by global warming.

Ice shelves have shown significant changes in Antarctic, the break-up of Larsen-B and widespread thinning along the Amundsen Sea coast, indicating increased basal melting due to increased ocean heat fluxes in the cavities below the ice shelves. Tributary glaciers near Larsen-B and the thinning ice shelves exhibit strong increases in glacier speed with possible consequences for the adjacent part of the ice sheet. This is also true for the Jakobshavn Isbrae in western Greenland, whose floating tongue thinned drastically during the past 4 years.

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. Nevertheless, the data that are available portray a rather consistent picture of a cryosphere in decline over the 20th century.

[INSERT FIGURE 4.8.1 HERE]

4.8.2 Cryospheric Contributions to Sea Level Change

The most important cryospheric contributions to sea level change arise from changes of the ice on land, e.g., glaciers, ice caps, ice sheets and permafrost. The floating ice shelves and sea ice impact on sea level locally in a similar way as buoyancy fluxes through heating, cooling, evaporation, or precipitation elsewhere on the ocean. 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 from Dyurgerov and Meier (2005) showed that all glaciers outside Greenland and Antarctica contributed ~0.4 mm/yr during 1960–1998, increasing to more than double from 1992–2003 (Table 4.5.3).

Greenland contributed 0.1–0.2 mm/yr to sea level rise between 1993 and 2003, increasing during this time. Contributions from Antarctica were probably small since the mass loss in West Antarctica is roughly balanced by the mass gain in East Antarctica. Both ice sheets combined provided a small contribution of 0.1 mm/year for the past decade, increasing to 0.2 mm/year over the past 5 years.

The total volume of ground ice in permafrost in the Northern Hemisphere ranges from 11,370 to 36,550 km^3, which corresponds to 3–10 cm sea-level equivalent (Zhang et al., 1999). A recent study indicates that thawing of permafrost in Russian Arctic drainage basin might in part have contributed to the runoff increase (Zhang et al., 2005). However, the contribution to sea level rise may be small.

This suggests that the current total ice contribution to sea level rise is approximately 1 mm/yr.

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