Ice-Sheet and Sea-Level Changes

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Future sea-level rise is an important issue related to the continuing buildup of atmospheric greenhouse gas concentrations. The Greenland and Antarctic ice sheets, with the potential to raise sea level ~70 meters if completely melted, dominate uncertainties in projected sea-level change. Freshwater fluxes from these ice sheets also may affect oceanic circulation, contributing to climate change. Observational and modeling advances have reduced many uncertainties related to ice-sheet behavior, but recently detected, rapid ice-marginal changes contributing to sea-level rise may indicate greater ice-sheet sensitivity to warming than previously considered.

Because a heavy concentration of the population lives along coastlines, even small amounts of sea-level rise would have substantial societal and economic impacts through coastal erosion, increased susceptibility to storm surges, groundwater contamination by salt intrusion, and other effects. Over the last century, sea level rose ~1.0 to 2.0 mm/year, with water expansion from warming contributing 0.5 ± 0.2 mm (steric change) (*1*, *2*) and the rest from the addition of water to the oceans (eustatic change) due mostly to melting of land ice (*2*). By the end of the 21st century, sea level is projected to rise by 0.5 ± 0.4 m in response to additional global warming (*2*), with potential contributions from the Greenland and Antarctic ice sheets dominating the uncertainty of that estimate.

These projections emphasize surface melting and accumulation in controlling ice-sheet mass balance, with different relative contributions for warmer Greenland and colder Antarctica (*3*). The Greenland Ice Sheet may melt entirely from future global warming (*4*), whereas the East Antarctic Ice Sheet (EAIS) is likely to grow through increased accumulation for warmings not exceeding ~5°C (*5*). The future of the West Antarctic Ice Sheet (WAIS) remains uncertain, with its marine-based configuration raising the possibility of important losses in the coming centuries (*2*). Despite these uncertainties, the geologic record clearly indicates that past changes in atmospheric CO_2 were correlated with substantial changes in ice volume and global sea level (Fig. 1).

Recent observations of startling changes at the margins of the Greenland and Antarctic ice sheets indicate that dynamical responses to warming may play a much greater role in the future mass balance of ice sheets than previously considered. Models are just beginning to include these responses, but if they prove to be important, sea-level projections may need to be revised upward. Also, because sites of global deepwater formation occur immediately adjacent to the Greenland and Antarctic ice sheets, any notable increase in freshwater fluxes from these ice sheets may induce changes in ocean heat transport and thus climate. Here, we review these new developments in understanding ice-sheet mass balance and discuss their possible implications to future sea level and climate.

Paleoglaciology

The record of past glacial changes provides important insight to the behavior of large ice sheets during warming. At the last glacial maximum about 21,000 years ago, ice volume and area were more than twice modern values (*6*). Deglaciation was forced by warming from changes in Earth's orbital parameters, increasing greenhouse gas concentrations, and other attendant feedbacks. Deglacial sea-level rise averaged 10 mm/year, but with variations including two extraordinary episodes at 19,000 years before present (19 kyr B.P.) and 14.5 kyr B.P. (Fig. 2), when peak rates potentially exceeded 50 mm/year (*7*–*9*). Each of these "meltwater pulses" added the equivalent of 1.5 to 3 Greenland Ice Sheets to the oceans over a period of one to five centuries.

The freshwater fluxes associated with these events apparently induced large changes in ocean circulation and attendant heat transport. An important component of the ocean's overturning circulation involves deepwater formation in the North Atlantic Ocean and around the Antarctic continent, particularly in the Weddell and Ross Seas. Accordingly, partial collapse of northern ice into the North Atlantic Ocean at 19 kyr B.P. may have weakened North Atlantic deepwater formation, causing widespread cooling (*9*). In contrast, a large contribution of Antarctic ice to the event of 14.5 kyr B.P. (*10*) would have freshened the Southern Ocean, perhaps strengthening the Atlantic meridional overturning circulation (AMOC) and causing widespread warming (*11*).

Ice-Sheet Mass Balance

Ice-sheet mass balance can be estimated by taking the difference between ice input and output fluxes or by monitoring changes in ice-sheet elevation as a proxy for volume changes. Input, primarily from precipitation, can be estimated from field measurements and by atmospheric modeling. Output, primarily from surface melt, sub–ice-shelf melt, or iceberg calving, can be calculated from melt models or ice-velocity measurements from interfero-

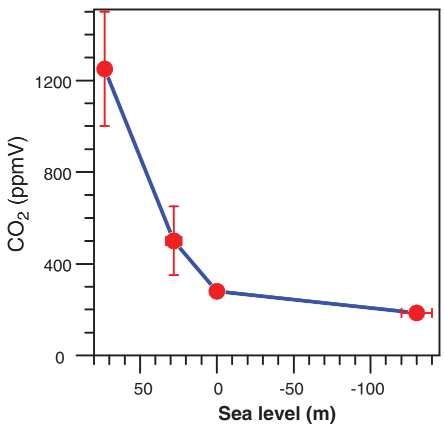

Fig. 1. Relation between estimated atmospheric CO_2 and the ice contribution to eustatic sea level indicated by geological archives and referenced to modern (pre-Industrial Era) conditions [CO_2 = 280 parts per million by volume (ppmV), eustatic sea level = 0 m]. The most recent time when no permanent ice existed on the planet (sea level = +73 m) occurred >35 million years ago when atmospheric CO_2 was 1250 ± 250 ppmV (*54*). In the early Oligocene (~32 million years ago), atmospheric CO_2 decreased to 500 ± 150 ppmV (*54*), which was accompanied by the first growth of permanent ice on the Antarctic continent, with an attendant eustatic sea-level lowering 45 ± 5 m (*55*). The most recent time of low atmospheric CO_2 (185 ppmV) (*56*) corresponds to the Last Glacial Maximum 21,000 years ago, when eustatic sea level was –130 ± 10 m (*8*). Error bars show means ± SD.

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metric synthetic-aperture radar (InSAR). Monitoring changing ice volume by repeat altimetry from aircraft or satellite is increasingly important, after correction for any isostatic adjustments of bedrock elevation in response to past ice-load changes and for changing density of the snow and ice column in response to changing climate. Although some altimetry data were collected in the 1970s, comprehensive mass-balance observations did not begin until the early 1990s, precluding separation of decadal or subdecadal variability from longer term trends. Nevertheless, observations have documented changes in Greenland and Antarctica including notable increases in ice discharge, especially since the mid- to late 1990s (*12*, *13*).

For Greenland, updated estimates based on repeat altimetry, and the incorporation of atmospheric and runoff modeling, indicate increased net mass loss, with most change toward the coasts (*13*). Between 1993 to 1994 and 1998 to 1999, the ice sheet was losing 54 ± 14 gigatons per year (Gt/year) of ice, equivalent to a sea-level rise of ~0.15 mm/year (where 360 Gt of ice = 1 mm sea level). The excess of meltwater runoff over surface accumulation was about 32 ± 5 Gt/year, leaving ice-flow acceleration responsible for loss of ~22 Gt/year. Despite highly anomalous excess snowfall in the southeast in 2002 to 2003, net mass loss over the 1997-to-2003 interval was higher than the loss between 1993 and 1999, averaging 74 ± 11 Gt/year or ~0.21 mm/year sea-level rise, with increases in both the excess of surface melt over snow accumulation (42 ± 6 Gt/year) and the ice-flow loss. Summers were warmer from 1997 to 2003 than from 1993 to 1999, which likely explains the increased surface melt (*13*). These results are broadly similar to those from a mesoscale atmospheric model used to simulate the surface mass balance of the Greenland Ice Sheet from 1991 to 2000 (*14*). Accounting for additional mass loss from iceberg discharge and basal melting (assumed constant) yielded an estimated net mass loss of 78 Gt/year. Large interannual variability did not obscure significant simulated trends toward increased melting and snowfall consistent with reconstructed warming, especially in west Greenland.

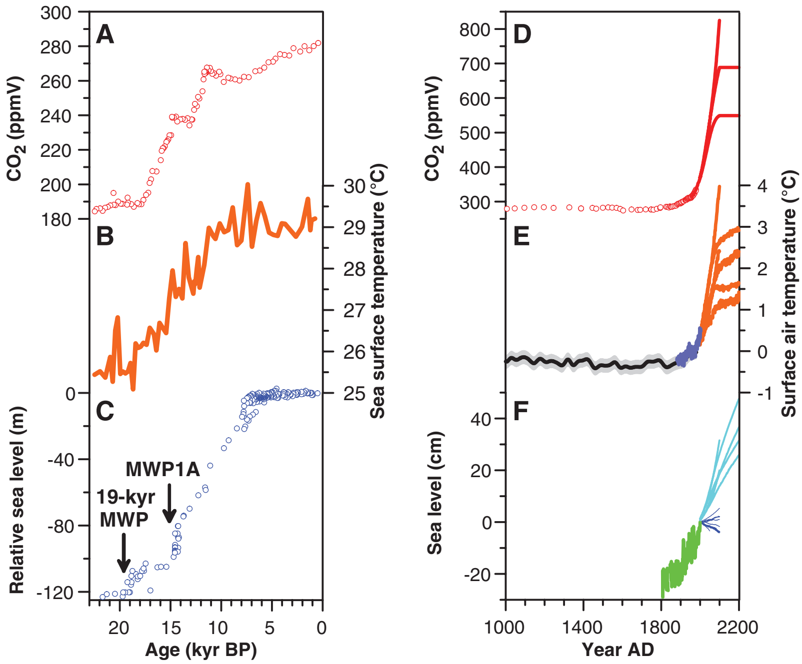

Fig. 2. Time series of key variables encompassing the last interval of significant global warming (last deglaciation) (left) compared with the same variables projected for various scenarios of future global warming (right). (**A**) Atmospheric CO_2 from Antarctic ice cores (*56*). (**B**) Sea surface temperature in the western equatorial Pacific based on Mg/Ca measured in planktonic foraminifera (*57*). (**C**) Relative sea level as derived from several sites far removed from the influence of former ice-sheet loading (*8*, *58*–*60*). MWP, meltwater pulse. (**D**) Atmospheric CO_2 over the past millennium (circles) and projections for future increases (solid lines). Records of atmospheric CO_2 are from Law Dome, Antarctica (*61*), and direct measurements since 1958 are from Mauna Loa, Hawaii (*62*). Also shown are three emission scenarios for time evolution of atmospheric CO_2 over the course of the 21st century and subsequent stabilization through the 22nd century (*63*). (**E**) Temperature reconstruction for Northern Hemisphere from 1000 to 2000 AD (*64*) (gray time series), global temperature based on historic measurements, 1880 to 2004 (*65*) (blue time series), and projected warming based on simulations with two global coupled three-dimensional (3D) climate models with the use of three emission scenarios (*66*) (orange time series). (**F**) Relative sea-level rise during the 19th and 20th centuries from tide gauge record at Brest, France (*67*) (green time series), projections for contributions from combined Greenland and Antarctic ice sheets (*3*) (dark blue time series), and projections for sea-level rise from thermal expansion based on climate simulations shown in (E) (light blue time series) (*66*).

In Antarctica, altimetry-derived estimates show thickening in EAIS (*15*) but thinning along the Amundsen Coast of WAIS (*15*–*17*). From 1992 to 2003, measured thickening (1.8 ± 0.3 cm/year) of the larger EAIS more than balances measured WAIS thinning (0.9 ± 0.3 cm/year); assigning snow density to the changes in EAIS owing to correlation with rising accumulation rate, but ice density to WAIS changes in light of probable dynamic contributions, yields a combined mass gain of ~33 ± 8 Gt/year (*15*). However, data gaps remain, including on the Antarctic Peninsula and near the South Pole, and additional uncertainty arises from lack of knowledge of changing density in upper layers. Mass-balance estimates covering an overlapping subset of drainage basins and times also suggest EAIS growth (20 ± 21 Gt/year) and show WAIS loss (44 ± 13 Gt/year) (*12*). Atmospheric modeling indicates that increasing snow accumulation has been important over the last decades (*15*, *18*), perhaps in response to weak warming, especially in coastal regions (*19*).

Modeling

Interpretation of past changes and projection of future changes requires modeling. Two major traditions of ice-deformation modeling have developed, reflecting the very different stress regimes for inland ice versus ice shelves. Current comprehensive ice-sheet models typically treat inland ice and ice shelves separately, with regime coupling, but this approach does not fully capture the transitional behavior of ice streams and outlet glaciers (*5*). Use of this approach is motivated in part by poor understanding of the basal boundary condition in transition zones. Computational limitations also dictate rather coarse grid spacings, numerically widening and slowing fast-flowing ice streams. Because slower flowing ice cannot contribute as rapidly to sea-level change, this grid coarsening can cause models to respond more slowly than actual ice sheets. Furthermore, actual ice sheets transmit longitudinal stress perturbations almost instantaneously, but inland-ice models do not. Accordingly, these ice-sheet models may underestimate rates of change.

Despite their limitations, coupled-regime models show substantial skill in simulating ongoing changes, except in the rapidly changing marginal regions. Recent coupled-regime simulations (*3*) with mass balance driven by climate-model output suggest that 20th-century surface forcing should have caused slight inland thickening in Antarctica and Greenland, and coastal thinning in Greenland. These results show relatively little long-term trend in Greenland, but a long-term Antarctic thinning trend from the end of the last ice age, especially for WAIS. The modeled behavior in Greenland agrees with available data within stated errors (*13*), except for the highly variable ice-marginal changes discussed below. The modeled long-term Antarctic trend shows poorer agreement with the limited data (*16*); this may reflect the short time span of the data or uncertainty in timing of ice-age forcing. The simulated Antarctic response to 20th-century forcing matches many observations in areas controlled mainly by changes in accumulation rate. However, rapid changes in some regions such as the Amundsen Coast are not simulated, in part because the oceanic forcing thought responsible for these shifts was not included.

Rapid Ice-Marginal Changes

The theory that ice shelves or ice tongues buttress fast-flowing ice streams and outlet glaciers, preventing faster flow and ice-sheet shrinkage or collapse (*20*), was disputed by subsequent work [such as (*21*)] but is again supported by recent observations and modeling.

On Greenland's west coast, Jakobshavn Isbrae, which drains about 6% of the ice-sheet area, experienced slight slowing from 1985 to 1992 but remained among the fastest glaciers on Earth. Jakobshavn has subsequently nearly doubled its flow speed and thinned rapidly, with the speedup extending ~30 km inland (*22*, *23*). Increased surface melting alone cannot explain this thinning. Instead, Jakobshavn Isbrae's acceleration in association with the loss of its floating ice tongue suggests a dynamic thinning following loss of restraint to flow provided by the ice tongue (Fig. 3) (*22–24*).

Although Jakobshavn Isbrae is the most notable example, laser altimeter surveys over a 5-year interval document that the lower sections of many other Greenland outlet glaciers also thinned (*25*). As for Jakobshavn Isbrae, surface melting cannot account for much of this thinning. The largest changes are found near the fronts of fast-moving outlet glaciers that feed small ice shelves or tongues that have retreated in response to atmospheric or ocean warming, suggesting that warming-induced reduction of ice-shelf restraint triggered flow acceleration. Taken together, accelerated discharge of documented Greenland outlet glaciers may have contributed up to ~0.09 mm/year to sea level since the mid-1990s (*13*, *23*).

Altimetry surveys and InSAR data document recent acceleration of the WAIS contribution to sea-level rise as a result of rapid ice-marginal changes. Along the Antarctic Peninsula, warming over the last few decades has caused retreat or near-total loss of several ice shelves, at least some of which had persisted for millennia (*26*). Ice shelves are susceptible to attack by warming-induced increases of meltwater ponding in crevasses that cause hydrologically driven fracturing (*27*) and by warmer subshelf waters that increase basal melting (*28*). Responses to ice-shelf breakup have been noteworthy. Collapse of the Larsen B Ice Shelf in 2002 was followed by speedup of its major tributary glaciers, by twofold to eightfold where they entered the former ice shelf; speedup decreased inland but was recognizable for roughly 10 km inland and contributed about 0.07 mm/year to sea-level rise; adjacent glaciers still buttressed by shelf ice changed little (*29*, *30*). Loss of the Larsen A Ice Shelf, north of the Larsen B, seems to have caused acceleration of tributary glaciers (*30*), and observed acceleration of tributary flow to the former Wordie Ice Shelf on the other side of the Antarctic Peninsula may have been linked to loss of that ice shelf as well (*31*). Models indicate that geometric and other factors contribute to the magnitude and speed of tributary-glacier response to ice-shelf reduction (*32*), so the range of observed responses is not surprising.

Recent changes in glaciers along the Amundsen Coast of WAIS are also contributing to sea-level rise, with discharge in excess of accumulation accounting for 0.13 ± 0.02 mm/year (*33*) to 0.24 mm/year (*34*). In Pine Island Bay of this coast, ice-shelf thinning at rates locally exceeding 5 m/year (*33*) was accompanied by grounding-line retreat of 1.2 km/year in the early 1990s (*35*). Apparently in response, large glaciers feeding Amundsen Coast ice shelves have thinned and accelerated by up to 26% over the last three decades, with perturbations extending more than 200 km inland (*17*, *34*, *36*). Although acceleration of Amundsen Coast glaciers increased mass flux to ice shelves, the shelves have thinned, suggesting increased basal melting likely from increased penetration of relatively warm Circumpolar Deep Water (*37*).

Not all ice-dynamical anomalies are causing thinning. On the Siple Coast of WAIS, thinning has switched to thickening as Whillans ice stream (ice stream B) slowed between 1974 and 1997, causing the ice-sheet contribution to sea-level rise in 1997 to be 0.03 mm/year smaller than if 1974 velocities had been maintained (*38*). Large dynamical changes of opposing signs have affected this region over the last millennium (*39*), however, and features of the region may predispose it to ice-flow "noise" (*40*).

The association of outlet-glacier acceleration, dynamical thinning, and ice-shelf changes affecting several ice streams implicates a response to ice-shelf changes rather than individual dynamical explanations such as periodic surging. An ice-sheet model including lateral drag and longitudinal stress gradients applied to Pine Island Glacier simulates instantaneous acceleration extending ~100 km inland in response to ice-shelf reduction, followed by diffusive-advective thinning up to 200 km inland, in good agreement with observations (*41*) and with results from other models that include nonlocal stresses (Fig. 4) (*32*, *42*). Similarly detailed

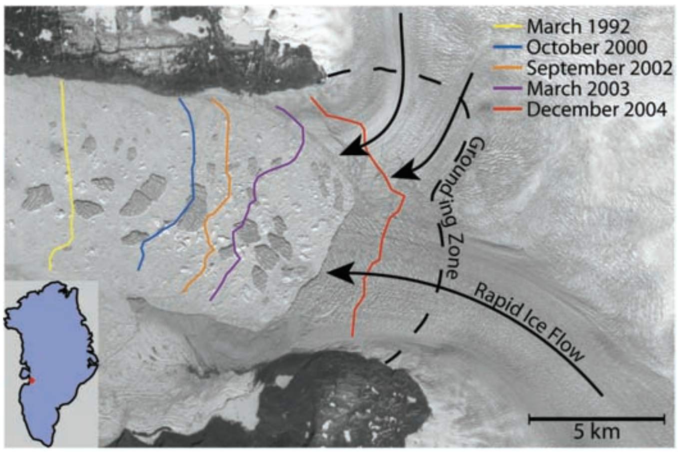

Fig. 3. Section of a May 2003 Landsat image acquired after the nearly complete disintegration of the floating ice tongue of Jakobshavn Isbrae, Greenland's largest outlet glacier. The black dashed line shows the approximate grounding zone (*68*). The color lines show the location of the ice tongue's front at several times. Short-term oscillations (not shown) were superimposed on the general trend. The ice tongue's breakup coincided with rapid thinning upstream of the floating ice (up to 15 m/year) and with a near doubling of the glacier's speed (up to ~13 km/year) (*22*, *23*).

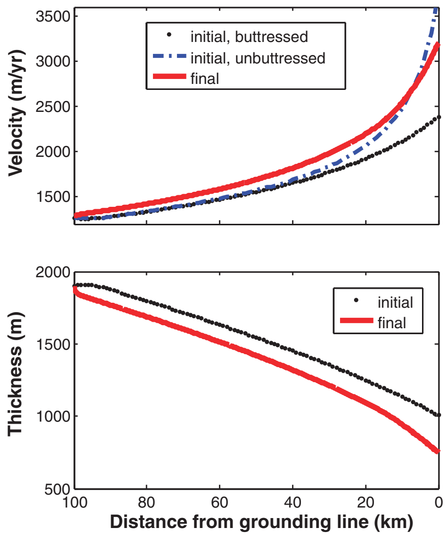

Fig. 4. Modeled response of an idealized version of Pine Island Glacier, West Antarctica, to loss of a small ice shelf (one resisting half of the tendency for ice spreading at the grounding line), following (*32*). In this model, response is limited to the ice stream itself and cannot propagate into the ice sheet beyond the ice stream. The near-instantaneous increase in velocity following ice-shelf loss is physical but is not simulated in older models lacking longitudinal stresses. The subsequent velocity evolution is largely a result of the thinning and stress reduction in response to that near-instantaneous speedup.

modeling has not been conducted for the recent changes of Jakobshavn Isbrae and the tributaries to the former Larsen B ice shelf, but all appear to be changing at rates not fully captured by models that exclude longitudinal stress gradients.

The data summarized above indicate a contribution to sea-level rise from ice-flow changes in marginal regions of roughly ⅓ mm/year, with evidence of increased discharge since the mid-1970s and especially since the mid-1990s. This dynamic imbalance is of comparable magnitude to the direct effect of recent surface mass-balance changes, of the same sign for Greenland but of opposite sign for Antarctica (*13*, *14*, *18*, *43*). The recently detected glacier accelerations are too young, however, and the observational record is too short to evaluate whether they represent short-term fluctuations or are part of a longer term trend that might scale with future climatic warming. Slight deceleration of portions of the fastest glaciers flowing into the former Larsen B ice shelf only 1 year after the ice shelf broke up (*29*) and a known record of variability of Jakobshavn Isbrae with slight thickening occurring only a decade ago (*22*, *23*, *44*) suggest that these events may just represent fast adjustments to marginal fluctuations. Alternatively, oceanic erosion of ice margins, especially in the Antarctic, may well continue in the future, in which case the eventual response could be far greater (*34*, *41*). Ice-flow perturbations may also affect basal conditions in poorly understood ways, with both positive and negative feedbacks possible and potentially large (*40*).

Whereas we have focused on inland response to changes in floating ice, an additional process that may prove important in Greenland is meltwater penetration to the bed, providing near-instantaneous communication between surface forcing and basal-ice dynamics. This is known to occur through ~1 km of cold ice in western Greenland, with modest (order of 10%) increase in speed from enhanced basal lubrication (*45*). In addition, one case of meltwater penetration through Ryder Glacier in northern Greenland was observed to cause a speed-up of more than threefold persisting several weeks (*46*). Meltwater penetration likely occurs through water-filled fractures, especially if fed by surface lakes (*47*). If meltwater access to the bed moves inland with warming, as seems likely, fracture penetration to and along the bed may cause rapid thawing and ice-flow speedup, thus shortening the ice-sheet response time to global warming and accelerating sea-level rise (*48*).

Future Behavior

Predictions of ice-sheet contributions to sea-level rise have relied on long integrations of ice-sheet models that extend well into the 21st century and beyond (*2*, *4*). These predict that up to the year 2100, warming-induced ice-sheet

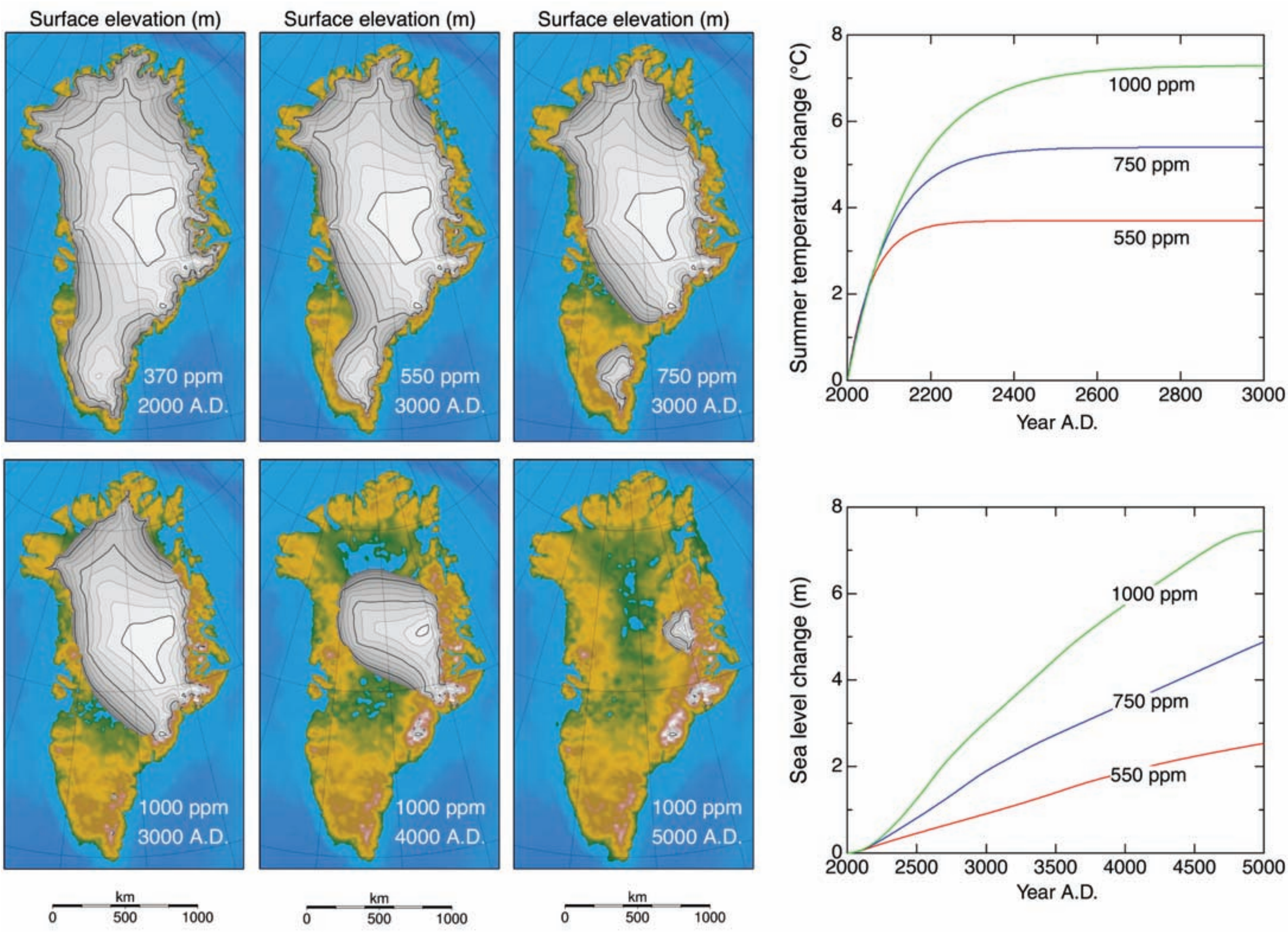

Fig. 5. Future evolution of the Greenland Ice Sheet calculated from a 3D ice-sheet model forced by three greenhouse gas stabilization scenarios. The warming scenarios correspond to the average of seven IPCC models in which the atmospheric carbon dioxide concentration stabilizes at levels between 550 and 1000 ppm after a few centuries (*4*) and is kept constant after that. For a sustained average summer warming of 7.3°C (1000 ppm), the Greenland Ice Sheet is shown to disappear within 3000 years, raising sea level by about 7.5 m. For lower carbon dioxide concentrations, melting proceeds at a slower rate, but even in a world with twice as much CO_2 (550 ppm or a 3.7°C summer warming) the ice sheet will eventually melt away apart from some residual glaciation over the eastern mountains. The figure is based on the models discussed in (*5*).

growth in Antarctica will offset enhanced melting in Greenland (*3*). For the full range of climate scenarios and model uncertainties, average 21st-century sea-level contributions are –0.6 ± 0.6 mm/year from Antarctica and +0.5 ± 0.4 mm/year from Greenland, resulting in a net contribution not significantly different from zero, but with uncertainties larger than the peak rates from outlet glacier acceleration during the past 5 to 10 years.

Looking further into the future, inland-ice models raise concerns about the Greenland Ice Sheet (Fig. 5). At present, mass loss by surface-meltwater runoff is similar to iceberg-calving loss plus sub–ice-shelf melting, with total loss only slightly larger than snow accumulation. For warming of more than about 3°C over Greenland, surface melting is modeled to exceed snow accumulation (*4*), and the ice sheet would shrink or disappear. For the most extreme Intergovernmental Panel on Climate Change (IPCC) warming scenario, one modeling study found 7 m of sea-level rise from Greenland in about 1000 years (*4*, *49*). This loss of the Greenland Ice Sheet would be irreversible without major cooling (*50*). However, increased Greenland meltwater may suppress the AMOC (*51*), causing regional cooling and an attendant decrease in Greenland melting. In contrast, important mass loss from surface melting of Antarctic ice is not expected in existing scenarios, although grounding-line retreat along the major ice shelves is modeled for basal melting rates >5 to 10 m/year, causing the demise of WAIS ice shelves after a few centuries and retreat of coastal ice toward more firmly grounded regions after a few millennia (*5*, *52*), with implied rates of sea-level rise of up to 3 mm/year. If large and rapid mass losses from WAIS occurred, any attendant freshening of Antarctic Intermediate Water may strengthen the AMOC (*11*).

Because the models used in these projections lack some of the physical processes that might explain the rapid rates of ongoing coastal changes and lack the oceanic forcing responsible for inducing these changes, previous estimates of sea-level change may become lower limits if ongoing speedups are sustained and eventually become more widespread. Progress is being made in ice-flow models (*32*, *41*, *53*), but no model including all relevant forces has yet been produced, and comprehensive ice-sheet integrations with such a model do not seem imminent. Nonetheless, the recent observations discussed here reveal that rapid dynamic changes can be important, contributing a notable fraction of ongoing sea-level rise and potentially becoming dominant over ice-sheet surface mass-balance changes in the future.

Summary

Ice sheets now appear to be contributing modestly to sea-level rise because warming has increased mass loss from coastal areas more than warming has increased mass gain from enhanced snowfall in cold central regions. At present, thickening on the EAIS appears to be nearly balanced by WAIS thinning along the Amundsen Coast, much of which reflects recent changes. With an Antarctic Ice Sheet not far from balance despite large regional imbalances, Greenland presently makes the largest contribution to sea-level rise. Ice-sheet models that have supported the IPCC effort do not include the full suite of physical processes implicated in the ongoing changes, however, and so are not able to assess whether these ongoing changes represent minor perturbations before stabilization or a major change that may affect sea level notably. Fundamental shortcomings in available data sets as well as models preclude confident projection of rapid future changes, and this difficulty is compounded by possible interactions between freshwater fluxes from ice sheets, ocean circulation, and climate. The major challenges then are to acquire the observations necessary to characterize rapid dynamic changes, and to incorporate those data into improved models, allowing more reliable predictions of ice contributions to sea-level change over the coming decades and centuries.

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