Chapter 2. Rapid Changes in Glaciers and Ice Sheets and their Impacts on Sea Level

Lead Author: Konrad Steffen,* CIRES, University of Colorado, Boulder, CO

Contributing Authors: Peter U. Clark,* Department of Geosciences, Oregon State University, Corvallis

J. Graham Cogley, Department of Geography, Trent University, Peterborough, Ontario

David Holland, Courant Institute of Mathematical Sciences, New York University, NY

Shawn Marshall,* Department of Geography, University of Calgary

Eric Rignot, UCI/ES, Irvine, CA, JPL/Caltech, Pasadena, CA, and and Centro de Estudios Cientificos, Valdivia, Chile

Robert Thomas, EG&G Services, NASA/GSFC/Wallops Flight Facility, Wallops Island, VA, and Centro de Estudios Cientificos, Valdivia, Chile

*SAP 3.4 FACA Committee Member

Key Findings

- Since the mid-19th century, small glaciers (sometimes called “glaciers and ice caps”; see Box 2.1 for definitions) have been losing mass at an average rate equivalent to 0.3-0.4 millimeters per year of sea level rise.
- The best estimate of the current (2007) mass balance of small glaciers is about -400 gigatonnes per year, or nearly 1.1 millimeters sea level equivalent per year.
- The mass balance loss of the Greenland Ice Sheet during the period with good observations decreased from 100 gigatonnes per year (Gt a^{-1}) in the mid 1990’s to more than 200 Gt a^{-1} for the most recent observations in 2006. Much of the loss is by increased summer melting as temperatures rise, but an increasing proportion is by enhanced ice discharge down accelerating glaciers.
- The mass balance for Antarctica is a net loss of about 100 Gt a^{-1} in the mid 1990s, increasing to almost 200 Gt a^{-1} in 2006. There is little surface melting in Antarctica, and the substantial ice losses from West Antarctica and the

Antarctic Peninsula are very likely caused by increasing ice discharge as glacier velocities increase.

- During the last interglacial period (~120 thousand years ago) with similar carbon dioxide levels to pre-industrial values and Arctic summer temperatures warmer than today, sea level was 4-6 meters above present, and sea level rise averaged 10-20 millimeters per year during the deglaciation period after the last ice age with large "meltwater fluxes" exceeding sea level rise of 50 millimeters per year lasting several centuries.
- The cause and mechanism of these meltwater fluxes is not well understood, yet the rapid large loss of ice likely had an effect on ocean circulation that resulted in a forcing of the global climate.
- The potentially sensitive regions for rapid changes in ice volume are those with ice masses grounded below sea level such as the West Antarctic Ice Sheet, with 5 to 6 meters sea level equivalent, or large glaciers in Greenland like the Jakobshavn Isbræ, also known as Jakobshavn Glacier and Sermeq Kujalleq (in Greenlandic) with an over-deepened channel reaching far inland; total breakup of Jakobshavn Isbræ ice tongue in Greenland as well as other tidewater glaciers and ice cap outlets were preceded by its very rapid thinning.
- Several ice shelves in Antarctica are thinning, and their area declined by more than 13,500 square kilometers in the last 3 decades of the 20th century, punctuated by the collapse of the Larsen A and Larsen B ice shelves, soon followed by several-fold increases in velocities of their tributary glaciers.
- The interaction of warm waters with the periphery of the large ice sheets represents a strong potential cause of abrupt change in the big ice sheets, and future changes in ocean circulation and ocean temperatures will very likely produce changes in ice-shelf basal melting, but the magnitude of these changes cannot currently be modeled or predicted. Moreover, calving, which

can originate in fractures far back from the ice front, and ice-shelf breakup are very poorly understood.

- Existing models suggest that climate warming would result in increased melting from coastal regions in Greenland and an overall increase in snowfall. However, they are incapable of realistically simulating the outlet glaciers that discharge ice into the ocean and cannot predict the substantial acceleration of some outlet glaciers that we are already observing.

Recommendations

- Maintain and extend established programs, both governmental and university-based, of mass-balance measurements on small glaciers, and complete the World Glacier Inventory through programs such as the Global Land Ice Measurements from Space (GLIMS) program.
- Maintain climate networks on ice sheets to detect regional climate change and calibrate climate models.
- Utilize existing satellite interferometric synthetic aperture radar (InSAR) data to measure ice velocity, and develop and implement an InSAR mission to allow frequent and comprehensive observations of flow rates in glaciers and ice sheets worldwide.
- Use observations of the time-varying gravity field from satellites such as GRACE, and urgently plan for an appropriate follow-on mission with finer spatial resolution, to contribute to estimating changes in ice sheet mass and data continuity.
- Survey changes in ice-sheet topography using satellite radar (e.g., Envisat and Cryosat-2) and laser (e.g., ICESat-1/2) altimeters, and plan follow-on laser-altimeter missions with a wide-swath altimeter.
- Sustain aircraft observations of surface elevation, ice thickness, and basal characteristics to ensure that such information is acquired at high spatial resolution along specific routes, such as glacier flow lines, and along

transects close to the grounding lines. Observations of ice thickness along these specific routes are particularly important and needed urgently.

- Improve coverage of longer term (centennial to millennial) records of ice sheet and ocean history from geological observations.
- Support field, theoretical, and computational investigations of processes beneath and along ice shelves and beneath glaciers, especially near to the grounding lines of the latter, with the goal of understanding recent increases in mass loss.
- Support a major effort to develop ice-sheet models on a par with current models of the atmosphere and ocean. Particular effort is needed with respect to the modeling of ocean/ice-shelf interactions, of surface mass balance from climatic information, and of all (rather than just some, as now) of the forces which drive the motion of the ice.

1. Summary

1.1 Paleorecord

The most recent time with no ice on the globe was 35 million years ago during a period when the atmospheric carbon dioxide (CO_2) was 1,250± 250 parts per million by volume (ppmV) and a sea level +73 meters (m) higher than today. During the last interglacial period (~120 thousand years ago, ka) with similar CO_2 levels to pre-industrial values and Arctic summer temperatures warmer than today, sea level was 4-6 m above present. Most of that sea level rise (SLR) is believed to have originated from the Greenland Ice Sheet. Sea level rise averaged 10-20 millimeters per year (mm a^{-1}) during the deglaciation period after the last ice age with large "meltwater fluxes" exceeding SLR of 50 mm a^{-1} lasting several centuries. Each of these meltwater fluxes added 1.5–3 times the volume of the current Greenland Ice Sheet (7 m) to the oceans. The cause, ice-sheet source, and mechanism of the meltwater fluxes is not well understood, yet the rapid loss of ice must have had an effect on ocean circulation resulting in a forcing of the global climate.

1.2 Ice Sheets

Rapid changes in ice-sheet mass have surely contributed to abrupt changes in climate and sea level in the past. The mass balance of the Greenland Ice Sheet decreased in the late 1990s to -100 gigatonnes per year (Gt a^{-1}) or even less than -150 Gt a^{-1} for the most recent observations in 2006. It is extremely likely that the Greenland Ice Sheet is losing mass and very likely on an accelerated path since the mid-1990s. The mass balance for Antarctica as a whole is close to balance, but with a likely net loss since 2000 at rates of a few tens of gigatonnes per year. The largest losses are concentrated along the Amundsen and Bellinghausen sectors of West Antarctica and the northern tip of the Antarctic Peninsula. The potentially sensitive regions for rapid changes in ice volume are those with ice masses grounded below sea level such as the West Antarctic Ice Sheet, with 7 m sea level equivalent (SLE), or large glaciers in Greenland like the Jakobshavn, also known as Jakobshavn Isbræ and Sermeq Kujalleq (in Greenlandic), with an over-deepened channel reaching far inland. There are large mass-budget uncertainties from errors in both snow accumulation and calculated ice losses for Antarctica (~±160 Gt a^{-1}) and for Greenland (~±35 Gt a^{1}). Mass-budget uncertainties from aircraft or satellite observations (i.e., radar altimeter, laser altimeter, gravity measurements) are similar in magnitude. Most climate models suggest that climate warming would result in increased melting from coastal regions in Greenland and an overall increase in snowfall. However, they do not predict the substantial acceleration of some outlet glaciers that we are observing. This results from a fundamental weakness in the existing models, which are incapable of realistically simulating the outlet glaciers that discharge ice into the ocean.

Observations show that Greenland is thickening at high elevations, because of the increase in snowfall, which was predicted, but that this gain is more than offset by an accelerating mass loss, with a large component from rapidly thinning and accelerating outlet glaciers. Although there is no evidence for increasing snowfall over Antarctica, observations show that some higher elevation regions are also thickening, likely as a result of high interannual variability in snowfall. There is little surface melting in Antarctica, and the substantial ice losses from West Antarctica and the Antarctic Peninsula are very likely caused by increased ice discharge as velocities of some glaciers increase. This is of particular concern in West Antarctica, where bedrock beneath the ice sheet is deep below sea level, and outlet glaciers are to some extent "contained" by the ice shelves into which they flow. Some of these ice shelves are thinning, and some have totally broken up, and these are the regions where the glaciers are accelerating and thinning most rapidly.

1.3 Small Glaciers

Within the uncertainty of the measurements, the following generalizations are justifiable. Since the mid-19th century, small glaciers have been losing mass at an average rate equivalent to 0.3-0.4 mm a^{-1} of sea level rise. The rate has varied. There was a period of reduced loss between the 1940s and 1970s, with the average rate approaching zero in about 1970. We know with very high confidence that it has been accelerating. The best estimate of the current (2007) mass balance is near to -380 to -400 Gt a^{-1}, or nearly 1.1 mm SLE a^{-1}; this may be an underestimate if, as suspected, the inadequately measured rate of loss by calving outweighs the inadequately measured rate of gain by "internal"* accumulation. Our physical understanding allows us to conclude that if the net gain of radiative energy at the Earth's surface continues to increase, then so will the acceleration of mass transfer from small glaciers to the ocean. Rates of loss observed so far are small in comparison with rates inferred for episodes of abrupt change during the last few hundred thousand years. In a warmer world the main eventual constraint on mass balance will be exhaustion of the supply of ice from glaciers, which may take place in as little as 50-100 years.

1.4 Causes of Change

Potential causes of the observed behavior of ice bodies include changes in snowfall and/or surface melting, long-term response to past changes in climate, and changes in ice dynamics. Smaller glaciers appear to be most sensitive to radiatively induced changes in melting rate, but this may be because of inadequate attention to the dynamics of tidewater glaciers (see Box 2.1 for definitions). Recent observations of the ice sheets have shown that changes in dynamics can occur far more rapidly than previously suspected. There has been a significant increase in meltwater production on the Greenland Ice Sheet for the

* Refreezing at depth of percolating meltwater in spring and summer, and of retained capillary water during winter. Inability to measure these gains leads to a potentially significant systematic error in the net mass balance.

1998-2003 time period compared to the previous three decades, but this loss was partly compensated by increased precipitation. Total melt area is continuing to increase during summer and fall and has already reached up to 50% of the Greenland Ice Sheet; further increase in Arctic temperatures will continue this process and will add additional runoff. Recent rapid changes in marginal regions of both ice sheets show mainly acceleration and thinning, with some glacier velocities increasing more than twofold. Most of these glacier accelerations closely followed reduction or loss of ice shelves. Total breakup of Jakobshavn Isbræ ice tongue in Greenland was preceded by its very rapid thinning. Thinning of more than 1 meter per year (m a^{-1}), and locally more than 5 m a^{-1}, was observed during the past decade for many small ice shelves in the Amundsen Sea and along the Antarctic Peninsula. Significant changes in ice shelf thickness are most readily caused by changes in basal melting. Recent data show a high correlation between periods of heavy surface melting and increase in glacier velocity. A possible cause is rapid meltwater drainage to the glacier bed, where it enhances lubrication of basal sliding. Although no seasonal changes in the speeds were found for the rapid glaciers that discharge most ice from Greenland, meltwater remains an essential control on glacier flow and an increase in meltwater production in a warmer climate could likely have major consequences of increased flow rates and ice mass loss.

1.5 Ocean Influence

The interaction of warm waters with the periphery of the large ice sheets represents one of the most significant possibilities for abrupt change in the climate system. Mass loss through oceanic melting and iceberg calving accounts for more than 95% of the ablation from Antarctica and 40-50% of the ablation from Greenland. Future changes in ocean circulation and ocean temperatures will produce changes in basal melting, but the magnitude of these changes is currently not well modeled or predicted. The susceptibility of ice shelves to high melt rates and to collapse is a function of the presence of warm waters entering the cavities beneath ice shelves. Ocean circulation is driven by density contrasts of water masses and by surface wind forcing. For abrupt climate change scenarios, attention should be focused on the latter. A change in wind patterns could produce large and fast changes in the temperatures of ocean waters. A thinning ice shelf results in glacier ungrounding, which is the main cause of the glacier acceleration because it has a large effect on the force balance near the ice front. Calving, which can originate in fractures far back from the ice front, is very poorly understood. Antarctic ice-shelf area declined by more than 13,500 square kilometers (km^2) in the last 3 decades of the 20th century, punctuated by the collapse of the Larsen A and Larsen B ice shelves. Ice shelf viability is compromised if mean annual air temperature exceeds −5°C. Observations from the last decade have radically altered the thinking on how rapidly an ice sheet can respond to perturbations at the marine margin. Several-fold increases in discharge followed the collapse of ice shelves on the Antarctic Peninsula; this is something models did not predict *a priori*. No ice sheet model is currently capable of capturing the glacier speedups in Antarctica or Greenland that have been observed over the last decade.

1.6 Sea Level Feedback

The primary factor that raises concerns about the potential of abrupt changes in sea level is that large areas of modern ice sheets are currently grounded below sea level. An important aspect of these marine-based ice sheets which has long been of interest is that the beds of ice sheets grounded below sea level tend to deepen inland, either due to overdeepening from glacial erosion or isostatic adjustment. Marine ice sheets are inherently unstable, whereby small changes in climate could trigger irreversible retreat of the grounding line (locations along the coast where the ice is no longer ground supported and begins to float). For a tidewater glacier, rapid retreat occurs because calving rates increase with water depth. In Greenland, few outlet glaciers remain below sea level very far inland, indicating that glacier retreat by this process will eventually slow down or halt. A notable exception may be Greenland's largest outlet glacier, Jakobshavn Isbræ, which appears to tap into the central core of Greenland that is below sea level. Given that a grounding line represents the point at which ice becomes buoyant, then a rise in sea level will cause grounding line retreat. This situation thus leads to the potential for a positive feedback to develop between ice retreat and sea level rise. In considering various stabilizing factors, however, we conclude that, provided there is no rapid loss of ice shelves and attendant sea level rise, sea level forcing and feedback is unlikely to be a significant determinant in causing rapid ice-sheet changes in the coming century.

2. What is the Record of Past Changes in Ice Sheets and Global Sea Level?

2.1 Reconstructing Past Changes in Ice Sheets

There are several methods available to reconstruct past changes in ice-sheet area and mass, each with their own strengths and shortcomings. Terrestrial records provide information of former ice-sheet extent, whereby temporary stabilization of an ice margin may be recorded by an accumulation of sediment (moraine) that may be dated by isotopic methods (e.g., ^{10}Be, ^{14}C, etc.). These records are important in identifying the last maximum extent and retreat history of an ice sheet (e.g., *Dyke, 2004*), but most terrestrial records of glaciation prior to the Last Glacial Maximum (LGM) ~21,000 years ago have been removed by erosion, limiting the application of these records to times since the LGM. Moreover, in most cases they only provide information on extent but not thickness, so that potential large changes in volume are not necessarily captured by these records.

Application of this strategy to the retreat of the West Antarctic Ice Sheet (WAIS) from its LGM position provides important context for understanding current ice dynamics. *Conway et al. (1999)* dated recession of the WAIS grounding line in the Ross Sea embayment and found that modern grounding-line retreat is part of an ongoing recession that has been underway for the last ~9,000 years. *Stone et al. (2003)* took a slightly different approach to evaluating WAIS deglaciation whereby they determined the rate of lowering of the ice-sheet surface by dating recessional features preserved on a mountain slope that projected upwards through the ice sheet. Their results complemented those of *Conway et al. (1999)* in showing ice-sheet thinning for the last ~10,000 years that may still be underway. These results are important not only in providing constraints on long-term changes against which to evaluate short-term controls on ice-sheet change but also in providing important benchmarks for modeling ice sheet evolution. Nevertheless, the spatial coverage of these data from Antarctica remains limited, and additional such constraints are needed.

Another strategy for constraining past ice-sheet history is based on the fact that the weight of ice sheets results in isostatic compensation of the underlying solid Earth, generally referred to as glacial isostatic adjustment (GIA). Changes in ice-sheet mass cause vertical motions that may be recorded along a formerly glaciated coastline where the global sea level serves as a datum. Since changes in ice mass will also cause changes in local (due to gravity) and global (due to volume) sea level, the changes in sea level at a particular coastline record the difference between vertical motions of the land and sea, commonly referred to as near-field relative sea level (RSL) changes. Models that incorporate the physical properties of the solid Earth invert the RSL records to determine the ice-loading history required to produce the isostatic adjustment preserved by these records (e.g., *Peltier, 2004*). Because of the scarcity of such near-field RSL sites from the Antarctic continent, *Ivins and James (2005)* constructed a history of Antarctic ice mass changes from geologic evidence of ice-margin and ice-thickness changes, such as described above (*Conway et al., 1999; Stone et al., 2003*). This ice-load history was then used to derive a model of present-day GIA.

Regardless how it is derived, the GIA process must be accounted for when using satellite altimetry and gravity data to infer changes in ice mass (e.g., *Velicogna and Wahr, 2006b*) (see Sec. 3). Given the poor constraints from near-field RSL records and geologic records (and their dating) of ice limits and thicknesses for Antarctica, as well as uncertainties in properties of the solid Earth used in these models, uncertainties in this GIA correction is large (*Velicogna and Wahr, 2006; Barletta et al., 2008*). Accordingly, improvements in understanding present-day GIA are required to improve ice-mass estimates from altimetry and gravity data.

2.2 Reconstructing Past Sea Level

Sea level is a dynamic feature of the Earth system, changing at all time scales in response to tectonics and climate. As discussed above, changes that occur locally, due to regional uplift or subsidence, relative to global sea level are referred to as relative sea level (RSL) changes, whereas changes that occur globally are referred to as eustatic changes. On time scales greater than 100,000 years, eustatic changes occur primarily from changes in ocean-basin volume induced by variations in the rate of sea-floor spreading. On shorter time scales, eustatic changes occur primarily from changes in ice volume, with secondary contributions (order of 1 m) associated with changes in ocean temperature or salinity (steric changes). Changes in global ice volume also cause global changes in RSL in response to the redistribution of mass between land to sea and attendant isostatic compensation and gravitational reequilibration. This GIA process must be accounted for in determining eustatic changes from geomorphic records of former sea level. Because the effects of the GIA process diminish with distance from areas of former glaciation, RSL records from far-field sites provide a close approximation of eustatic changes.

An additional means to constrain past sea level change is based on the change in the ratio of ^{18}O to ^{16}O of seawater (expressed in reference to a standard as $\delta^{18}O$) that occurs as the lighter isotope is preferentially removed and stored in growing ice sheets (and vice versa). These $\delta^{18}O$ changes are recorded in the carbonate fossils of microscopic marine organisms (foraminifera) and provide a near-continuous time series of changes in ice volume and corresponding eustatic sea level. However, because changes in temperature also affect the $\delta^{18}O$ of foraminifera through temperature dependent fractionation during calcite precipitation, the $\delta^{18}O$ signal in marine records reflects some combination of ice volume and temperature. Figure 2.1 shows one attempt to isolate the ice-volume component in the marine $\delta^{18}O$ record (*Waelbroeck et al., 2002*). Although to a first order this record agrees well with independent estimates of eustatic sea level, this approach fails to capture some of the abrupt changes in sea level that are documented by paleoshoreline evidence (*Clark and Mix, 2002*), suggesting that large changes in ocean temperature may not be accurately captured at these times.

2.3 Sea Level Changes During the Last Glacial Cycle

The record of past changes in ice volume provides important insight to the response of large ice sheets to climate change. Our best constraints come from the last glacial cycle (125,000 years ago to the present), when the combination of paleoshorelines and the global $\delta^{18}O$ record provides reasonably well-constrained evidence of changes in eustatic sea level (Fig. 2.1). Changes in ice volume over this interval were paced by changes in the Earth's orbit around the sun (orbital timescales, 10^4-10^5 a), but amplification from changes in atmospheric CO_2 is required to explain the synchronous and extensive glaciation in both polar hemispheres. Although the phasing relationship between sea level and atmospheric CO_2 remains unclear (*Shackleton, 2000; Kawamura et al., 2007*), their records are coherent and there is a strong positive relation between the two (Fig. 2.2).

A similar correlation holds for earlier times in Earth history when atmospheric CO_2 concentrations were in the range of projections for the end of the 21st century (Fig. 2.2). The most recent time when no permanent ice existed on the planet (sea level = +73 m) occurred >35 million years ago when atmospheric CO_2 was 1,250±250 ppmV (*Pagani et al., 2005*). In the early Oligocene (~32 million years ago), atmospheric CO_2 decreased to 500±50 ppmV (*Pagani et al., 2005*), which was accompanied by the first growth of permanent ice on the Antarctic continent, with an attendant eustatic sea level lowering of 45±5 m (*DeConto and Pollard, 2003*). The fact that sea level projections for the end of the 21st century (*Meehl et al., 2007; Rahmstorf, 2007; Horton et al., 2008*) are far below those suggested by this relation (Fig. 2.2) reflects the long response time of ice sheets to climate change. With sufficient time at elevated atmospheric CO_2 levels, sea level will continue to rise as ice sheets continue to lose mass (*Ridley et al., 2005*). What remains unclear, however, is what the response time of large ice sheets is. If the ice-dynamical changes observed over the last few years (see Sec. 3) are sustained under global warming, the response time will be significantly shorter.

During the last interglaciation period (LIG), from ~129,000 years ago to at least 118,000 years ago, CO_2 levels were similar to pre-industrial levels (*Petit et al., 1999; Kawamura et al., 2007*), but large positive anomalies in early-summer solar radiation driven by orbital changes caused Arctic summer temperatures to be warmer than they are today (*Otto-Bleisner et al.,* 2006). Corals on tectonically stable coasts indicate that sea level during the LIG was 4 to 6 m above present (Fig. 2.1) (*Stirling et al., 1995, 1998; Muhs et al., 2002*), and ice-core records (*Koerner, 1989; Raynaud et al., 1997*) and modeling (*Cuffey and Marshall, 2000; Otto-Bliesner et al., 2006*) indicate that much of this rise originated from a reduction in the size of the Greenland Ice Sheet, although some contribution from the Antarctic Ice Sheet may be required as well.

At the last glacial maximum, about 21,000 years ago, ice volume and area were more than twice modern, with most of the increase occurring in the Northern Hemisphere (*Clark and Mix, 2002*). Deglaciation was forced by warming from changes in the Earth’s orbital parameters, increasing greenhouse gas concentrations, and attendant feedbacks. The record of deglacial sea level rise is particularly well-constrained from paleoshoreline evidence (Fig. 2.3). Deglacial sea-level rise averaged 10-20 mm a^{-1}, or at least 5 times faster than the average rate of the last 100 years (Fig. 2.1), but with variations including two extraordinary episodes at 19,000 thousand years before present (19 ka BP) and 14.5 ka BP, when peak rates potentially exceeded 50 mm a^{-1} *(Fairbanks, 1989; Yokoyama et al., 2000; Clark et al., 2004*) (Fig. 2.3), or five times faster than projections for the end of this century (*Rahmstorf, 2007*). Each of these "meltwater pulses" added the equivalent of 1.5 to 3 Greenland ice sheets (~7 m) to the oceans over a one- to five-century period, clearly demonstrating the potential for ice sheets to cause rapid and large sea level changes. A third meltwater pulse may have occurred ~11,700 years ago (*Fairbanks, 1989*), but the evidence for this event is less clear (*Bard et al., 1996; Bassett et al., 2005*).

Recent analyses indicate that the earlier 19-ka event originated from Northern Hemisphere ice (*Clark et al., 2004*). The ~20-m sea level rise ~14,500 years ago (*Fairbanks, 1989; Hanebuth et al., 2000*), commonly referred to as meltwater pulse (MWP) 1A, indicates an extraordinary episode of ice-sheet collapse, with an associated freshwater flux to the ocean of ~0.5 sverdrup (Sv) over several hundred years. The timing, source and climatic effect of MWP-1A, however, remain widely debated. In one scenario, the event was triggered by an abrupt warming (start of the Bølling warm interval) in the North Atlantic region, causing widespread melting of Northern Hemisphere ice sheets (*Fairbanks et al., 1992; Peltier, 2005*). In another scenario, MWP-1A largely originated from the Antarctic Ice Sheet (*Clark et al., 1996, 2002; Bassett et al., 2005*), possibly in response to the ~3,500-year warming in the Southern Hemisphere that preceded the event (*Blunier and Brook, 2001; Clark et al., 2004*). Although the cause of these events has yet to be established, their occurrences following hemispheric warming may implicate short-term dynamic processes activated by that warming, similar to those now being identified around Greenland and Antarctica.

Direct evidence from terrestrial geologic records of one scenario versus the other, however, thus far remains inconclusive. Well-dated terrestrial records of deglaciation of Northern Hemisphere ice sheets, which largely constrain changes in area only, show no acceleration of ice-margin retreat at this time (e.g., *Dyke, 2004; Rinterknecht et al., 2006*), leading some to conclude that the event occurred largely by ice-sheet deflation with little response of the margin (*Simms et al., 2007*). The record of deglaciation of the Antarctic Ice Sheet is less well constrained, and available evidence presents conflicting results, from no contribution (*Ackert et al., 2007; Mackintosh et al., 2007*), to a small contribution (*Heroy and Anderson, 2007; Price et al., 2007*), to a dominant contribution (*Bassett et al., 2007*).

The large freshwater fluxes that these events represent also underscore the significance of rapid losses of ice to the climate system through their effects on ocean circulation. An important component of the ocean's overturning circulation involves formation of deepwater at sites in the North Atlantic Ocean and around the Antarctic continent, particularly the Weddell and Ross Seas. The rate at which this density-driven thermohaline circulation occurs is sensitive to surface fluxes of heat and freshwater. Eustatic rises associated with the two deglacial meltwater pulses correspond to freshwater fluxes $\geq$ 0.25 Sv, which according to climate models would induce a large change in the thermohaline circulation (*Stouffer et al., 2006; Weaver et al., 2003*).

3. The current state of glaciers, ice caps, and ice sheets

Rapid changes in ice sheet mass have surely contributed to abrupt climate change in the past, and any abrupt change in climate is sure to affect the mass balance (see Box 2.2) of at least some of the ice on Earth.

3.1 Mass-Balance Techniques

Traditional estimates of the surface mass balance are from repeated measurements of the exposed length of stakes planted in the snow or ice surface. Temporal change in this length, multiplied by the density of the mass gained or lost, is the surface mass balance at the location of the stake. (In principle the density of mass gained can be measured in shallow cores or snow pits; but in practice there can be considerable uncertainty about density; see, e.g., Sec. 3.1.2.2.) Various means have been devised to apply corrections for sinking of the stake bottom into the snow, densification of the snow between the surface and the stake bottom, and the refreezing of surface meltwater at depths below the stake bottom. Such measurements are time consuming and expensive, and they need to be supplemented at least on the ice sheets by model estimates of precipitation, internal accumulation, sublimation, and melting. Regional atmospheric climate models, calibrated by independent in situ measurements of temperature and pressure (e.g., *Steffen and Box, 2001; Box et al., 2006*) provide estimates of snowfall and sublimation. Estimates of surface melting/evaporation come from energy-balance models and degree-day or temperature-index models (reviewed in, e.g., *Hock, 2003*), which are also validated using independent in situ measurements. Within each category there is a hierarchy of models in terms of spatial and temporal resolution. Energy-balance models are physically based, require detailed input data and are more suitable for high resolution in space and time. Degree-day models are advantageous for the purposes of estimating worldwide glacier melt, since the main inputs of temperature and precipitation are readily available in gridded form from Atmosphere-Ocean General Circulation Models (AOGCMs).

Techniques for measuring total mass balance include:

- the mass-budget approach, comparing gains by surface and internal accumulation with losses by ice discharge, sublimation, and meltwater runoff;
- repeated altimetry, or equivalently levelling or photogrammetry, to measure height changes, from which mass changes are inferred;
- satellite measurements of temporal changes in gravity, to infer mass changes directly.

All three techniques can be applied to the large ice sheets; most studies of ice caps and glaciers are annual (or seasonal) mass-budget measurements, with recent studies also using multi-annual laser and radar altimetry. The third technique is applied only to large, heavily glaciated regions such as Alaska, Patagonia, Greenland, and Antarctica. Here, we summarize what is known about total mass balance, to assess the merits and limitations of different approaches to its measurement, and to identify possible improvements that could be made over the next few years.

3.1.1 Mass Balance

Snow accumulation is estimated from stake measurements, annual layering in ice cores, sometimes with interpolation using satellite microwave measurements (*Arthern et al., 2006*), or meteorological information *(Giovinetto and Zwally, 2000*) or shallow radar sounding (*Jacka et al., 2004*), or from regional atmospheric climate modeling (e.g., *van de Berg et al., 2006; Bromwich et al., 2004*). The state of the art in estimating snow accumulation for periods of up to a decade is rapidly becoming the latter, with surface data being used mostly for validation, not to drive the models. This is not surprising given the immensity of large ice sheets and the difficulty of obtaining appropriate spatial and temporal sampling of snow accumulation at the large scale by field parties, especially in Antarctica.

Ice discharge is the product of velocity and thickness, with velocities measured in situ or remotely, preferably near the grounding line where velocity is almost depth independent. Thickness is measured by airborne radar, seismically, or from measured surface elevations assuming hydrostatic equilibrium, for floating ice near grounding lines. Velocities are measured by ground-based survey, photogrammetry or with satellite sensors; the latter are mostly imaging radars operating interferometrically. Grounding lines are poorly known from *in situ* measurement or visible-band imagery but can be mapped very accurately with satellite interferometric imaging radars.

Meltwater runoff (large on glaciers and ice caps, and near the Greenland coast and parts of the Antarctic Peninsula, but small or zero elsewhere) is traditionally inferred from stake measurements but more and more from regional atmospheric climate models validated with surface observations where available (e.g., *Hanna et al., 2005; Box et al., 2006*). The typically small mass loss by melting beneath grounded ice is also estimated from models.

Mass-budget calculations involve the comparison of two very large numbers, and small errors in either can result in large errors in estimated total mass balance. For example, total accumulation over Antarctica, excluding ice shelves, is about 1,850 Gt a^{-1} (*Vaughan et al., 1999; Arthern et al., 2006; van de Berg et al., 2006*), and 500 Gt a*–1* over

Greenland (*Bales et al., 2001*). Associated errors are difficult to assess because of high temporal and spatial variability, but they are probably about ±5% (20-25 Gt a^{-1}) for Greenland. The errors for Antarctica (*Rignot et al., 2008*) range from 5% in dry interior basins to 20% in wet coastal basins. The overall uncertainty on the 2055 Gt a^{-1} accumulation is 122 Gt a^{-1} or 6%.

Broad interferometric SAR (InSAR) coverage and progressively improved estimates of grounding-line ice thickness have substantially improved ice-discharge estimates, yet incomplete data coverage and residual errors imply errors on total discharge of 2% (*Rignot et al., 2008*). Consequently, assuming these errors in both snow accumulation and ice losses, current (2006) mass-budget uncertainty is ~ ±92 Gt a^{-1} (*Rignot et al., 2008*) for Antarctica and ±35 Gt a^{-1} for GreenlandMoreover, additional errors may result from accumulation estimates being based on data from the past few decades; at least in Greenland, we know that snowfall is increasing with time. Similarly, it is becoming clear that glacier velocities can change substantially over quite short time periods (*Rignot and Kanagaratnam, 2006*), and the time period investigated (last decade) showed an increase in ice velocities, so these error estimates might well be lower limits.

3.1.2 Repeated Altimetry

Rates of surface-elevation change with time (dS/dt) reveal changes in ice-sheet mass after correction for changes in depth/density profiles and bedrock elevation, or for hydrostatic equilibrium if the ice is floating. Satellite radar altimetry (SRALT) has been widely used (e.g., *Shepherd et al., 2002; Davis et al., 2005; Johannessen et al., 2005; Zwally et al., 2005*), together with laser altimetry from airplanes (Arendt e*t al., 2002; Krabill et al., 2000*), and from NASA's ICESat (*Zwally et al., 2002a; Thomas et al., 2006*). Modeled corrections for isostatic changes in bedrock elevation (e.g, *Peltier, 2004*) are small (a few millimeters per year) but with errors comparable to the correction. Those for near-surface snow density changes (*Arthern and Wingham, 1998; Li and Zwally, 2004*) are larger (1 or 2 cm a^{-1}) and also uncertain.

3.1.2.1 Satellite Radar Altimetry

Available SRALT data are from altimeters with a beam width of 20 km or more, designed and demonstrated to make accurate measurements over the almost flat, horizontal ocean. Data interpretation is more complex over sloping and undulating ice-sheet surfaces with spatially and temporally varying dielectric properties. Errors in SRALT-derived values of dS/dt are typically determined from the internal consistency of the measurements, often after iterative removal of dS/dt values that exceed some multiple of the local value of their standard deviation. This results in small error estimates (e.g., *Zwally et al., 2005, Wingham et al., 2006*) that are smaller than the differences between different interpretations of essentially the same SRALT data (*Johannessen et al., 2005; Zwally et al., 2005*). In addition to processing errors, uncertainties result from the possibility that SRALT estimates are biased by the effects of local terrain or by surface snow characteristics, such as wetness (*Thomas et al., in press*). Observations by other techniques reveal extremely rapid thinning along Greenland glaciers that flow along depressions where dS/dt cannot be inferred from SRALT data, and collectively these glaciers are responsible for most of the mass loss from the ice sheet (*Rignot and Kanagaratnam, 2006*), implying that SRALT data underestimate near-coastal thinning rates significantly. Moreover, the zone of summer melting in Greenland progressively increased between the early 1990s and 2005 (*Box et al., 2006*), probably raising the radar reflection horizon within near-surface snow by a meter or more over a significant fraction of the ice-sheet percolation facies (*Jezek et al., 1994*). Comparison between SRALT and laser estimates of dS/dt over Greenland show differences that are equivalent to the total mass balance of the ice sheet (*Thomas et al., 2007*)

3.1.2.2 Aircraft and Satellite Laser Altimetry

Laser altimeters provide data that are easier to validate and interpret: footprints are small (about 1 m for airborne laser, and 60 m for ICESat), and there is negligible laser penetration into the ice. However, clouds limit data acquisition, and accuracy is affected by atmospheric conditions and particularly by laser-pointing errors. The strongest limitation by far is that existing laser data are sparse compared to SRALT data.

Airborne laser surveys over Greenland in 1993-94 and 1998-89 yield elevation estimates accurate to ~10 cm along survey tracks (*Krabill et al., 2002*), but with large gaps between flight lines and an incomplete coverage of the glaciers. ICESat orbit-track separation is also quite large compared to the size of a large glacier, particularly in southern Greenland and the Antarctic Peninsula where rapid changes are occurring, and elevation errors along individual orbit tracks can be large (many tens of centimeters) over sloping ice. Progressive improvement in ICESat data processing is reducing these errors and, for both airborne and ICESat surveys, most errors are independent for each flight line or orbit track, so that estimates of dS/dt averaged over large areas containing many survey tracks are affected most by systematic ranging, pointing, or platform-position errors, totaling probably less than 5 cm. In Greenland, such conditions typically apply at elevations above 1,500-2,000 m. dS/dt errors decrease with increasing time interval between surveys. Nearer the coast there are large gaps in both ICESat and airborne coverage, requiring dS/dt values to be supplemented by degree-day estimates of anomalous melting *(Krabill et al., 2000, 2004*). This increases overall errors and almost certainly underestimates total losses because it does not take full account of dynamic thinning of unsurveyed outlet glaciers.

In summary, dS/dt errors cannot be precisely quantified for either SRALT data, because of the broad radar beam, limitations with surface topography at the coast, and time-variable penetration, or laser data, because of sparse coverage. The SRALT limitations discussed above will be difficult to resolve. Laser limitations result primarily from poor coverage and can be partially resolved by increasing spatial resolution.

All altimetry mass-balance estimates include additional uncertainties in:

1. The density (rho) assumed to convert thickness changes to mass changes. If changes are caused by recent changes in snowfall, the appropriate density may be as low as 300 kilograms per cubic meter (kg m^{-3}); for long-term changes, it may be as high as 900 kg m^{-3}. This is of most concern for high-elevation regions with small dS/dt, where the simplest assumption is rho = 600±300 kg m^{-3}. For a 1-cm a^{-1} thickness change over the million square kilometers of Greenland above 2,000

m, uncertainty would be ±3 Gt a^{-1}. Rapid, sustained changes, commonly found near the coast, are almost certainly caused by changes in melt rates or glacier dynamics, and for which rho is ~900 kg m^{-3}.

2. Possible changes in near-surface snow density. Densification rates are sensitive to snow temperature and wetness. Warm conditions favor more rapid densification (*Arthern and Wingham, 1998; Li and Zwally, 2004*), and melting is likely to be followed by refreezing as ice. Consequently, recent Greenland warming probably caused surface lowering simply from this effect. Corrections are inferred from largely unvalidated models and are typically <2 cm a^{-1}, with unknown errors. If overall uncertainty is 5 mm a^{-1}, associated mass-balance errors are approximately ±8 Gt a^{-1} for Greenland and ±60 Gt a^{-1} for Antarctica.
3. The rate of crustal uplift. This is inferred from glacio-isostatic models and has uncertain errors. An overall uncertainty of 1 mm a^{-1} would result in mass-balance errors of about ±2 Gt a^{-1} for Greenland and ±12 Gt a^{-1} for Antarctica.
4. There is evidence for large interannual to decadal changes in snowfall and hence accumulation in Antarctica (*Monaghan et al., 2006*) and also a lack of overall trend in net accumulation over the entire continent. This makes it particularly difficult to estimate the mass balance of interior regions because satellite missions have been collecting data for merely 10-15 years. Such investigation clearly requires several decades of data to provide meaningful results.

3.1.3 Temporal Variations in Earth's Gravity

Since 2002, the GRACE satellite has measured Earth's gravity field and its temporal variability. After removing the effects of tides, atmospheric loading, spatial and temporal changes in ocean mass, etc., high-latitude data contain information on temporal changes in the mass distribution of the ice sheets and underlying rock. Because of its high altitude, GRACE makes coarse-resolution measurements of the gravity field and its changes with time. Consequently, resulting mass-balance estimates are also at coarse resolution – several hundred kilometers. But this has the advantage of covering entire ice sheets, which is extremely difficult using other techniques. Consequently, GRACE estimates include mass changes on the many small ice caps and isolated glaciers that surround the big ice sheets; the former may be quite large being strongly affected by changes in the coastal climate. Employing a surface mass concentration (mascon) solution technique, *Luthcke et al. (2006)* computed multi-year time series of GRACE-derived surface mass flux for Greenland and Antarctica coastal and interior ice sheet sub-drainage systems as well as the Alaskan glacier systems. These mascon solutions provide important observations of the seasonal and inter-annual evolution of the Earth's land ice.

Error sources include measurement uncertainty, leakage of gravity signal from regions surrounding the ice sheets, interannual variability in snowfall, melt and ice dynamics, and causes of gravity changes other than ice-sheet changes. Of these, the most serious are the gravity changes associated with vertical bedrock motion. *Velicogna and Wahr (2005)* estimated a mass-balance correction of 5±17 Gt a–1 for bedrock motion in Greenland, and a correction of 173±71 Gt a–1 for Antarctica (*Velicogna and Wahr, 2006a*), which may be under-estimated (*Horwath and Dietrich, 2006*) or quite reasonable (*Barletta et al., 2008*). Although other geodetic data (variations in length of day, polar wander, etc.) provide constraints on mass changes at high latitudes, unique solutions are not yet possible from these techniques. One possible way to reduce uncertainties significantly, however, is to combine time series of gravity measurements with time series of elevation changes, records of rock uplift from GPS receivers, and records of snow accumulation from ice cores. Yet, this combination requires years to decades of data to provide a significant reduction in uncertainty (see point 4 above).

3.2 Mass Balance of the Greenland and Antarctic Ice Sheets

Ice locked within the Greenland and Antarctic ice sheets (Table 2.1) has long been considered comparatively immune to change, protected by the extreme cold of the polar regions. Most model results suggested that climate warming would result primarily in increased melting from coastal regions and an overall increase in snowfall, with net 21st century effects probably a small mass loss from Greenland and a small gain in Antarctica, and little combined impact on sea level (*Church et al., 2001*). Observations generally confirmed this view, although Greenland measurements during the 1990s (*Krabill et al., 2000; Abdalati et al., 2001*) began to suggest that there might also be a component from ice-dynamical responses, with very rapid thinning on several outlet glaciers. Such responses had not been seen in prevailing models of glacier motion, primarily determined by ice temperature and basal and lateral drag, coupled with the enormous thermal inertia of a large glacier.

Increasingly, measurements in both Greenland and Antarctica show rapid changes in the behavior of large outlet glaciers. In some cases, once-rapid glaciers have slowed to a virtual standstill, damming up the still-moving ice from farther inland and causing the ice to thicken (*Joughin et al., 2002; Joughin and Tulaczyk, 2002*). More commonly, however, observations reveal glacier acceleration. This may not imply that glaciers have only recently started to change; it may simply mean that major improvements in both quality and coverage of our measurement techniques are now exposing events that also occurred in the past. But in some cases, changes have been very recent. In particular, velocities of tributary glaciers increased markedly very soon after ice shelves or floating ice tongues broke up (e.g., *Scambos et al., 2004; Rignot et al., 2004a*). Moreover, this is happening along both the west and east coasts of Greenland (*Joughin et al., 2004; Howat et al., 2005; Rignot and Kanagaratnam,* 2006) and in at least two locations in Antarctica (*Rignot et al.,* 2002; *Joughin et al., 2003; Scambos et al., 2004; Rignot et al., 2004a*). Such dynamic responses are not explainable in large-scale ice sheet predictive models, nor is the forcing thought responsible for initiating them included in these ice sheet evolutive models.

3.2.1 Greenland

Above ~2,000 m elevation, near-balance between about 1970 and 1995 (*Thomas et al., 2001*) shifted to slow thickening thereafter (*Thomas et al., 2001, 2006; Johannessen et al., 2005; Zwally et al., 2005*). Nearer the coast, airborne laser altimetry (ATM) surveys supplemented by modeled summer melting show widespread thinning (*Krabill et al., 2000, 2004*), resulting in net loss from the ice sheet of 27±23 Gt a–1, equivalent to ~0.08 mm a–1 sea level equivalent (SLE) between 1993-94 and 1998-89 doubling to 55±23 Gt a–1 for 1997-2003[†]. However, the airborne surveys did not include some regions where

[†] Note that these values differ from those in the Krabill et al. publications primarily because they take account of possible surface lowering by accelerated snow densification as air temperatures rise; moreover, they probably underestimate total losses because the ATM surveys undersample thinning coastal glaciers.

other measurements show rapid thinning, so these estimates represent lower limits of actual mass loss.

More recently, four independent studies also show accelerating losses from Greenland:

(1) Analysis of gravity data from GRACE show total losses of 75±20 Gt a^{-1} between April 2002 and April 2004 rising to 223±33 Gt a^{-1} between May 2004 and April 2006 (*Velicogna and Wahr, 2005, 2006a*). (2) Other analyses of GRACE data show losses of 129±15 Gt a^{-1} for July 2002 through March 2005 (*Ramillien et al., 2006*), (3) 219±21 Gt a^{-1} for April 2002 through November 2005 (*Chen et al., 2006*), and (4) 101±16 Gt a^{-1} for July 2003 to July 2005 (*Luthcke et al., 2006*). Although the large scatter in the estimates for similar time periods suggests that errors are larger than quoted, these results show an increasing trend in mass loss.

Interpretations of SRALT data from ERS-1 and 2 (*Johannessen et al., 2005; Zwally et al., 2005*) show quite rapid thickening at high elevations, with lower elevation thinning at far lower rates than those inferred from other approaches that include detailed observations of these low-elevation regions. The *Johannessen et al. (2005)* study recognized the unreliability of SRALT data at lower elevations because of locally sloping and undulating surface topography. *Zwally et al. (2005)* attempted to overcome this by including dS/dt estimates for about 3% of the ice sheet derived from earlier laser altimetry, to infer a small positive mass balance of 11±3 Gt a–1 for the entire ice sheet between April 1992 and October 2002.

Mass-budget calculations for most glacier drainage basins indicate total ice-sheet losses increasing from 83±28 Gt a^{-1} in 1996 to 127±28 Gt a^{-1} in 2000 and 205±38 Gt a^{-1} in 2005 (*Rignot and Kanagaratnam, 2006*). Most of the glacier losses are from the southern half of Greenland, especially the southeast sector, center east, and center west. In the northwest, losses were already significant in the early 1990s and did not increase in recent decades. In the southwest, losses are low but slightly increasing. In the north, losses are very low, but also slightly increasing in the northwest and northeast.

Comparison of 2005 ICESat data with 1998-89 airborne laser surveys shows losses during the interim of 80±25 Gt a^{-1} (*Thomas et al., 2006*), and this is probably an underestimate because of sparse coverage of regions where other investigations show large losses.

The pattern of thickening/thinning over Greenland, derived from laser-altimeter data, is shown in Figure 2.4, with the various mass-balance estimates summarized in Figure 2.5. It is clear that the SRALT-derived estimate differs widely from the others, each of which is based on totally different methods, suggesting that the SRALT interpretations underestimate total ice loss for reasons discussed in Section 3.1.1. Here, we assume this to be the case, and focus on the other results shown in Figure 2.5, which strongly indicate net ice loss from Greenland at rates that increased from at least 27 Gt a–1 between 1993-94 and 1998-99 to about double between 1997 and 2003, to more than 80 Gt a–1 between 1998 and 2004, to more than 100 Gt a–1 soon after 2000, and to more than 200 Gt a-1 after 2005. There are insufficient data for any assessment of total mass balance before 1990, although mass-budget calculations indicated near overall balance at elevations above 2,000 m and significant thinning in the southeast (*Thomas et al., 2001*).

3.2.2 Antarctica

Determination of the mass budget of the Antarctic ice sheet is not as advanced as that for Greenland. Melt is not a significant factor, but uncertainties in snow accumulation are larger because fewer data have been collected, and ice thickness is poorly characterized along outlet glaciers. Instead, ice elevations, which have been improved with ICESat data, are used to calculate ice thickness from hydrostatic equilibrium at the glacier grounding line. The grounding line position and ice velocity are inferred from Radarsat-1 and ERS-1/2 InSAR. For the period 1996-2000, *Rignot and Thomas (2002)* inferred East Antarctic growth at 20±1 Gt a^{-1}, with estimated losses of 44±13 Gt a^{-1} for West Antarctica, and no estimate for the Antarctic Peninsula, but the estimate for East Antarctica was based on only 60% coverage. Using improved data for 1996-2004 that provide estimates for more than 85% of Antarctica (and which were extrapolated on a basin per basin basis to 100% of Antarctica), *Rignot et al. (2008)* found an ice loss of 106±60 Gt a^{-1} for West Antarctica, 28±45 Gt a^{-1} for the Peninsula, and a mass gain of

4±61 Gt a^{-1} for East Antarctica in year 2000. In year 1996, the mass loss for West Antarctica was 83±59 Gt a^{-1}, but the mass loss increased to 132±60 Gt a^{-1} in 2006 due to glacier acceleration. In the Peninsula, the mass loss increased to 60±46 Gt a^{-1} in 2006 due to the massive acceleration of glaciers in the northern Peninsula following the break up of the Larsen B ice shelf in year 2002. Overall, the ice sheet mass loss nearly doubled in ten years, nearly entirely from West Antarctica and the northern tip of the Peninsula, while little changes have been found in East Antarctica. Other mass-budget analyses indicate thickening of drainage basins feeding the Filchner-Ronne ice shelf from portions of East and West Antarctica (*Joughin and Bamber, 2005*) and of some ice streams draining ice from West Antarctica into the Ross Ice Shelf (*Joughin and Tulaczyk, 2002*), but mass loss from the northern part of the Antarctic Peninsula (*Rignot et al., 2005*) and parts of West Antarctica flowing into the Amundsen Sea (*Rignot et al., 2004b*). In both of these latter regions, losses are increasing with time.

Although SRALT coverage extends only to within about 900 km of the poles (Fig. 2.6), inferred rates of surface elevation change (dS/dt) should be more reliable than in Greenland, because most of Antarctica is too cold for surface melting (reducing effects of changing dielectric properties), and outlet glaciers are generally wider than in Greenland (reducing uncertainties associated with rough surface topography). Results show that interior parts of East Antarctica monitored by ERS-1 and ERS-2 thickened during the 1990s, equivalent to growth of a few tens of gigatonnes per year, depending on details of the near-surface density structure (*Davis et al., 2005; Wingham et al., 2006; Zwally et al., 2005*), but *Monaghan et al. (2006)* and *van den Broeke et al. (2006)* show no change in accumulation over a longer time period in this region, suggesting that SRALT may be biased by the large decadal variability in snowfall in Antarctica With ~80% SRALT coverage of the ice sheet, and interpolating to the rest, *Zwally et al. (2005)* estimated a West Antarctic loss of 47±4 Gt a^{-1}, East Antarctic gain of 17±11 Gt a^{-1}, and overall loss of 30±12 Gt a^{-1}, excluding the Antarctic Peninsula, a large fraction of the coastal sectors, and with error estimates neglecting potential uncertainties. *Wingham et al. (2006)* interpret the same data to show that mass gain from snowfall, particularly in the Antarctic Peninsula and East Antarctica, exceeds dynamic losses from West Antarctica. More importantly, however, *Monaghan et al. (2006)* and *van den Broeke et al. (2006)* found very strong decadal variability in Antarctic accumulation, which suggests that it will require decades of data to separate decadal variations from long-term trends in accumulation, for instance, associated with climate warming.

The present ice mass balance of Antarctica and its deglaciation history from the Last Glacial Maximum are still poorly known. It has been shown recently that the uplift rates derived from Global Positioning System (GPS) can be employed to discriminate between different ice loading scenarios. There is general agreement that Antarctica was a major participant in the last glacial age within the West Antarctic Ice Sheet (WAIS), perhaps contributing more than 25 m to rising sea level during the last 21,000 years (*Clark et al., 2002*). The main controversy is whether or not the dominant Antarctic melt contribution to sea level rise is relatively young, perhaps related to Hypsithermal period warming events during the Holocene (10–8 to 6–4 ka), or older, corresponding to the initial collapse phase (21–14 ka) of Northern Hemispheric ice sheets (*Peltier, 1998*). Post-glacial rebound rates are not well constrained and are an error source for ice mass-balance assessment with GRACE satellite data. Analyses of GRACE measurements for 2002-05 show the ice sheet to be very close to balance with a gain of 3±20 Gt a-1 (*Chen et al, 2006*) or net loss sheet ranging from 40±35 Gt a^{-1} (*Ramillien et al., 2006*) to 137±72 Gt a^{-1} (*Velicogna and Wahr, 2006b*), primarily from the West Antarctic Ice Sheet.

Taken together, these various approaches indicate a likely net loss of 100 Gt a^{-1} in the mid 1990s growing to 200 Gt a^{-1} in mid 2000s.

The largest losses are concentrated along the Amundsen and Bellingshausen sectors of West Antarctica, in the northern tip of the Antarctic Peninsula, and to a lesser extent in the Indian Ocean sector of East Antarctica.

A few glaciers in West Antarctic are losing a disproportionate amount of mass. The largest mass loss is from parts of the ice sheet flowing into Pine Island Bay, which represents enough ice to raise sea level by 1.2 m.

In East Antarctica, with the exception of glaciers flowing into the Filchner/Ronne, Amery, and Ross ice shelves, nearly all the major glaciers are thinning, with those draining the Wilkes Land sector losing most mass. Like much of West Antarctica, this sector is grounded well below sea level.

There are insufficient observations to provide reliable estimates of mass balance before 1990, yet there is evidence for long-term loss of mass from glaciers draining the Antarctic Peninsula (*Pritchard and Vaughan, 2007*) and speed up of Pine Island Glacier and neighbors since at least the 1970s (*Joughin and Bohlander, 2003*) In addition, balancing measured sea-level rise since the 1950s against potential causes such as thermal expansion and non-Antarctic ice melting leaves a “missing” source equivalent to many tens of gigatonnes per year.

3.3 Rapid Changes of Small Glaciers

3.3.1 Introduction

Small glaciers are those other than the two ice sheets. Mass balance is a rate of either gain or loss of ice, and so a change in mass balance is an acceleration of the process. Thus we measure mass balance in units such as kg m^{-2} a^{-1} (mass change per unit surface area of the glacier; 1 kg m^{-2} is equivalent to 1 mm depth of liquid water) or, more conveniently at the global scale, Gt a^{-1} (change of total mass, in gigatonnes per year). A change in mass balance is measured in Gt a^{-2}, gigatonnes per year per year: faster and faster loss or gain.

3.3.2 Mass Balance Measurements and Uncertainties

Most measurements of the mass balance of small glaciers are obtained in one of two ways. *Direct* measurements are those in which the change in glacier surface elevation is measured directly at a network of pits and stakes. Calving is treated separately. In *geodetic* measurements, the glacier surface elevation is measured at two times with reference to some fixed external datum. Recent advances in remote sensing promise to increase the contribution from geodetic measurements and to improve spatial coverage, but at present the observational database remains dominated by direct measurements. The primary source for these is the World Glacier Monitoring Service (*WGMS; Haeberli et al., 2005*). *Kaser et al. (2006)* (see also *Lemke et al., 2007*; Sec. 4.5) present compilations which build on the WGMS dataset and extend it significantly.

In Figure 2.3 (see also Table 2.2), the three spatially corrected curves agree rather well, which motivated *Kaser et al. (2006)* to construct their consensus estimate of mass balance, denoted MB. The arithmetic-average curve C05a is the only curve extending before 1961 because measurements are too few at those times for area-weighting or spatial interpolation to be practicable. The early measurements suggest weakly that mass balances were negative. After 1961, we can see with greater confidence that mass balance became less negative until the early 1970s, and that thereafter it has been growing more negative.

The uncorrected C05a, a simple arithmetic average of all the measurements, generally tracks the other curves with fair accuracy. Apparently spatial bias, while not negligible, is of only moderate significance. However the C05a estimate for 2001-04 is starkly discordant. The discordance is due in large part to the European heat wave of 2003 and to under-representation of the high Arctic latitudes, where measurements are few and 2003 balances were only moderately negative. It illustrates the extent to which spatial bias can compromise global estimates. The other curves, C05i, DM05 and O04, each attempt to correct carefully for spatial bias.

Mass-balance measurements at the glacier surface are relatively simple, but difficulties arise with contributions from other parts of the glacier. Internal accumulation is one of the most serious problems. It happens in the lower percolation zones of cold glaciers (those whose internal temperatures are below freezing) when surface meltwater percolates beneath the current year's accumulation of snow. Internal accumulation is impractical to measure and is difficult to model with confidence. It is a plausible conjecture that there are many more cold glaciers than temperate glaciers (in which meltwater can be expected to run off rather than to refreeze).

The calving of icebergs is a significant source of uncertainty. Over a sufficiently long averaging period, adjacent calving and noncalving glaciers ought not to have very different balances, but the time scale of calving is quite different from the annual scale of surface mass balance, and it is difficult to match the two. Tidewater glaciers tend to evolve by slow growth (over centuries) alternating with brief (decades-long) episodes of rapid retreat. Many tidewater glaciers are undergoing such retreat at present, but in general they are under-represented in the list of measured glaciers. The resulting bias, which is known to be opposite to the internal-accumulation bias, must be substantial.

We can draw on geodetic and gravimetric measurements of multidecadal mass balance to reinforce our understanding of calving rates. To illustrate, *Larsen et al. (2007)* estimated the mass balance in southeastern Alaska and adjacent British Columbia as -16.7±4.4 Gt a^{-1}. Earlier, *Arendt et al. (2002)* measured glaciers across Alaska by laser altimetry and estimated an acceleration in mass loss for the entire state from 52±15 Gt a^{-1} (mid-1950s to mid-1990s) to 96±35 Gt a^{-1} (mid-1990s to 2001). These are significantly greater losses than the equivalent direct estimates, and much of the discrepancy must be due to under-representation of calving in the latter. This under-representation is compounded by a lack of basic information. The extent, and even the total terminus length, of glacier ice involved in calving is not known, although a substantial amount of information is available in scattered sources.

Global mass-balance estimates suffer from uncertainty in total glacierized area, and the rate of shrinkage of that area is not known accurately enough to be accounted for. A further problem is delineating the ice sheets so as to avoid double-counting or omitting peripheral ice bodies.

Measured glaciers are a shifting population. Their total number fluctuates, and the list of measured glaciers changes continually. The commonest record length is 1 year; only about 50 are longer than 20 years. These difficulties can be addressed by assuming that each single annual measurement is a random sample. However, the temporal variance of such a short sample is difficult to estimate satisfactorily, especially in the presence of a trend.

On any one glacier, a small number of point measurements must represent the entire glacier. It is usually reasonable to assume that the mass balance depends only on the surface elevation, increasing from net loss at the bottom to net gain above the equilibrium line altitude. A typical uncertainty for elevation-band averages of mass balance is ±200 kilograms per square meter per year (kg m-2 a^{-1}), but measurements at different elevations are highly correlated, meaning that whole-glacier measurements have intrinsic uncertainty comparable with that of elevation-band averages.

At the global scale, the number of measured glaciers is small by comparison with the total number of glaciers. However the mass balance of any one glacier is a good guide to the balance of nearby glaciers. At this scale, the distance to which single-glacier measurements yield useful information is of the order of 600 km. Glacierized regions with few or no measured glaciers within this distance obviously pose a problem. If there are no nearby measurements at all, we can do no better in a statistical sense than to set the regional average equal to the global average, attaching to it a suitably large uncertainty.

3.3.3 Historical and Recent Balance Rates

To extend the short time series of measured mass balance, *Oerlemans et al. (2007)* have tried to calibrate records of terminus fluctuations (i.e., of glacier length) against the direct measurements by a scaling procedure. This allowed them to interpret the terminus fluctuations back to the mid-19th century in mass-balance units. Figure 2.7 shows modeled mass loss since the middle of the 19th century, at which time mass balance was near to zero for perhaps a few decades. Before then, mass balance had been positive for probably a few centuries. This is the signature of the Little Ice Age, for which there is abundant evidence in other forms. The balance implied by the *Oerlemans et al. (2007)* reconstruction is a net loss of about 110 to 150 Gt a^{-1} on average over the past 150 years. This has led to a cumulative rise of sea level by 50-60 mm.

It is not possible to detect mass-balance acceleration with confidence over this time span, but we do see such an acceleration over the shorter period of direct measurements (Fig. 2.7). This signature matches well with the signature seen in records of global average surface air temperature (*Trenberth et al., 2007*). Temperature remained constant or decreased slightly from the 1940s to the 1970s and has been increasing since. In fact, mass balance also responds to forcing on even shorter time scales. For example, there is a detectable small-glacier response to large volcanic eruptions. In short, small glaciers have been evolving as we would expect them to when subjected to a small but growing increase in radiative forcing.

At this point, however, we must recall the complication of calving, recently highlighted by *Meier et al. (2007)*. Small glaciers interact not only with the atmosphere but also with the solid earth beneath them and with the ocean. They are thus subject to additional forcings which are only indirectly climatic. *Meier et al. (2007)* made some allowance for calving when they estimated the global total balance for 2006 as -402±95 Gt a^{-1}, although they cautioned that the true magnitude of loss was probably greater.

"Rapid" is a relative term when applied to the mass balance of small glaciers. For planning purposes we might choose to think that the 1850-2000 average rate of *Oerlemans et al. (2007)* is "not very rapid". After all, human society has grown accustomed to this rate, although it is true that the costs entailed by a consistently non-zero rate have only come to be appreciated quite recently. But a loss of 110 to 150 Gt a^{-1} can be taken as a useful benchmark. It is greater in magnitude than the net loss of 54±82 Gt a^{-1} estimated by *Kaser et al. (2006)* for 1971-75 and significantly less than the *Kaser et al. (2006)* net loss of 354±70 Gt a^{-1} for 2001-04. So in the last three decades the world's small glaciers have moved from losing mass at half the benchmark rate to rates two or three times faster than the benchmark rate. As far as the measurements are able to tell us, this acceleration has been steady.

What can we say about extreme rates in the past? We have to rely on estimated changes of temperature. *Severinghaus et al. (1998)* estimated a mean-annual warming rate at the abrupt (decadal- to century-scale) termination of the Younger Dryas episode, ~11.64 ka, of order 0.1-1.0 Kelvin (K) a^{-1}, while *Denton et al. (2005)* argued that the total summer warming during this event was about 4 K. *Huber et al. (2006)* gave a typical warming rate for the onset of Dansgaard-Oeschger events during the last glacial period of 0.05 K a^{-1}. The small glaciers of the time are unlikely to have had a role in forcing these shifts, but they must have responded to them and probably provided the leading edge of the response.

Figure 2.8 shows accordance between balance and temperature. Each degree of warming yields about another -300 Gt a^{-1} of mass loss beyond the 1961-90 average, -136 Gt a^{-1}. This suggestion is roughly consistent with the current warming rate, about 0.025 K a^{-1}, and balance acceleration, about -10 Gt a^{-2} (Fig. 2.8). The warming rate is not very much less than the extreme rates of the previous paragraph, and it may be permissible to extrapolate (with caution, because we are neglecting the sensitivity of mass balance to change in precipitation and also the sensitivity of dB/dT, the change in mass balance per degree of warming, to change in the extent and climatic distribution of the glaciers). For example, at the end of the Younger Dryas, small glaciers could have contributed at least 1,200 Gt a^{-1} [4 K × (300) Gt a^{-1} K^{-1}] of meltwater if we adopt the summer warming rate of the previous paragraph

Such large rates, if reached, could readily be sustained for at least a few decades during the 21st century. At some point the total shrinkage must begin to impact the rate of loss (we begin to run out of small-glacier ice). Against that certain development must be set the probability that peripheral ice caps would also begin to detach from the ice sheets, thus "replenishing" the inventory of small glaciers. *Meier et al. (2007)*, by extrapolating the current acceleration, estimated a total contribution to sea level of 240±128 mm by 2100, implying a negative balance of 1,500 Gt a^{-1} in that year. These figures assume that the current acceleration of loss continues. Alternatively, if loss continues at the current rate of 400 Gt a^{-1}, the total contribution is 104±25 mm. In contrast *Raper and Braithwaite (2006)*, who allowed for glacier shrinkage, estimated only 97 mm by 2100. Part of the difference is due to their exclusion of small glaciers in Greenland and Antarctica. If included, and if they were assumed to contribute at the same rate as the other glaciers, these would raise the *Raper-Braithwaite (2006)* estimate to 137 mm.

3.4 Causes of Changes

Potential causes of the observed behavior of the ice sheets include changes in snowfall and/or surface melting, long-term responses to past changes in climate, and changes in the dynamics, particularly of outlet glaciers, that affect total ice discharge rates. Recent observations have shown that changes in dynamics can occur far more rapidly than previously suspected, and we discuss causes for these in more detail in Section 4.

3.4.1 Changes in Snowfall and Surface Melting

Recent studies find no continent-wide significant trends in Antarctic accumulation over the interval 1980-2004 (*van den Broeke et al., 2006; Monaghan et al., 2006*), and surface melting has little effect on Antarctic mass balance. Modeling results indicate probable increases in both snowfall and surface melting over Greenland as temperatures increase (*Hanna et al., 2005; Box et al., 2006*). An update of estimated Greenland Ice Sheet runoff and surface mass balance (i.e., snow accumulation minus runoff) results presented in *Hanna et al. (2005)* shows significantly increased runoff losses for 1998-2003 compared with the 1961-90 climatologically "normal" period. But this was partly compensated by increased precipitation over the past few decades, so that the decline in surface mass balance between the two periods was not statistically significant. Data from more recent years, extending to 2007, however, suggest a strong increase in the net loss from the surface mass balance. However, because there is summer melting over ~50% of Greenland already (*Steffen et al., 2004b*), the ice sheet is particularly susceptible to continued warming. Small changes in temperature substantially increasing the zone of summer melting, and, a temperature increase by more than 3ºC would probably result in irreversible loss of the ice sheet (*Gregory et al., 2004*). Moreover, this estimate is based on imbalance between snowfall and melting and would be accelerated by changing glacier dynamics of the type we are already observing.

In addition to the effects of long-term trends in accumulation/ablation rates, mass-balance estimates are also affected by inter-annual variability. This increases uncertainties associated with measuring surface accumulation/ablation rates used for mass-budget calculations, and it results in a lowering/raising of surface elevations measured by altimetry (e.g., *van der Veen, 1993*). *Remy et al. (2002)* estimate the resulting variance in surface elevation to be around 3 m over a 30-year time scale in parts of Antarctica. This clearly has implications for the interpretation of altimeter data.

3.4.2 Ongoing Dynamic Ice Sheet Response to Past Forcing

The vast interior parts of an ice sheet respond only slowly to climate changes, with time scales up to 10,000 years in central East Antarctica. Consequently, current ice-sheet response does includes a component from ongoing adjustment to past climate changes.

Model results [e.g., *Huybrechts (2002)* and *Huybrechts et al. (2004)*] show only a small long-term change in Greenland ice-sheet volume, but Antarctic shrinkage of about 90 Gt a^{-1}, concomitant with the tail end of Holocene grounding-line retreat since the Last Glacial Maximum. This places a lower bound on present-day ice sheet losses.

3.4.3 Dynamic Response to Ice-Shelf Break-Up

Recent rapid changes in marginal regions of both ice sheets include regions of glacier thickening and slowdown but mainly acceleration and thinning, with some glacier velocities increasing more than twofold. Most of these glacier accelerations closely followed reduction or loss of ice shelves. Such behavior was predicted almost 30 years ago by *Mercer (1978)*, but was discounted, as recently as the IPCC Third Assessment Report (*Church et al., 2001*) by most of the glaciological community, based largely on results from prevailing model simulations. Considerable effort is now underway to improve the models, but it is far from complete, leaving us unable to make reliable predictions of ice-sheet responses to a warming climate if such glacier accelerations were to increase in size and frequency. It should be noted that there is also a large uncertainty in current model predictions of the atmosphere and ocean temperature changes which drive the ice-sheet changes, and this uncertainty could be as large as that on the marginal flow response.

Total breakup of Jakobshavn Isbræ ice tongue in Greenland was preceded by its very rapid thinning, probably caused by a massive increase in basal melting rates (*Thomas et al., 2003*). Despite an increased ice supply from accelerating glaciers, thinning of more than 1 m a^{-1}, and locally more than 5 m a^{-1}, was observed between 1992 and 2001 for many small ice shelves in the Amundsen Sea and along the Antarctic Peninsula (*Shepherd et al., 2003; Zwally et al., 2005*). Thinning of ~1 m a^{-1} *(Shepherd et al., 2003*) preceded the fragmentation of almost all (3,300 km^2) of the Larsen B ice shelf along the Antarctic Peninsula in fewer than 5 weeks in early 2002 (*Scambos et al., 2003*), and the correlation between long melt seasons and ice shelf break-up was highlighted by *Fahnestock et al. (2002)*. A southward-progressing loss of ice shelves along the Antarctic Peninsula is consistent with a thermal limit to ice-shelf viability (*Mercer*, 1978; *Morris and Vaughan, 1994*). found that no ice shelves exist on the warmer side of the –5°C mean annual isotherm, whereas no ice shelves on the colder side of the –9°C isotherm have broken up. Before the 2002 breakup of Larsen B ice shelf, local air temperatures increased by more than 1.5°C over the previous 50 years (*Vaughan et al., 2003*), increasing summer melting and formation of large melt ponds on the ice shelf. These may have contributed to breakup by draining into and wedging open surface crevasses that linked to bottom crevasses filled with seawater (*Scambos et al., 2000*).

Most ice shelves are in Antarctica, where they cover an area of ~1.5×10^6 km^2 with nearly all ice streams and outlet glaciers flowing into them. The largest ones in the Weddell and Ross Sea Embayments also occupy the most poleward positions and are currently still far from the viability criteria cited above. By contrast, Greenland ice shelves occupy only a few thousand square kilometers, and many are little more than floating glacier tongues. Ice shelves are nourished by ice flowing from inland and by local snow accumulation, and mass loss is primarily by iceberg calving and basal melting. Melting of up to tens of meters per year has been estimated beneath deeper ice near grounding lines (*Rignot and Jacobs, 2002*). Significant changes in ice-shelf thickness are most readily caused by changes in basal melting or iceberg calving.

Ice-shelf basal melting depends on temperature and ocean circulation within the cavity beneath (*Jenkins and Doake, 1991*). Isolation from direct wind forcing means that the main drivers of below-ice-shelf circulation are tidal and density (thermohaline) forces, but lack of knowledge of bathymetry below the ice has hampered the use of three-dimensional models to simulate circulation beneath the thinning ice shelves as well as a lack of basic data on changes in ocean thermal forcing.

If glacier acceleration caused by thinning ice shelves can be sustained over many centuries, sea level will rise more rapidly than currently estimated. A good example are tidewater glaciers as discussed in Section 3.3.2. But such dynamic responses are poorly understood and, in a warmer climate, the Greenland Ice Sheet margin would quickly retreat from the coast, limiting direct contact between outlet glaciers and the ocean. This would remove a likely trigger for the recently detected marginal acceleration. Nevertheless, although the role of outlet-glacier acceleration in the longer term (multidecade) evolution of the ice sheet is hard to assess from current observations, it remains a distinct possibility that parts of the Greenland Ice Sheet may already be very close to their threshold of viability.

3.4.4 Increased Basal Lubrication

Observations on some glaciers show seasonal variations in ice velocity, with marked increases soon after periods of heavy surface melting (e.g., *O'Neel et al., 2001*). Similar results have also been found on parts of the Greenland ice sheet, where ice is moving at ~100 m a-1 (*Zwally et al., 2002b*). A possible cause is rapid meltwater drainage to the glacier bed, where it enhances lubrication of basal sliding. If so, there is a potential for increased melting in a warmer climate to cause an almost simultaneous increase in ice-discharge rates. However, there is little evidence for seasonal changes in the speeds of the rapid glaciers that discharge most Greenland ice. In northwest, northeast, southeast, and central west Greenland, *Rignot and Kanagaratnam (2006)* found a 8-10% increase in monthly velocity over the summer months compared to the winter months, so that abundance of meltwater in the summer is not providing a significant variation in ice discharge compared to the yearly average. However, this does not mean that a doubling of the meltwater production could only drive a 16-20% increase in speed. Meltwater remains an essential control on glacier flow as many studies of mountain glaciers have shown for many decades, so it is quite likely that an increase in meltwater production from a warmer climate could likely have major consequences on the flow rates of glaciers.

4. Potential Mechanism of Rapid Ice Response

4.1 Ocean-Ice Interactions

The interaction of warm waters of the global ocean with the periphery of the large ice sheets represents one of the most significant possibilities for abrupt change in the climate system. Ocean waters provide a source of energy that can drive high melt rates beneath ice shelves and at tidewater glaciers. Calving of icebergs at glacier termini is an additional mechanism of ice loss and has the capacity to destabilize an ice front. Mass loss through oceanic melting and iceberg calving accounts for more than 95% of the ablation from Antarctica and 40-50% of the ablation from Greenland. As described in the previous section, we have seen evidence over the last decade or so, largely gleaned from satellite and airborne sensors, that the most evident changes in the ice sheets have been occurring at their periphery. Some of the changes, for example in the area of the Pine Island Glacier, Antarctica, have been attributed to the effect of warming ocean waters at the margin of the ice sheet (*Payne et al., 2004*). There does not yet exist, however, an adequate observational database against which to definitively correlate ice shelf thinning or collapse with warming of the surrounding ocean waters.

4.1.1 Ocean Circulation

To understand how changes in ocean temperature can impact ice shelves and tidewater glaciers, it is necessary first to understand properties of the global ocean circulation. The polar oceans receive warm salty water originating in the nonpolar oceans. In the North Atlantic Ocean, the northward flowing extension of the Gulf Stream ultimately arrives in the vicinity of the Greenland Ice Sheet, at depth. In the Southern Ocean, the southward extension of the North Atlantic Deep Waters ultimately arrive in the vicinity of the Antarctic Ice Sheet, again at depth. The polar oceans themselves produce cold, fresh water, and salty waters are denser than the cold, fresh waters. The result is that the warm, salty waters are found at depths of several hundred meters in the polar oceans, having subducted beneath the cold, fresh surface polar waters.

Despite the potential of the warm, deep waters to impact the basal melting of ice shelves, little observational progress has been made in studying these waters, nor is there any information on the pre-instrumental (geologic) record of these waters. The main obstacle to progress has been that no sustained observation program can provide a regional and temporal view of the behavior of these deep waters. Instead, for the most part, we have only scattered ship-based observations, poorly sampled in time and space of the locations and temperatures of the deep waters. Limited observations have established that warm, deep waters are present near some Antarctic ice shelves (e.g., Pine Island Glacier, *Jacobs et al., 1996*) and not near others (e.g., Ross Ice Shelf, *Jacobs and Giulivi, 1998*). Greenland's ice shelves follow similarly with some having warm, deep waters present (e.g., Jakobshavn Isbræ, *Holland et al., 2007a*) and others much less so (e.g., Petermann Gletscher, *Steffen et al., 2004a*).

The nature of the circulation of ocean waters beneath an ice shelf can be broadly classified into two regimes. In one regime, only cold ocean waters (i.e., near the freezing point) are found in front of and beneath an ice shelf. These waters produce little melting of the ice shelf base, as for instance, the base of the Ross Ice Shelf, which is estimated to melt at about 0.2 m a^{-1} (*Holland et al., 2003*). In a second regime, warm waters (i.e., a few degrees above the freezing point) are found in front of and beneath the ice shelf. Here, the melt rate can be one-hundred-fold stronger, up to 20 m a^{-1}, as for example at the base of the Pine Island Glacier (*Jacobs et al., 1996*). This nonlinear sensitivity of basal mass balance to ocean temperature has recently been highlighted (*Holland et al., 2007b*), as well as the sensitivity of melt rate to the geometry of the environment. The presence of warm water in the vicinity of an ice shelf is a necessary condition for high melting, but it is not sufficient by itself. Additional factors such as the details of the bathymetry can be equally important, as for example, a submarine sill can block access of warm waters while a submarine canyon can facilitate the exchange of warm, deep waters into a cavity beneath an ice shelf. Recent years have seen an increase in the collection of bathymetric data around the Greenland and Antarctic continental shelves, and in some instances even beneath the ice shelves.

4.1.2 Ice-Pump Circulation

The manner in which ocean waters circulate beneath an ice shelf has loosely become known as the 'ice-pump' circulation (*Lewis and Perkins, 1986*). The circulation can be visualized as dense, salty water (either cold or warm), entering an ice shelf cavity and flowing toward the back of the cavity, to the grounding line where the ice shelf first goes afloat on the ocean. Here at the grounding line, the ice shelf is at its greatest thickness. Because the freezing point of seawater decreases as ocean depth increases, the invading ocean waters have an ever increasing thermal head with respect to the ice as the depth of the ice increases. The thermal head determines the amount of melting at the grounding line. An end result of melting is a cooled and freshened ocean water mass at the grounding line. An empirical consequence of the equation of state for seawater is that this water mass will always be less dense than the source waters that originally fed into the ice shelf cavity. These light waters subsequently flow upward along the ice shelf base as a kind of upside-down gravity current, a flow feature termed a plume. As the waters rise, the depth-dependent freezing point also rises, and at some point the rising waters can actually become supercooled with respect to the local freezing point. In this instance some of the meltwaters refreeze to the base of the ice shelf, forming so-called marine ice, in contrast to the meteoric ice (also called snow/ice) that feeds the ice shelf from the inland ice sheet. It is the manner in which ocean waters can melt the deep ice and refreeze ice at shallow depths that has given rise to the term 'ice pump'. In the case of warm waters in the cavity beneath the ice shelf, the term ice pump is a misnomer, as there may be no refreezing of ice whatsoever, just melting. These under-ice circulation processes are clearly important to the stability of ice shelves or ice tongues, but it is difficult to yet predict their impact on Antarctica and Greenland in the coming decades. Future changes in ocean circulation and ocean temperatures will produce changes in basal melting, but the magnitude of these changes is currently not modeled or predicted.

4.2 Ice Shelf Processes

4.2.1 Ice Shelf Basal Melting

A nonlinear response of ice shelf melting to increasing ocean temperatures is a central tenet in the scenario for abrupt climate change arising from ocean–ice-shelf interaction. The nonlinear response is a theoretical and computational result; observations are yet inadequate to verify this conclusion. Nonetheless, the basis of this result is that the melt rate at the base of an ice shelf is the product of the thermal head and the velocity of the ocean waters at the base. The greater the thermal head or the velocity, then the greater the melt rate. A key insight from the theoretical and modeling research is that as the ocean water temperature is increased, the buoyancy of the plume beneath the ice shelf is increased because greater melting is initiated by the warmer waters. A more buoyant plume rises faster, and causes greater melting, and becomes more buoyant. This positive feedback is a key nonlinear response mechanism of an ice shelf base to warming ocean waters.

The susceptibility of ice shelves to high melt rates and to collapse is a function of the presence of warm waters entering the ice shelf cavities. But the appearance of such warm waters does not actually imply that the global ocean needs to warm. It is true that observational evidence (*Levitus et al., 2000*) does indicate that global ocean has warmed over the past decades, and that the warming has been modest (approximately 0.5° C globally). While this is one mechanism for creating warmer waters to enter a cavity beneath the ice shelf, a more efficient mechanism for melting is not to warm the global ocean waters but to redirect existing warm water from the global ocean toward ice shelf cavities. Ocean circulation is driven by density contrasts of water masses and by surface wind forcing. Subtle changes in surface wind forcing (*Toggweiler and Samuels, 1995*) may have important consequence for the redistribution of warm water currents in polar oceans A change in wind patterns (i.e., a relatively fast process) could produce large and fast changes in the temperatures of ocean waters appearing at the doorstep of the ice shelves.

4.2.2 Ice Shelf Thinning

Changes in the geometry of ice shelves or floating ice tongues can cause a dynamic response that penetrates hundreds of kilometers inland. This can be triggered through high rates of basal melt or through a calving episode, providing the perturbation impacts the ice sheet grounding zone (*Thomas et al., 2005; Payne et al., 2004; Pattyn et al., 2006*). Grounding-zone thinning can induce rapid and widespread inland ice response if fast-flowing ice streams are present. This has been observed in the Pine Island and Thwaites Glacier systems (*Rignot et al., 2002; Shepherd et al., 2002*). Glacier discharge also increased on the Antarctic Peninsula following the 2002 collapse of the Larsen B ice shelf *(Rott et al., 2002; DeAngelis and Skvarca, 2003; Rignot et al., 2004a*).

Whether or not a glacier will stabilize following a perturbation depends to a large degree on whether it is grounded or floating. Flow rates of more than 300 tidewater glaciers on the Antarctic Peninsula increased by an average of 12% from 1992 to 2005 (*Pritchard and Vaughan, 2007*). Pritchard and Vaughan interpret this as a dynamic response to thinning at the ice terminus. Glaciers in contact with the ocean are likely to see an ongoing response to ice-shelf removal.

A thinning ice shelf results in glacier ungrounding, which is the main cause of the glacier acceleration because it has a large effect on the force balance near the ice front (*Thomas,* *2004*). This effect also explains the retreat of Pine Island Glacier (*Thomas et al., 2005*) and the recent acceleration and retreat of outlet glaciers in east Greenland.

4.2.3 Iceberg Calving

Calving is the separation of ice blocks from a glacier at a marginal cliff. This happens mostly at ice margins in large water bodies (lakes or the ocean), and the calved blocks become icebergs. The mechanism responsible for iceberg production is the initiation and propagation of fractures through the ice thickness. Calving can originate in fractures far back from the ice front (*Fricker et al., 2005*). This process is incompletely understood, partly because of the difficulty and danger of making observations.

While it is not clear that calving is a deterministic process (because outcome cannot be predicted exactly from knowledge of initial condition), some internal (ice dynamical) and external influences on calving rates have been qualitatively elucidated. Internal dynamical controls are related to the stiffness and thickness of ice, longitudinal strain rates, and the propensity for fractures to form and propagate. High rates of ice flow promote longitudinal stretching and tensile failure. External influences on calving rates include ocean bathymetry and sea level, water temperature, tidal amplitude, air temperature, sea ice, and storm swell.

These variables may have a role in a general “calving law” that can be used to predict calving rates. Such a law does not yet exist but is important because calving has the capacity to destabilize an ice front. Acceleration of Jakobshavn Isbræ beginning in 2000 has been interpreted as a response to increased calving at the ice front and collapse of the floating tongue following very rapid thinning (*Thomas, 2004; Joughin et al., 2004*).

The external variables that trigger such an event are not well understood. Increased surface melting due to climatic warming can destabilize the ice front and lead to rapid disintegration of an entire ice shelf (*Scambos et al., 2004*). Penetration of surface meltwater into crevasses deepens the fissures and creates areas of weakness that can fail under longitudinal extension.

A number of small ice shelves on the Antarctic Peninsula collapsed in the last three decades of the 20th century. Ice shelf area declined by more than 13,500 km^2 in this period, punctuated by the collapse of the Larsen A and Larsen B ice shelves in 1995 and 2002 (*Scambos et al., 2004*). This was possibly related to atmospheric warming in the region, estimated to be about 3°C over the second half of the 20th century. *Vaughan and Doake (1996)* suggest that ice shelf viability is compromised if mean annual air temperature exceeds −5°C. Above this temperature, meltwater production weakens surface crevasses and rifts and may allow them to propagate through the ice thickness. It is also likely that thinning of an ice shelf, caused by increased basal melting, preconditions it for breakup. Consequently, warming of ocean waters may also be important. The Weddell Sea warmed in the last part of the 20th century, and the role that this ocean warming played in the ice shelf collapses on the Antarctic Peninsula is unknown. Warmer ocean temperatures cause an increase in basal melt rates and ice-shelf thinning. If this triggers enhanced extensional flow, it might cause increased crevassing, fracture propagation, and calving.

Similarly, the impacts of sea ice and iceberg-clogged fjords are not well understood. These could damp tidal forcing and flexure of floating ice tongues, suppressing calving. *Reeh et al. (1999)* discuss the transition from tidewater outlets with high calving rates in southern Greenland to extended, floating tongues of ice in north Greenland, with limited calving flux and basal melting representing the dominant ablation mechanism. Permanent sea ice in northeast Greenland may be one of the factors enabling the survival of floating ice tongues in the north (*Higgins, 1991*). This is difficult to separate from the effects of colder air and ocean temperatures.

4.3 Ice Stream and Glacier Processes

Ice masses that are warm based (at the melting point at the bed) can move via basal sliding or through deformation of subglacial sediments. Sliding at the bed involves decoupling of the ice and the underlying till or bedrock, generally as a result of high basal water pressures (*Bindschadler, 1983*). Glacier movement via sediment deformation involves viscous flow or plastic failure of a thin layer of sediments underlying the ice (*Kamb, 1991; Tulaczyk et al., 2001*). Pervasive sediment deformation requires large supplies of basal meltwater to dilate and weaken sediments. Sliding and sediment deformation are therefore subject to similar controls; both require warm-based conditions and high basal water pressures, and both processes are promoted by the low basal friction associated with subglacial sediments. In the absence of direct measurements of the prevailing flow mechanism at the bed, basal sliding and subglacial sediment deformation can be broadly combined and referred to as *basal flow*.

4.3.1 Basal Flow

Basal flow can transport ice at velocities exceeding rates of internal deformation: 100s to more than 10,000 meters per year, and glacier surges, tidewater glacier flow, and ice stream motion are governed by basal flow dynamics (*Clarke, 1987*). Ice streams are responsible for drainage of as much as 90% of West Antarctica (*Paterson, 1994*), leading to a low surface profile and a mobile, active ice mass that is poorly represented by ice sheet models that cannot portray these features.

Glaciers and ice sheets that are susceptible to basal flow can move quickly and erratically, making them intrinsically less predictable than those governed by internal deformation. They are more sensitive to climate change because of their high rates of ice turnover, which gives them a shorter response time to climate (or ice-marginal) perturbations. In addition, they may be directly responsive to increased amounts of surface meltwater production associated with climate warming.

This latter process is crucial to predicting dynamic feedbacks to the expanding ablation area, longer melt season, and higher rates of surface meltwater production that are predicted for most ice masses.

Although basal meltwater has traditionally been thought to be the primary source of subglacial water, models have shown that supraglacial streams with discharges of over 0.15 $m^3 s^{-1}$ can penetrate down through 300 m of ice to reach bedrock, via self-propagation of water-filled crevasses (*Arnold and Sharp, 2002*). There are several possible subglacial hydrological configurations: ice-walled conduits, bedrock conduits, water film, linked cavities, soft-sediment channels, porous sediment sheets, and ordinary aquifers (*Mair et al., 2001; Flowers and Clarke, 2002*).

Modern interest in water flow through glaciers can be dated from a pair of theoretical papers published in 1972. In one of these, *Shreve (1972)* discussed the influence of ice pressure on the direction of water flow through and under glaciers, and in the other *Röthlisberger (1972)* presented a theoretical model for calculating water pressures in subglacial conduits. Through a combination of these theoretical considerations and field observations, it is concluded that the englacial drainage system probably consists of an arborescent network of passages. The millimeter-sized finger-tip tributaries of this network join downward into ever larger conduits. Locally, moulins provide large direct connections between the glacier surface and the bed. Beneath a valley glacier the subglacial drainage is likely to be in a tortuous system of linked cavities transected by a few relatively large and comparatively straight conduits. The average flow direction in the combined system is controlled by a combination of ice-overburden pressure and bed topography, and in general is not normal to contours of equal elevation on the bed. Although theoretical studies usually assume that subglacial conduits are semicircular in cross section, there are reasons for believing that this ideal is rarely realized in nature. Much of the progress in subglacial hydrology has been theoretical, as experimental techniques for studying the englacial hydraulic system are few, and as yet not fully exploited, and observational evidence is difficult to obtain.

How directly and permanently do these effects influence ice dynamics? It is not clear at this time. This process is well known in valley glaciers, where surface meltwater that reaches the bed in the summer melt season induces seasonal or episodic speedups (*Iken and Bindschadler, 1986*). Speedups have also been observed in response to large rainfall events (e.g., *O'Neel et al., 2005*).

4.3.2 Flow Acceleration and Meltwater

Summer acceleration has also been observed in the ablation area of polar icefields (*Copland et al., 2003*), where meltwater ponds drain through moulins and reach the bed through up to 200 m of cold ice (*Boon and Sharp, 2003*). The influx of surface meltwater triggers a fourfold speedup in flow in the lower ablation area each year. There is a clear link between the surface hydrology, seasonal development of englacial drainage connections to the bed, and basal flow, at least at this site.

It is uncertain whether surface meltwater can reach the bed through thick columns of cold ice. Cold ice is impermeable on the intergranular scale (*Paterson, 1994*). However, water flowing into moulins may carry enough kinetic and potential energy to penetrate to the bed and spread out over an area large enough to affect the basal velocity. *Zwally et al. (2002a)* record summertime speedup events near the western margin of the Greenland Ice Sheet, associated with the drainage of large supraglacial lakes in a region where the ice sheet is several hundred meters thick. It is unknown whether the meltwater penetrated all the way to the bed, but this is interpreted to be the cause of the summer speedups and is consistent with observations on valley glaciers.

These observations are unequivocal but the speedups are modest (10%) and localized. Alternative interpretations of the *Zwally et al. (2002a)* data have also been proposed. The region may be influenced by seasonal acceleration at the downstream ice margin or through accelerated summer flow in nearby Jakobshavn Isbræ, rather than local supraglacial lake drainage. Recent summer speedups in Jakobshavn Isbræ are believed to be a response to marine conditions (summer calving, seasonal sea ice, and basal melting on the floating ice tongue).

More studies like that of *Zwally et al. (2002a)* are needed to determine the extent to which supraglacial water actually reaches the bed and influences basal motion. At this time it is still unclear how influential surface meltwater is on polar icefield dynamics, but it may prove to be an extremely important feedback in icefield response to climate change, as it provides a direct link between surface climate and ice dynamics. A modeling study by *Parizek and Alley (2004)* that assumes surface-meltwater-induced speedups similar to those observed by *Zwally et al. (2002a)* found this effect to increase the sensitivity of the Greenland Ice Sheet to specified warmings by 10-15%. This is speculative, as the actual physics of meltwater penetration to the bed and its influence on basal flow are not explicitly modeled or fully understood.

4.4 Modeling

4.4.1 Ice-Ocean Modeling

There has been substantial progress in the numerical modeling of the ice-shelf–ocean interaction over the last decade. A variety of ocean models have now been adapted so that they can simulate the interaction of the ocean with an overlying ice shelf (see *ISOMIP Group, 2007,* for summary of modeling activities). The present state of the art in these simulations is termed as static-geometry simulations, as the actual shape of the ice-shelf cavity does not change during these simulations. Such static geometry simulations are a reasonable first step in advancing understanding of such a complex system. Steps are now being taken to co-evolve the ocean and ice shelf (*Grosfeld and Sandhager, 2004; Walker and Holland, 2007*) in what can be termed as dynamic-geometry simulations. It is only the latter type of simulations that can ultimately provide any predictive capability on abrupt change in global sea level as resulting from changing ocean temperatures in cavities beneath the ice shelf. The scientific community presently does not possess an adequate observational or theoretical understating of this problem. Progress is being made, but given the relatively few researchers and resources tackling the problem, the rate of progress is slow. It is conceivable that changes are presently occurring or will occur in the near term (i.e., the present century) in the ice-shelf–ocean interaction that we are not able to observe or model.

4.4.2 Ice Modeling

The extent of impact of ice-marginal perturbations depends on the nature of ice flow in the inland ice. Ice dynamics in the transition zone between inland and floating ice – the grounding zone – are complex, and few whole-ice-sheet models have rigorously addressed the mechanics of ice flow in this zone. *MacAyeal (1989)* introduced a model of ice shelf-ice stream flow that provides a reasonable representation of this transition zone, although the model has only been applied on regional scales. This model, which has had good success in simulating Antarctic ice-stream dynamics, assumes that ice flux is dominated by flow at the bed and longitudinal stretching, with negligible vertical shear deformation in the ice.

The Western half of Antarctica contains enough ice to raise sea levels by about 6 m. It also rests on bedrock below sea level, which leaves it vulnerable to irreversible shrinkage if the rate of ice flow from the grounded ice sheet into the surrounding ice shelves were to increase, causing partial flotation and hence retreat of the grounded ice sheet. A hotly debated hypothesis in glaciology asserts that a marine ice sheet is susceptible to such irreversible shrinkage if its grounding line rests on an upward-sloping bed, because a small retreat in grounding line position should lead to increased discharge, which leads to further retreat and so on. The key to this hypothetical positive feedback is that discharge through the grounding line - where grounded ice lifts off the bed to become an ice shelf - must increase with water depth there. The assertion that this is the case has been around for over 30 years but has not previously been proven. *Schoof (2007)* has been able to use the boundary layer theory to show that the positive feedback does indeed exist.

Recent efforts have explored higher order simulations of ice sheet dynamics, including a full-stress solution that allows modeling of mixed flow regimes *(Pattyn, 2002; Payne et al., 2004*). The study by *Payne et al. (2004)* examines the inland propagation of grounding-line perturbations in the Pine Island Glacier. The dynamic response has two different time scales: an instantaneous mechanical response through longitudinal stress coupling, felt up to 100 km inland, followed by an advective-diffusive thinning wave propagating upstream on a decadal time scale, with a new equilibrium reached after about 150 years. These modeling results are consistent with observations of recent ice thinning in this region.

Full-stress solutions have yet to be deployed on continental scales (or applied to the sea-level question), but this is becoming computationally tractable. Improvements may also be possible through nested modeling, with high-resolution grids and high-order physics in regions of interest. Moving-grid techniques for explicit modeling of the ice sheet - ice shelf grounding zone are also needed *(Vieli and Payne, 2005*). The current suite of models does not handle this well. Most regional-scale models that focus on ice shelf dynamics use fixed grounding lines, while continental-scale ice sheet models distinguish between grounded and floating ice, but the grounding zone falls into the horizontal grid cell where this transition occurs. At model resolutions of 10s of kilometers, this does not capture the details of grounding line migration. *Vieli and Payne (2005)* show that this has a large effect on modeled ground-line stability to external forcing.

Observations from the last decade have radically altered the thinking on how rapidly an ice sheet can respond to perturbations at the marine margin. Several-fold increases in discharge followed the collapse of ice shelves on the Antarctic Peninsula, with accelerations of up to 800% following collapse of the Larsen B ice shelf (*Scambos et al., 2004; Rignot et al., 2004a*). The effects on inland ice flow are rapid, large, and propagate immediately over very large distances. This is something models did not predict *a priori*, and the modeling community is now scrambling to catch up with the observations. No whole-ice sheet model is presently capable of capturing the glacier speedups in Antarctica or Greenland that have been observed over the last decade. This means that we have no real idea of how quickly or widely the ice sheets will react if they are pushed out of equilibrium.

4.5 Sea-Level Feedback

Perhaps the primary factor that raises concerns about the potential of abrupt changes in sea level is that large areas of modern ice sheets are currently grounded below sea level (i.e., the base of the ice sheet occurs below sea level) (Fig. 2.10). Where it exists, it is this condition that lends itself to many of the processes described in previous sections that can lead to rapid ice-sheet changes, especially with regard to atmosphere-ocean-ice interactions that may affect ice shelves and calving fronts of tidewater glaciers.

An equally important aspect of these marine-based ice sheets which has long been of interest is that the beds of ice sheets grounded below sea level tend to deepen inland, either due to overdeepening from glacial erosion or isostatic adjustment. The grounding line is the critical juncture that separates ice that is thick enough to remain grounded from either an ice shelf or a calving front. In the absence of stabilizing factors, this configuration indicated that marine ice sheets are inherently unstable, whereby small changes in climate could trigger irreversible retreat of the grounding line (*Hughes, 1973; Weertman, 1974; Thomas and Bentley, 1978*). For a tidewater glacier, rapid retreat occurs because calving rates increase with water depth (*Brown et al., 1983*). Where the grounding line is fronted by an unconfined ice shelf, rapid retreat occurs because the extensional thinning rate of an ice shelf increases with thickness, such as would accompany grounding-line retreat (*Weertman, 1974*).

The amount of retreat clearly depends on how far inland glaciers remain below sea level. Of greatest concern is West Antarctica, where all the large ice streams are grounded well below sea level, with deeper trenches lying well inland of their grounding lines (Fig. 2.10). A similar situation applies to the entire Wilkes Land sector of East Antarctica. In Greenland, few outlet glaciers remain below sea level very far inland, indicating that glacier retreat by this process will eventually slow down or halt. A notable exception may be Greenland's largest outlet glacier, Jakobshavn Isbræ, which appears to tap into the central core of Greenland that is below sea level (Fig. 2.10).

Several factors determine the position of the grounding line, and thus the stability of marine ice sheets. On time scales that may lead to rapid changes, the two most important of these are the backstress provided by ice-shelf buttressing and sea level (*Thomas and Bentley, 1978*). Given that a grounding line represents the point at which ice becomes buoyant, then a rise in sea level will cause grounding line retreat (and vice versa). Following some initial perturbation, this situation thus leads to the potential for a positive feedback to develop between ice retreat and sea level rise. Recent studies from West Antarctica, however, suggest that for some geological situations, the sensitivity of grounding line retreat to sea level rise may be less important than previously considered. *Anandakrishnan et al. (2007)* documented formation of a wedge of subglacial sediment at the grounding line of the Whillans Ice Stream, resulting in ice to be substantially thicker there than floating ice in hydrostatic equilibrium. *Alley et al. (2007)* showed with numerical ice-flow models that a grounding line sitting on a sedimentary wedge is immune to sea-level changes of up to 10 m. Because the wedges develop by accumulation of debris delivered to the grounding line from a subglacial deforming sediment layer, this stabilizing mechanism only applies to those places where such a process is operating. Today, this likely applies to the Siple Coast ice streams and perhaps those flowing into the Ronne Ice Shelf. It is not clear, however, that it applies to ice streams flowing into other Antarctic ice shelves or to the outlet glaciers draining Greenland.

Of these two factors, the buttressing force of the ice shelf is likely more important than sea level in affecting grounding-line dynamics. If this force is greater than that just caused by seawater pressure, then the grounding line is vulnerable to ice-shelf changes. For thick grounding lines, such as characterize most outlet glaciers and ice streams draining Greenland and Antarctica today, this vulnerability far exceeds that associated with feasible sea-level changes expected by the end of this century (0.5-1.0 m) (*Rahmstorf, 2007*), particularly in the context of the likelihood of substantial climate change that would affect the ice shelves in the same timeframe. In considering the wedge-stability factor as well, we thus conclude that, in the absence of rapid loss of ice shelves and attendant sea level rise, sea level forcing and feedback are unlikely to be significant determinants in causing rapid ice-sheet changes in the coming century.

Box 2.1—Glaciers: Some Definitions

Glaciers are bodies of ice resting on the Earth's solid surface (Box 2.1 Fig. 1). We distinguish between *ice sheets* (Box 2.1 Fig. 2), which are glaciers of near-continental extent and of which there are at present two, the Antarctic Ice Sheet and the Greenland Ice Sheet, and *small glaciers*, sometimes also referred to as *glaciers and ice caps* (Box 2.1 Fig. 2). There are several hundred thousand small glaciers. They are typically a few hundred meters to a few tens of kilometers long, while the ice sheets are drained by ice streams many tens to hundreds of kilometers long. In terms of volume, the ice sheets dwarf the small glaciers. If they all melted, the equivalent sea level rise would be 57 m from Antarctica and 7 m from Greenland but only 0.5 m from the small glaciers. Of the Antarctic total, about 7 m would come from West Antarctica, which may be especially vulnerable to abrupt changes.

Ice at the Earth's surface is a soft solid because it is either at or not far below its melting point. It therefore deforms readily under stress, spreading under its own weight until a balance is achieved between mass gains, mainly as snowfall, in the cold interior or upper parts of the glacier, and mass loss in the lower parts by melting or right at sea level by the calving of icebergs. The glacier may, however, keep spreading when it reaches sea level, and in this case it has a floating tongue or, when several glaciers are involved, a buttressing *ice shelf* (Box 2.1 Fig. 3), the weight of which is supported not by the solid earth but by the ocean. A glacier which reaches sea level is called a *tidewater glacier*.

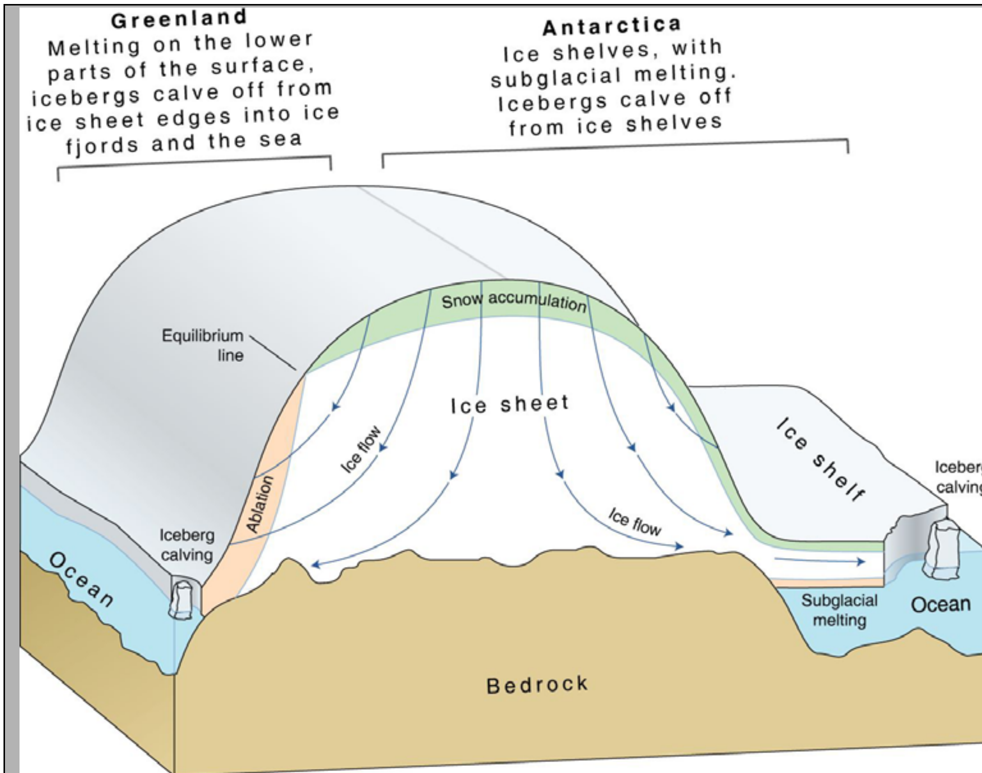
Ice shelves, which are mostly confined to Antarctica, are typically a few hundred meters thick and must not be confused with sea ice, typically a few meters thick. They are a critical part of the picture because they can lose mass not just by melting at their surfaces and by calving but also by melting at their bases. Increased basal melting, due for example to the arrival of warmer seawater, can “pull” more ice across the grounding line.

The *grounding line* separates the grounded inland ice from the floating shelf or tongue ice. It is also where the ice makes its contribution to sea level change. When it begins to float, it displaces seawater whether or not it becomes an iceberg.

There is another crucial role for ice shelves, for they appear to be thermally unstable – there are no ice shelves where the annual average temperature is higher than about minus 5°C. Recently several “warm” ice shelves have collapsed dramatically, and their disintegration has been followed by equally dramatic acceleration of tributary glaciers across what was once the grounding line, where the grounded ice calves directly into the ocean at a far greater rate than before ice-shelf breakup.

Ice streams are rapid flows of ice with walls of slower ice, and are the principal means by which ice is evacuated from the interiors of the ice sheets and supplied to the larger ice shelves. Similar flows with walls of rock are called *outlet glaciers*, although this term is sometimes used quite loosely.

Box 2.1 Figure 1. Glaciers are slow-moving rivers of ice, formed from compacted layers of snow, that slowly deforms and flows in response to gravity. Glacier ice is the largest reservoir of freshwater, and second only to oceans the largest reservoir of total water. Glaciers cover vast areas of polar regions and are restricted to the mountains in mid latitudes. Glaciers are typically a few hundred meters to a few tens of kilometers long; most of the glaciers in mid latitudes have been retreating in the last two centuries (Rhône Glacier, Switzerland, photograph courtesy of K. Steffen, CIRES, University of Colorado at Boulder.)



Box 2.1 Figure 2. The ice cover in Greenland and Antarctica has two components – thick, grounded, inland ice that rests on a more or less solid bed, and thinner floating ice shelves and glacier tongues. An ice sheet is actually a giant glacier, and like most glaciers it is nourished by the continual accumulation of snow on its surface. As successive layers of snow build up, the layers beneath are gradually compressed into solid ice. Snow input is balanced by glacial outflow, so the height of the ice sheet stays approximately constant through time. The ice is driven by gravity to slide and to flow downhill from the highest points of the interior to the coast. There it either melts or is carried away as icebergs which also eventually melt, thus returning the water to the ocean whence it came. Outflow from the inland ice is organized into a series of drainage basins separated by ice divides that concentrate the flow of ice into either narrow mountain-bounded outlet glaciers or fast-moving ice streams surrounded by slow-moving ice rather than rock walls. In Antarctica, much of this flowing ice has reached the coast and has spread over the surface of the ocean to form ice shelves that are floating on the sea but are attached to ice on land. There are ice shelves along more than half of Antarctica's coast, but very few in Greenland (UNEP Maps and Graphs; K. Steffen, CIRES, University of Colorado at Boulder.).

Box 2.1 Figure 3. An ice shelf is a thick, floating platform of ice that forms where a glacier or ice sheet flows down to a coastline and onto the ocean surface. Ice shelves are found in Antarctica, Greenland, and Canada. The boundary between the floating ice shelf and the grounded (resting on bedrock) ice that feeds it is called the grounding line. The thickness of modern-day ice shelves ranges from about 100 to 1,000 meters. The density contrast between solid ice and liquid water means that only about 1/9 of the floating ice is above the ocean surface. The picture shows the ice shelf of Petermann Glacier in northwestern Greenland (right side of picture) with a floating ice tongue of 60 km in length and 20 km wide. Glaciers from the left are merging with the ice shelf. (Petermann Glacier, northwest Greenland, photograph courtesy of K. Steffen, CIRES, University of Colorado at Boulder.)

Box 2.2—Mass Balance, Energy Balance, and Force Balance

The glaciological analyses which we summarize here can all be understood in terms of simple arithmetic.

To determine the mass balance, we add up all the gains of mass, collectively known as accumulation and dominated by snowfall, and all the losses, collectively known as ablation and dominated by melting and calving. The difference between accumulation and ablation is called, by long-established custom, the total mass balance, although the reader will note that we really mean "mass imbalance." That is, there is no reason why the difference should be zero; the same is true of the energy balance and force balance.

The mass balance is closely connected to the energy balance. The temperature of the glacier surface is determined by this balance, which is the sum of gains by the absorption of radiative energy, transfer of heat from the overlying air, and heat released by condensation, and losses by radiative emission, upward transfer of heat when the air is colder than the glacier surface, and heat consumed by evaporation. A negative energy balance means that the ice temperature will drop. A positive energy balance means either that the ice temperature will rise or that the ice will melt.

Ice deformation or dynamics is the result of a balance of forces, which we determine by arithmetic operations comparable to those involved in the mass and energy balances. Shear forces, proportional to the product of ice thickness and surface slope, determine how fast the glacier moves over its bed by shear deformation where the ice is frozen to the bed, or by basal sliding where the bed is wet. Spreading forces, determined by ice thickness, are resisted by drag forces at the glacier bed and its margins, and by forces transmitted upstream from its floating tongue or ice shelf as this pushes seaward past its margins and over locally shoaling seabed. The sum of these forces determines the speed at which the ice moves, together with its direction. However, we must also allow for ice stiffness, which is strongly affected by its temperature, with cold ice much stiffer (more sluggish) than ice near its melting point.

The temperature becomes still more important when we consider basal drag, which is high for a dry-based glacier (one frozen to its bed), but can be very small for wet-based glaciers where their beds have been raised to the melting point by heat conducted from the Earth's interior and frictional heat generated on the spot. Once the bed is at the melting point, any further gain of heat yields meltwater. One of glaciology's bigger surprises is that large parts of the ice sheets, whose surfaces are among the coldest places on Earth, are wet-based.

The varying pressure of basal meltwater on the moving ice can alter the force balance markedly. Its general impact is to promote basal sliding, by which mechanism the glacier may flow much more rapidly than it would by shear deformation alone. Basal sliding, in conjunction with the presence of a porous reservoir for meltwater where the bed consists of soft sediment rather than rock, plays a major role in the behavior of ice streams.

There are subtle links between the mass balance and the force balance. The ice flows from where there is net accumulation to where there is net ablation, and the changing size and shape of the glacier depend on the interplay of dynamics and climate, the latter including the climate of the ocean.

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Tables

Table 2.1 Summary of the recent mass balance of Greenland and Antarctica. (*) 1 km^3 of ice = ~0.92 Gt; (#) Excluding ice shelves; SLE = sea level equivalent.

	Greenland	**Antarctica**
Area (10^6 km^2)	1.7	12.3
Volume (10^6 km^3)*	2.9 (7.3 m SLE)	24.7 (56.6 m SLE)
Total accumulation (Gt a^{-1})#	500 (1.4 mm SLE)	1850 (5.1 mm SLE)
Mass Balance	Since ~1990: Thickening above 2,000 m, at an accelerating rate; thinning at lower elevations also accelerating to cause a net loss from the ice sheet of perhaps >100 Gt a^{-1} after 2000.	Since early 1990s: slow thickening in central regions and southern Antarctic Peninsula; localized thinning at accelerating rates of glaciers in Antarctic Peninsula and Amundsen Sea region. Probable net loss, but close to balance.

Table 2.2 Global small-glacier mass balance for different periods. Consensus estimates (*Kaser et al., 2006*), including small glaciers in Greenland and Antarctica, of global average specific mass balance (*b*); global total mass balance (*B*), equal to $A \times b$ where $A = 785 \times 10^9$ m^2 is the areal extent of small glaciers; and the sea level equivalent (*SLE*), equal to $-B/(\rho_w AO)$, where $\rho_w = 1{,}000$ kg m^{-3} and ocean area $AO = 362 \times 10^{12}$ m^2.

Period	**b (kg m-2 a-1)**	**B (Gt a-1)**	**SLE (mm a-1)**
1961-2004	-231±101	-182±78	0.50±0.22
1961-1990	-173±89	-136±70	0.37±0.19
1991-2004	-356±121	-280±95	0.77±0.26
2001-2004	-451±89	-354±70	0.98±0.19

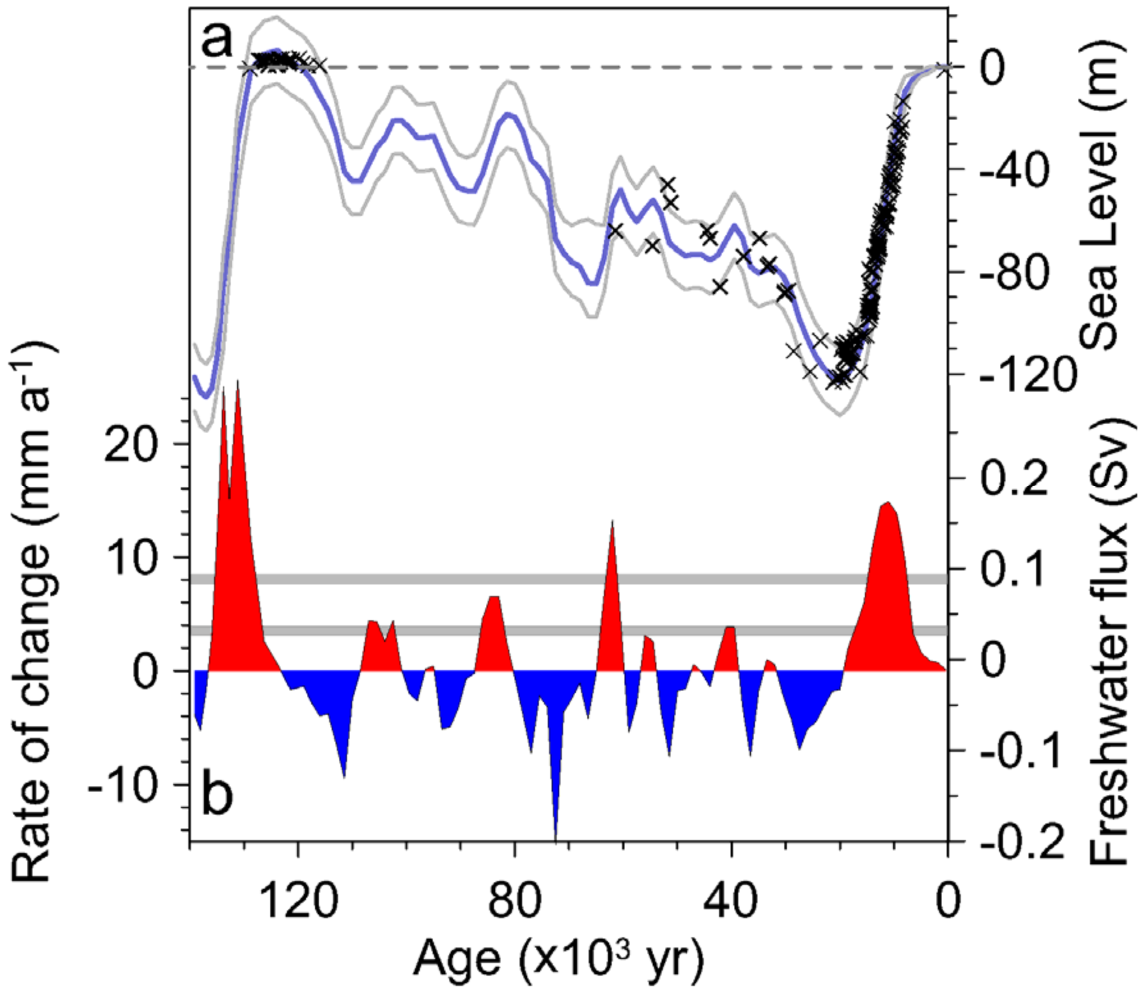

Figure 2.1 (a) Record of sea-level change over the last 130,000 years. Thick blue line is reconstruction from $\delta^{18}O$ records of marine sediment cores through regression analyses (*Waelbroeck et al., 2002*), with ±13 m error shown by thin gray lines. The × symbols represent individually dated shorelines from Australia (*Stirling et al., 1995, 1998*), New Guinea (*Edwards et al., 1993; Chappell, 2002; Cutler et al., 2003*), Sunda Shelf (*Hanebuth et al., 2000*), Bonaparte Gulf (*Yokoyama et al., 2000*), Tahiti (*Bard et al., 1996*), and Barbados (*Peltier and Fairbanks, 2006*). (b) Rate of sea level change (mm a^{-1}) and equivalent freshwater flux (Sv, where 1 Sv = 106 m^3 s^{-1} = 31,500 Gt a^{-1}) derived from sea-level record in (a). Horizontal gray bars represent average rates of sea level change during the 20th century (lower bar) and projected for the end of the 21st century (upper bar) (*Rahmstorf, 2007*).

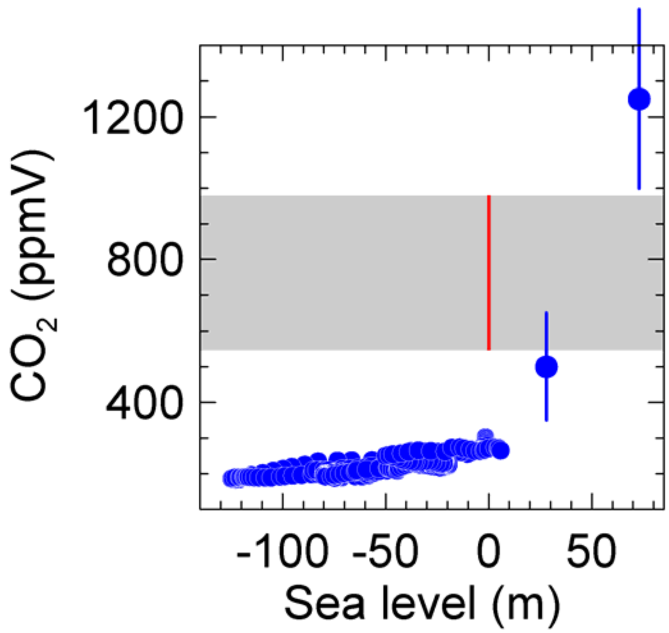

Figure 2.2. Relation between estimated atmospheric CO_2 and the ice contribution to eustatic sea level indicated by geological archives and referenced to modern (pre-industrial era) conditions [CO_2 =280 parts per million by volume (ppmV), eustatic sea level = 0 m]. Horizontal gray box represents range of atmospheric CO_2 concentrations projected for the end of the 21st century based on IPCC emission scenarios (lower end is B1 scenario, upper end is A1F1 scenario) (*Nakicenovic et al., 2000*). The vertical red bar represents the IPCC Fourth Assessment Report (AR4) estimate of sea level rise by the end of the 21st century (*Meehl et al., 2007*). The difference between the IPCC AR4 estimate and the high paleo-sea levels under comparable atmospheric CO_2 levels of the past (blue dots with vertical bar given as uncertainties) largely reflects the long response time of ice sheets. A central question raised by the dynamical changes in ice sheets described in this chapter (and that are not included in the IPCC AR4 estimates) is how much they will reduce the ice-sheet response time to climate change.

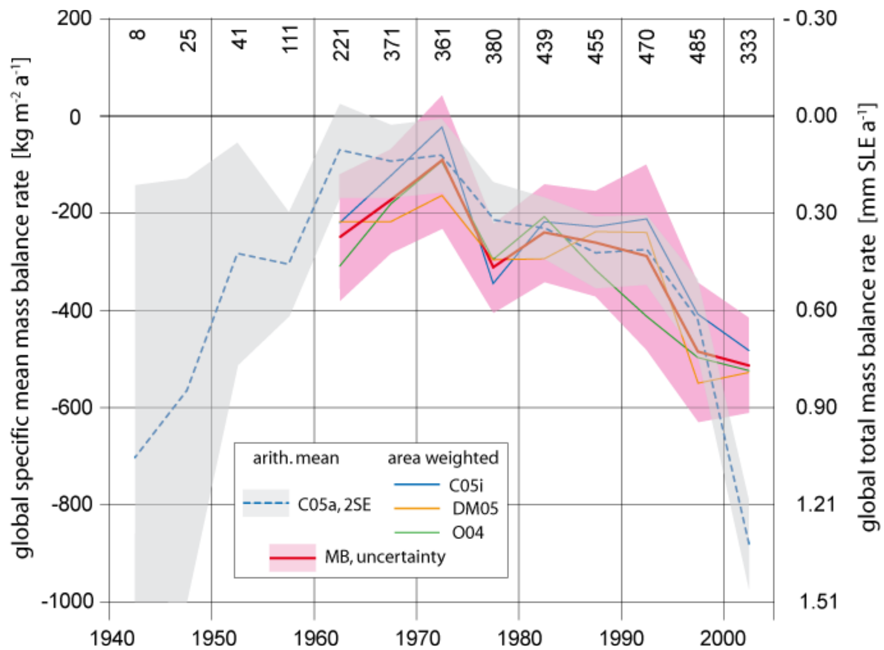

Figure 2.3. Pentadal average mass-balance rates of the world's glaciers and ice caps, excluding Greenland and Antarctica, for the last half century. Specific mass balance (left axis) is converted to total balance and to sea level equivalent (right axis) as described in Table 2.2. C05a: an arithmetic mean over all annual measurements within each pentad, with confidence envelope shaded grey and number of measurements given at top of graph. C05i, DM05, O04: independently obtained spatially corrected series. MB: arithmetic mean of C05i, DM05 and O04, with confidence envelope shaded red. See *Kaser et al. (2006)* for sources and uncertainties; the latter are "2-sigma-like". Estimates are incomplete for the most recent pentad. Copyright American Geophysical Union, 2006; reprinted with permission.

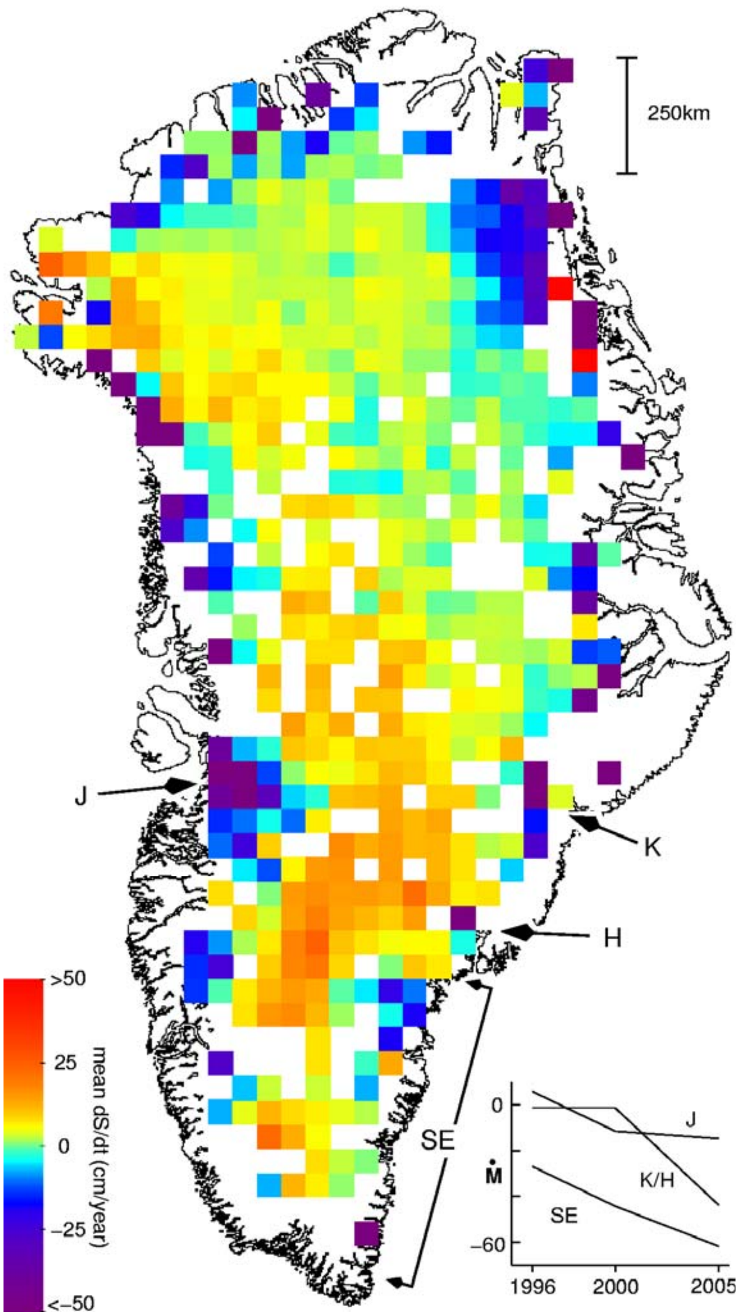

Figure 2.4. Rates of elevation change (dS/dt) for Greenland derived from comparisons at more than 16,000 locations where ICESat data from Oct/Nov and May/June 2004 overlay ATM surveys in 1998/9, averaged over 50-km grid squares. Locations of rapidly thinning outlet glaciers at Jakobshavn (J), Kangerdlugssuaq (K), Helheim (H), and along the southeast coast (SE) are shown, together with plots showing their estimated mass balance ($\dot{M}$ Gt a^{-1}) versus time (*Rignot and Kanagaratnam, 2006*).

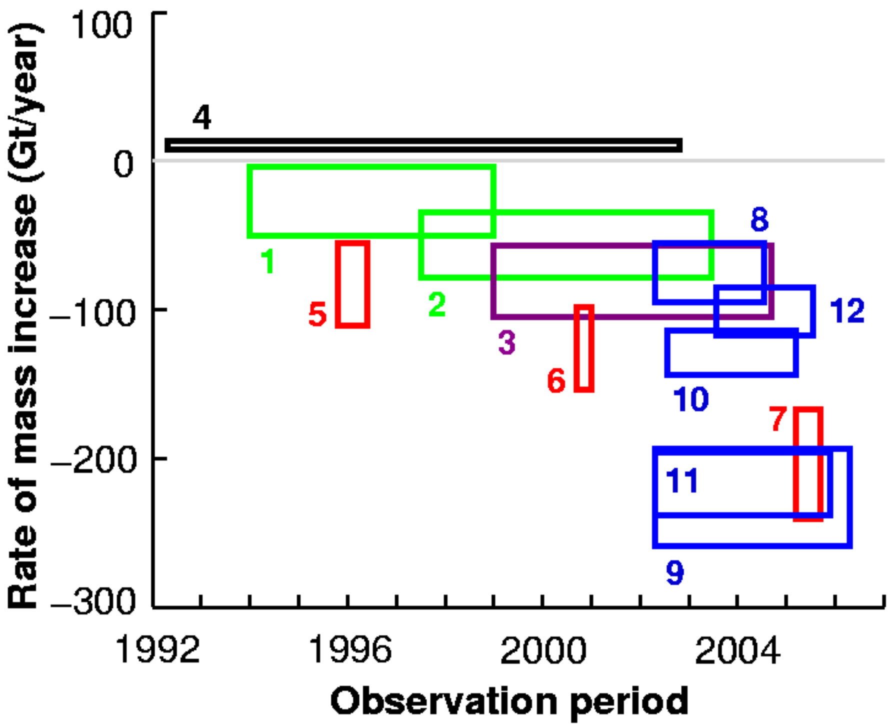

Figure 2.5. Mass-balance estimates for the entire Greenland ice sheet: green—airborne laser altimetry (ATM); purple—ATM/ICESat (summarized in *Thomas et al., 2006*); black—Satellite Radar Altimetry (SRALT) (4: *Zwally et al., 2005*); red—mass budget (5,6,7: *Rignot and Kanagaratnam, 2006*); blue—GRACE (8 and 9: *Velicogna and Wahr, 2005, 2006a*; 10: *Ramillien et al., 2006*; 11*: Chen et al., 2006*; 12*: Luthcke et al., 2006*). The ATM results were supplemented by degree-day estimates of anomalous melting near the coast (*Krabill et al., 2000,; 2004*), and probably underestimate total losses by not taking full account of dynamic thinning of outlet glaciers (*Abdalati et al., 2001*). SRALT results seriously underestimate rapid thinning of comparatively narrow Greenland glaciers, and may also be affected by progressively increased surface melting at higher elevations.

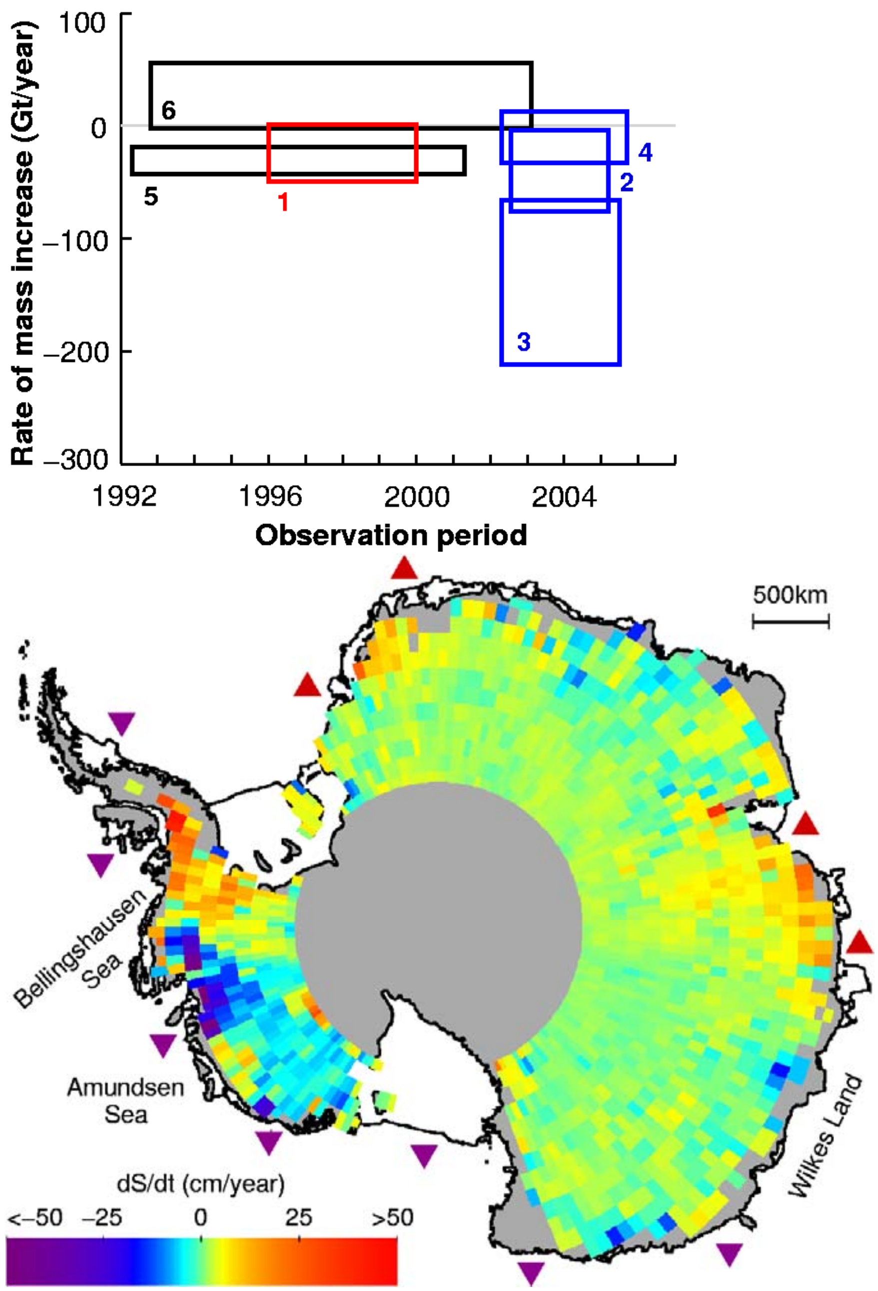

Figure 2.6. Rates of elevation change (dS/dt) derived from ERS radar-altimeter measurements between 1992 and 2003 over the Antarctic Ice Sheet (*Davis et al., 2005*). Locations of ice shelves estimated to be thickening or thinning by more than 30 cm a-1 (*Zwally et al., 2005*) are shown by purple triangles (thinning) and red triangles (thickening). Inset shows mass-balance estimates for the ice sheet: red—mass budget (1: *Rignot and Thomas, 2002*); blue—GRACE (2: *Ramillien et al., 2006*; 3: *Velicogna and Wahr, 2006b*; 4: *Chen et al., 2006*); black—ERS SRALT (5: *Zwally et al., 2005*; 6: *Wingham et al., 2006*).

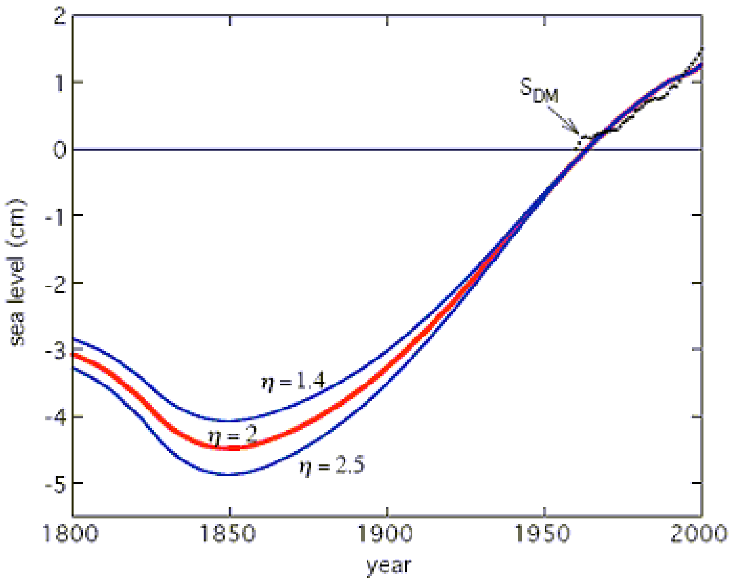

Figure 2.7. Reconstruction of the cumulative glacier contribution to sea level change relative to an arbitrary zero in 1961 (*Oerlemans et al., 2007*). The three smooth curves represent different choices for η, a parameter which regulates the conversion of normalized glacier length to volume. *SDM* (dots) is the cumulative contribution estimated directly from measurements. Copyright of the author; reprinted with permission.

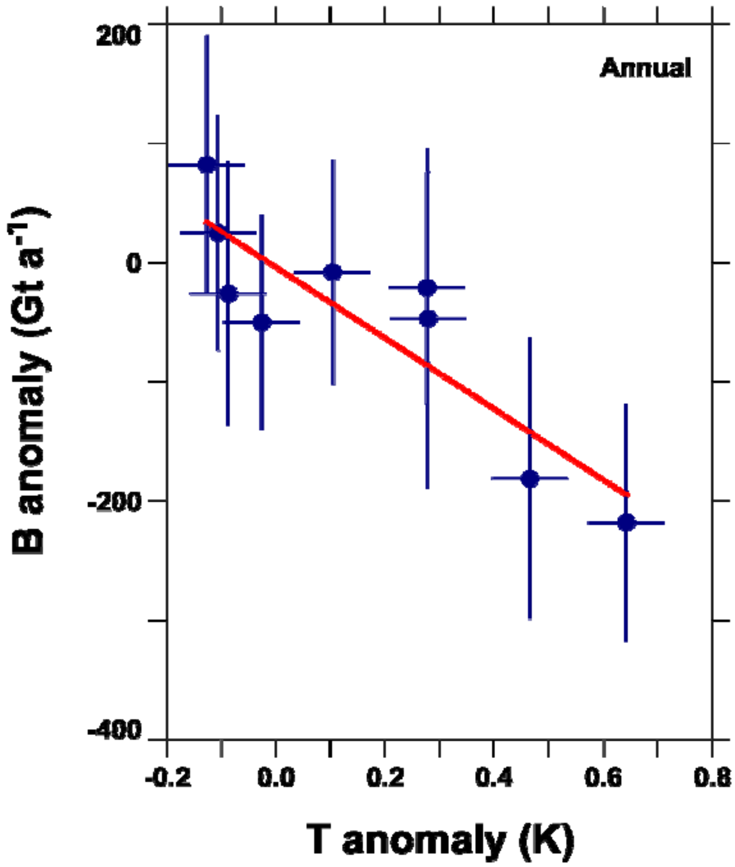

Figure 2.8. Correlation of the anomaly (relative to the 1961-1990 average) in pentadal mean annual mass balance *B* (*Kaser et al., 2006*) with the corresponding anomaly in *T*, surface air temperature over land (CRUTEM3*; Trenberth et al., 2007*). The fitted line suggests a proportionality *dB/dT* of -297±133 Gt a^{-1} K^{-1} for the era of direct balance measurements (1961-2004).

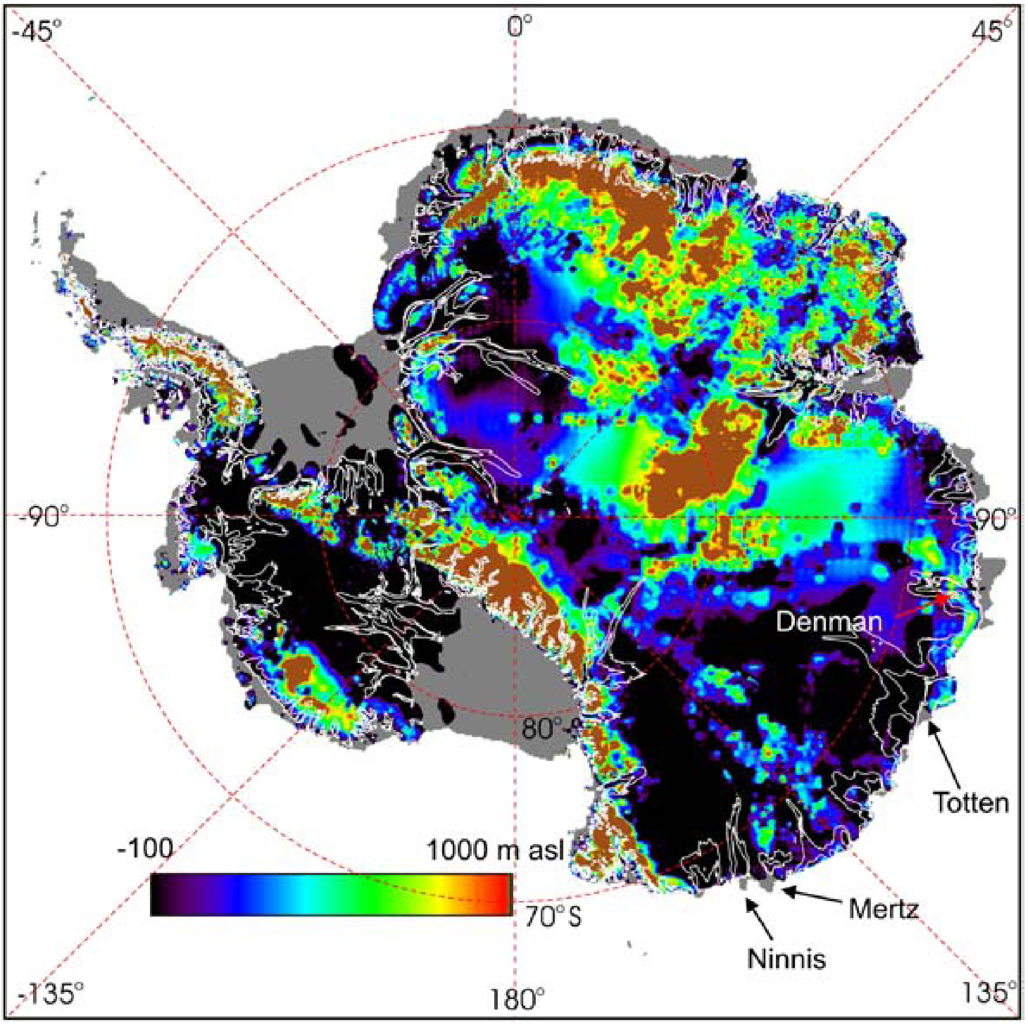

Figure 2.9. Bedrock topography for Antarctica highlighting areas below sea level (in black), fringing ice shelves (in dark grey), and areas above sea level (in rainbow colors). Areas of enhanced flow are identified by contours (in white) of estimated steady-state velocities, known as balance velocities. From *Bamber et al. (2007)*.

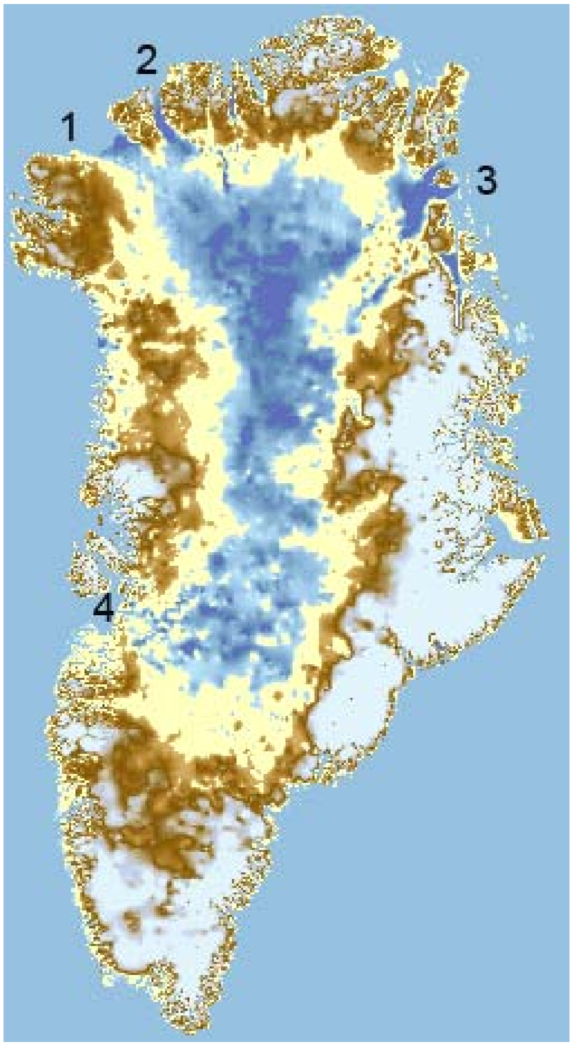

Figure 2.10. Bedrock topography for Greenland; areas below sea level are shown in blue. Note the three channels in the north (1: Humboldt Glacier; 2: Petermann Glacier; 3: 79-North Glacier or Nioghalvfjerdsfjorden Glacier) and at the west coast (4: Jakobshavn Isbrae) connecting the region below sea level with the ocean (Russell Huff and Konrad Steffen, CIRES, University of Colorado at Boulder.)