Glacier science and environmental change

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Numerical modelling of polar ice sheets through time

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80.1 Introduction

Ice sheets respond dynamically to changes in boundary conditions, such as climate variations, basal thermal conditions, and isostatic adjustments of the underlying bedrock. These cause the ice sheets to evolve towards a new equilibrium. Long response time-scales of up to 10^4 years are involved, determined by the ratio of ice thickness to yearly mass turnover, physical and thermal processes at the bed, and processes affecting ice viscosity and mantle viscosity. The response of the ice sheets is further complicated by feedback processes which may amplify or mitigate the ice sheet's adjustment to the forcing or by internal instabilities that may cause rapid changes in ice volume due to changes in the dynamic flow regime. A primary motivation for developing numerical models of ice flow is to gain a better understanding of the spatial and temporal behaviour of ice sheets and glaciers and to predict their response to external forcing. Modelling ice-sheet dynamics presents a powerful framework to investigate the complex interactions between the ice sheets and the climate system in a quantitative way, in past as well as future environments. Ice-flow models are commonly based on fundamental physical laws and assumptions thought to describe glacier flow.

At the top end of the class of ice-sheet models are so-called three-dimensional thermomechanical models, which are able to describe the time-dependent flow and shape of real ice sheets. These models are akin to general circulation models developed in other branches of climate science. Their development closely follows technical progress in such fields as computer power, ice-core and sediment drilling, remote sensing and geophysical dating techniques, which are providing both the required calculating means and the necessary data to feed and validate such models. Models of this type have been applied to the existing ice sheets of Greenland and Antarctica, and to those which covered the continents of the Northern Hemisphere during the Quaternary ice ages. Typical studies have concentrated on mechanisms and thresholds of ice-sheet inception during the Tertiary (Huybrechts, 1994a; DeConto & Pollard, 2003), ice-sheet form and extent during glacial–interglacial cycles (Marshall *et al*., 2000; Ritz *et al*., 2001; Charbit *et al*., 2002; Huybrechts, 2002), and the response of the polar ice sheets to future climatic warming (Huybrechts & de Wolde, 1999; Van de Wal *et al*., 2001). In this context, the key interactions being investigated are between the effects of a change in climate on the accumulation and ablation fields and the ice sheet's response in terms of changed geometry and flow, including the ice sheet's contribution to the worldwide sea-level stand. Related work has considered the ice sheets as a boundary condition for other components of the Earth's geophysical system, providing changes in surface loading for isostasy and gravity models (Le Meur & Huybrechts, 2001; Tarasov & Peltier, 2004), or providing changes in freshwater fluxes for ocean models, especially to investigate changes of the thermohaline circulation of the North Atlantic Ocean (Schmittner *et al*., 2002; Fichefet *et al*., 2003). Three-dimensional thermomechanical ice-sheet models have also been used to investigate the potential for internally generated flow instability (Payne, 1995; Payne & Dongelmans, 1997; Marshall & Clarke, 1997a). In this application, the crucial interactions are between the thermal and flow regimes. In addition, models of the Greenland and Antarctic ice-sheets are in use to assist with the location and dating of ice cores (Greve, 1997b; Huybrechts *et al*., 2004a), estimating internal distributions of passive tracers such as oxygen isotope ratios (Clarke & Marshall, 2002), yield information about fields that are inaccessible for direct observation such as at the ice-sheet base (Huybrechts, 1996), or assess the component of their present-day evolution due to adjustment to past climate changes (Huybrechts & Le Meur, 1999).

In this chapter the discussion concentrates on three-dimensional whole ice-sheet models applied to the ice sheets of Antarctica and Greenland. This has the advantage that the quality of the model simulations can be assessed against available observations. Also, the range of physical characteristics, climate regimes, and flow mechanisms encountered in both polar ice sheets is probably a good representation of the range of behaviour which occurred in the palaeo-ice-sheets at various times during their evolution. The Antarctic ice sheet is located in a very cold climate, where almost no surface melting occurs and precipitation amounts are limited by low air temperature. Therefore virtually all Antarctic ice is eventually transported into floating ice shelves that experience melting or freezing at their underside and eventually break up to form icebergs. External forcing of such a cold ice sheet is mainly through changes in accumulation rate and basal melting rates below the ice shelves. The ice-sheet extent is limited mainly by the depth of the surrounding ocean and by the capacity of the ocean to float the ice shelves, touching upon the crucial issue of grounding-line dynamics. An important characteristic of the West Antarctic ice sheet is that it is a marine ice sheet, resting on a bed far below sea level. Much of its ice transport towards the coast occurs in ice streams, which are distinct fast-flowing features that rest on smooth sedimentary beds, have very flat surface profiles, and are sharply bordered by relatively stagnant ice at their sides. Their ice flux is dominated by basal flow, mostly through deformation of the underlying water-saturated sediments (Alley *et al.*, 1986a). By contrast, the Greenland ice sheet is situated in a much warmer climate, with a temperature difference of 10–15°C in the annual mean. Summer temperatures are high enough to initiate widespread summer melting. All around the ice-sheet margin, mean annual ablation exceeds the accumulation. A negative surface budget results at elevations below about 1000 m in the north and 1600–1800 m in the southwest. High coastal temperatures do not favour ice shelves, but there are a few along the north and northeast coast. The Greenland ice sheet loses mass by calving of icebergs, mostly at grounding lines in a tidewater environment, and by meltwater runoff from the surface, in roughly equal shares.

80.2 Building a three-dimensional thermomechanical ice-sheet model

Planform time-dependent modelling of ice sheets largely stems from early work by Mahaffy (1976) and Jenssen (1977), extending on the pioneering Antarctic study of Budd *et al.* (1971). These papers develop work by Nye (1957) on what has become known as the shallow-ice approximation (Hutter, 1983). This approximation recognizes the disparity between the vertical and horizontal length scales of ice flow, and implies grounded ice flow by simple shear. This means that the gravitational driving stress is balanced by shear stresses and that transverse and longitudinal strain rate components are neglected. Although the assumption is not valid at all places in the ice sheet, such as at the ice divide or near the ice-sheet margin (Baral *et al.*, 2001), it has shown general applicability in large-scale ice-sheet modelling as long as surface slopes are evaluated over horizontal distances an order of magnitude greater than ice thickness. Under the shallow-ice approximation, both components of the horizontal velocity can be represented as algebraic functions of the local ice geometry (surface slope and ice thickness), which greatly simplifies the numerical solution. The model by Mahaffy (1976) was vertically integrated and was developed as a computer program to find the heights of an arbitrary ice sheet on a rectangular grid. It incorporated Glen's flow law (Glen, 1955) for polycrystalline ice deformation by dislocation creep. That is an empirical relation derived from laboratory tests in analogy with the behaviour of metals at temperatures near to their melting point, and is most commonly used in ice flow modelling. It considers ice as a non-Newtonian viscous fluid, relating strain rates to stresses raised mostly to the third power. However, in ice the rate of deformation for a given stress is also a strong function of temperature. For the range of temperatures found in natural ice masses (−50°C to 0°C), the effective viscosity changes by more than three orders of magnitude. The first model that dealt with the flow–temperature coupling in a dynamic fashion was developed by Jenssen (1977). Jenssen introduced a stretched vertical coordinate, transformed the relevant continuity and thermodynamic equations, and presented a framework to solve the system numerically.

At the heart of three-dimensional thermomechanical models is the simultaneous solution of two evolutionary equations for ice thickness and temperature, together with diagnostic representations of the ice velocity components. These express fundamental conservation laws for momentum, mass and heat, supplemented with a constitutive equation (the flow law). Plate 80.1 shows the structure of one such model as it was described in Huybrechts (1992), and further refined in Huybrechts & de Wolde (1999) and Huybrechts (2002). This model was first developed for the Antarctic ice sheet. It solves the thermomechanically coupled equations for ice flow in three subdomains, namely the grounded ice sheet, the floating ice shelf, and a stress transition zone in between at the grounding line. The flow within the three subdomains is coupled through the continuity equation for ice thickness, from which the temporal evolution of ice-sheet elevation and ice-sheet extent can be calculated by applying a flotation criterion. Grounded ice flow is assumed to result both from internal deformation and from basal sliding over the bed in those areas where the basal temperature is at the pressure melting point and a lubricating water layer is present. Ice deformation in the ice-sheet domain results from vertical shearing, most of which occurs near to the base. For the sliding velocity, a generalized Weertman relation is adopted, taking into account the effect of the subglacial water pressure. Ice shelves are included by iteratively solving a coupled set of elliptic equations for ice-shelf spreading in two dimensions, including the effect of lateral shearing induced by sidewalls and ice rises. At the grounding line, longitudinal stresses are taken into account in the effective stress term of the flow law. These additional stress terms are found by iteratively solving three coupled equations for depth-averaged horizontal stress deviators. Calving of ice shelves is ignored. Instead, the ice shelves extend to the edge of the numerical grid but this has little influence on the position of the grounding line. The temperature dependence of the rate factor in Glen's flow law is represented by an exponential Arrhenius equation.

The distinction between ice-sheet flow and ice-shelf flow was also made in the Antarctic model developed by Ritz *et al.* (2001), but they introduced an additional subdomain representing a 'dragging ice shelf' to incorporate ice-stream dynamics. In their model, inland ice is differentiated from an ice stream zone by the magnitude of basal drag. This is based on the observation that ice stream zones are characterized by low surface slopes and thus low driving stresses, but yet have fast sliding, as is the case at the Siple Coast in West Antarctica. Ritz *et al.* (2001) treat these zones as semi-grounded ice shelves by replacing the shallow ice approximation by the set of ice-shelf equations to which basal drag is added. Gross model behaviour turns out to be quite similar to that of the Huybrechts model, except that the West Antarctic ice sheet has a lower surface slope near the grounding line and that the break in the slope now occurs further upstream at the place where the dragging ice shelf joins the inland ice subject to the shallow ice approximation. One consequence is that grounding-line retreat in the Ritz model occurs more readily in response to rising sea levels, and that the West Antarctic sheet contains less additional ice for an expanded grounding line.

Three-dimensional flow models applied to the Greenland ice sheet have been developed along similar lines, except that ice flow is only considered in the grounded ice domain and ice shelves are not dealt with (Ritz *et al*., 1997; Greve, 1997a; Huybrechts & de Wolde, 1999). The position of the calving front in these models is not predicted from a self-consistent treatment of calving dynamics, as its physics are poorly understood and a convincing calving relation does not exist. Instead, these models prescribe the position of the coastline, beyond which all ice is removed as calf ice. Another characteristic of most Greenland models concerns the incorporation of variable ice fabric in the flow law to account for different ice stiffnesses as established for Holocene and Wisconsin ice in Greenland ice cores. This is achieved by prescribing a variable enhancement factor in the rate factor of the flow law, and necessitates the simultaneous calculation of ice age to track the depth of the Holocene/Wisconsin boundary. The model developed by Greve (1997a) furthermore incorporates polythermal ice and considers the possibility of a temperate basal ice layer, in which the water content and its impact on the ice viscosity are computed.

In whole ice-sheet models it is necessary to take into account the isostatic adjustment of the bedrock to the varying ice load. Early studies considered a damped return to local isostatic equilibrium (e.g. Oerlemans & van der Veen, 1984) but in most recent models the bedrock model consists of a rigid elastic plate (lithosphere) that overlies a viscous asthenosphere. This means that the isostatic compensation not only considers the local load, but integrates the contributions from more remote locations, giving rise to deviations from local isostasy. For an appropriate choice of the viscous relaxation time, this treatment produces results close to those from a sophisticated self-gravitating spherical visco-elastic earth model, while at the same time being much more efficient in terms of computational overhead (Le Meur & Huybrechts, 1996). Another feature common to most thermomechanical models is the inclusion of heat conduction in the bedrock, which gives rise to a variable geothermal heat flux at the ice-sheet base depending on the thermal history of the ice and rock.

Interaction with the atmosphere and the ocean in large-scale ice-sheet models is effectuated by prescribing the climatic input, consisting of the surface mass-balance (accumulation minus ablation), surface temperature, and, if applicable, the basal melting rate below the ice shelves. Changes in these fields are usually heavily parametrized in terms of air temperature. Precipitation rates are often based on their present distribution and perturbed in different climates according to temperature sensitivities derived from ice cores or climate models. This is principally because of a lack of a better method and implies that interaction between the pattern of precipitation and an evolving ice sheet cannot be accounted for properly. Meltwater runoff, if any, is usually obtained from the positive degree-day method (Braithwaite & Olesen, 1989; Reeh, 1991). This is an index method providing the bulk melting rate depending on air temperature only, but is very efficient in its use and generally gives very acceptable results (Ohmura, 2001). Models of this type are usually driven by time series of regional temperature changes (available from ice-core studies) and by the eustatic component of sea-level change, relative to present values.

Ice-sheet models are typically implemented using finite-difference techniques on a regular grid of nodes in the two horizontal dimensions, and using a stretched co-ordinate system in the vertical. Horizontal grid resolutions are mostly in the range of 20 to 50 km with between 20 and 100 layers in the vertical, concentrated towards the base where the bulk of the velocity shear takes place. Finite element implementations exist (e.g. Hulbe & MacAyeal, 1999) although these are often performed on a regular grid (Fastook & Prentice, 1994). Recent model applications have used much improved compilations of crucial input data such as bed elevation that became available on high-resolution grids from the BEDMAP (Lythe *et al*., 2001) and PARCA projects (Bamber *et al*., 2001a; Gogineni *et al*., 2001).

80.3 Model applications

Three-dimensional models of the Antarctic and Greenland ice sheets have been used to address two main issues: the expansion and contraction of these ice sheets during the glacial–interglacial cycles, and the likely effects of greenhouse-induced polar warming.

80.3.1 Model validation

Before models can have any predictive capabilities, it is necessary to confirm that they are a realistic representation of the real-world system. One often distinguishes between the steps of calibration and validation. The usual practice is to first vary a few adjustable parameters to give a qualitatively best fit with observations. This mainly concerns the multiplier (‘enhancement factor’) in the rate factor of the flow law and/or the basal sliding parameter, which are chosen to give a good representation of the present-day ice-sheet configuration, preferably after spinning up over the glacial cycles as the ice sheets are currently not in steady state. Often it is also necessary to adjust the values of the degree-day factors in the melt-and-runoff parametrization in order to have the modelled ice margin coincide as closely as possible with its observed location. Another parameter available for tuning is the geothermal heat flux. Its value is not very well defined and therefore can be adjusted to obtain a good fit with measured borehole temperatures where these are available. The problem with using observations of ice thickness and basal temperature to calibrate a model is that these fields are strictly speaking no longer available to validate the model. This problem is difficult to avoid because of the paucity of suitable test data (Van der Veen & Payne, 2004, who prefer the term ‘confirmation’ rather than ‘validation’), leaving only velocity as an independent field for model verification (or confirmation). Another way of confirming time-dependent models is to compare simulations of past behaviour against the geological record but this procedure is also by itself not fully conclusive as different parameter combinations may yield the same result, and the geological record is often ambiguous. Nevertheless, these seem to be the only options at hand to assess the performance of current ice sheet models.

Plate 80.2 shows examples of fundamental output fields available for model testing. As far as can be judged from available data, the predicted fields of vertically averaged velocity and basal temperature look very reasonable. In Greenland most of the base in central areas appears to be frozen to bedrock, with homologous basal temperatures typically between 4 and 8°C below the pressure melting point. Temperate ice is mainly confined to the coastal region and a number of fast-flowing outlet glaciers where dissipation rates are highest. More widespread basal melting also occurs in the northeastern and central-western parts of the Greenland ice sheet. In Antarctica, bottom ice at pressure melting point is widespread in both West and East Antarctica. Both heat dissipation at the base and the insulating effect of thick interior ice in combination with low vertical advection rates play a role here. Also for Antarctica, basal temperatures can be verified in only a few deep boreholes. However, the pattern of pressure melting shown in Plate 80.2 generally can be well correlated with radio-echo sounding data indicative of basal water (Siegert, 2000). The coolest basal layers are found above the Gamburtsev Mountains and the fringing mountain ranges, where the ice is thinnest. It should, however, be kept in mind that a strong control on basal temperatures is exerted by the geothermal heat flux, a parameter already used for tuning the model. Moreover, its value is known to have a large spatial variation and this is not included in current models because of lack of data.

Perhaps the best independent test to confirm models is to compare their velocity fields with observations. This option has not been fully exploited yet. Remote sensing techniques using satellite-derived information make it possible to obtain good representations of surface velocities, but ice-sheet-wide maps have so far not been published. At this stage, modelled velocities can be compared with so-called balance velocities. These are also modelled velocities based on the assumption of stationary downhill flow, which are believed to be correct to within 25% of reality (Bamber *et al*., 2000a). Gross comparison of the overall patterns of modelled velocities in Plate 80.2 with balance velocities (Budd & Warner, 1996; Joughin *et al*., 1997; Huybrechts *et al*., 2000) is certainly favourable, but the details differ. In particular the flow concentration in narrow outlet glaciers and ice streams does not occur to the same degree in the modelled fields. This is a matter of model resolution, but also the diffusive properties of the model physics and the numerical scheme, insufficient basal flow, and the neglect of additional terms in the force balance play a role. Features such as the northeast Greenland ice stream and the flow over Lake Vostok are missing altogether from the fields shown in Plate 80.2, chiefly because the specific mechanisms thought responsible for their formation are not included in the models.

80.4 Glacial cycle simulations

80.4.1 The Antarctic ice sheet

Long integrations of the Antarctic ice sheet during the last glacial cycles were analysed in Budd *et al*. (1998), Huybrechts & de Wolde (1999), Ritz *et al*. (2001) and Huybrechts (2002). Plate 80.3 (left panel) shows the evolution of key glaciological variables over the last four glacial cycles in a typical run with the Huybrechts model, with forcing derived from the Vostok ice core (Petit *et al*., 1999) and the SPECMAP sea-level stack (Imbrie *et al*., 1984). In line with the generally accepted view, volume changes are largely concentrated in the West Antarctic and Peninsula ice sheets. These are caused by a repeated succession of areal expansion and contraction of grounded ice close to the continental break during glacial maxima. Around the East Antarctic perimeter, grounding-line advance was limited because of the proximity of the present-day grounding line to the continental shelf edge. In these models, glacial–interglacial fluctuations are mainly controlled by changes in the global sea-level stand and dynamic processes in the ice shelves. This supports the hypothesis that the Antarctic ice sheet basically follows glacial events in the Northern Hemisphere by means of sea-level teleconnections. Typical glacial–interglacial volume changes correspond to global sea-level contributions of about 20 m. Freshwater fluxes originating from the Antarctic ice sheet are an important output because of their role in modulating the deep-water circulation of the ocean. Model predictions displayed in Plate 80.2 show that these are fairly constant in time and are almost entirely dominated by the iceberg flux. During the last two glacial–interglacial transitions meltwater peaks occurred about three times larger than the normal background fluxes. During interglacials, melting from below the ice shelves is also an important contribution but surface runoff always remained negligible.

According to the model, surface elevations over most of West Antarctica and the Antarctic Peninsula were, at the Last Glacial Maximum (LGM), up to 2000 m higher than at present in direct response to the grounding-line advance (Plate 80.4). Over central East Antarctica, surface elevations at the LGM were 100–200 m lower because of the lower accumulation rates (Huybrechts, 2002). A characteristic of this model is that most of the Holocene grounding-line retreat in West Antarctica occurs after 10 kyr BP and lags the eustatic forcing by up to 10 kyr. This behaviour is related to the existence of thresholds for grounding-line retreat, and to the offsetting effect of the late-glacial warming leading to enhanced accumulation rates and a thickening at the margin. The late timing is in line with recent geological evidence (Ingolffson *et al*., 1998; Conway *et al*., 1999) and is supported by some interpretations of relative sea-level data (Tushingham & Peltier, 1991), but other inferences have been made. The implication is an ongoing shrinking of the Antarctic ice sheet at the present time equivalent to a global sea-level rise of about 2.5 cm per century (Huybrechts & de Wolde, 1999). An important unknown regarding the glacial history of the West Antarctic ice sheet is whether widespread ice-streaming comparable to the present Siple Coast continued to exist at LGM, in which case surface elevations may have been substantially lower than shown in Plate 80.4, and the contribution to the global sea-level lowering was less by perhaps several metres (Huybrechts, 2002).

80.4.2 The Greenland ice sheet

Similar results from glacial cycle simulations of the Greenland ice sheet are shown in Plate 80.3 (right panel) and Plate 80.5. Here the temperature forcing was assembled from the GRIP $\delta^{18}O$ record (Dansgaard *et al*., 1993) for the most recent 100 kyr, and from the Vostok record (Petit *et al*., 1999) for the period before that to circumvent known defects in the GRIP record during the last interglacial. The most conspicious feature over the last two glacial cycles concerns the fate of the Greenland ice sheet during the Eemian interglacial, when temperatures peaked up to 7°C higher than today and global sea levels are believed to have been 6 m higher. At this time, the model indicates that massive marginal melting caused the ice sheet to shrink to a central-northern dome that existed together with small pockets of residual mountain glaciation over the southeastern highlands. Nevertheless, the ice sheet did not disappear entirely, as evident from the retrieval of pre-Eemian ice from central Greenland ice cores. This behaviour was confirmed by Cuffey & Marshall (2000), although Huybrechts (2002) found that the Greenland minimum is not very well constrained: for plausible combinations of climate conditions and only small shifts in the duration and magnitude of the peak warming, the Eemian ice sheet could have varied from just a little smaller than today to only a small single dome in central-north Greenland. During glacial periods, when melting is unimportant, the Greenland ice sheet expanded beyond the present coastline to cover most of the continental shelf, with an implied contribution to the LGM sea level lowering of about 3 m (Plate 80.5). The evolution of freshwater fluxes displayed in Plate 80.2 shows an interesting contrast with the Antarctic ice sheet. Whereas iceberg calving is also the dominant component during glacial periods, most of the interglacial retreat is caused by surface runoff. A striking feature of the temporal evolution of freshwater fluxes are the recurrent meltwater peaks of up to two times larger than the present-day surface runoff during glacial times. These can be correlated with the warm interstadials punctuating the Dansgaard–Oeschger events and with the warm interval which occurred prior to the Younger Dryas cold period.

80.4.3 Response of the polar ice sheets to future climatic warming

80.4.3.1 Response during the 21st century

Three-dimensional ice-sheet modelling studies all indicate that on time-scales less than a century the direct effects of changes in the surface mass-balance dominate the response. This means that the response is largely static, and thus that the ice flow on this time-scale does not react much to changes in surface mass balance. Greenland studies by Van de Wal & Oerlemans (1997) and Huybrechts & de Wolde (1999) found that ice-dynamics counteract the direct effect of mass-balance changes by between 10 and 20%. The mechanism arises because surface slopes at the margin are steepened in response to the increased melting rates. This causes the ice to flow more rapidly from the accumulaton to the ablation zone, leading to a dynamic thickening below the equilibrium line. The higher surface level of the ablation zone in turn leads to less melting than would be the case if ice dynamics were not included. Because of its longer response time-scales, the Antarctic ice sheet hardly exhibits any dynamic response on a century time-scale, except when melting rates below the ice shelves are prescribed to rise by in excess of $1\,\mathrm{m\,yr^{-1}}$ (O'Farrell *et al*., 1997; Warner & Budd, 1998; Huybrechts & de Wolde, 1999).

These responses should be considered in addition to the long-term background trend as a result of ongoing adjustment to past environmental changes as far back as the last glacial period. The IPCC Third Assessment Report estimates the latter contribution to be between 0 and $0.5\,\mathrm{mm\,yr^{-1}}$ of equivalent sea-level rise for both polar ice sheets combined (Church *et al*., 2001a). Three-dimensional modelling studies which analyse the imbalance pattern resulting for the present-day in glacial cycle simulations typically find a long-term sea-level evolution of between 1 and 4 cm per century for Antarctica but a negligible contribution of only a few millimetres per century for Greenland (Huybrechts & de Wolde, 1999; Huybrechts & Le Meur, 1999). Another component to the current and future evolution of ice sheets are the effects of 'unexpected ice-dynamic responses' which may or may not be related to contemporary climate changes, and which find their origin in variations at the ice-sheet base or at the grounding line. Examples are the measured thinning of the Pine Island and Thwaites sectors of the West Antarctic ice sheet (Shepherd *et al*., 2002), the oscillatory behaviour of the Siple Coast ice streams (Joughin *et al*., 2002), or the surging behaviour of some Greenland outlet glaciers (Thomas *et al*., 2000a). Such mechanisms are hard to predict and currently are not incorporated in any large-scale model of the polar ice sheets.

Plate 80.6 shows an example of a series of ice-sheet simulations predicting 20th and 21st century volume changes. Boundary conditions of temperature and precipitation were in these experiments derived by perturbing present-day climatologies according to the geographically and spatially dependent patterns predicted by the T106 ECHAM4 model (Wild *et al*., 2003) for a doubling of CO_2 under the IS92a scenario. To generate time-dependent boundary conditions, these patterns were scaled with the area-average changes over the ice sheets as a function of time for available AGCM results. Typically, mass-balance changes cause a Greenland contribution to global sea level rise of +2 to +7 cm between 1975 and 2100, and an Antarctic contribution of between −2 and −14 cm. This differential response is because increased marginal melting on Greenland is predicted to outweigh the effect of increased precipitation, whereas a warmer atmosphere over Antarctica is expected to lead to more precipitation, but still negligible surface melting. For the majority of the driving AGCMs, the Antarctic response is larger than for Greenland, so that the combined sea-level contribution from mass-balance changes alone is negative. However, when the background trend is taken into account, the sea-level contribution from both polar ice sheets is not significantly different from zero (Huybrechts *et al*., 2004b), strengthening earlier conclusions that Antarctica and Greenland may well balance one another on a century time-scale.

The results shown in Plate 80.6 were used as the base for the IPCC TAR projections of sea-level rise from the polar ice sheets. To do that, they were regressed against global mean temperature to enable further scaling to take into account the complete range of IPCC temperature predictions for the most recent SRES emission scenarios. Taking into account the background evolution and various sources of uncertainties, this yielded a predicted Antarctic contribution to global sea-level change between 1990 and 2100 of between −19 and +5 cm, which range can be considered as a 95% confidence interval (Church *et al.*, 2001). For Greenland, the range was −2 to +9 cm. Most of this spread came from the climate sensitivity of the forcing AGCMs, and less from the emission scenario or the uncertainty in the ice-sheet models. These numbers should be compared with the predicted contributions to 21st century sea level rise of between +11 and +43 cm from thermal expansion of the sea water and of between +1 and +23 cm from melting of mountain glaciers and small ice caps, based on the same set of AGCMs. Taking into account all sources and uncertainties, the IPCC TAR predicts a sea-level rise from 1990 to 2100 of between 9 and 88 cm, with a central estimate of 48 cm (Church *et al.*, 2001).

80.4.3.2 Response during the third millennium and beyond

Beyond the 21st century, the approximate balance between both polar ice sheets is, however, unlikely to hold. If greenhouse warming conditions were to be sustained after the year 2100, the picture is expected to change drastically. In particular the Greenland ice sheet is very vulnerable to a climatic warming. For an annual average warming over Greenland of more than about 2.7°C, mass-balance models predict that ablation will exceed accumulation (Huybrechts *et al.*, 1991; Janssens & Huybrechts, 2000). Under these circumstances, the ice sheet must contract, even if iceberg production is reduced to zero as it retreats from the coast. For a warming of 3°C, the ice sheet loses mass slowly and may be able to approach a new steady state with reduced extent and modified shape if this results in less ablation. For greater warming, mass is lost faster and the Greenland ice sheet eventually melts away, except for residual glaciers at high altitudes. Two powerful positive feedbacks may accelerate the melting process: lower ice-sheet elevations lead to higher surface temperatures, and land-surface changes from ice to tundra further increase summer temperatures (Toniazzo *et al.*, 2004). Huybrechts & de Wolde (1999) find the Greenland ice sheet to contribute about 3 m of sea level rise by the year 3000 for a sustained warming of 5.5°C. For a warming of 8°C, they calculate a contribution of about 6 m. Greve (2000) reports that loss of mass would occur at a rate giving a sea-level rise of between 1 mm yr^{-1} for a year-round temperature perturbation of 3°C to as much as 7 mm yr^{-1} for a warming of 12°C. Gregory *et al.* (2004) have investigated the development of Greenland's temperature using IPCC scenarios in which atmospheric CO_2 stabilizes at different levels over the next few centuries. They find that the 2.7°C threshold is passed in all but one of 35 combinations of AOGCM and stabilization level; the warming exceeds 8°C in many cases and continues to rise after 2350 for the higher concentrations. The conclusion is that the Greenland ice sheet is likely to be eliminated over the course of the next millennia, unless drastic measures are taken to curb the predicted warming. Even if atmospheric composition and the global climate were to return to pre-industrial conditions, the ice-sheet might not be regenerated, implying that the sea-level rise could be irreversible (Gregory *et al.*, 2004; Toniazzo *et al.*, 2004).

On centennial to millennial time-scales, Antarctic model predictions demonstrate how several mechanisms depending on the strength of the warming come into play. For warmings below about 5°C, runoff remains insignificant and there is hardly any change in the position of the grounding line (Huybrechts & de Wolde, 1999). For larger warmings, however, significant surface melting occurs around the ice-sheet edge and basal melting increases below the ice shelves, causing the ice shelves to thin. When rapid ice-shelf thinning occurs close to the grounding line, grounding-line retreat is induced. In large-scale ice-sheet models, this occurs in two ways: steeper gradients across the grounding zone cause larger driving stresses, and higher deviatoric stress gradients across the grounding zone lead to increased strain rates, and hence a speed-up of the grounded ice and subsequent thinning. In the model studies performed by the Australian group (Budd *et al.*, 1994; O'Farrell *et al.*, 1997; Warner & Budd, 1998), large increases in bottom melting are the dominant factor in the longer-term response of the Antarctic ice sheet, even for moderate climate warmings of a few degrees. Budd *et al.* (1994) found that without increased accumulation, the increased basal melt of 10 m yr^{-1} would greatly reduce ice shelves and contribute to a sea-level rise of over 0.6 m by 500 yr, but no drastic retreat of the grounding line. With a similar model but different climatic forcing, O'Farrell *et al.* (1997) find a sea-level rise of 0.21 m after 500 yr for a transient experiment with basal melt rates evolving up to 18.6 m yr^{-1}. In the study by Warner & Budd (1998), a bottom melt rate of 5 m yr^{-1} causes the demise of WAIS ice shelves in a few hundred years and removal of the marine portions of the West Antarctic ice sheet and a retreat of coastal ice towards more firmly grounded regions elsewhere over a time period of about 1000 years. Predicted rates of sea-level rise in these studies are up to between 1.5 and 3.0 mm yr^{-1} depending on whether accumulation rates increase together with the warming. Although these are large shrinking rates, obtained under severe conditions of climate change, they cannot be considered to support the concept of a catastrophic collapse or strongly unstable behaviour of the WAIS, which is usually defined to mean its demise within several centuries, implying sea-level rises in excess of 10 mm yr^{-1} (Oppenheimer, 1998; Vaughan & Spouge, 2002). It should, however, be noted that the mechanics of grounding-line migration are not fully understood, and that none of these three-dimensional models adequately includes ice streams, which may be instrumental in controlling the behaviour and future evolution of the ice sheet in West Antarctica.

80.5 Conclusions and future outlook

Three-dimensional ice-sheet modelling significantly contributes to a better understanding of the polar ice sheets and their interactions with the climate system. Current models available to the community are able to predict the spatial and temporal ice-sheet response to changes in environmental conditions with increasing confidence. Large-scale models perform best over interior portions of continentally-based ice sheets, where ice deformation is well understood, obeys a simple force balance, and can be reliably modelled taking into account the flow law of ice. In some instances, when the basal ice has developed a strong fabric, making the ice anisotropic, or when crystal properties have introduced gradients in hardness, the resulting effects usually can be handled satisfactorily by prescribing a (variable) enhancement factor in the flow law.

Shortcomings in these models require further investigation in two main fields: incorporation of more appropriate physics and incorporation of improved boundary data. In particular basal sliding, marine ice dynamics and iceberg calving remain problematic. These processes are not easily quantified and are typically highly parametrized. Fast glacier conditions at the base are poorly understood, and so is the development of ice streams in marine-based ice sheets. Processes related to bed roughness, till rheology, and basal water pressure are all thought to be important elements but a realistic basal boundary condition for use in numerical models has not yet been developed. A credible treatment will need to include subglacial hydrology and the geological controls on soft-sediment deformation. The physics of grounding-zone migration is subgrid scale and is also not yet portrayed reliably in current models. In the grounding zone, a change takes place between flow dominated by basal stress to a basal stress-free regime, with flow primarily driven by longitudinal extension rather than vertical shear deformation. The spatial scale over which this transition takes place is unclear, however, and is therefore included in current models in an *ad hoc* way, if at all. The classic example where many of these problems converge is the Siple Coast area of the West Antarctic ice sheet, which is characterized by extensive ice streaming, low surface slopes, and a seemingly smooth transition into a floating ice shelf. Iceberg calving may be an even greater challenge to model in large-scale treatments. Calving at marine margins is related to fracture dynamics and temperature and stress fields in the ice, but the process is not well understood and therefore impossible to model with confidence. A proper treatment of calving is nevertheless warranted because ice-front degradation into bordering marine waters is the dominant means of ablation in Greenland tidewater glaciers and Antarctic ice shelves.

Although ice-sheet evolution is sensitive to several glaciological controls, long-term variability is dictated by climate and mass-balance related boundary conditions. Uncertainties in parametrizations have a large impact, particularly with respect to ice-sheet ablation. The mass-balance calculation is also sensitive to model resolution, as topographic detail is important in high-relief areas and ablation is typically concentrated in a narrow band at the ice-sheet margin. Present-day atmospheric boundary conditions such as mean annual air temperature and snow accumulation are known to a level of accuracy commensurate with that required by ice-sheet models but their patterns of change in past as well as future climates are poorly constrained. Even more troublesome is the melt rate from the underside of the ice shelves, which may affect grounding lines but for which we have very limited data. The same is also true of the geothermal heat warming at the ice-sheet base, which exerts a crucial control on the spatial extent of basal melting, but for which there are very few data.

Current developments in large-scale ice-sheet modelling mainly occur along two lines: incorporation of ice-sheet models in climate or earth system models of varying complexity, and refinements of the ice dynamics at the local scale using higher-order representations of the force balance. Interactive coupling of ice-sheet models with atmosphere and ocean models enables mass-balance changes over the ice sheets to be prescribed more directly and the effects of circulation changes to be dealt with more properly. If the coupling is two-way, the approach can additionally take into account the effect of ice-sheet changes on its own forcing. Such coupled experiments have only just begun but are likely to highlight interesting behaviour. Recent examples include the effects of freshwater fluxes on the circulation of the North Atlantic Ocean (Schmittner *et al*., 2002; Fichefet *et al*., 2003), the enhancing effect of ice-dammed lakes on ice-sheet growth (Krinner *et al*., 2004), or the climate feedbacks resulting from Greenland ice-sheet melting (Ridley *et al*., 2004). A second line of current research concerns the nesting of detailed higher-order flow models (e.g. Pattyn, 2003) into large-scale models to study the flow at high spatial resolution for which the usual assumptions made in zero-order models are known to break down. First attempts in this direction for limited inland areas near ice divides were presented in Greve *et al*. (1999) and further explored for the purpose of ice-core dating and interpretation in Huybrechts *et al*. (2004a).

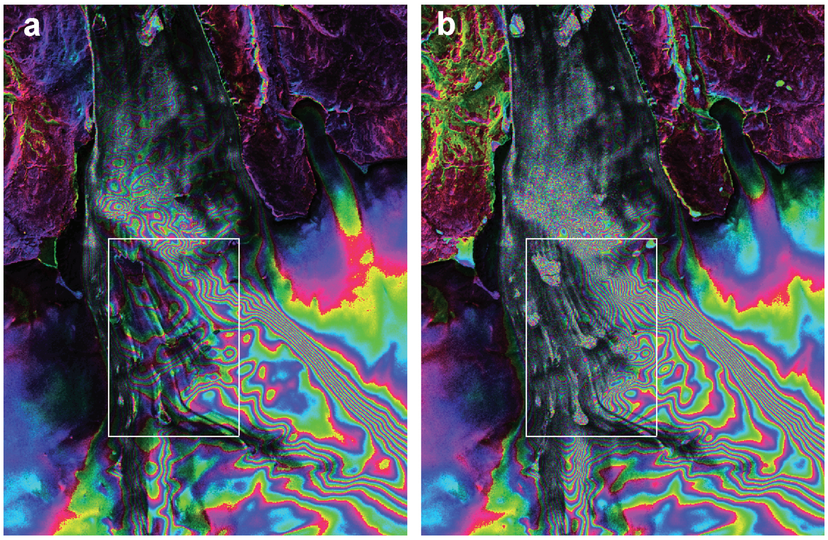

Plate 74.2 Interferograms from the fast moving area of the Ryder Glacier displayed as hue-saturation-value images with value (brightness) determined by the SAR amplitude, hue determined by the interferometric phase, and saturation held constant. Each fringe (yellow–red transition) represents 2.8 cm of displacement directed toward or away from the radar. (a) Interferogram for the interval 21–22 September 1995. (b) Interferogram for the interval 26–27 October 1995. The much denser fringes, particularly on the lower portions of the glacier (white box), indicate a dramatic change in velocity over the September observation.

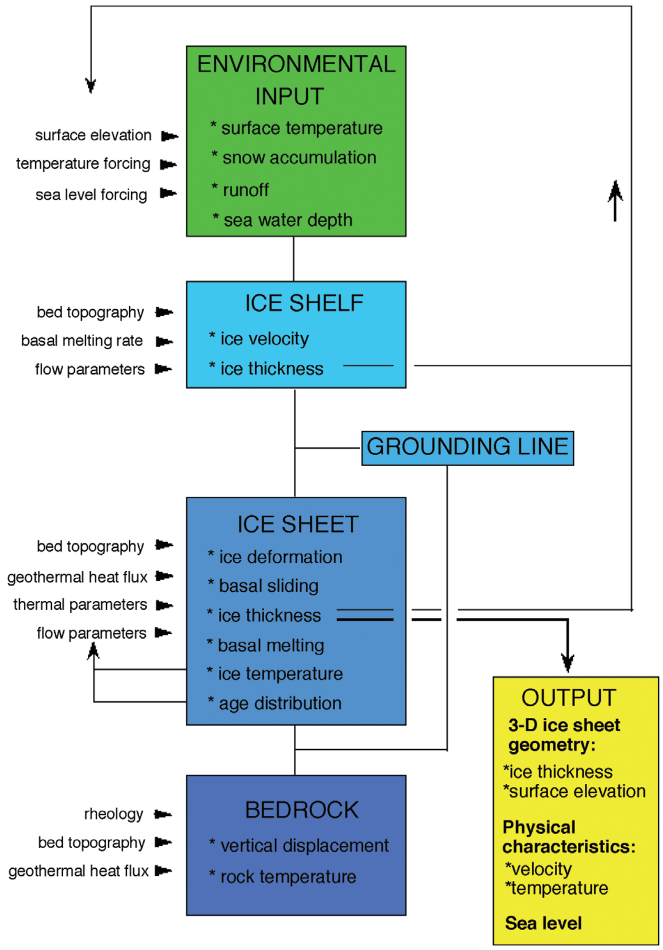

Plate 80.1 Structure of a comprehensive three-dimensional ice-sheet model applied to the Antarctic ice sheet. The inputs are given at the left-hand side. Prescribed environmental variables drive the model, which has ice shelves, grounded ice and bed adjustment as major components. The position of the grounding line is not prescribed, but internally generated. Ice thickness feeds back on surface elevation, an important parameter for the calculation of the mass balance. The model essentially outputs the time-dependent ice-sheet geometry and the coupled temperature and velocity fields. Three-dimensional models applied to the Northern Hemisphere ice sheets are similar, but do not include ice-shelf flow and explicit grounding-line dynamics. (After Huybrechts, 1992.)

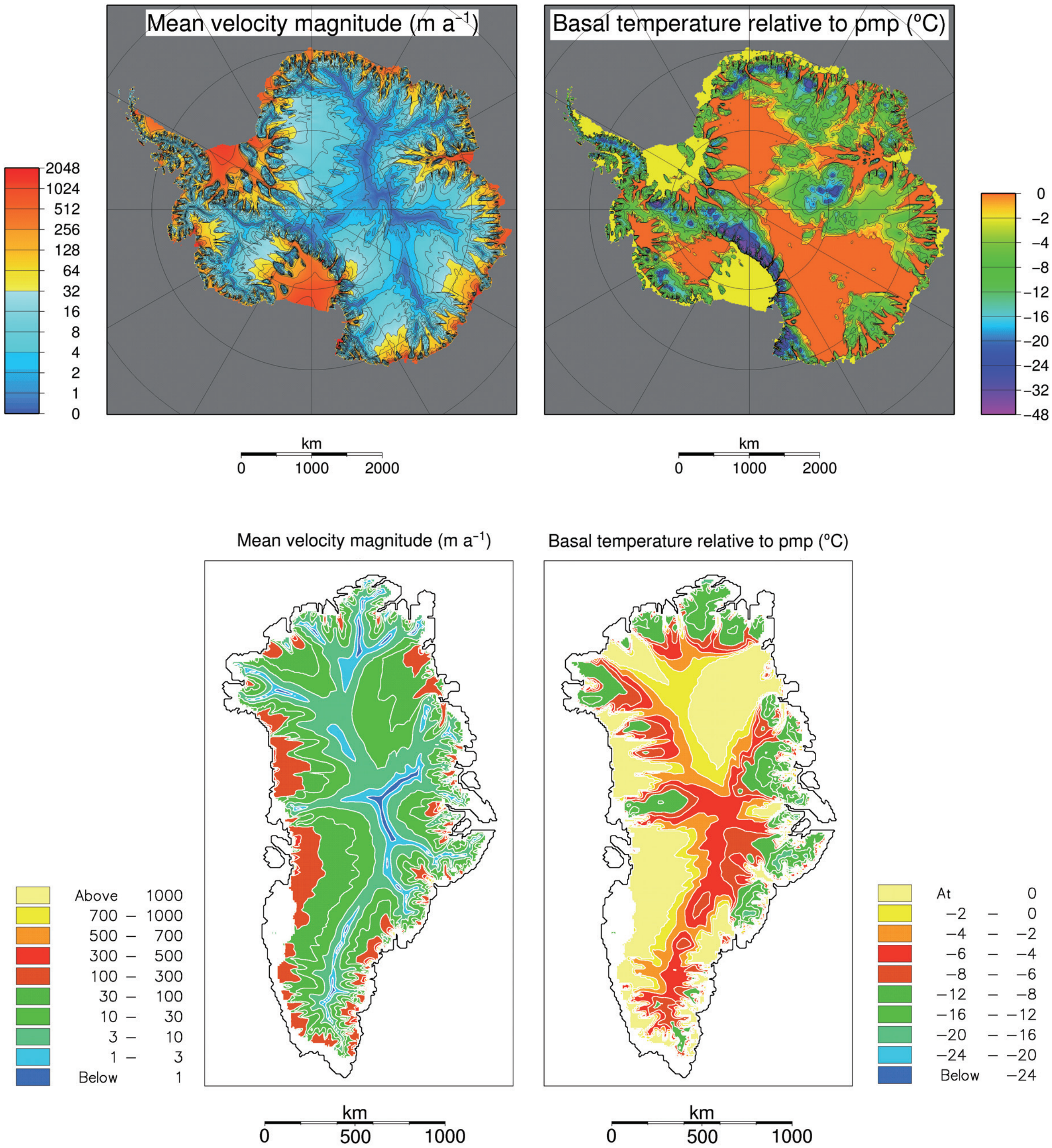

Plate 80.2 Present-day vertically averaged ice velocities and basal temperature field as simulated by three-dimensional models applied to the Antarctic (upper pictures) and Greenland (lower pictures) ice sheets. The orange, respectively yellow, colours in the pictures at the right are areas where the basal ice is at the pressure melting point and basal sliding occurs. These fields were obtained from model versions implemented at 10 km resolution spun up over several glacial cycles. Basal temperatures were obtained for a uniform geothermal heat flux of 50.4 W m^{-2} for Greenland and 54.6 W m^{-2} for Antarctica. Note that despite what the figures may suggest, the Antarctic ice sheet is about eight times larger than the Greenland ice sheet.

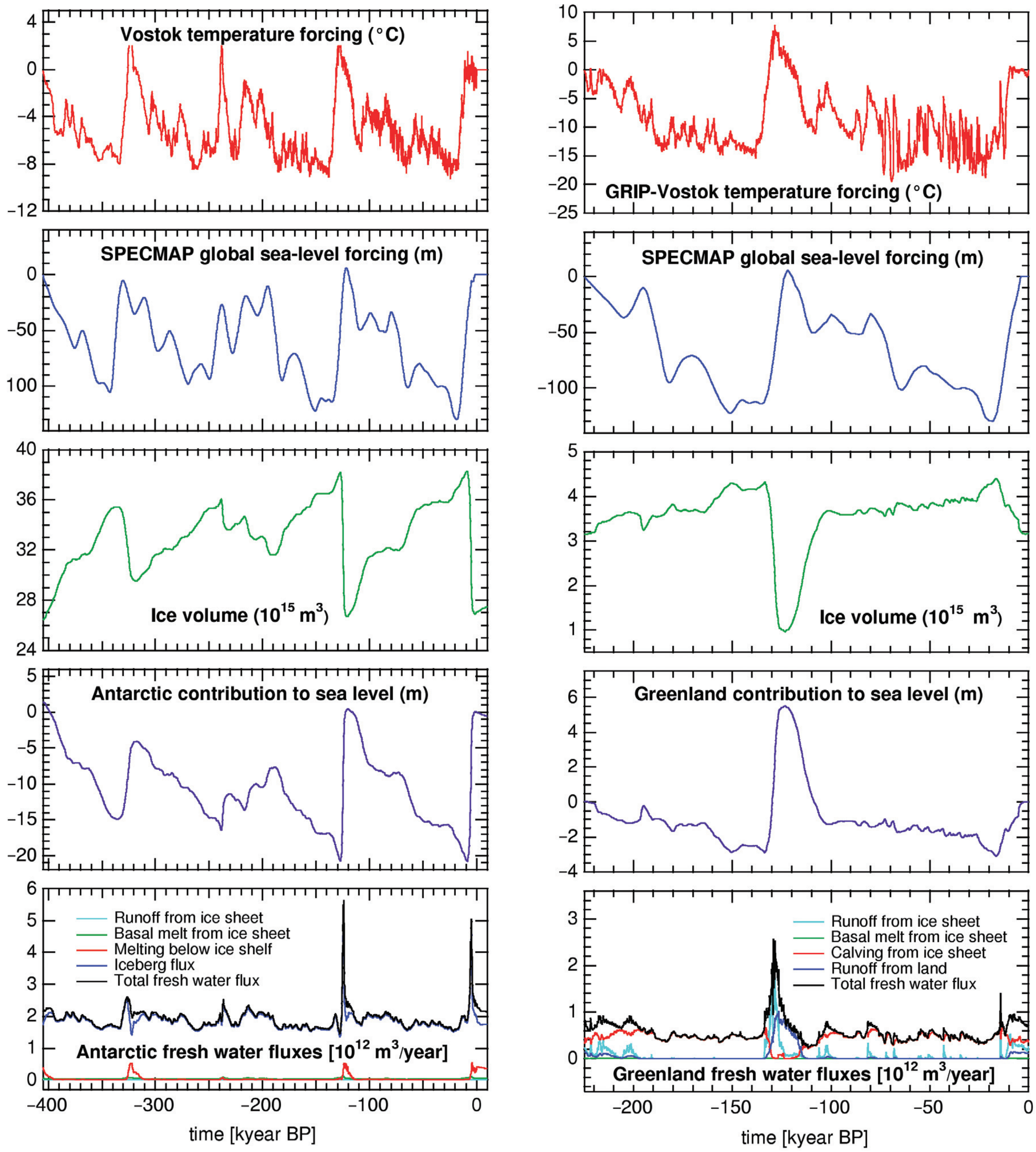

Plate 80.3 Forcing (mean annual air temperature and eustatic sea level) and predicted evolution of key glaciological variables (ice volume, contribution to sea level, freshwater fluxes into the ocean) in typical three-dimensional model experiments over the last few glacial cycles. (Based on the ice-sheet model experiments described in Huybrechts, 2002.)

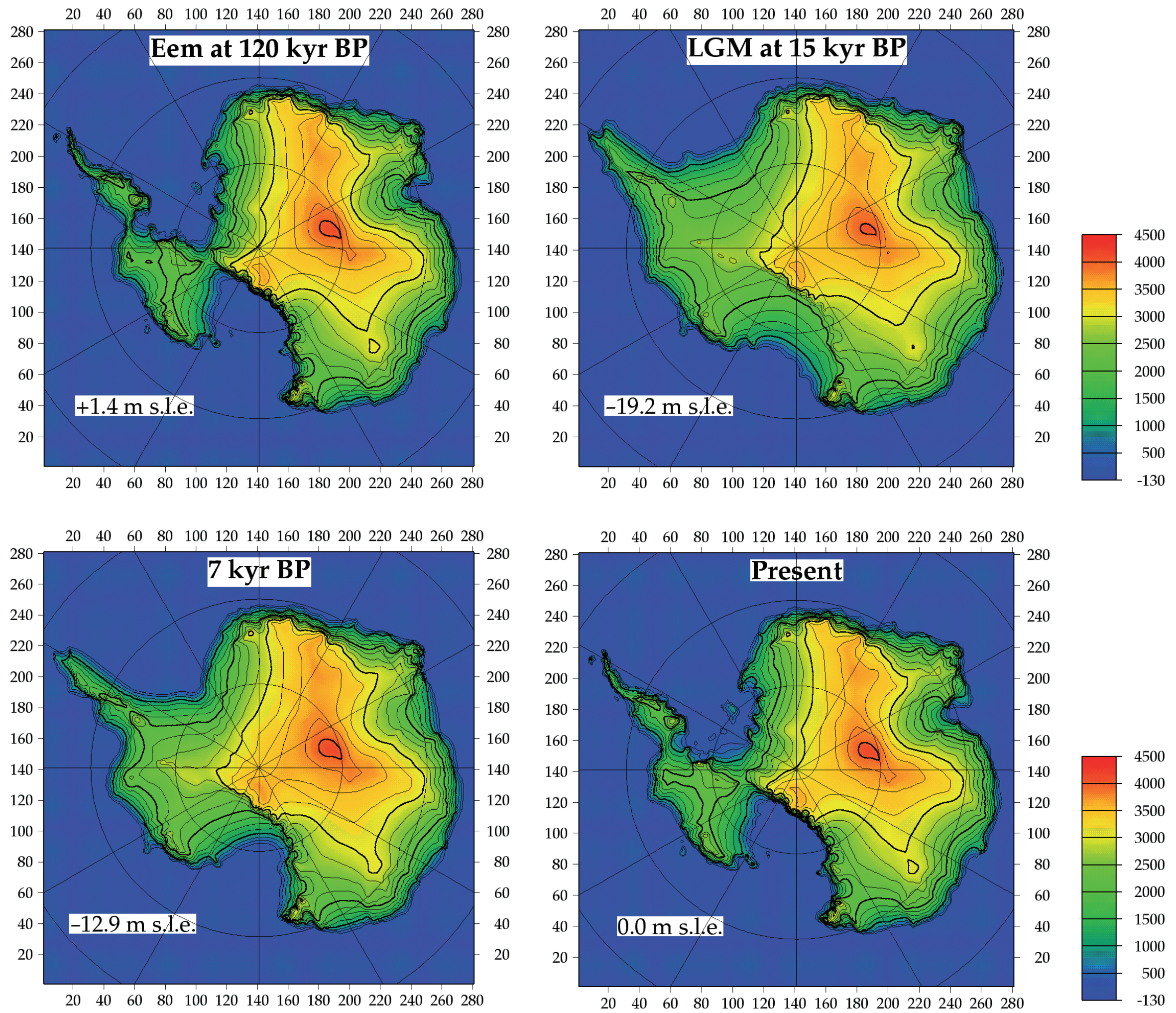

Plate 80.4 Modelled extent and surface topography of the Antarctic ice sheet at a few selected times during the last glacial cycle. In line with glacial–geological evidence, the most pronounced changes take place in the West Antarctic ice sheet. In East Antarctica, variations in ice-sheet geometry are comparably small. A main characteristic of the model is the late Holocene retreat of the grounding line in West Antarctica, still continuing today. Contour interval is 250 m; the lowest contour approximately coincides with the grounding line. (Modified after Huybrechts, 2002.)

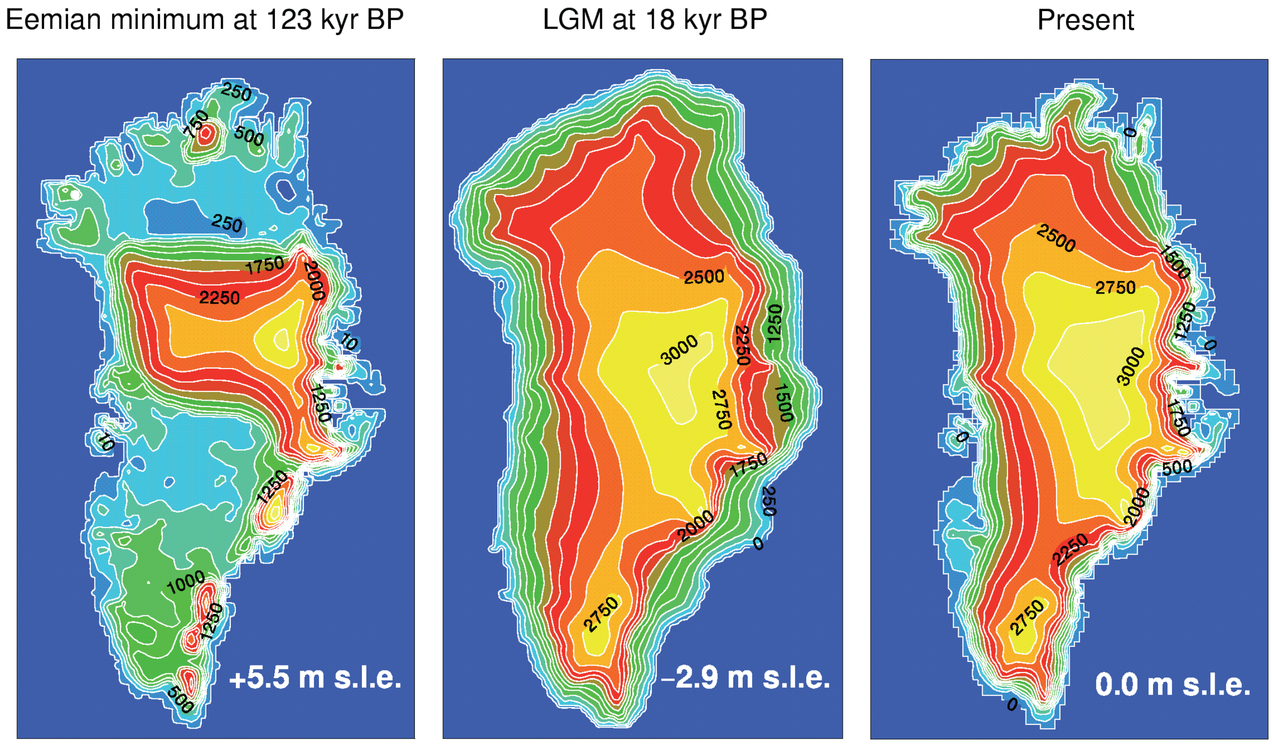

Plate 80.5 Snapshots of Greenland's ice-sheet evolution at three intervals during the last glacial cycle. According to the model, the ice sheet retreated to a small central dome during the Eemian warm period before expanding over most of the continental shelf at the Last Glacial Maximum. Implied global sea-level changes are between –3 m and +6 m. (Modified after Huybrechts, 2002.)

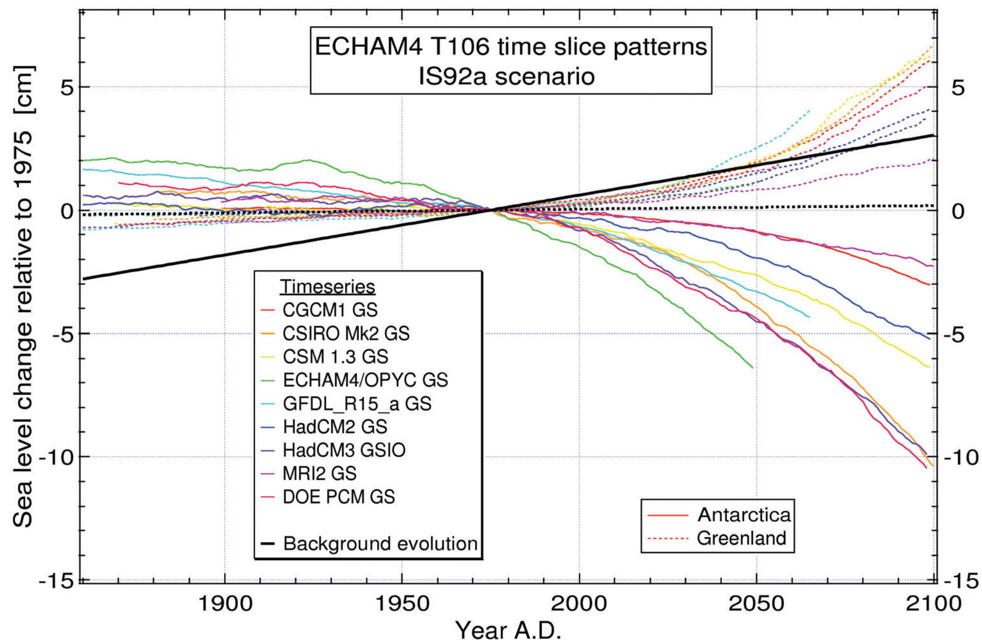

Plate 80.6 Volume changes of the Greenland and Antarctic ice sheets in greenhouse warming experiments expressed in equivalent global sea-level changes. The climatic forcing was derived from scaling time slices from a high-resolution AGCM (ECHAM4) with a suite of lower-resolution AOGCMs. On these short time-scales, the ice-sheet response is entirely dominated by the direct effect of mass-balance changes. The background trend resulting from past environmental changes is shown separately by the thick black lines. The stippled lines refer to the Greenland ice sheet; the full lines are for the Antarctic ice sheet. These experiments were at the base of the polar ice-sheet component to the global sea-level projections of the IPCC Third Assessment Report (Church *et al.*, 2001). (From Huybrechts *et al.*, 2004b.)

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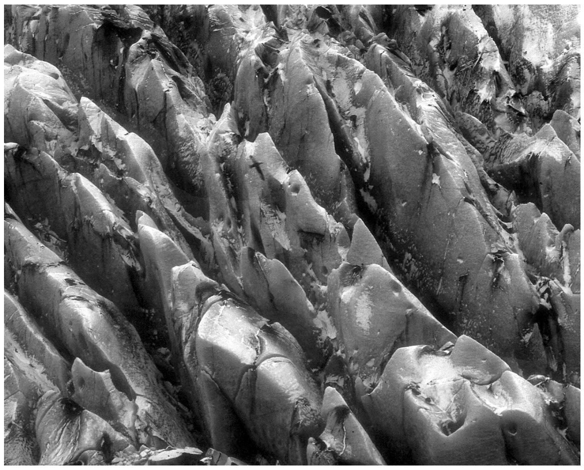

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