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The Earth's Cryosphere: Current State and Recent Changes

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Abstract. The Earth continues to have a third of the ice that it had at the peak of the last ice age, although that ice continues to decrease, as it has, overall, for the past 18,000 years. Over the last 100 years, the retreat signal has been especially strong in ice shelves of the Arctic and along the Antarctic Peninsula, with a more mixed signal elsewhere. For instance, since the early 1990s the massive Greenland and Antarctic ice sheets have thinned along the coasts but thickened in the interior, and since the late 1970s sea ice has decreased in the Arctic but increased (slightly) in the Antarctic. Major difficulties in the interpretations of the climate record come from the high interannual variability of most cryosphere components and the lack of consistent long-term global data records, the latter problem now being slowly remedied, in part, through satellite technology.

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1. INTRODUCTION

Throughout the period of human existence, and for many millions of years before, the Earth has had a substantial cryosphere, with many and varied impacts on the Earth's climate and its ecosystems. As the ice cover advanced equatorward and retreated poleward during the glacial/interglacial cycles, coastlines changed, surface topography changed, and atmospheric and ocean circulations were forced to adjust. Plant and animal species throughout the immediate region of the ice cover were forced also to adjust, mostly by moving out of the way of the advancing ice and then, eventually, reoccupying areas after the ice retreated from them.

Quite counter to common perceptions, the Earth still retains about one-third of the ice cover that it had at the peak of the last Ice Age approximately 18,000-20,000 years ago (1). Scientists have known for some time the basic qualitative impacts of the cryosphere, defined as encompassing the frozen-water (ice or snow) and frozen ground (permafrost) on or just beneath the Earth's land and ocean surfaces. Ice and snow are strong reflectors of solar radiation, reducing the amount of solar energy that is absorbed within the Earth system. They are strong insulators, restricting exchanges between the atmosphere and the underlying surface. They store water, in some locations in vast quantities, with the overall volume of land ice on Earth today being enough to raise sea level worldwide by approximately 70-80 m. They provide recreation and tourist possibilities and sometimes severe transportation hazards. Atmospheric circulation and composition are affected by the topography of the land ice and by sea ice's restriction of evaporative and other chemical fluxes between the ocean and the atmosphere. Ocean circulations and salinity distributions are affected by the discharge of salt to the underlying ocean as sea ice forms and as melt water flushes through the ice, and also by the transport of relatively fresh water through ice dynamics. Icebergs present major hazards to ships in their vicinity although also provide scenic beauty and sometimes much-needed fresh water. Permafrost affects the solidity of the ground, the ability of plant roots to penetrate the ground, and flows of heat and chemicals within the ground and between the ground and the atmosphere. As permafrost decays, considerable structural damage can result to buildings, roads, and other infrastructure built above the permafrost. Additional examples and details of the impacts of land ice (2, 3), sea ice (4, 5), lake and river ice (6, 7), ice shelves (8, 9), icebergs (10, 11), snow cover (12, 13), and permafrost (14, 15) abound.

This article reviews the current understanding of the extent of today's cryosphere and the changes occurring in it (Section 2), then broadens the discussion to the coupled climate system, the issue of causation, and the inadequacy of the data record (Section 3).

2. THE CRYOSPHERE TODAY AND ITS RECENT CHANGES

The Earth's cryosphere includes the following diverse components: snow cover; land ice, formed through the accumulation and consolidation of snow; sea ice, lake ice, and river ice, formed through the freezing of liquid water in oceans and seas, lakes, and rivers, respectively; ice shelves, sometimes formed as former land ice flows outward over water (losing its classification as land ice but remaining attached to the land ice upstream) and other times formed from sea ice and snow; icebergs, formed from the breaking off of land ice or shelf ice into bodies of water; and permafrost, formed by the freezing of ground materials where the ground temperature remains below 0°C for two or more years. Land ice is further divided into ice sheets (covering continent or near-continent sized areas) and mountain glaciers, the latter including ice caps, which cover large portions of mountains or even groups of mountains. This section is subdivided according to these main cryosphere components, with the text in each case first describing the particular component, then describing the recent changes occurring in it.

2.1 Land Ice

The overwhelming majority of the Earth's ice resides in its two remaining major ice sheets, those overlying the continent of Antarctica and the island of Greenland. These ice sheets and the Earth's mountain glaciers are important not just because of the space they occupy and the effects of that occupation but also because as land ice decreases much of the reduced mass is added to the oceans, raising sea levels and potentially causing worldwide coastal damage. Consequently, particular interest lies in the mass balance of the various land ice masses. Mass balance is positive if the accumulation of mass, generally through added snowfall, exceeds the loss of mass, generally through melt, iceberg calving, and flow of land ice into ice shelves; mass balance is negative if the loss of mass exceeds the accumulation.

2.1.1 *Antarctic Ice Sheet*

The Antarctic ice sheet (Figure 1) is by far the largest mass of ice on Earth, spreading over an area of 12.4×10^6 km^2 and containing approximately 25.7×10^6 km^3 of ice (16), with an average ice thickness of approximately 2 km and much of the ice sheet having a thickness exceeding 3 km. Despite the fact that this massive amount of ice exists because of the accumulation and consolidation of snowfall, the precipitation rate in Antarctica is sufficiently low that much of the continent is technically a desert (17). However, the temperatures are cold enough [averaging about -34°C (18)] that almost all the precipitation is snow and there is little summer melting (17), with the result that the precipitation that falls tends to accumulate.

The Antarctic ice sheet is far from static, as snowfall adds mass, the ice flows in response to gravity and other influences, substantial pieces frequently calve off at its outer margins, and it experiences localized melt. As a result, its size and the geography of its outer margins are constantly changing, even if by only small amounts. Satellite mapping reveals a complex pattern of surface ice velocities and ice-margin advances and retreats (19).

The ice sheet is divided into two uneven masses by the Transantarctic Mountains, with the bulk of the ice contained in the East Antarctic ice sheet, largely in the Eastern Hemisphere, and the remainder contained in the West Antarctic ice sheet, largely in the Western Hemisphere. Besides size, another key contrast between the East and West portions of the Antarctic ice is the fact that the East Antarctic ice is largely grounded above sea level while the West Antarctic ice is largely grounded below sea level, making the West Antarctic ice sheet likely more unstable and more susceptible to surges (20). The East Antarctic ice sheet is believed to have been relatively stable for millions of years, whereas the West Antarctic ice sheet is believed to have disappeared at least once within the last 600,000 years (18).

Drainage from the Antarctic ice sheet is predominantly through glaciers and ice streams, most of which flow into ice shelves (1, 17). Ice streams are typically 100s of km long and dozens of km wide, and sometimes flow many times faster than an average glacier (18). This flow, however, can be decidedly intermittent, as illustrated by a 2003 report of a "stick-slip" behavior of the Whillans Ice Stream flowing into the Ross Ice Shelf. This ice stream experienced quiet periods lasting 6-18 hours interrupted by rapid bursts (or slips) of ice movement at rates on the order of 1 m h^{-1}, lasting 10-30 minutes and felt to be tidally controlled (21).

In view of its size, estimating overall (e.g., mass balance) changes in the Antarctic ice sheet has been difficult, although satellite altimetry and imaging provide a means for eventual comprehensive records. Currently, it is uncertain even whether the Antarctic ice sheet as a whole is in a state of positive mass balance, negative mass balance, or equilibrium (e.g., 22, 23). However, satellite radar altimeter data covering 77.1% of the ice sheet from mid-April 1992 to mid-April 2001 indicate that over that 9-year period the East Antarctic ice sheet had a positive mass balance of 17 ± 11 Gt year^{-1} and the West Antarctic ice sheet had a negative mass balance of -47 ± 4 Gt year^{-1}, combining to an overall negative mass balance of -30 ± 12 Gt year^{-1}, equivalent to 0.09 mm year^{-1} sea level rise, for the 77.1% of the Antarctic ice sheet covered by the satellite data (24).

Because of the importance of ice streams to ice sheet drainage and the possibility that the West Antarctic ice sheet might be unstable, particular attention has been paid to the West Antarctic ice streams. In the latter part of the 20th century, much concern and effort centered on the flow of ice into the Ross Ice Shelf; and the inflowing Ross Ice Streams (originally lettered A-F from south to north along the eastern boundary of the Ross Ice Shelf; later renamed after individuals) were estimated in 1987 as having a negative mass balance of -23 ± 15 km^3 year^{-1} (25). More recent and spatially comprehensive estimates show instead a positive mass balance amounting to approximately 26.8 Gt of ice per year for these same ice streams (26). This shift from a negative to a positive mass balance lessens concern that decay in this region is likely to precipitate a massive outflow and significant sea level rise. Concern subsequently has centered on the region of the Thwaites and Pine Island glaciers (100°W-110°W), which show evidence of thinning, at an overall rate of about 1 m year^{-1} during the 1990s, and associated retreat of the grounding line (where the ice sheet loses contact with the underlying ground), at a rate of about 1.2 km year^{-1} (27). The Pine Island Glacier, with the largest discharge of any West Antarctic ice stream, at 75 Gt year^{-1}, thinned by up to 1.6 m year^{-1} between 1992 and 1999 (28). Total mass losses for the Pine Island and Smith/Thwaites systems over the period from mid-April 1992 to mid-April 2001 are estimated at 17.2 ± 1.4 Gt year^{-1} and 45.6 ± 1.4 Gt year^{-1}, respectively, from satellite radar altimetry (24).

Along the Antarctic Peninsula, jutting north toward South America, glacier retreat has been dominant over the past several decades. Of 244 peninsula glaciers considered by Cook et al. (29), 87% retreated overall from the mid-20th century (average record start: 1953) through 2004, with advances more common through 1964 and retreats more common since 1964, and with a general geographic tendency for the predominance of retreat to advance southward. Still, the individual glaciers fluctuate between advances and retreats, and the retreat percentages for shorter time periods are less than the 87% figure for the full period. In the most recent 5-year period of the record, 2000-2004, 75% of the glaciers were in retreat (29).

2.1.2 *Greenland Ice Sheet*

The Greenland ice sheet (Figure 2) covers an area of 1.7 x 10^6 km^2 and contains approximately 3 x 10^6 km^3 of ice (on the order of one-tenth the volume of the Antarctic ice), with an average ice thickness of about 1.6 km (17). Averaged over the ice sheet, the snow accumulation each year is approximately 0.3 m, more than twice the spatially averaged accumulation on Antarctica (17). Furthermore, roughly half the Greenland ice sheet experiences summer melting at the upper surface (17). Another contrast with the Antarctic ice sheet is that around much of the island, the Greenland ice sheet does not extend to the coast and so in those locations does not have ice shelves, although a few ice shelves exist along the north and northeast coasts. Much of the flow from Greenland emerges through outlet glaciers that tend to be much narrower but flow much faster than the primary outlet glaciers in Antarctica (17). In fact, the Jakobshavn glacier on the west coast of Greenland (69°N, 50°W) has the fastest average speed of any land ice on Earth, with an annual velocity of 8 km $year^{-1}$ (1, 30).

During recent years, repeat aircraft surveys have indicated that the Greenland ice sheet has, overall, been losing mass at low elevations while being in approximate balance for elevations above about 2000 m. From aircraft laser altimeter surveys undertaken in 1993/94 and 1998/99, at low elevations some local thickening occurred but thinning predominated, with the thinning exceeding 1 m $year^{-1}$ in many areas and reaching 10 m $year^{-1}$ near the terminus of the Kangerdlugssuaq Glacier in southeastern Greenland (31). In contrast, at elevations above 2000 m, the pattern of thinning and thickening was more evenly mixed, with thickening of approximately 14 ± 7 mm $year^{-1}$ in the north and thinning of approximately 11 ± 7 mm $year^{-1}$ in the south (32). Repeat aircraft surveys in 2003 yielded further low-elevation thinning between 1997 and 2003, with an increase in the rate of ice loss from about 60 km^3 $year^{-1}$ for the 1993/4-1998/9 period to about 80 km^3 $year^{-1}$ for the 1997-2003 period (33). Combining satellite radar altimeter data for 1978-1988, aircraft laser altimeter data for 1993-1999, and volume budget calculations representing the past few decades, Thomas et al. (34) find overall thickening in the southwest of the ice sheet, thinning in the southeast, and a balance to within 10 mm $year^{-1}$ for high elevations.

Satellite radar altimeter data, covering 90% of the Greenland ice sheet over the period mid-April 1992 to mid-October 2002 and used in conjunction with airborne laser altimeter surveys of the margins, confirm thinning at the edges of the ice sheet. However, these data yield a growth at the higher elevations that actually outweighs the low-elevation thinning, producing a positive mass balance for the ice sheet as a whole. Specifically, the overall mass balance for Greenland over the 10.5 years of the satellite data record is calculated to be +11 ±

3 Gt $year^{-1}$, equivalent to a 0.03 mm $year^{-1}$ sea level fall (24). Both the thinning at the margins, from increased melt, and the thickening inland, from increased snowfall, are viewed as expected responses to climate warming (24). The distribution of melt on the ice sheet has been mapped for the years 1979-1999 using satellite passive-microwave data, and a positive trend of approximately 1% $year^{-1}$ has been found in the total melt area (35).

2.1.3 *Mountain Glaciers*

Mountain glaciers exist scattered around the globe in all latitude zones, with increasingly high elevations generally required for the ice in progressively lower latitudes. Areally, the land ice outside of the Greenland and Antarctic ice sheets totals approximately 785,000 km^2 (3). Of this, approximately 582,000 km^2 are in the Northern Hemisphere, 202,000 km^2 in the Southern Hemisphere. In the Northern Hemisphere, the distribution breaks down as: 151,433 km^2 in the Canadian Arctic Archipelago, 120,680 km^2 in Asia, 92,386 km^2 in the islands of the Arctic Ocean, 76,200 km^2 in Greenland outside of the Greenland ice sheet, 74,600 km^2 in Alaska, 49,660 km^2 in North America outside of Alaska and the Canadian Archipelago, and 17,290 km^2 in Europe. In the Southern Hemisphere, the distribution is: 169,000 km^2 in Antarctica outside of the Antarctic ice sheet, 25,000 km^2 in South America, 7,000 km^2 on subantarctic islands, 1,160 km^2 in New Zealand, 6 km^2 in Africa, and 3 km^2 in New Guinea, Irian Jaya (3). To gain a sense of how much ice is involved, consider that even in Africa, with its slim 6 km^2 areal total, the south-facing wall of ice of Mount Kilimanjaro's northern icefield is still imposing, at approximately 25 m high (36).

Broad ranging historical information, drawings, photographs, and satellite imagery of many of the Earth's glaciers can be found in the 11-volume *Satellite Image Atlas of Glaciers of the World* (2). A further effort currently underway to inventory and assess the state of the

Earth's land ice is the international project entitled Global Land Ice Measurements from Space (GLIMS). The GLIMS project is centered largely on satellite data from the Enhanced Thematic Mapper Plus on Landsat 7 and the Advanced Spaceborne Thermal Emission and Reflection Radiometer on Terra and aims to create a much more comprehensive and homogeneous data base on the Earth's glaciers than currently exists (37). In the meantime, the existent record is far from complete or homogeneous.

Few of the Earth's more than 160,000 glaciers have any scientific records, and even fewer have records appropriate for studies of climate change. For instance, only 246 glaciers have mass-balance records for any year within the extended period 1946-1995, only 86 of those have a record length of at least 10 years (38), and only about 40 glaciers have continuous mass balance measurements for a period of at least 20 years (39). Furthermore, glacier measurements have often been done based more on ease of access or the importance of the specific glacier for the hydroelectric industry than on the representativeness of the selected glaciers (38, 39). Nonetheless, despite the limited measurements, the evidence that exists suggests widespread retreat of glaciers both over the past few decades and over the past 100-200 years. For the glaciers with data, most have lost mass, overall over the past several decades, although all glaciers fluctuate in their behaviors and at any time some glaciers are gaining mass while others are losing mass.

Despite the absence of a globally comprehensive record, several studies have attempted a global view by examining selected records from around the globe. Oerlemans (40) analyzes records from 169 glaciers, 93 from the European Alps and the rest more widely distributed, and reports a decrease in average length of the glaciers starting in about 1800 and continuing nearly unabated to the end of his record in 2000. Of the 36 glaciers with records for the 1860-

1900 period, 35 of them retreated during that period; and of the 144 glaciers with records for 1900-1980, 142 of them retreated during that period (40). Braithwaite (38) examines the available mass-balance records for 11 regions around the globe for the years 1980-1995, finding predominantly negative mass balances (except in Scandinavia, the Caucasus, and the Altai mountains) but no trend toward either more or less negative mass balances over the course of the 15 years. Dyurgerov and Meier (41) examine the period 1961-1997, finding considerable variability both spatially and temporally but overall a global reduction in volume of mountain glaciers averaging 100 km^3 $year^{-1}$. They note in particular the contrast between glaciers in dry regions, which tended to contribute to the overall negative mass balance, and glaciers in moist maritime regions, many of which instead grew over the 1961-1997 period.

Regional studies reveal additional details, although with a mixture of temporal and spatial scales. Aircraft surveys of 67 glaciers in Alaska and nearby Canada indicate an average thickness change of -0.52 m $year^{-1}$ between the mid-1950s and the mid-1990s, with fewer than 5% of the glaciers experiencing thickening. Later repeat aircraft surveys of 28 of the 67 glaciers indicate continued and accelerated thinning, with an average rate of 1.8 m $year^{-1}$ thinning from the mid-1990s to 2000-2001 (39). The latter result, when extrapolated to all the Alaskan and nearby Canadian glaciers corresponds to a sea level equivalent of 0.27 ± 0.10 mm $year^{-1}$ (39). Further east, repeat aircraft surveys over the major ice caps of the Canadian Arctic Archipelago in 1995 and 2000 reveal thickening (averaging about 5 cm $year^{-1}$) or little change above 1600 m elevation but an outweighing of this by thinning, in general, below 1600 m, with a total regional mass balance for the archipelago ice estimated at -25 km^3 $year^{-1}$, equivalent to 0.064 mm $year^{-1}$ sea level rise (42). The most pronounced thinning was on southern Baffin Island's Barnes ice cap, at over 1 m $year^{-1}$ (42).

For the period 1961-1990, the glaciers in Asia, North America, and the Arctic islands, overall, were found to be losing mass (8, 43), but the glaciers of Europe, overall, were found to be growing, especially in the western mountain regions of Scandinavia and in portions of Switzerland (43). Iceland and Scandinavian glaciers gained mass in the 1960s to 1990s, then started losing mass (44). From a longer perspective, the Europe volume of the *Satellite Image Atlas of Glaciers of the World* reveals overall retreat of European glaciers in the 19th and 20th centuries, although with each glacier and regional grouping of glaciers showing its own pattern of fluctuating advances and retreats. For instance, Italian glaciers predominantly retreated from the 1930s to the 1960s, then advanced afterwards (2).

In the tropics, Qori Kalis, the largest outlet glacier of the Quelccaya ice cap in Peru, retreated markedly from 1963 to 2000, and the retreat accelerated over that time period, with the exception of a stall in 1991-1993 thought to have been influenced by the 1991 eruption of Mount Pinatubo (45). In Africa, the ice fields of Mount Kilimanjaro, Africa's highest mountain, lost 80% of their area between 1912 and 2000, with a remaining ice cover of only 2.6 km^2 by 2000 (36, 46); and the ice fields of Mount Kenya lost nearly 40% of their area between 1963 and 1987 (47). More generally, tropical glaciers in South America, Africa, and New Guinea appear to have experienced widespread although not uniform retreat since the mid-19th century, near the end of the Little Ice Age (LIA) (36, 48). Retreat, overall, appears to have been strong in the second half of the 19th century, to have slowed in the early 20th century, with some glaciers advancing back close to their LIA extents, and to have dominated again in the 1930s-1950s. The 1960s and 1970s witnessed some notable advances, followed by a reemergence of dominant retreat in the 1980s and 1990s (48).

2.2 Floating Ice

Volume-wise, the Earth's floating ice is quite minor versus its land ice; and because the floating ice is already displacing the same volume of water as the volume it would have when melted, changes in floating ice are not relevant to sea level. However, areally, floating ice is considerably more expansive than land ice, directly affecting a much larger area in terms of the reflection of solar radiation and the insulation between the underlying surface and the atmosphere; and the floating ice has major impacts on local and regional climate, ecosystems, and human activities despite its lack of impact on global sea level. Of the floating-ice components of the climate system, by far the most areally expansive is sea ice.

2.2.1 *Sea Ice*

Sea ice covers substantial ocean areas in both polar regions at all times of the year, although with large seasonal cycles and quite noticeable variability from one year to another. In the Northern Hemisphere, ice extents (areal coverages of ice) typically range from a minimum of approximately 7 x 10^6 km^2 in September to a maximum of approximately 15 x 10^6 km^2 in March, with the summertime ice extending over most of the Arctic Ocean and much of the Canadian Archipelago and the northernmost Greenland Sea and the wintertime ice extending throughout the Arctic Ocean, Canadian Archipelago, Hudson Bay, Baffin Bay, and Kara Sea and also covering much of the Sea of Okhotsk, Bering Sea, Davis Strait, Labrador Sea, Greenland Sea, and Barents Sea (Figure 3; 49). In the Southern Hemisphere, ice extents typically range from a minimum of approximately 3 x 10^6 km^2 in February to a maximum of approximately 18 x 10^6 km^2 in September (49). The summertime Southern Hemisphere ice is confined largely to the western Weddell Sea, the southern Bellingshausen and Amundsen Seas, and the southeastern Ross Sea, with much lesser ice amounts around the coast of East Antarctica. The wintertime Southern Hemisphere ice surrounds the Antarctic continent, extending northward beyond 55°S in the far eastern Weddell Sea and to 45° - 55°S around most of the rest of the continent (Figure 3).

Sea ice thicknesses are not nearly as well known as the ice distributions and extents, the latter being readily determined on a routine basis from satellite data (49). The ice thicknesses, however, are, like the ice extents, also critical for ice volume and mass determinations. Ice thicknesses traditionally have been obtained largely from in situ and submarine data, providing measurements that are quite nonuniformly distributed in both space and time. With that qualifier, sea ice thicknesses in the central Arctic probably average somewhere in the range 2-4 m, and ice thicknesses in the surrounding seas and bays, where the ice tends to be seasonal rather than year-round, are predominantly less than 1 m, as are the sea ice thicknesses in the Southern Hemisphere, where over 80% of the ice cover is seasonal (e.g., 50, 51, 52, 53).

In the Northern Hemisphere, sea ice coverage has decreased, overall, in the past several decades. The retreat in areal coverage since the late 1970s has averaged approximately 36,000 km^2 $year^{-1}$, or 3% per decade, for the ice cover as a whole (e.g., 49, 54, 55) and approximately 7-9% for multiyear ice (ice that has survived a summer melt period) and late-summer ice (e.g., 56, 57, 58). Correspondingly, the length of the sea ice season has decreased (59) and the length of the melt season has increased (60).

Northern Hemisphere sea ice, overall, has also thinned, although the magnitude of the thinning is less certain than the retreat, in view of the spatially and temporally incomplete thickness measurements and the added complications produced by the highly uneven nature of the ice cover and the fact that the ice floes are in motion. Submarine-based estimates of thinning of the Arctic Ocean sea ice include values as high as 40% from the period 1958-1976 to the period 1993-1997 (61), reduced to approximately 32% for the same period after incorporation of additional submarine tracks (62), but also include suggestions of no thinning between 1991 and 1997 (51) and substantial uncertainty in all periods. Surface-based measurements from Russian drifting ice stations suggest a thinning of approximately 10 cm (less than 4%) over the 20-year period 1971-1990 (63). In view of the uncertainties and wide range of published estimates, Holloway and Sou (64) examined the ice thinning issue with a coupled ice/ocean/snow model, concluding that the large thickness decreases reported in some studies probably were not representative of the Arctic as a whole. Still, although the magnitude varies depending on location and time, the preponderance of evidence suggests that there has been at least some Arctic sea ice thinning over the past several decades.

In contrast to the Northern Hemisphere sea ice, the Southern Hemisphere sea ice, which underwent a rapid retreat in the 1970s (65), experienced a slight areal advance of approximately 11,000 km^2 $year^{-1}$, or approximately 1% per decade, since the late 1970s, as determined from satellite passive-microwave observations (49, 66, 67). In line with the advance, the Southern Hemisphere sea ice, overall, has also experienced a lengthening of the sea ice season, although spatial contrasts are marked, with a noticeable shortening of the sea ice season in the vicinity of the Antarctic Peninsula and extending well to its west, throughout the Bellingshausen and Amundsen seas (5). The limited amount of thickness data available for the Southern Hemisphere sea ice prevents determination of reliable ice thickness trends.

2.2.2 *Lake Ice and River Ice*

Although the lake ice most familiar to most people is visible, seasonal, wintertime ice formed on lakes immediately in contact with the atmosphere, by far the thickest lake ice on Earth is probably on Lake Vostok at the bottom of the Antarctic ice sheet at about 77°S, 105°E. Lake Vostok is known so far only through remote sensing and is estimated through remote sensing to have an area comparable to Lake Ontario (68) and a lake-ice thickness of approximately 210 m (69).

More directly visible, ice exists year-round on a few surface lakes in Antarctica and in the northern Arctic and, more commonly, routinely forms each autumn/winter on lakes and rivers throughout the high and mid-latitudes of the Northern Hemisphere. Freeze-up on the seasonally ice-covered Northern Hemisphere lakes and rivers occurs as early as early September in the far north, with progressively later freeze-up dates typical as one moves farther south. However, freeze-up timings depend additionally on factors other than latitude, with, for instance, deeper lakes generally taking longer to cool to the freezing point than shallow lakes, other factors being equal (6). Ice seasons of more than 100 days are common on North American rivers as far south as 42°N and on Asian rivers as far south as 30°N (7).

Lake and river ice thicknesses vary considerably, with the greatest thicknesses for seasonally ice-covered lakes and rivers, in the early 21st century, tending to be on the order of 1.8-2.6 m in the Canadian High Arctic (6). Much greater ice thicknesses, even as high as 19 m, can occur on perennially frozen surface lakes (6).

All lakes and rivers on which ice forms have significant interannual variability in their ice covers. However, analyses of the available records reveal multi-decadal trends on many lakes and rivers, both individually and regionally. For the period 1893-1985, significant trends were found toward later freeze-up dates and earlier breakup dates for river ice in western Siberia, the European portion of the Former Soviet Union, and the region of the Black Sea (7). The timing of break-up on many of the western Siberian and European Former Soviet Union rivers has advanced, on average, by 7-10 days per 100 years (7). In the case of the Lower Don

River, the length of the ice season has been reduced by a full month (7). However, rivers in central and eastern Siberia show generally weaker but still significant trends in the opposite direction: toward earlier freeze-up dates and later break-up dates (7).

In North America, analysis of 39 lake and river records for the 150-year period 1846-1995 yields shorter ice seasons for 38 of the 39 records, with 14 of the 15 freeze-up records indicating later freeze-up and all 24 of the break-up records indicating earlier breakup (70). Linear trends show freeze-up dates later, overall, by 5.8 ± 1.9 days $(100\ \text{years})^{-1}$ and breakup dates earlier by 6.5 ± 1.4 days $(100\ \text{years})^{-1}$, for a decrease of 12.3 days $(100\ \text{years})^{-1}$ in the duration of the ice cover (70). Another analysis of selected long-term lake-ice breakup records in North America obtained quite different results, finding progressively later breakup dates from 1870 until 1940, followed by a stabilizing of the breakup dates from 1940 through 1971 (71). Still, from the majority of records available, overall the tendency from the late 19th century to the end of the 20th century appears to have been toward shorter lake-ice and river-ice seasons in North America as well as in Eurasia (e.g., 7, 70). In the case of the perennially ice-covered far northern lakes, two along the north coast of Ellesmere Island actually lost their ice covers temporarily in the summers of 1998 and 2000 and others in the vicinity experienced a considerable reduction in their ice areas (72).

On a shorter, more recent time frame, Pavelsky & Smith (73) examine 1992-2003 satellite data for the four largest rivers flowing into the Arctic Ocean. While not finding strong trends for any of the four rivers, they do find systematic internannual variations that show a strong positive correlation between the Ob' and Yenisey Rivers, in Asia, and a strong negative correlation between the Asian Lena River and the North American Mackenzie River, with some suggestion of a connection, in the latter case, with the Pacific Decadal Oscillation.

2.2.3 *Ice Shelves*

At many locations around the Antarctic continent, totaling approximately 40% of the coastline (8), the massive Antarctic ice sheet extends outward over the surrounding oceans, forming glacially-fed ice shelves (Figure 4). The largest of these is the Ross Ice Shelf, at approximately 525,000 km^2 in area and with thicknesses varying from approximately 1000 m at its landward edge to approximately 250 m at its seaward edge (74). The second largest is the Ronne-Filchner Ice Shelf, also exceeding 400,000 km^2 in area and having thicknesses of hundreds of meters. Altogether, ice shelves cover approximately 11% of the area of Antarctica (75).

In the Arctic, there are no ice shelves comparable in size to the large Antarctic ice shelves, as the Arctic's main ice sheet, overlying Greenland, generally does not extend outward to the coast. In fact most of the Arctic ice shelves, along the northern Ellesmere Island and Greenland coasts, are not predominantly glacially fed, instead being formed from landfast sea ice, bottom accretion of additional sea ice, and overlying accumulation of snow (8, 76, 77). The largest of the Arctic ice shelves is the Ward Hunt Ice Shelf, located at 83°N, 74°W on the north coast of Ellesmere Island and formed from sea ice (77). Smaller ice shelves on Ellesmere Island include some that are largely glacially fed, such as the Milne Ice Shelf, and others that have substantial glacial feeding and substantial sea ice growth, such as the Alfred Ernest Ice Shelf (76). When summed, however, all of the Ellesmere ice shelf areas total less than 1350 km^2, miniscule in comparison to the Antarctic's Ross and Ronne-Filchner ice shelves (76).

The Arctic ice shelves constitute one segment of the cryosphere in which a 20th century trend is unquestionable: these ice shelves underwent substantial decay (8). At the start of the

20^{th} century, an Ellesmere Ice Shelf estimated at 7500 km^2 - 8900 km^2 in area extended along 500 km of the northern Ellesmere Island coast and was described in detail by Robert Peary and other explorers (8, 76). This ice shelf had largely disintegrated by the 1960s and by the end of the century was reduced to less than 20% of its former size, the remnants consisting of the Ward Hunt and other remaining small ice shelves along the coast (8, 76). The Ward Hunt Ice Shelf underwent a massive calving event in 1961-1962, losing half its mass in those two years (76). Although the ice shelf stabilized for a while after 1982, it experienced substantial further break up in 2000-2002, recorded by satellite synthetic aperture radar, helicopter transects, and in situ measurements, as initial small fractures lengthened, widened, and split the ice shelf (77).

In the Southern Hemisphere, substantial ice shelf decay has occurred since about 1970 along the Antarctic Peninsula, with a loss of over 13,500 km^2 in ice shelf area (78, 79), although has not occurred elsewhere around the continent. Along the west coast of the Peninsula, the retreats were preceded in some cases by advances in the mid-20^{th} century, as the Müller Ice Shelf expanded rapidly from the mid-1940s to the mid-1950s before decaying in extent by over a third from the mid-1950s to the mid-1990s, and the Wordie Ice Shelf experienced an expansion from the late 1930s to the mid-1960s (78) before decaying by over 60%, from approximately 2000 km^2 in 1966 to only 700 km^2 in 1989 (8). In contrast, the George VI Ice Shelf, farther south along the Peninsula than the Wordie or Müller shelves, retreated between 1949 and 1974 but remained largely stable from 1974 to the mid-1990s (78). The Wilkins Ice Shelf lost 1100 km^2 in the single month of March 1998 (79).

On the east coast of the Antarctic Peninsula, the Prince Gustav Channel Ice Shelf in the far north decayed by well over 50% and the Larsen A Ice Shelf to its south decayed by approximately 50% between the mid-1940s and the mid-1990s (78). Larsen A lost an additional 1,300 km^2 as it fractured and disintegrated in a 50-day period in early 1995, with 1000s of icebergs generated (78). The larger Larsen B Ice Shelf, just south of Larsen A, also suffered substantial losses in January 1995 (75), then collapsed further in March 2002, losing over 12,000 km^2 of ice (80).

One feared consequence of the loss of ice shelves is that these shelves might be buttressing land ice masses upstream and that the loss of the ice shelves might precipitate a massive outflow of land ice, causing enhanced sea level rise (81). This is of particular relevance in the Antarctic case in view of the massive amount of land ice. Although the inflow to the ice shelves along the Antarctic Peninsula is from Peninsula glaciers and not the bulk of the Antarctic ice sheet, and there appears to be no immediate threat of massive decay of the larger and more critical Ross and Ronne-Filchner ice shelves, still it is sobering to note that tributary glaciers into the Larsen A Ice Shelf increased in velocity by up to three-fold in the 1995-1999 period after the January 1995 Larsen A collapse (82) and that tributary glaciers into the Larsen B Ice Shelf increased in velocity by factors of 2-6 after the March 2002 Larsen B collapse (9, 79). The importance of the ice shelves in buttressing the upstream ice has been controversial (e.g., 83), but the events after the Larsen A and Larsen B collapses, including not just the acceleration of ice flow into the regions of collapse but also the lack of acceleration of two glaciers flowing into the remnant portions of Larsen B, i.e., glaciers still buttressed by the ice shelf, strongly support the importance of the buttressing effect (9).

2.2.4 *Icebergs*

Thousands of icebergs calve into the ocean each year in both the Arctic and the Antarctic, amounting to over 300 km^3 of ice per year in the Arctic and an estimated 2,600 km^3

of ice per year in the Antarctic (84, 85). Depending on circumstances, the calving process can occur quite rapidly or can extend over many years, one example of the latter occurring along the Ninnis Glacier tongue in East Antarctica, which had a noticeable fracture in 1989 but did not complete the calving process until January 2000, when an iceberg of area approximately 800 km^2 broke off (11). Debris within icebergs, useful in paleoclimate studies of marine sediments, tends to be concentrated in the bottom 3 m of the ice (84).

The largest Antarctic icebergs are comparable in size to a small state in the United States, with many longer than 100 km. An iceberg larger than Delaware broke off the Ronne Ice Shelf in 1998 (Iceberg A-138, approximately 150 km x 50 km) (85), and a much larger iceberg broke off the Ross Ice Shelf in March 2000 (Iceberg B-15, approximately 295 km long, up to 40 km wide, and 10,000 km^2 in area; Figure 5) (10). B-15 is one of the largest icebergs ever recorded and is estimated to have originally had a thickness of approximately 400 m and a total fresh water content of 500-1000 trillion gallons (85). This iceberg precipitated a cascading sequence of impacts on the local environment and ecosystem that illustrates the impacts a large iceberg can have. By blocking the normal drift of sea ice, B-15 resulted in heavy sea ice concentrations on its windward side, causing a substantial delay in the annual springtime phytoplankton bloom and thereby a substantial reduction in annual primary production, which spatially averaged approximately 40% below normal for the southwest Ross Sea. Animals up through the food chain were affected, with local penguin populations having to alter their diet as well as their seasonal migration trajectories due to the iceberg presence and the altered sea ice coverage (10). Although icebergs divide and decay with time, they can remain sizeable for many years, as illustrated by Iceberg B-9B, which calved from the Ross Ice Shelf in 1987 and was measured at approximately 97 km x 20-35 km in 1997 (11).

Similarly to the case with Arctic versus Antarctic ice shelves, Arctic icebergs are considerably smaller than the largest Antarctic icebergs, rarely being longer than a few 10s of km. Among the Arctic icebergs are relatively large and flat versions called ice islands. These most commonly derive from the Ellesmere Ice Shelf and are used on occasion for scientific and military research stations (8). The ice islands vary in area from a few hundred m^2 to a few hundred km^2 (76). One of the smallest ever recorded was a 10 m x 11 m ice island off the north coast of Alaska, while one of the largest, named T-2, was 31.1 km x 33.4 km (76). Jeffries (76) estimates a total of approximately 600 ice islands in the Arctic and surrounding waters from 1946 to 1992. Because of the nearly enclosed nature of the Arctic Ocean, the Arctic icebergs can float within the Arctic Basin for decades before emerging through Fram Strait into the warmer waters of the Greenland Sea and North Atlantic.

Of most concern to humans are the icebergs that drift far enough south in the North Atlantic to reach the vicinity of major shipping routes. These often derive from the west coast of Greenland rather than Ellesmere Island or the east coast of Greenland. Because of their importance as potential hazards, there is a much better record of the icebergs near and upstream of major shipping routes than the icebergs elsewhere around the globe. In particular, the United States Coast Guard's International Ice Patrol (IIP) has since 1913, following the 1912 sinking of the *Titanic*, tracked icebergs that drift south of 52°N, the full patrol area being 40-52°N, 39-57°W (www.uscg.mil/lantarea/iip/General/mission.shtml), and icebergs in this vicinity have been tracked from radar tracking stations and satellite-tracked beacons (86). On average, approximately 600 sizeable icebergs arrive in the IIP patrol area each year, although the recorded numbers have varied from 0 in 1996 to over 2,000 in 1984 (85, 86). Icebergs south of 48°N tend to be most numerous in the April-June time frame (86). Many of the icebergs reaching the IIP patrol area formed by calving from West Greenland into Baffin Bay/Davis Strait, sometimes flowing northward in the narrow, north-flowing West Greenland Current before turning west and then flowing south in the south-flowing western Baffin Bay and Labrador currents (86). Typical autumn and winter drift speeds of the south-flowing icebergs are 5-20 km day^{-1} (86).

When particularly large or numerous icebergs form, they often receive media attention. However, it is not known whether the frequency of iceberg calvings in the past several years is unusual in any way (high or low), in view of the meagerness of the pre-satellite record in areas other than the IIP patrolled region of the North Atlantic. In the IIP region, there has been no prominent trend toward either more or fewer icebergs over the course of the nine-decade record (86). The decades of the 1950s and 1960s, near the middle of the record, had the fewest icebergs; and the decades after that had a few years with particularly high counts but also more years with low counts (under 200) than in the first half of the century (86).

2.3 Snow Cover on Land

Like sea ice, snow generally is a thin covering that spreads over large areas, reflecting solar radiation, insulating the underlying surface from the atmosphere, and influencing the regional climate and ecosystems in numerous ways. Snow in the Southern Hemisphere is dominated by the perennial snow cover on the Antarctic continent, covering approximately 13 x 10^6 km^2. The Northern Hemisphere contains much less perennial snow but approximately 98% of the Earth's seasonal snow (87).

The seasonal cycle of continental snow coverage in the Northern Hemisphere is huge, typically ranging from a summer minimum of approximately 4 x 10^6 km^2 in August to a winter maximum of approximately 46 x 10^6 km^2 in January and February (Figure 6) (88, 89). Over half (60-65%) of the Northern Hemisphere winter snow is in Eurasia, whereas most of the mid-summer snow is in Greenland (88). Large interannual variability exists, with, for instance, the Northern Hemisphere February snow extent varying between 40 x 10^6 km^2 and 50 x 10^6 km^2 depending on year (15). Atmospheric circulation plays a major role in influencing snow extent and its variability, at least over North America in winter, although its impact has been found to be considerably less in springtime, during the ablation season (88).

Northern Hemisphere snow distributions have been monitored from satellite visible and/or infrared data since 1966, originally on a weekly basis and now on a daily basis (89). Snow water equivalent, important for local fresh water supplies and flooding considerations, additionally requires snow thickness information, and this has been estimated for the period since October 1978 with satellite microwave data (89). Snow cover is in several ways more difficult to observe from satellites than sea ice is. When visible and/or infrared data are used, both snow and sea ice are obscured in the presence of a heavy cloud cover, but continental snow has the added complication that vegetation and manmade structures can obscure the snow cover even in the absence of clouds. Furthermore, when microwave data are used, continental snow has the disadvantage that the background land is composed of markedly different natural and manmade surfaces, all compounding the microwave signal. Nonetheless, despite the complications, satellite data remain the primary means of obtaining systematic large-scale data coverage.

Over the period of the satellite record, overall snow coverage in the Northern Hemisphere has declined, although with considerable interannual variability. Walsh et al. (15) indicate an approximate 10% decrease from 1972 to 2003, with decreases greater in Eurasia but present also in North America and with decreases greatest in spring and summer and insignificant in autumn and winter. Over shorter periods, Groisman et al. (90) report an approximate 10% decline in snow coverage from 1973 to 1992, and Armstrong & Brodzik (87) report an approximate 20% decline (0.2% year^{-1}) from late 1978 to early 1999. The calculated 21-year trend in annual Northern Hemisphere snow extent from late 1978 to early 1999 is higher, at 59,000 km^2 year^{-1}, or 0.26% year^{-1}, when determined from satellite visible data than when determined from satellite passive-microwave data, at 35,000 km^2 year^{-1}, or 0.17% year^{-1} (87), reflecting uncertainties in the data sets, with, for instance, the passive-microwave data being less able to detect very thin snow (87) and the visible data being less able to detect snow under clouds.

The spatially comprehensive satellite record of North American snow extent for the period 1972-1994 was extended back to the start of the 20th century by Frei et al. (91) by determining regions of coherent interannual fluctuations, then using linear regression to expand the local information from limited station data (in the United States and southern Canada) to regional coverage. Results suggest, for example, that February snow extent decreased from 1900 to 1930, then increased from the 1930s to about 1980, prior to again decreasing in the 1980s and 1990s. Results vary by month, with March snow extents increasing from the mid-1920s to the early 1950s, and decreasing since then. November snow extent increased slightly from the 1960s to 1990s, and March snow extent decreased from the 1950s to the 1990s (91).

Over Russia, ground data from 119 irregularly spaced stations in the extensive region encompassing 30°-140°E and 50°-70°N yielded an overall 2.4% per decade increase in snow depth for the study period 1936-1983, with snow depths tending to increase north of 60°N, at an overall rate of 4.7% per decade, and to decrease south of 60°N, at the much lower overall rate of 0.8% per decade (92). In contrast, over much of Canada snow depths during winter and early-spring decreased over the 1946-1995 period, and over most of western Canada snow cover duration during spring and summer decreased over the same period (93).

2.4 Permafrost

Permafrost underlies approximately one-fourth of the surface land area in both the Northern and Southern Hemispheres. In the Southern Hemisphere, this is almost entirely in Antarctica, underneath the Antarctic ice sheet. In the Northern Hemisphere, permafrost underlies the Greenland ice sheet, much of the ice of the Himalayas, Rocky Mountains, and other alpine regions, and also vast areas of unglaciated terrain (14), for a total areal coverage of approximately 26 x 10^6 km^2 (15). Permafrost regions are typically divided into two zones: a zone of continuous permafrost, where permafrost is pervasive throughout the region, and a zone of discontinuous permafrost, where the permafrost is intermixed with unfrozen ground (14). The boundaries of the continuous and discontinuous permafrost zones follow annual temperature isotherms fairly closely, with, for instance, the boundary between the two zones typically having mean annual air temperatures between -6°C and -9°C (14). Precipitation, however, is also a factor (94), as a thick snow cover insulates the ground, making it less likely to form permafrost, and high summer rainfall contributes to local permafrost degradation.

Continuous permafrost underlies most (including the entire northern half) of the Greenland ice sheet, with discontinuous permafrost underlying the rest of the ice sheet.

Continuous permafrost also spreads over much of Asian Russia, the northern 25% of Alaska, and northernmost Canada. Discontinuous permafrost spreads over much of the rest of Asian Russia, much of Mongolia, some of China, almost all of Alaska other than the region of continuous permafrost, and much of Canada (14).

Permafrost thicknesses in the continuous zone are hundreds of m in much of the Canadian north (94) and range up to 1500 m in parts of Siberia (95). Thicknesses in the discontinuous zone generally decrease equatorward (94).

Throughout the unglaciated permafrost regions, there is a ground layer between the permafrost and the upper ground surface that freezes in winter and thaws in summer; this layer of summer thaw is termed the "active layer" and is typically between about 2 cm and 2 m thick (14). As permafrost thaws, the thawed portion generally becomes part of the active layer above the permafrost, so that the active layer thickens (e.g., 96), although in some regions there is a permanently thawed layer (a "talik") between the permafrost and the active layer (14).

Significant warming and degradation of Northern Hemisphere permafrost appear to have occurred over the last 50 years (15, 97), with reports of permafrost warming and/or degradation in Alaska (98), western Canada (94), Europe (99), and Russia (96). Beilman et al. (94) report an average 39 km northward movement of the southern boundary of the discontinuous permafrost zone in western Canada in the 20th century; and Frauenfeld et al. (96) report a 20 cm deepening of the active layer (due to degradation of the top of the underlying permafrost layer) in Russia over the period 1956-1990.

However, permafrost warming has not occurred everywhere. Serreze et al. (100) report a mixture of warming and cooling. For instance, near-surface permafrost temperatures increased in northern Russia from 1970 to 1990 but decreased in northern Quebec from the mid-1980s to the mid-1990s. Permafrost surface temperatures in northern Alaska and far eastern Russia show a cyclical variation with a period of a decade or longer, although superimposed on a 20th century warming trend (100). Nelson (95) also mentions local regions of permafrost cooling, in addition to the dominant warming trends.

In considering permafrost temperature changes, it is important to recognize that atmospheric warming does not necessarily translate to permafrost warming and degradation, as snow cover insulation is an additional major factor in permafrost development and maintenance. If snow cover decreases, insulation decreases and the underlying permafrost could get colder even with a rise in air temperatures (e.g., 101). In fact, the warming trend found in permafrost in Switzerland displays high-amplitude interannual fluctuations that correspond more closely with snow cover differences than with air temperature differences (99).

3. DISCUSSION

3.1 The Coupled Climate System and the Issue of Causation

The changes in the cryosphere are integrally coupled with changes in other elements of the climate system. Several reviews have examined the coherent nature of a multitude of changes in the Arctic environment (e.g., 100, 102, 103). There is widespread recognition that many changes are occurring and many factors are involved but that cause and effect are rarely straightforward (in some instances resembling a "chicken and egg" dilemma). Serreze et al. (100) report overall Arctic warming, sea ice decreases, snow cover decreases, glacier mass reductions, terrestrial precipitation increases, northward migration of the tree line, a lengthening of the growing season, and a pattern of permafrost warming in Alaska and Russia but cooling in eastern Canada. However, given the temporal and spatial patterns of the changes, the limitations of many of the data sets, and the deficiencies of the global climate models used to examine climate change, they are unwilling to interpret the changes explicitly as signals of enhanced greenhouse gas warming. Other possible key factors in the changes include solar variability, the El Niño/Southern Oscillation, the North Atlantic Oscillation, and the broader Arctic Oscillation (100).

Many additional studies also either attribute various of the cryosphere changes reported in Section 2 to one named oscillation or another or at least report strong correlations with these oscillations (e.g., 8, 70, 73, 88, 103). Furthermore, some cases that can appear to be a response to warming are not, such as the marked glacier retreat on Mount Kilimanjaro since the mid-19th century, which was apparently predominantly due to a sharp drop in atmospheric moisture, starting around 1880, and was not accompanied by local warming (36). Glacier retreats elsewhere in the tropics since the mid-19th century might also be caused in substantial part by reductions in moisture (48); and permafrost thawing in Russia over the period 1956-1990 is more strongly correlated with snow depth than with air temperature, thicker snow producing more insulation, a deeper active layer, and reduced permafrost (96).

Still, warming, which quite directly increases the likelihood of melt and decreases the likelihood of freezing, is at least a contributing factor in many of the reductions seen in the cryosphere in recent decades and has been asserted as the primary cause in many studies (e.g., 40, 79, 94). It has also been implicated in some cryosphere increases, in particular in the growth of the inland ice of the Greenland ice sheet from mid-April 1992 to mid-October

2002, through the mechanism of increased ocean evaporation leading to increased snowfall (24).

3.2 The Inadequacy of the Record and Considerations for Future Research

Some of the possible climate connections mentioned in the previous section will likely be eliminated, as invalid, once longer and more spatially complete records are obtained and the correlations are recalculated. In fact, the inadequacy of the observational record is a major issue for each of the cryosphere components:

- In the case of land ice (Section 2.1), few of the Earth's approximately 160,000 glaciers have any mass balance records, and the mass balances of the Greenland and Antarctic ice sheets – critical in view of the approximately 68-70 m of sea level equivalent that they contain (16, 27) – remain uncertain even as to the sign of the balance. The Intergovernmental Panel on Climate Change 2001 report suggests that Greenland provided a slight positive contribution of 0.0-0.1 mm year^{-1} to the 20th century's estimated 1.0-2.0 mm year^{-1} sea level rise and that Antarctica had a countering effect, equivalent to a lowering of sea level by 0.0-0.2 mm year^{-1} (16). Satellite altimeter data suggest that more recently the respective balances instead show Antarctica with a slight positive contribution (0.09 mm year^{-1} from mid-April 1992 to mid-April 2001, over 77.1% of the ice sheet) and Greenland with a slight negative contribution (0.03 mm year^{-1} from mid-April 1992 to mid-October 2002, over 90% of the ice sheet), although aircraft altimeter data over Greenland show positive sea level contributions of 0.15 mm year^{-1} for the 1993/94-1998/9 period and 0.2 mm year^{-1} for the 1997-2003 period (Sections 2.1.1 and 2.1.2; 24, 33). The contrasts highlight not only the uncertainties but also the possibility that mass balance might vary considerably from year to year, including changes in sign, in which case mass balance measurements might need to be made routinely over the full ice sheets for meaningful mass balance studies.

- In the case of floating ice (Section 2.2), satellite data have provided a global record of sea ice distributions and extents since the 1970s, but there are far fewer data on sea ice thicknesses or on ice distributions and extents prior to the satellite record. Some lakes and rivers have much longer records, some exceeding 100 years, but the distribution of those is sparse. Nearly century-long iceberg records exist for the IIP patrol area in the North Atlantic, but iceberg records elsewhere are quite limited. Ice shelves are now being well monitored from satellites, although continual long-term records are not available.

- In the case of snow cover on land (Section 2.3), satellite records, while hugely valuable, are hindered in several ways: snow cover can be hidden, e.g., under a forest canopy; cloud cover limits the visible data; the variety of background surfaces limits the microwave data; the microwave data fail to register very thin snow and are more appropriate, with current algorithms, for dry than for wet snowpacks. Long-term in situ records are far more numerous than for sea ice, but are, like those for lake and river ice, quite sparsely distributed.

- In the case of permafrost (Section 2.4), the difficulties are even greater than for the other cryosphere components, because of its being under the ground and hidden from view.

For the future, current satellite technology allows far more consistent and spatially comprehensive records for all cryosphere components except permafrost than were feasible in the past. This includes some possibilities for which satellite data to date have not been a primary resource, such as monitoring river ice breakup through visible and near-infrared satellite imagery (73). It also includes new instrumentation, as illustrated by the first satellite laser altimeter for Earth observations, which was launched in January 2003 and is now being used not just for its primary objective of obtaining ice sheet elevations (Figures 1-2) but also for measurements of sea ice thickness (104).

The satellite data alone, however, are not adequate for comprehensive climate-change studies. They need to be complemented by in situ measurements and by records for the pre-satellite era. Furthermore, the data need to be combined in systematic global studies that establish the interannual variabilities and spatial differences in a way that allows the trends to be cleanly identified, spatially and temporally, lessening the confusion regarding whether calculated trends differ because of the locations, the time periods, or the techniques. The cryosphere is a vital, continually-changing part of the Earth's climate system; it is important that we obtain a better handle on exactly what the changes are and how they tie to the rest of the climate system and to the future.

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FIGURE LEGENDS

Figure 1. Antarctic ice sheet elevations (m), as determined from data collected over the period February 2003 – June 2005 by the Geoscience Laser Altimeter System (GLAS) on the Ice, Cloud and land Elevation Satellite (ICESat). Contours are plotted at 10 m and at 500 m intervals from 500 m to 3500 m; dark grey shading is used to highlight the low elevations (10-500 m). The digital data were provided by Jay Zwally and John DiMarzio.

Figure 2. Greenland ice sheet elevations (m), as determined from data collected over the period February 2003 – June 2005 by the GLAS instrument on ICESat. Contours are plotted at 10 m and at 500 m intervals from 500 m to 3500 m; dark grey shading is used to highlight the low elevations (10-500 m). The digital data were provided by Jay Zwally and John DiMarzio.

Figure 3. Global sea ice distributions, showing maximum and minimum monthly average sea ice coverages in each hemisphere, averaged over the years 1979-2004, as derived from 1979-1987 data from the Scanning Multichannel Microwave Radiometer on the Nimbus 7 satellite and 1987-2004 data from the Special Sensor Microwave Imager on satellites of the Defense Meteorological Satellite Program.

Figure 4. Major Antarctic ice shelves. The background map was adapted from a CIA World Map database and was made available through the Interactive Data Language data visualization and analysis software application.

Figure 5. Formation of Iceberg B-15, calved from Antarctica's Ross Ice Shelf in March 2000. (a) The edge of the ice shelf on March 3, 2000, showing no evidence of a crack. (b) The edge of the ice shelf on March 28, 2000, showing B-15 fully formed and about to break away. The iceberg at this time was approximately 300 km x 40 km, over twice the size of Delaware. The

data are from the Moderate Resolution Imaging Spectroradiometer (MODIS) on the Terra satellite; and the images were provided courtesy of Jacques Descloitres and the MODIS Land Science Team.

Figure 6. Global snow cover in August and February 2005, as determined from data from the MODIS instrument on the Terra satellite. White indicates snow in both August and February, gray indicates snow in February but not August, and red indicates snow in August but not February. The Antarctic continent was not included in the processed data set, although is set to white because of its perennial snow cover. Also, the 5-km resolution of the mapped data limits the ability to see small snow fields. The digital data were provided by Dorothy Hall and the MODIS Snow and Ice Team.

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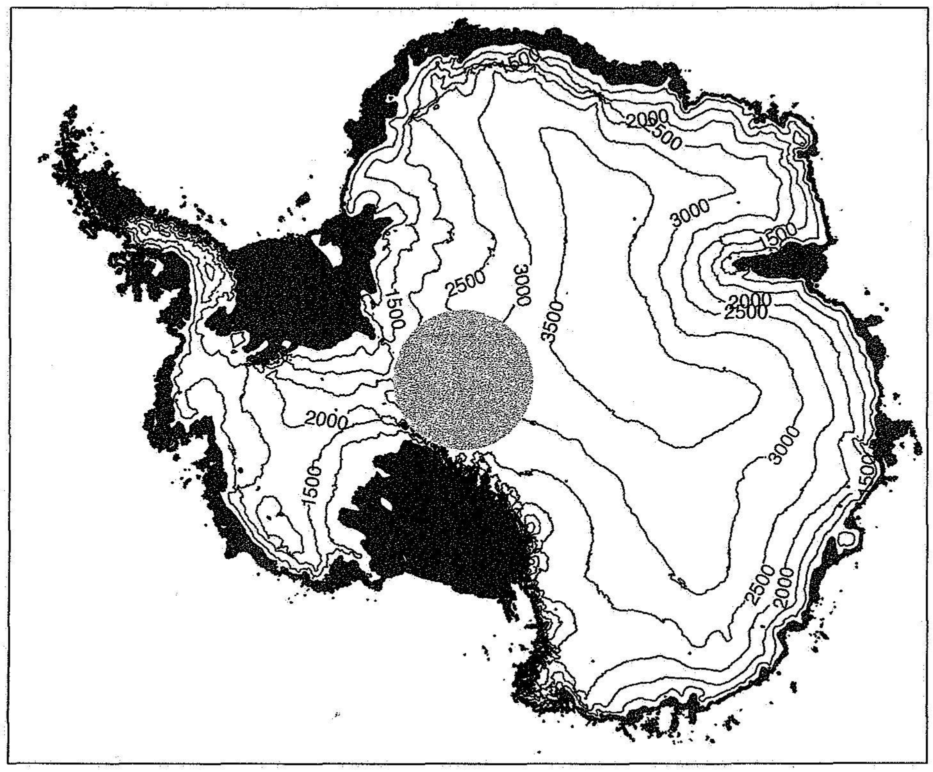
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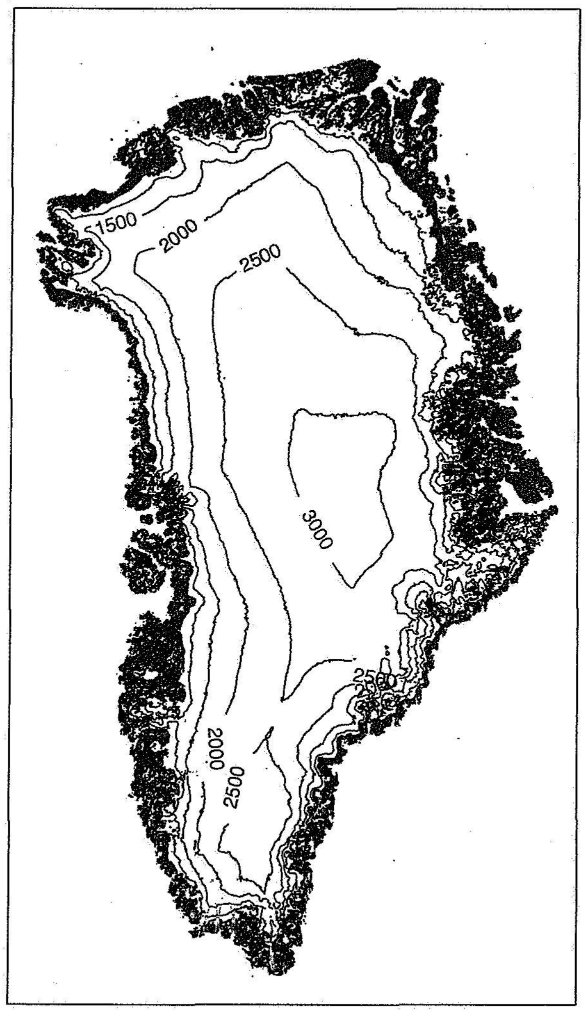
Figure 1



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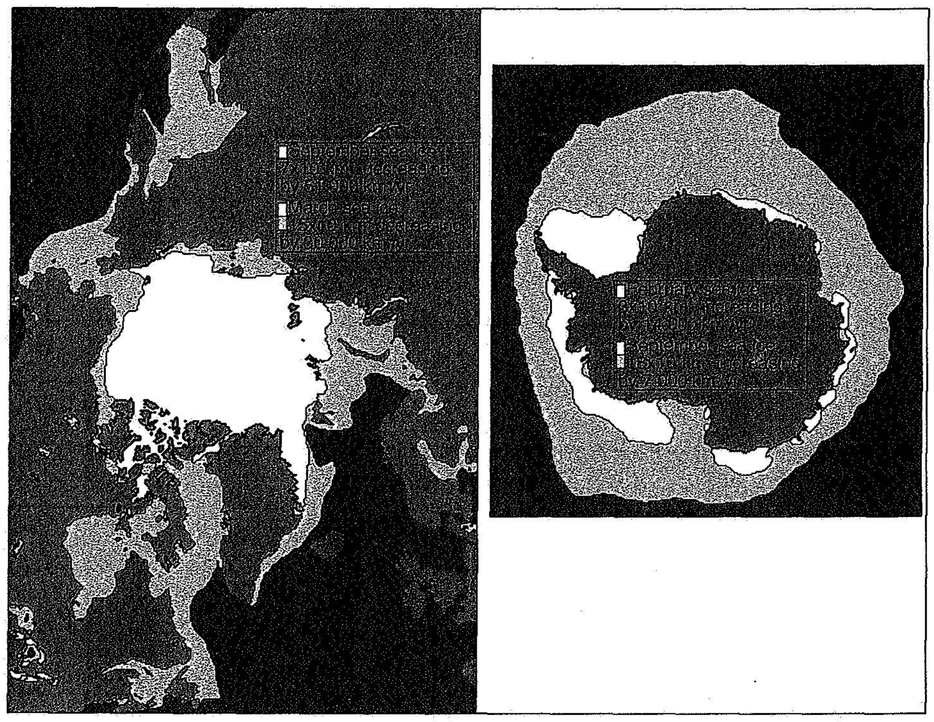

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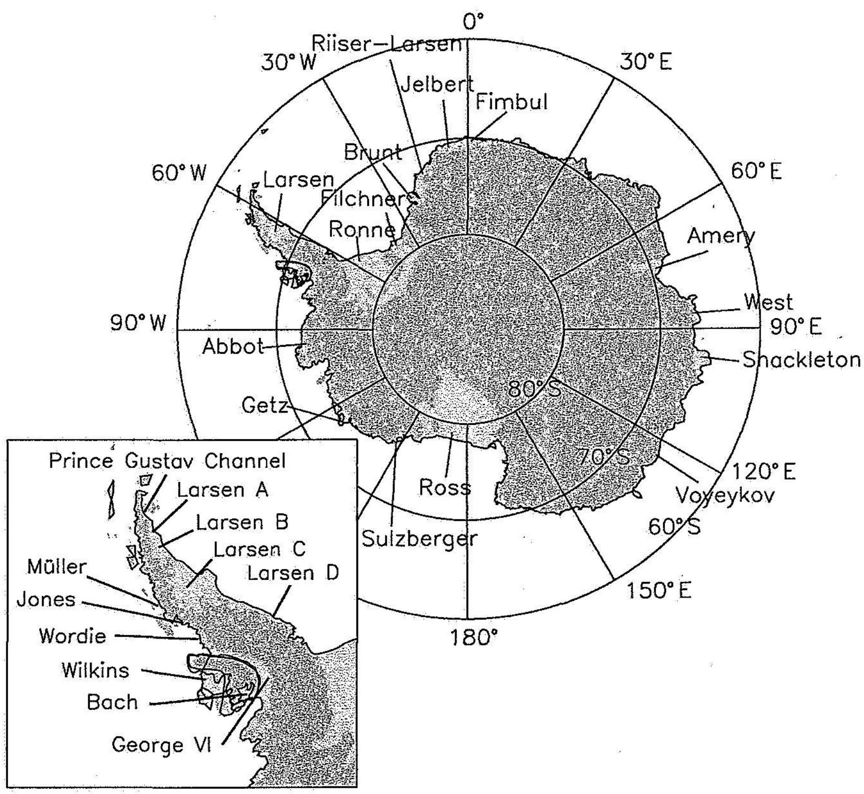
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Abbot
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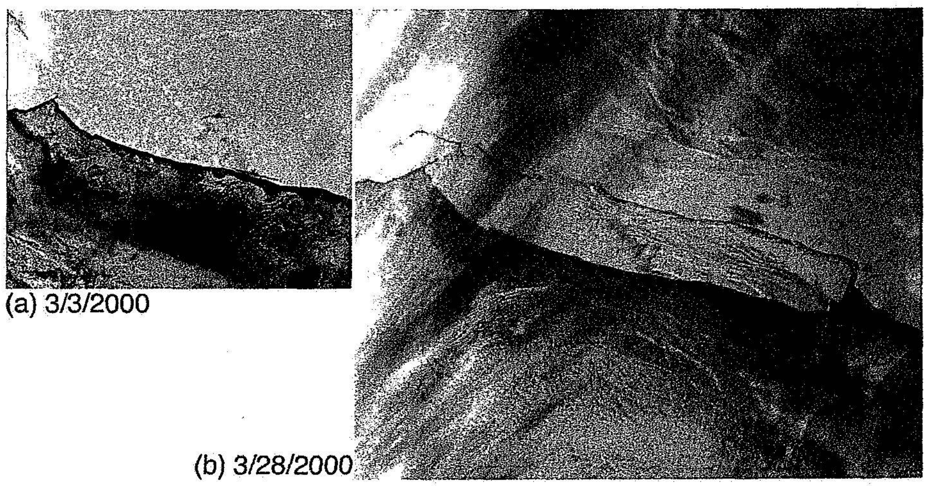

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Figure 5

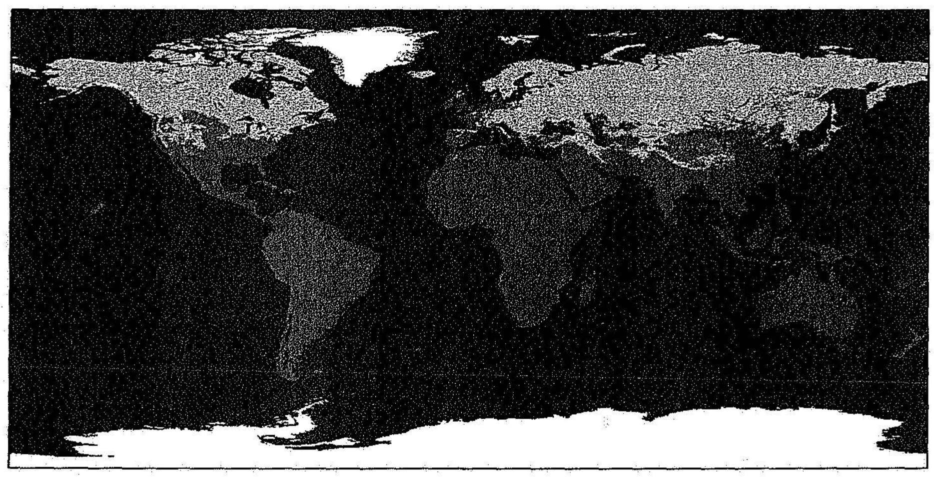

Figure 6