13 Flow and balance of the polar ice sheets

13.1 Introduction

The distribution of permanent ice masses over the globe is very uneven, but ice does occur permanently at all latitudes. Land ice covers approximately 10% of the Earth's land surface and can conveniently be grouped according to size and volume into a very small component consisting of glaciers and ice caps and into the two great ice sheets, which cover Greenland and Antarctica. Compared to the latter two, which contain over 99% of the global ice volume and roughly 80% of the Earth's fresh water resources, glaciers and ice caps seem almost negligible. They are however important factors influencing local climate variables as well as the seasonal behaviour of regional runoff and thus the hydrologic cycle.

The two large ice sheets on the other hand have a crucial impact on the global environment. Their surface is characterised by high albedo and highly effective longwave radiative properties and therefore modulate global atmospheric processes. Their high topography is an effective barrier to atmospheric circulation patterns. They lose meltwater and icebergs to the oceans and thus also influence oceanographic processes. The existence of large floating ice shelves in Antarctica is thought to play a major role in the formation of Antarctic bottom water, which in turn is an effective driver for global ocean circulation. As importantly, changes in their ice volume have a direct impact on the global sea level stand.

When considering glaciers and ice sheets in the climate context, respectively in budget terms one needs to be aware that there are varying time constants involved in their response to changes in climate variables. These time constants are first order functions of glacier or ice sheet sizes with small sizes related to rapid response. For the large ice sheets typical e-folding response time scales are of the order of 10^2 to 10^4 years. One implication of these long response times is that a true state of equilibrium is most likely never attained.

In this chapter the Greenland and Antarctic ice sheets are first discussed in terms of their physiogeographic setting and flow regime, and basic datasets documenting their geometry and surface climate are presented. Section 13.3 introduces the ice fluxes and flow velocities obtained under the assumption that the ice sheets are in balance with their current pattern of mass gain and mass loss at the surface. A discussion on the current mass budget and on the methods that have been used to determine the current evolution of the ice sheets follows in Section 13.4. The chapter closes with an outlook on how future developments in observations and modelling may ultimately contribute to confidently answer the fundamental question of the present-day balance of the Greenland and Antarctic ice sheets.

13.2 Physiogeographic setting and characteristics

13.2.1 Configuration

Present-day continental ice cover is dominated by the Antarctic ice sheet which is responsible for 90% of the volume and 85% of the area of all land ice. The Antarctic ice sheet rests on a continental land-mass centred over the pole that is almost twice as large as Australia. It is formed of several dynamically distinctive elements. The major part is the large East Antarctic ice sheet, which constitutes a vast, relatively flat dome with a maximum elevation of 4030 m and a maximum known thickness of 4776 m. Several long ridges define large-scale drainage basins which control the outward flow of the ice. Towards the coast, where the ice is thinner, the ice flow is impeded by mountain ranges, and the ice is chanelled into distinctive fast-moving outlet glaciers several hundred kilometers in length and tens of kilometers wide. Many outlet glaciers flow at speeds of hundreds of meters per year. The Transantarctic Mountains divide the East Antarctic ice sheet from the West Antarctic ice sheet and the Antarctic Peninsula.

The West Antarctic ice sheet has a more complicated configuration. Much of the ice sheet rests on bedrock below sea level and is therefore termed a marine ice sheet. It basically consists of three domes but nowhere does the ice surface rise above 2400 m. The West Antarctic ice sheet largely drains into two big ice shelves in the Ross and Weddell Seas. Into the Ross ice shelf, this drainage largely takes place through ice streams, which rest on smooth sedimentary beds, have very flat surface profiles, and are sharply bordered by relatively stagnant ice at their sides. They flow at speeds up to several km per year. The ice shelves are up to 1000 m thick and float in equilibrium with the ocean water. Smaller ice shelves fringe the Antarctic coastline elsewhere. Mass is lost from the ice shelves by bottom melting and by calving of icebergs from their ice fronts. In Antarctica, there is almost no loss of ice by surface melting.

The Greenland ice sheet is the only ice mass of significance in the Arctic and is about ten times smaller in volume than the Antarctic ice sheet. It is a relict from the ice ages that overlies a bowl-shaped continent almost completely fringed by coastal mountains. The ice sheet consists of a northern dome and a southern dome, with maximum elevations of 3230 m and 2850 m, respectively, linked by a long saddle with elevations around 2500 m. Higher coastal temperatures for Greenland do not favour ice shelves, but there are a few along the north and northeast coast. Most Greenland outlet glaciers are narrower, by an order of magnitude, than their Antarctic counterparts, but some reach speeds that are an order of magnitude larger. Jakobshavn Isbræ on the western side is the fastest glacier on Earth and exhibits velocities up to 10 km per year or 30 m per day. The Greenland ice sheet loses mass by calving of icebergs and meltwater runoff from the surface, in roughly equal shares.

13.2.2 Geometric datasets

The polar ice sheets were among the last features on the surface of the Earth to be explored and investigated during the course of the 20th century. The Greenland ice sheet is surveyed best. Because of the inaccessibility and vast size of Antarctica, significant portions of the ice sheet, especially in the interior, still remain unexplored. Here, only aircraft and satellite-borne observations are able to approach continent-wide coverage. In this section we present datasets on the form of both ice sheets as they have been compiled from various sources and were subsequently adapted for modelling the Greenland and Antarctic ice sheets on 20 km grids (e.g. [02Huy]). These datasets are also used as input to calculate the balance flow further below.

13.2.2.1 Surface elevation

Surface elevations over the ice sheets have traditionally been determined by pressure altimetry, geometrical levelling, and by combined airborne barometric altimetry and radar distance measurement. These methods had a precision of typically several tens of meters at best and had only a limited coverage. Gains in accuracy by at least an order of magnitude have been accomplished during the last two decades from satellite radar altimetry and airborne laser altimetry. The widest coverage has come from satellites, especially from ERS-1, which extends to 81.5° of latitude, covering almost all of Greenland but still leaving a gap over central Antarctica. The datasets presented in Fig. 13.1 originate from the best Digital Elevation Models (DEM's) presently available as they were compiled by different authors from most available data. Surface elevation over Antarctica derives from the Liu et al. [99Liu] DEM, based mostly on ERS-1 data but supplemented with airborne altimetry and digital cartographic data south of 81.5°S and over high surface slopes where data are available and where the radar signal is seriously degraded. Absolute vertical accuracy over the interior ice sheet is considered better than 15 m. The Greenland surface elevation dataset is based on a 2 km DEM published by Ekholm [96Ekh]. This dataset was also compiled from a wide selection of data sources, mainly from satellite radar altimetry from ERS-1 and GEOSAT, but with contributions from airborne radar altimetry, airborne laser altimetry, and photogrammetric and manual map scannings at the margin. The mean accuracy of the Greenland ice surface elevations is estimated to be 12-13 m.

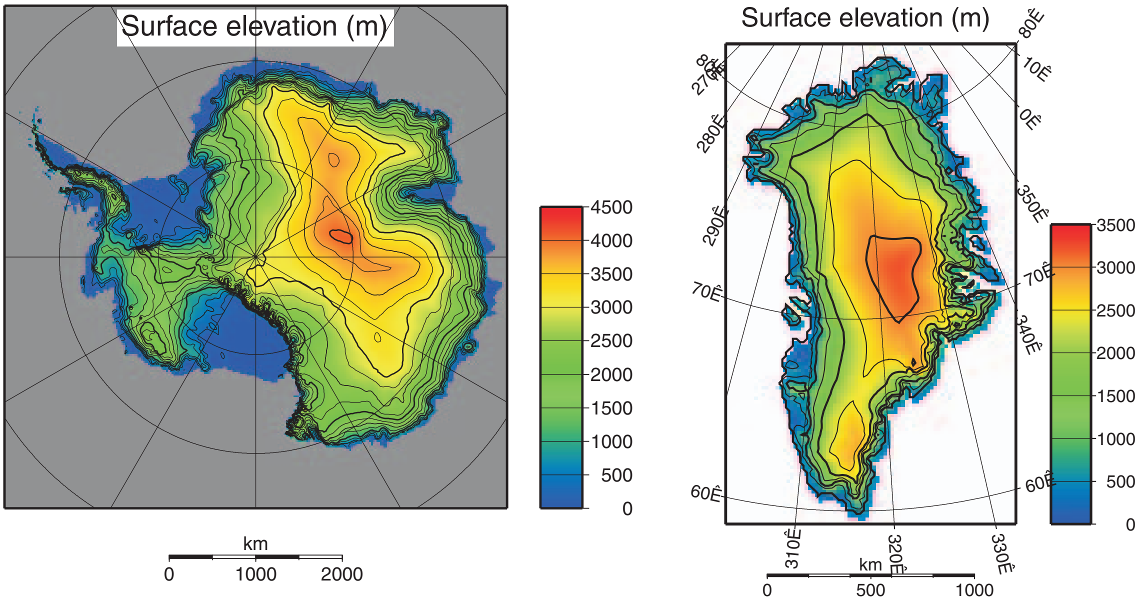

Fig. 13.1 Surface elevation of the Antarctic ice sheet (left panel) and the Greenland ice sheet (right panel).

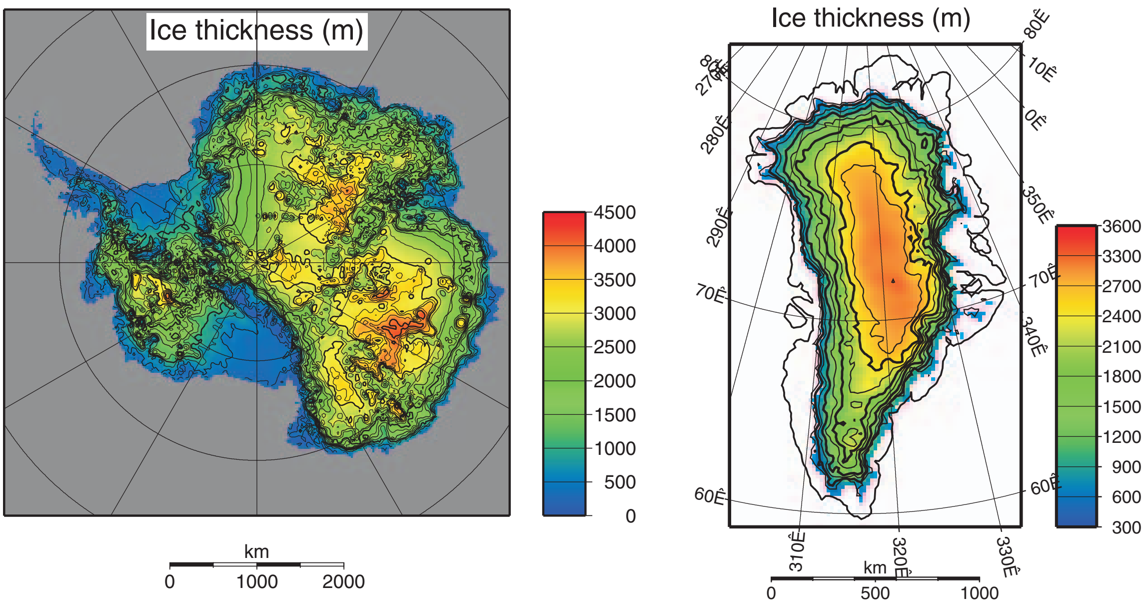

Fig. 13.2 Ice thickness of the Antarctic ice sheet (left panel) and the Greenland ice sheet (right panel).

13.2.2.2 Ice thickness

By far the bulk of ice-thickness data over the polar ice sheets result from airborne radar (radio-echo) sounding. A compilation of all the ice-thickness soundings collected in Antarctica (including the old seismic soundings of pre-radar days) was undertaken within the SCAR-sponsored BEDMAP project [01Lyt]. There are still large gaps and for roughly one quarter of the total area of Antarctica (interior East Antarctica in particular), there are no ice thickness data available. The thickness data shown in Fig. 13.2 for Antarctica are primarily based on the BEDMAP data, but were amended for part of Dronning Maud Land with more recent measurements from the EPICA pre-site surveys conducted by the Alfred Wegener Institute [01Ste]. The coverage of the Greenland ice sheet is far more comprehensive, thanks in part to the

PARCA programme [01Tho1]. The Greenland ice thickness data derive from a compilation by Gogineni et al. [01Gog] with improvements from [01Bam], as well as some other model-based corrections by one of the authors (PH) to fill gaps at the ice-sheet margin where no flight lines are available.

13.2.2.3 Bedrock elevation

The bedrock topography datasets presented in Fig. 13.3 were obtained by subtracting grounded ice thickness from surface elevation for the datasets discussed above. The result was subsequently matched with topographical datasets for ice-free regions and for the bathymetry below the ice shelves and the surrounding oceans. These maps provide a unique insight into the subice topography of both ice sheets. They reveal the remarkably flat bedrock surface below the central Greenland ice sheet and the existence of major mountain chains below the East Antarctic ice sheet, amongst many other features of large interest to better understand the geology and glaciology of these polar areas.

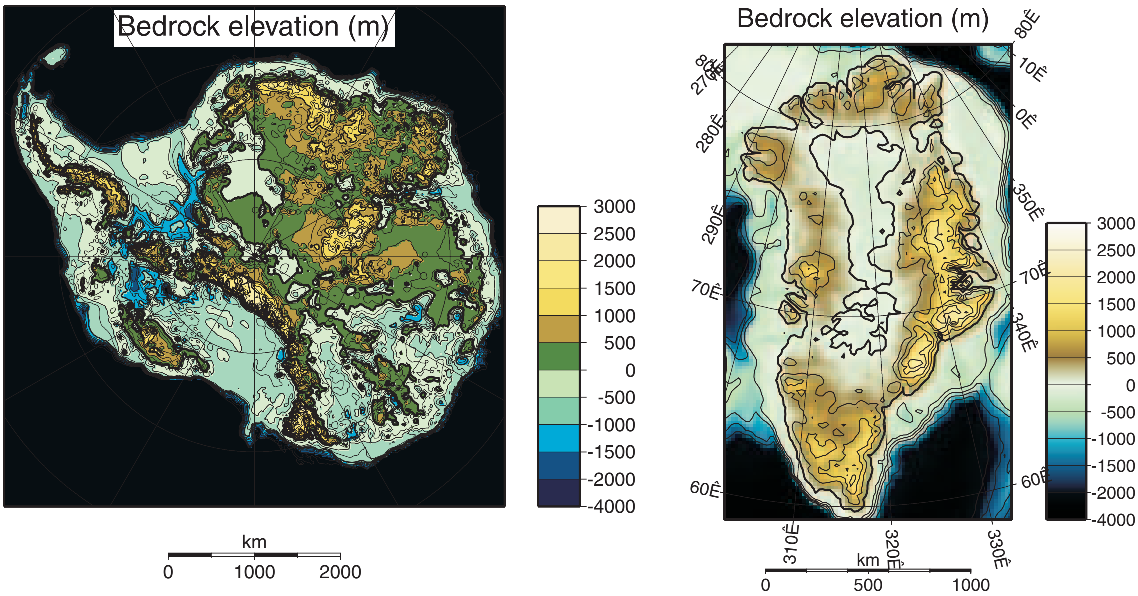

Fig. 13.3 Bedrock elevation of the Antarctic ice sheet (left panel) and the Greenland ice sheet (right panel).

13.2.3 Sizes and volumes

At present it is not yet possible to precisely determine the total permanently ice covered area because there are not enough data available. Satellite remote sensing, and especially the availability of radar imagery, has greatly improved the figures for the area, but the lack of data for ice thickness makes the estimates for total volume less reliable. However we seem to have convergence towards the figures given in Table 13.1. Based on these numbers one can derive that total melting of all glaciers and ice caps would raise global sea-level by about 50 cm. The numbers for the Antarctic and Greenland ice sheets are 61.1 m and 7.2 m, respectively.

Table 13.1 Area and volume of present-day land ice cover, quantifying the relative importance of glaciers, ice caps and the Greenland and Antarctic ice sheets ([00Huy]; [01Chu]).

	Glaciers	Ice caps	Glaciers and ice caps	Greenland ice sheet	Antarctic ice sheet
Number	> 160 000	70			
Area (10^6 km^2)	0.43	0.24	0.68	1.71	13.856
Volume (10^6 km^3)	0.08	0.10	0.18 ± 0.04	2.85 ± 5%*	26.37 ± 5%*

* figures include local glaciers and ice caps; the error estimate is an upper bound

13.2.4 Surface climate

13.2.4.1 Mean annual surface temperature

There are only very few permanently occupied weather stations in Greenland and Antarctica and they mostly are at coastal sites. In the interior of the ice sheets only recently some automatic weather stations have been established which provide temperature information. These observational datasets can however be augmented by temperature data taken in boreholes at 10 m depth. These 10 m temperatures can be taken as a very good approximation to mean annual temperatures at screen level and are routinely taken by glaciologists drilling shallow holes for ice or firn cores along traverse routes. Thus sufficient data points have become available to allow the generation of reliably interpolated maps of this climatic variable. The maps shown in Fig. 13.4 have been reconstructed from combining such interpolated representations of available meteorological and glaciological data together with parameterisations based on the same data in terms of latitude and surface elevation [99Huy]. The average error on these representations is believed to be less than 1-2 °C over both ice sheets. The observed surface temperatures tend to be up to 10 °C lower than those shown in Fig. 17.2 (Kottek and Hantel, chapter 17) for the interior of the ice sheets. Possibly the ERA-40 temperatures have been obtained for a heavily smoothed ice-sheet topography which is lower than reality, or the ERA-40 re-analysis data fail to sufficiently capture the strong surface temperature inversion observed over the East Antarctic ice sheet in particular. As evident from the maps in Fig. 13.4 there is a clear temperature contrast between the two ice sheets with a typical altitude-corrected temperature difference of about 10-15 °C in the annual mean. This temperature difference can partly be explained by the position of both ice sheets relative to the pole but also reflect the heat sink conditions of the Antarctic continent brought about by clear and dry atmospheric conditions and the thermal isolation from the rest of the climate system by the Circumpolar Ocean and the position of the polar front. In summer surface temperatures over Greenland rise well above the freezing point with the 0 °C-isotherm up at 1500 m elevation in the south and 500 m in the north [87Ohm]. Over Antarctica, mean surface temperature generally remains below freezing also at sea level during summer, with the exception of the northern tip of the Antarctic Peninsula and some ice-free areas around the perimeter of East Antarctica [84Sch]. An important implication of this climatic contrast is the occurrence of widespread summer melting on the Greenland ice sheet but the virtual absence of any meltwater runoff on the Antarctic ice sheet.

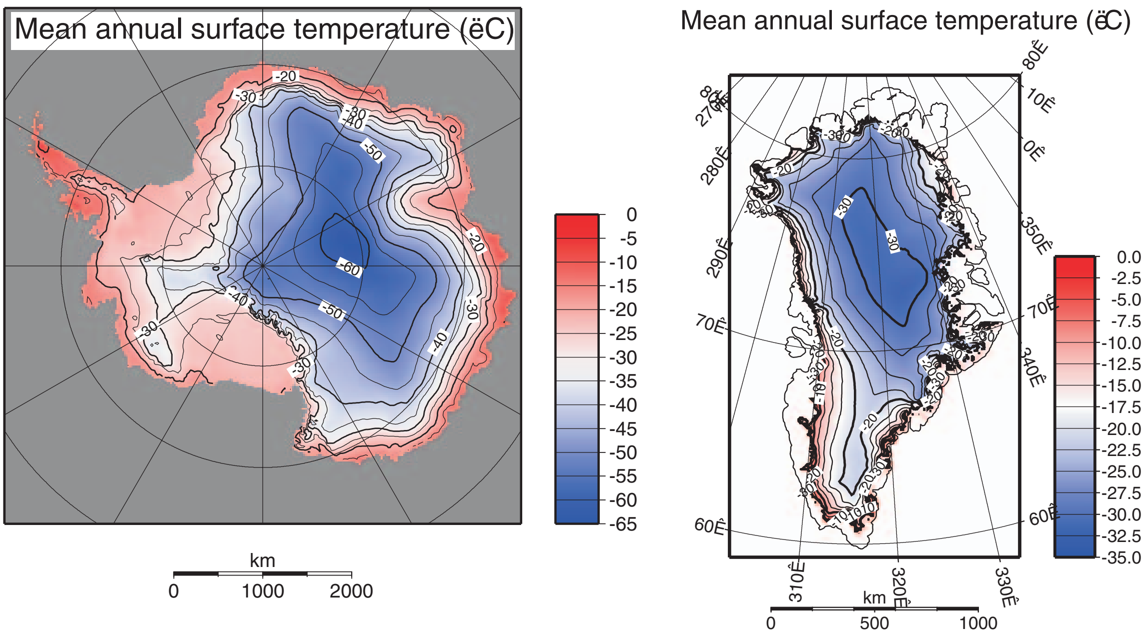

Fig. 13.4 Mean annual surface temperature over the Antarctic and Greenland ice sheets representing the climatological mean for the second half of the 20th century.

13.2.4.2 Precipitation, accumulation, and surface mass balance

The distribution of precipitation is mostly based on accumulation rates over ice-covered areas and recorded precipitation at coastal weather stations. Accumulation rates on the ice sheet are usually obtained by means of stratigraphic analyses of pit walls and of firn and shallow ice cores, but their coverage is very uneven. In such cores, absolute dating is often possible using the 1963 beta horizons resulting from the 1961 and 1962 atmospheric atomic tests, and using volcanic horizons. Other studies have derived accumulation rates from internal layering in shallow sections sounded by snow-penetrating radar [99Ric]. Accumulation rates have also been derived from satellite microwave emission data [99Vau]. This method, however, seems to work accurately only in the dry snow zones of an ice sheet, hampering its applicability in the lower parts of the Greenland ice sheet [95Zwa].

The distributions presented in Fig. 13.5 were derived from compilations by Ohmura and Reeh [91Ohm] for Greenland and from Giovinetto and Zwally [00Gio] for Antarctica. They were further amended with accumulation data for Dronning Maud Land (East Antarctica) obtained during the EPICA pre-site survey [00Huy] and the results of shallow ice cores taken during AWI oversnow traverses in north Greenland during the mid-1990s. Over Greenland, the general trend is a decrease of precipitation from south to north from about 2500 mm per year in the southeast to less than 150 mm per year in interior northeast Greenland. The southern high precipitation zone is largely determined by the Icelandic low and the resulting onshore flow which is forced to ascend the surface of the ice sheet. Similar orographic lifting of moist maritime air masses explains the high accumulation rates observed along the coast of Antarctica, in particular along the northwestern side of the Antarctic Peninsula. Away from the coast, however, much of Antarctica is a cold desert, with precipitation rates strongly limited by the moisture-carrying capacity of the overlying air. In fact, accumulation rates over the vast interior of the continent are only a few cm per year and this makes the central Antarctic plateau one of the driest places on Earth. The observed precipitation and snow accumulation distributions can be compared with the corresponding ERA-40 precipitation maps of Figure 17.118 (Kottek and Hantel, chapter 17). Apart from the heavy spatial smoothing applied to the ERA-40 data, the agreement is significantly better than for the surface temperature observations.

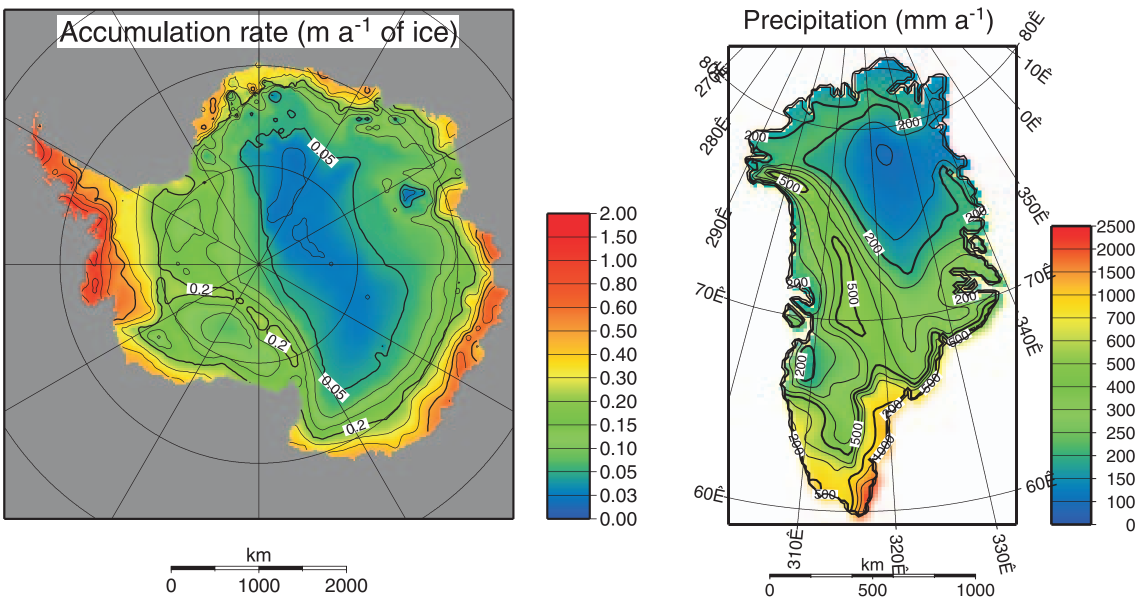

Fig. 13.5 Mean annual snow accumulation rate over the Antarctic ice sheet (in m per year of ice, left panel) and mean annual total precipitation rate over Greenland (in mm per year of water, right panel).

For Antarctica the pattern of net surface mass balance is nearly identical to the net precipitation rate (or snow accumulation rate) because significant meltwater runoff is lacking for present-day climatic conditions. Some surface melting does however occur during the summer along the coast, but most melt water soaks into the underlying snow and refreezes, and therefore does not contribute to the local surface mass balance. The Greenland ice sheet, on the other hand, has an ablation zone all around its perimeter where the mass balance is thus negative. Ablation rates have only been measured directly at a few selected sites. Since available records are short and characterised by a large interannual variability, the distribution of melt is better estimated from models. The map of net surface mass balance (accumulation minus ablation) presented in Fig. 13.6 was calculated from a melt-and-runoff model based on the positive degree-day method [00Jan]. The method takes into account the process of meltwater refeezing and meltwater retention in the snowpack by capillary forces, and distinguishes between snow melt and ice smelt and subsequent runoff. Ablation rates are highest over the southwestern part of the Greenland ice sheet, and reach typical values of the order of 5 m per year of ice equivalent. The equilibrium line which separates the ablation zone from the accumulation zone, ranges from slightly over 1500 m in the southwest to less than 1000 m along the northern coast.

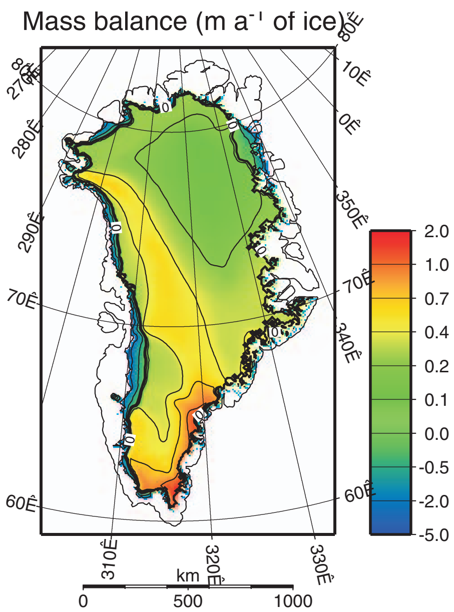

Fig. 13.6 Surface mass balance of the Greenland ice sheet as calculated from a degree-day mass-balance model.

13.2.5 Flow of ice sheets

Ice sheets and glaciers gain mass by accumulation of snow, which gradually densifies into ice, and lose mass by melting at the surface or base with subsequent runoff. Ice may also be removed by discharge into a floating ice shelf or glacier tongue. Net accumulation occurs at higher altitude, net ablation at lower altitude. To compensate for net accumulation and ablation, ice continuously flows downhill under the action of gravity. This ice flow results from internal deformation and from ice sliding over its bed where the basal temperature has reached the melting temperature and a lubricating water-saturated layer has formed. In ice sheets, the discharge at the margin is mostly chanelled in outlet glaciers and ice streams, which often lie in depressions and move much faster than the surrounding ice.

Basal sliding involves the down-slope movement of a column of ice without changing its vertical orientation. Internal deformation, on the other hand, involves vertical shear in ice velocity, with zero speed at the base and maximum speed at the surface. Whereas basal sliding depends to a large extent on the properties of the bed under the ice, internal deformation is the inherent manifestation of individual ice crystals subjected to stress. This deformation is reasonably well understood on the macro scale and can be reliably modelled taking into account the flow law of ice. In glacier ice, the response to stress is visco-elastic and thus non-linear. The rate of deformation for a given stress also depends on the temperature of the ice and the fabric of the ice. The warmer the ice, the easier it deforms. For the temperature range encountered in the polar ice sheets, three orders of magnitude are involved. In the flow law, this temperature effect is usually incorporated by adopting a temperature-dependent rate factor. Consequently, thermodynamics need to be included when modelling glacier flow. Generally, deeper layers are warmer than layers closer to the surface, and combined with greater shear stresses near the glacier base, this means that most vertical shear occurs in the lower ice layers. In some instances, the ice has developed a strong fabric, with the majority of crystal axes aligned in one preferred direction, making the ice 'soft' with respect to some stress and 'hard' with respect to other stresses. Such fabric development may influence the strain for a given stress by an additional factor 3 to 5.

Glacier flow is commonly modelled with Glen's flow law, which relates strain rates to stresses raised to the third power. Because glacier flow is sufficiently slow that accelerations can be neglected, Newton's second law of motion reduces to an equilibrium of forces. The action force is the gravitational driving stress which acts in the downslope direction. This action is opposed by resistive forces acting at the boundaries of the ice mass. These boundaries include the glacier bed (basal drag), the lateral margins (lateral drag), and the up- and down-glacial ends (gradients in longitudinal stress). In interior portions of ice sheets, the force balance is essentially between the driving stress and basal drag. In floating ice shelves there is negligible basal friction and the driving stress is balanced by gradients in longitudinal stresses and by lateral drag. Lateral drag and longitudinal stress gradients also play an important role in the fast outlet glaciers and ice streams which are responsible for the bulk of the ice discharge towards the margin. Because of the low driving stresses in the downstream portions of such ice streams, much, if not all, of the differential flow between the ice surface and the bedrock is caused by either basal sliding or by deformation of a subglacial till layer. Fast-glacier conditions at the base are however poorly understood. 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. Further discussions on the physics of ice flow can be found in texts written by [94Pat], [98Hoo], and [99Van].

13.3 Balance flow

A useful synoptic estimate of the flow field can be obtained by calculating so-called balance fluxes or balance velocities ([96Bud]; [97Jou]; [00Huy]; [00Bam]). The balance flux is the depth averaged ice flux required at any point to maintain the ice sheet in a state of balance, given a specified distribution of net surface flux. In other words, the balance flow is that flow for which the horizontal divergence of the depth-averaged ice volume flux of any vertical column equals the mass balance at the upper and lower surfaces of that column. The balance velocity is simply the balance flux divided by the local ice thickness. Thus, for a gate of width L oriented normal to the flow direction, balance velocity v_b can be calculated as:

$$v_b = \iint_S M(x,y)\,\mathrm{d}x\mathrm{d}y \Bigg/ \int_{-L/2}^{L/2} H\mathrm{d}l$$

where M (x, y) is the net surface mass balance field (accumulation minus ablation), H represents ice thickness over the gate (i.e. from $–L/2$ to $L/2$), and S is the catchment area upstream of the gate forming a closed basin with zero flux across its upstream boundaries. Apart from the assumption of stationarity, the main assumption underlying the method is that the flow is in the direction of steepest descent, which is typically calculated over a representative distance of about 10 ice thicknesses to conform with theory (e.g. [94Pat]. As input gridded datasets of surface elevation, ice thickness, and net surface balance are required. For the exercise in this book chapter, we employed the datasets presented above and used the method described in [00Huy].

The resulting balance flow fields for Antarctica and Greenland are shown in Figs. 13.7 and 13.8. A lot of detail can be seen, which primarily results from the elevation model, and to a lesser extent from the surface mass balance data. All major outlet glaciers and ice streams are clearly visible and a dendritic pattern of major flow paths can be traced well inland. While these fluxes and velocities assume steady state, and are subject to error from sparse accumulation data and limited ice thickness data, the pattern of ice flow is largely determined by surface elevations and therefore should be a very good reflection of reality. Since over most of the ice sheet the longer-term imbalance is expected to be quite small and below 10% of the local mass balance [01Chu], and an average uncertainty of 15% for the surface mass balance and the ice thickness grids can be assumed, the modelled balance flow magnitudes are believed to be correct to within 25% [00Bam]. Further comparisons with velocity fields inferred from satellite interferometry and in-situ observations are however needed to better validate the calculated patterns.

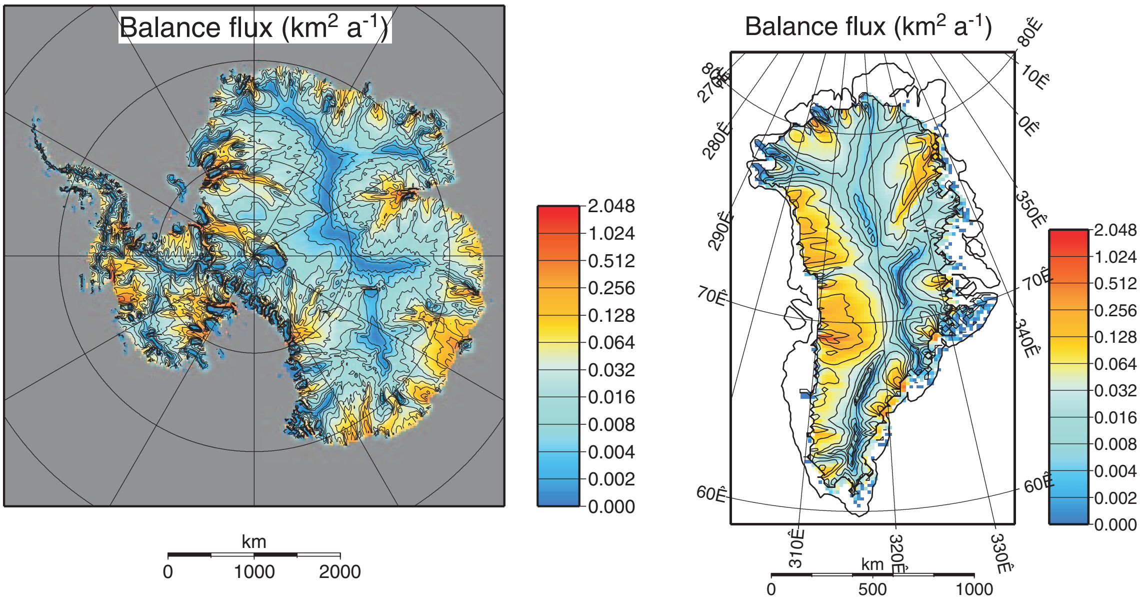

Fig. 13.7 Calculated balance fluxes over the grounded portions of the Antarctic ice sheet (left panel) and the Greenland ice sheet (right panel) employing the data presented in Sect. 13.2. They are presented in units of the transport quantity km^2 a^{-1}.

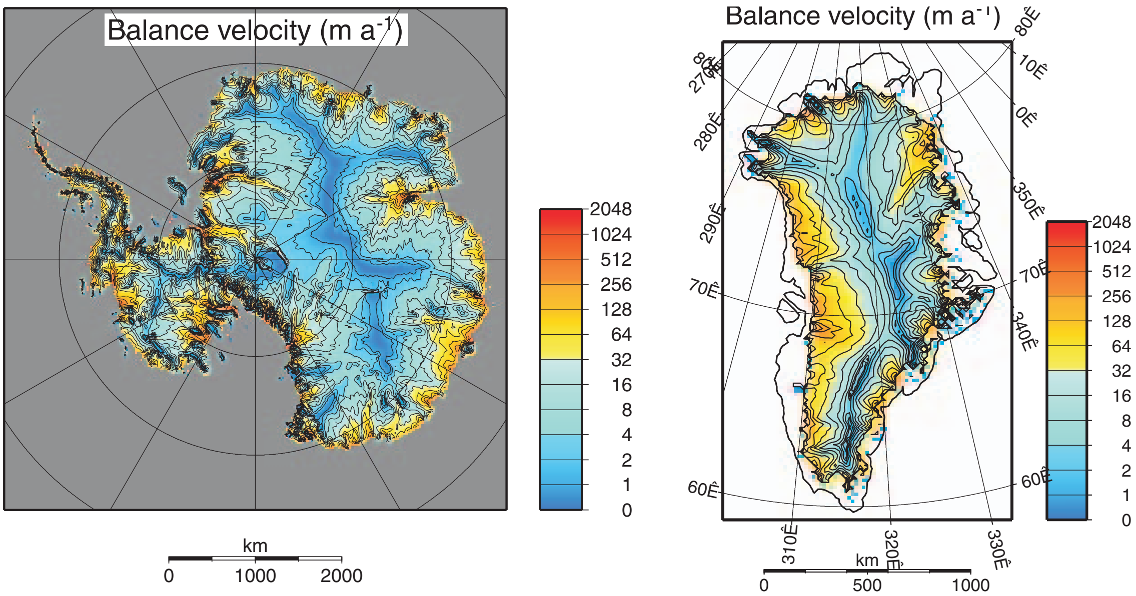

Fig. 13.8 Balance velocities calculated by dividing the balance fluxes shown in Fig, 13.7 by the ice thickness distribution shown in Fig. 13.2.

13.4 Mass budget

The average annual solid precipitation falling onto the Antarctic and Greenland ice sheets is equivalent to 5.1 and 1.4 mm of sea level, respectively [01Chu]. For glaciers and ice caps, the value is 1.9 mm per year. This input is approximately balanced by loss from melting and iceberg calving. If mass-balance terms do not balance, ice volume will change and the mass and shape of the ice sheets will adjust until a steady state is regained. Although the total volume contained in the Antarctic ice sheet is far larger than all of the other ice masses combined, the Greenland ice sheet and in particular the glaciers have a much more vigorous mass turnover, and therefore can react much more quickly to changes in their mass balance.

The traditional method to obtain the state of balance of the polar ice sheets, or parts thereof, is to estimate individual mass balance terms, and to make the budget. The basic equation, expressing the law of mass conservation, reads:

$$\frac{\partial V}{\partial a} = A - R - I - B$$

where $\partial V/\partial a$ is the annual change of ice volume, A is the annual surface accumulation, R is the annual loss by meltwater runoff, I is the annual loss by iceberg calving and B is the annual balance at the bottom of the ice shelves (melting minus freeze-on). Since ice can be considered incompressible, volume and mass are related one-to-one through a fixed ice density that is usually taken to be 910 kg m^{-3}. The mass balance is negative when the mass loss terms are larger than the mass gain term; in that case ice volume decreases with time.

Table 13.2 Current state of balance of the polar ice sheets (10^{12} kg/year). Numbers show the mean and standard deviation of published estimates of individual mass balance terms. After [01Chu].

	Accumulation	Runoff	Iceberg production	Ice shelf melting	Balance
Antarctic ice sheet	2246±86	10±10	2072±304	540±218	**–376±384**
Greenland ice sheet	520±26	297±32	235±33	32±3	**–44±53**

As shown in Table 13.2, the balance between net accumulation and attrition of ice is not the same for both ice sheets, on account of their different climatic regimes. For the Antarctic ice sheet, bottom melting and iceberg calving from the ice shelves are the major loss term, whereas for the Greenland ice sheet the surface runoff term is about as large as the iceberg production plus the ice shelf melting term. It is clear that large errors are associated with the budget method, the more so since the net balance needs to be obtained as a relatively small residual between large terms. The major error source lies with the determination of the mass loss terms, as recent accumulation estimates display a tendency for convergence towards a common value with a remaining error of perhaps only 5%. For Antarctica, there are only a few estimates available of the ice loss from the ice shelves. For sea level changes, ice discharge from the grounded ice sheet is the relevant parameter, but its value is likewise uncertain because of the difficulty of determining the position and thickness of the ice at the grounding line and the need for assumptions about the vertical distribution of velocity. For Greenland, runoff is an important term but net ablation has only been measured directly at a few locations and therefore has to be calculated from models, which have considerable sensitivity to the surface elevation data set and the parameters of the melt and refreezing methods used. Taken together, current global mass budget results suggest that the mass balance of both ice sheets lies between –35% and +5% of their mass input, or a combined imbalance equivalent to a sea-level contribution of between +2.4 and –0.2 mm/ year. This range is large compared to the central estimate of 1.5 mm/year of total sea level rise for the 20th century [01Chu], but not significantly different from zero (Table 13.2).

More recently, mass balance estimates have also been made for individual ice sheet sectors. Thomas et al. [00Tho] made a detailed comparison of the ice flux through a closed transect approximately following the 2000 m contour with the average accumulation rate derived from shallow ice cores. This revealed most of the accumulation zone to be near to equilibrium, albeit with somewhat larger positive and negative local imbalances. Both flow velocity and ice thickness are slowly changing quantities, and so the result is believed to provide a good estimate of the century time scale imbalance. A similar study comparing velocities derived from RADARSAT interferometry with a thickness map and an accumulation estimate found the Siple Coast ice streams of West Antarctica to have a positive balance of 25% of the accumulation rate [02Jou]. The implied thickening of the area was attributed to recent stoppage of Ice Stream C and slowdown of Whillans ice stream. [02Rig] looked at other drainage basins of the Antarctic ice sheet combining InSAR velocities to map grounding-line positions and outflow rates with basin-wide accumulation estimates. For 59% of Antarctica they find a slightly negative balance that is however not conclusive.

A more direct method to estimate the current evolution of the ice sheets is to monitor changes of their surface elevation. In the past ten years, important progress was achieved based on altimetric methods, both from satellites and from aircraft. Improvements of nearly two orders of magnitude have occurred in the accuracy of the localisation of satellite and aircraft platforms and the reduction of other error sources. Recent results from Greenland from 20 years of SEASAT/GEOSAT satellite radar altimetry data and 6 years of airborne laser altimetry data show a broad picture of a small thickening in the interior and a mixed pattern of a substantially larger thinning in the ablation area ([00Kra]; [01Tho2]). ERS-1/ERS-2 satellite data since 1992 indicate that much of interior East Antarctica is close to balance, but with substantially more negative trends in West Antarctica, largely located in the Pine Island and Thwaites Glacier basins ([98Win]; [02She]). Problems with current satellite data are missing data poleward of 72°N/S (SEASAT/ GEOSAT) or 82°N/S (ERS-1/ERS-2) as well as from the steeper parts at the margin, so that important areas with possibly large changes remain undetected. Another limitation is the short period over which satellite data are presently available. Altimetry records are at present too short to confidently distinguish between a short-term surface mass-balance variation and the longer-term ice-sheet dynamic imbalance. Also, altimetry only measures elevation changes, and these include a signal from both isostatic rebound and from a variable rate of snow compaction, which components need to be determined otherwise. This can be achieved by modelling efforts combining spaceborne gravity, altimetry, and GPS measurements [02Vel].

Despite all the uncertainties involved, it seems that a coherent picture of the current trend of the polar ice sheets is starting to emerge. The Greenland ice sheet appears to be close to an overall balance or is thinning slightly, but with a clear distinction between a small thickening in the accumulation zone together with a larger thinning in the ablation zone. For the Antarctic ice sheet the uncertainty is larger but the available evidence suggests that there is probably an overall negative balance. The East Antarctic ice sheet may be in balance or thickening slightly, whereas the West Antarctic ice sheet is likely thinning overall, most strongly in the Thwaites and Pine Island Glacier catchment areas but thickening elsewhere.

13.5 Outlook

Upcoming observations from spaceborne and airborne remote sensing instruments are expected to significantly reduce the uncertainty in determining the imbalances of the Antarctic and Greenland ice sheets. New data from the satellite altimetry missions ICESat (NASA) and CryoSat (ESA) will provide a detailed picture of the spatial distribution of surface elevation and its temporal evolution, and will cover polar areas beyond latitudes currently accessed. The laser altimeter on board ICESat will extend to 86° latitude and provide coverage of the steep, varying topography of the ice sheet margins where radar altimeter performance is significantly degraded. To establish long-term trends altimetry records should eventually cover several decades. However, at present the main obstacle to further geodetic constraint is the unknown snow accumulation fluctuation. Its spatial covariance is not well known, and this bears directly on the century-scale imbalance uncertainty [98Win]. It is important that recent statistics of accumulation fluctuation are obtained from extensive shallow coring of the ice sheets.

Additional constraints on the contemporary mass imbalances of the polar ice sheets are expected to come from recent and upcoming gravity field missions. GRACE and GOCE will provide the time-variant gravity field at an unprecedented spatial resolution at the scale of individual drainage basins. One problem however is the extent to which the gravity signal may be contaminated by postglacial rebound, requiring to combine the gravity data with independent observations of vertical motion.

The interferometric analysis of Synthetic Aperture Radar data from ERS1/ERS-2 and RADARSAT provides a tool for mapping the surface velocity field of the ice sheets. A comparison of these velocities with modelled balance velocities is expected to give valuable information of the mass balance status of the ice sheets. Furthermore, InSAR data should be further exploited to determine the position and flow velocities across grounding lines to improve estimates of the mass balance of individual drainage basins. More data are also needed from radio-echo sounding to improve our knowledge of the sub-ice topography in key areas.

A major benefit from the new satellite missions will be greatly improved datasets for ice-sheet model input and model validation. Surface elevation is an important boundary condition to derive the force balance and the distribution of stress and velocities with depth and is furthermore required to determine the bedrock elevation from measured ice thickness. Together with surface velocities and their directions as derived from InSAR, this constitutes a complete set of surface boundary conditions to study the dynamics of ice flow. In addition, large-scale numerical ice-sheet models have a role to play in the interpretation of the satellite observations. They provide an independent approach to determine the vertical component from isostatic rebound, which is required to transform surface elevation changes into ice thickness changes, and to transform relative sea-level changes from tide gauges into ocean level changes. They can also provide a glaciologically sound loading history which in combination with visco-elastic Earth models can be used to simulate the gravitational effects of isostatic land movements to distinguish them from ice mass changes. Improved modeling of the ice sheets during the glacial cycles, taking into account the improved boundary conditions, will lead to improved simulations of the current evolution of ice sheets. These are all elements required to increase the predictive power of ice-sheet models to better assess future changes of the Antarctic and Greenland ice sheets.

13.6 References

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