Changes in Sea Level

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SUMMARY

The purpose of this chapter is to assess the current state of knowledge regarding climate and sea level change, with special emphasis on scientific developments since IPCC (1990). The main focus is on changes that occur on the time-scale of a century. We thus look for evidence of sea level change during the last 100 years, examine the factors that could be responsible for such changes, and consider the possible changes in sea level during the next 100 years as a result of global warming.

With respect to the past, recent analyses suggest that:

- global mean sea level has risen 10–25 cm over the last 100 years. This range is slightly higher than that reported in IPCC (1990) (i.e., 10–20 cm). The higher estimate results largely from the use of geodetic and geodetic models for filtering out long-term vertical land movements, as well as from the greater reliance on the longest tide gauge records for estimating trends.

- there has been no detectable acceleration of sea level rise during this century. However, the average rise during the present century is significantly higher than the rate averaged over the last several thousand years, although century-time-scale variations of several decimetres almost certainly occurred within that longer period. The exact timing of the onset of the present, higher rate of sea level rise remains uncertain.

It is likely that the rise in sea level has been due largely to the concurrent increase in global temperature over the last 100 years. The possible climate-related factors contributing to this rise include thermal expansion of the ocean and melting of glaciers, ice caps and ice sheets. Changes in surface water and ground water storage may also have affected sea level. The assessment of the scientific evidence suggests that:

- global warming should, on average, cause the oceans to warm and expand, thus increasing sea level. The various models, from simple upwelling diffusion models to complex coupled atmosphere-ocean GCMs, all agree that oceanic thermal expansion is one consequence of global warming. The thermal expansion over the last 100 years is estimated to be 2–7 cm. Large-scale observations of changes in subsurface ocean temperatures are beginning to support these estimates.

- global warming should, on average, increase the melt rates of glaciers and ice caps, causing sea level to rise. Observational data indicates that, globally, there has been a general retreat of glaciers during this century. Based on both observations and models, recent analyses suggest that this enhanced melting may have increased sea level by about 2–5 cm over the last 100 years.

- with respect to the Greenland ice sheet, a warmer climate should increase the melt rates at the margins. The increase in melting should dominate over any increase in accumulation rates in the interior, causing sea level to rise. However, observational evidence is insufficient to say with any certainty whether the ice sheet is currently in balance or has increased or decreased in volume over the last 100 years.

- with respect to the Antarctic ice sheet, a warmer climate should increase the accumulation rates, causing sea level to fall. Here, too, the observational evidence is insufficient to say with any certainty whether the ice sheet is currently in balance or has increased or decreased in volume over the last 100 years.

- it is unclear how changes in surface water or ground water storage have affected sea level. Estimates vary widely of the net effects of activities (largely anthropogenic) such as dam construction and reservoir filling, which lower sea level, and ground water pumping, deforestation and wetland loss,
which tend to raise sea level. However, the potential future effect on sea level from such sources is probably relatively small, of the order of a few centimetres during the next century.

An exact accounting of the past sea level rise is difficult, particularly in the light of the large uncertainties associated with the mass balances of the ice sheets. However, the observed rise lies well within the combined ranges of uncertainty of the above factors.

Projections of future changes in sea level as a consequence of greenhouse-gas-induced warming were made for each of the six IPCC IS92 emission scenarios, with and without the effect of aerosol changes after 1990, for the period 1990 to 2100. In addition, high, middle and low estimates, using a range of parameter values based on key model uncertainties, were made for IS92a (the emission scenario most comparable to the IPCC (1990) Scenario A, the so-called "Business-as-usual" scenario). The results showed that:

- for Scenario IS92a, sea level is projected to be about 50 cm higher than today by the year 2100, with a range of uncertainty of 20–86 cm;

- for the range of emission scenarios IS92a–f using "best-estimate" model parameters, sea level is projected to be 38–55 cm higher than today by the year 2100;

- the extreme range of projections, taking into account both emission scenarios and model uncertainties, is 13–94 cm;

- most of the projected rise in sea level is due to thermal expansion, followed by increased melting of glaciers and ice caps. On this time-scale, the contributions made by the major ice sheets are relatively minor, but are a major source of uncertainty.

It is evident that the choice of emission scenario makes relatively little difference to the projected rise in sea level, especially for the first half of the next century. This is because much of the rise has already been determined by past changes in radiative forcing, due to lags in the response of the oceans and ice masses. For this same reason, in model simulations sea level continues to rise over many centuries even after concentrations of greenhouse gases are stabilised. In contrast, the scientific uncertainties - as reflected partly in intra-model uncertainties in the choice of individual model parameter values, and partly in inter-model uncertainties in the choice of methods for climate, glacier and ice sheet modelling - make a very large difference in the estimate of future sea level rise.

A major source of uncertainty concerns the polar ice sheets. Not only is there a lack of understanding of the current mass balance, but there is also considerable uncertainty regarding the possible dynamic responses on time-scales of centuries. Concern has been expressed that the West Antarctic Ice Sheet might "surge", causing a rapid rise in sea level. The current lack of knowledge regarding the specific circumstances under which this might occur, either in total or in part, limits the ability to quantify the risk. Nonetheless, the likelihood of a major sea level rise by the year 2100 due to the collapse of the West Antarctic Ice Sheet is considered low.

The changes in future sea level will not occur uniformly around the globe. Recent coupled atmosphere-ocean model experiments suggest that the regional responses could differ significantly, due to regional differences in heating and circulation changes. In addition, geological and geophysical processes cause vertical land movements and thus affect relative sea levels on local and regional scales. Finally, extreme sea level events - tides, waves and storm surges - could be affected by regional climate changes but are, at present, difficult to predict.

Overall, the basic understanding of climate-sea level relationships has not changed fundamentally since IPCC (1990). The estimates of global sea level rise presented here are lower than those presented in IPCC (1990), due primarily to significantly lower estimates of global temperature change which drive the projections of sea level rise. Thus, if global warming were to occur more rapidly than expected, the rate of sea level rise would consequently be higher.
7.1 Introduction

In terms of environmental and social consequences, sea level rise is arguably one of the most important potential impacts of global climate change. As in IPCC (1990), the primary purpose of this chapter is to assess what is known regarding how sea level has changed in the past and could change in the future, on time-scales of decades to centuries. The chapter begins by reviewing the evidence for trends in sea level over the past 100 years, based on tide gauge records. Next, the factors that could contribute to sea level change — namely, changes related to oceanic thermal expansion, glaciers and small ice caps, the large ice sheets of Greenland and Antarctica, and the possible changes in land-surface water and ground water storage — are examined. Future projections of global mean sea level rise are then made, on the basis of scenarios of future greenhouse gas emissions and projections of future global warming. The important factors that may cause such changes in sea level to vary spatially and temporally are considered, including geological, geophysical and dynamic effects.

Overall, this chapter finds that the major conclusions reached in IPCC (1990) remain qualitatively unchanged and are reinforced by the recent scientific literature. Nonetheless, improvements in observations and ocean-atmosphere modelling and in the calculation of future radiative forcing changes have led to revisions of the estimates of future sea level rise and their uncertainties.

7.2 How Has Sea Level Changed Over the Last 100 Years?

7.2.1 Sea Level Trends

Secular trends in “custatic” global sea level (i.e., corresponding to a change in ocean volume) over the past century have been studied by a large number of authors (as summarised in Chapter 9 and Table 9.1 of the 1990 IPCC Scientific Assessment and as described in several recent reviews — see Emery and Aubrey, 1991; Woodworth, 1993; Douglas, 1995). At this time-scale, evidence for such trends is obtained from tide gauge data.

All authors of recent global sea level change reviews have used the Permanent Service for Mean Sea Level (PSMSL) data set (Spencer and Woodworth, 1993). A central problem in identifying trends in eustatic sea level from tide gauge data is how to account for vertical land movements, which also affect relative sea level change as measured by tide gauges. At the global scale, a major source of vertical land movement derives from the continuing readjustment of the Earth’s crust since the glacial retreat marking the end of the last ice age. This “post-glacial rebound” (PGR) is the only globally coherent geological contribution to long-term sea level change about which we possess detailed understanding. Since IPCC (1990), Douglas (1991) has applied post-glacial rebound corrections from the ICE-3G model of Tushingham and Peltier (1991) to the observed tide gauge data (avoiding tide gauge records in areas of converging tectonic plates, since such processes are not represented by the PGR model) and produced a highly consistent set of long-term sea level trends. The value for the mean rate of sea level rise that he obtained from a global set of 21 such stations in nine oceanic regions with an average record length of 76 years during the period 1880–1980 was 1.8 ± 0.1 mm/year. Geodynamic models of post-glacial rebound have also been employed in global analyses by Peltier and Tushingham (1989, 1991) and Trupin and Wahr (1990), who obtained similar results. Uncertainties in global sea level trends determined from PGR models may include a range of uncertainty of approximately 0.5 mm/year, depending on the Earth structure parametrization employed by the model (Mitrovica and Davis, 1995).

Other authors have used geological data directly for sites adjacent to tide gauges, a procedure which, in principle, should accommodate other geological processes in addition to post-glacial rebound (e.g., Gornitz and Lebedeff, 1987). This sort of analysis has been conducted for the North Sea region of Europe, which has an extensive tide gauge and geological sea level data set, by Shennan and Woodworth (1992). Earlier analyses (e.g., Barnett, 1984) simply averaged tide gauge records on the implicit assumption that the effects of land movements would somehow average out. It is interesting that the results of most of these studies are within the range 1–2 mm/year, with some bias towards the lower end of the range (see Gornitz, 1995a, for a review).

Gornitz (1995a) has also pointed out that estimates using the longest records obtain the highest values, closer to 2 mm/year than 1 mm/year, and that correction for post-glacial rebound by means of the Peltier models gives higher values for the trend than those analyses which employed nearby geological data values directly. This is a result of the post-glacial rebound models producing smaller extrapolated sea level trends into the present day than the linear fit to nearby geological data procedure such as Gornitz and colleagues have employed. Hence, the corrected tide gauge trends using the post-glacial rebound models will be higher than those using the nearby geological observations. In the latter case, problems could arise from the large scatter in the geological data distributed in area around the gauge locations, and from the
linear extrapolation technique. Gornitz and Lebedeff (1987) found that in most cases a linear fit gives as good a
description of the extrapolated trend from the geological
data, as a higher order fit (see also discussion in Shennan
and Woodworth, 1992). However, this ignores the
additional information on the physics of the problem
contained in the Peltier post-glacial rebound models.

Peltier and Tushingham (1989) also commented that the
sea level trends depend strongly on the choice of minimum
record length and on the particular time interval selected.
Their sea level rise estimates, based on tide gauge record
lengths greater than 50 years, tended to lie closer to 2
mm/yr than those based on shorter record lengths. Six of
the longest tide gauge records (unadjusted for post-glacial
rebound or other geological effects) for each continent are
shown in Figure 7.1.

In summary, the best estimate based on recent analyses
is that sea level has risen about 18 cm over the last 100
years, with a range of uncertainty of 10–25 cm –
notwithstanding the fact that some authors (e.g., Gröger
and Pfleg, 1993; Pirazzoli, 1993) maintain that a global
figure for sea level rise cannot be estimated reliably,
stressing the limitations of the available data set. This
range is consistent with the 10–20 cm range given in IPCC
(1990). As geodynamic models and measurements of
vertical land movements become more reliable, confidence
in such estimates will increase.

7.2.2 Has Sea Level Rise Accelerated?
Archaeological and geological data suggest that global sea
levels have probably varied within a range of no more than a
few tens of centimetres over the past two millennia
(Flemming, 1969, 1993; Pirazzoli, 1977; Flemming and
Webb, 1986; Hofstede, 1991; Tanner, 1992; Varekamp et
al., 1992). The 10–25 cm rise over the past 100 years implies
a comparatively recent acceleration in the rate of sea level
change (Gornitz and Seeber, 1990; Shennan and
Woodworth, 1992; Gornitz, 1995b). However, the exact
timing of the onset of this acceleration remains uncertain.
The conclusions given in Woodworth (1990), Gornitz and
Solow (1991), and Douglas (1992) imply that the
acceleration probably began before the 1850s. But data for
the pre-instrumental period is sparse, at best. There is as yet
no evidence for any acceleration of sea level rise this century
(Woodworth, 1990; Gornitz and Solow, 1991; Douglas,
1992), nor would any necessarily be expected from the
observed climate change to date. Small accelerations have
been suggested over the past two or three centuries in
European sea level data (Mörner, 1973; Ekman, 1988;

The evidence, or lack of it, for sea level accelerations
over the past century depends critically on a small number of
long tide gauge records which is unlikely to be
supplemented significantly in the future. Nevertheless, the
search for old records (or “data archaeology”) can be
particularly rewarding when such data are found. For
example, Maul and Martin (1993) recently extended the
Key West, Florida data set back to 1846 and demonstrated
that no significant acceleration of sea level rise has
occurred since that time.

The main difficulties in determining a more robust
estimate for the sea level trends (let alone acceleration) are
the unequal geographical distribution of historical tide
gauge data and the considerable amount of typically
decalad variability present in all records. The former is
slowly being rectified with the development of a near-
global tide gauge network (GLOSS) and by means of
satellite radar altimetry (e.g., TOPEX/POSEIDON).

Atmospheric and oceanic dynamical processes account
for a large fraction of the interannual and interdecadal
variability. These fluctuations, 1–2 years to a decade in
duration and coherent over long distances, reflect changes
in temperature and salinity, currents, and coupled oceanic-
atmospheric forcing, such as the El Niño-Southern
Oscillation phenomenon in the Pacific Ocean (Komar
and Enfield, 1987), or the North Atlantic Oscillation in the
North Atlantic Ocean (Maul and Hanson, 1991). The
limitations imposed by large interannual sea level changes
in determining reliable sea level trends and accelerations
have been particularly emphasised by Douglas (1992)
and demonstrated by Sturges (1987) and Sturges and Hong
(1995). The understanding of such interannual processes
has been an aim of large international programmes such as
the Tropical Ocean Global Atmosphere (TOGA) project
and the World Ocean Circulation Experiment (WOCE),
and will continue to be the object of research based on the
results of a Global Ocean Observing System (GOOS). As
ocean modelling becomes more detailed, and as surface
and deep ocean observations become more routine,
increased understanding of interannual processes will
result, leading to an improved estimate of any underlying
long-term trend and acceleration.

7.3 Factors Contributing to Sea Level Change

7.3.1 Oceanic Thermal Expansion

At constant mass the volume of ocean water, and thus sea
level, varies with changes in sea water density (called
“stereic” changes). Density is related to temperature and
salinity, hence the dependency of sea level on temperature
Figure 7.1: Six long sea level records from major world regions: Takoradi (Africa), Honolulu (Pacific), Sydney (Australia), Bombay (Asia), San Francisco (North America) and Brest (Europe). Each record has been offset vertically for presentation purposes. The observed trends (in mm/yr) for each record over the 20th century are, respectively, 3.1, 1.5, 0.8, 0.9, 2.0 and 1.3. The effect of post-glacial rebound (lowering relative sea level) as simulated by the Peltier ICE-3G model is less than, or of the order of, 0.5 mm/yr at each site.
and salinity variations of the oceans. The main cause of steric change at the global scale is temperature change. As density is inversely related to temperature, the volume of the ocean must expand and sea level rise if the density of the sea water decreases because of rising temperature.

Marked changes in salinity can occur at the regional scale and modify sea water density and volume. In the short term, these effects are relatively minor at the global scale. However, the main anomalous buoyancy sources are located at high latitudes where sea water finds an effective pathway from the surface to the deeper layers of the ocean. Thus, in the longer term, regional salinity changes can affect the whole ocean circulation, the distribution of heat and oceanic expansion.

More generally, because of very complex physics, there are other important regional variations and time lags involved in sea level rise by thermal expansion. The redistribution of buoyancy from source regions to other parts of the oceans involves large-scale waves and transport by currents. The adjustment processes are very slow, because they are related to the internal modes of heat and mass redistribution within the ocean. Consequently, a sea level rise due to thermal heating must be out of equilibrium, especially for rapid climate change. This implies that the ocean will rise more rapidly in some areas than in other areas.

An important feedback mechanism between thermal expansion and increased sea level caused by increased melting of ice is linked to the above processes. Increased fresh water fluxes to the ocean have an impact on density stratification and thereby on the depths of convection of the surface waters. This, in turn, affects the circulation of the interior of the ocean and thus the rates of thermal expansion for the different layers of the world ocean.

To estimate oceanic expansion from observations, temperature and salinity have to be recorded over the water column for a long period of time. In the 1990 IPCC Scientific Assessment (Subsection 9.4.1), reference was made to the works of Roemmich (1985) concerning the Panulirus series of deep hydrographic stations off Bermuda over the period 1955–1981, and to Thomson and Tabata (1987) concerning the station PAPA steric height anomalies in the north-east Pacific Ocean over a 27-year period. It was noted that the conclusions drawn from these data sets were limited geographically and that additional ocean observations are required to discern large-scale changes. Since 1990, there has been a considerable increase in the quantity of high quality hydrography and tracer sections produced from programmes like WOCE. The analyses of these data sets in relation to previous hydrographic surveys are beginning to indicate large-scale, coherent changes in sub-surface ocean temperatures (see Chapter 3.2.4 for details). Estimates of the thermal expansion associated with some of these observed ocean temperature changes are similar in magnitude to changes in nearby sea level records (Roemmich, 1992; Salinger et al., 1996). However, these estimates are still regionally specific. Such observational consistency between thermal expansion and sea level observations will improve in the near future with the availability of the new WOCE data sets and the measurement of the ocean topography through satellite altimetry (Nerem, 1995).

The problem of a steric rise in sea level due to global warming has also been examined using a variety of models. These include upwelling diffusion models, two-dimensional models, subduction models, ocean general circulation models (OGCMs), and coupled atmosphere-ocean models (AOGCMs).

The advantage of the simpler upwelling diffusion climate models is that they are computationally efficient so that the sensitivity of results to model parameters and to uncertainties in greenhouse gas emissions can easily be examined. For example, the projections of oceanic thermal expansion made by IPCC (1990) (Warrick and Oort, 1990) were derived with the use of the simple upwelling diffusion-energy balance climate model of Wigley and Raper (1987, 1993). This model represents the vertical profile of area-averaged ocean temperature, which changes in response to changes in radiative forcing at the surface. Thermal expansion of sea water, and thus sea level change, is diagnosed from the temperature profiles through the use of expansion coefficients. For the period 1880 to 1990, the estimated range of sea level rise due to thermal expansion using this model is about 3.1 cm to 5.7 cm, with a best estimate of 4.3 cm (consistent with model parameters discussed in Chapter 6 and in Section 7.5.2), when the model is forced with the estimated radiative forcing changes from increases in greenhouse concentrations over the same period. A slightly more complex, 2-dimensional variant of this kind of model, using a prescribed ocean circulation and surface forcing based on observed sea surface temperature (de Wolde et al., 1995), leads to similar values: 2.2 cm to 5.1 cm with a best estimate 3.5 cm. The major disadvantage of these simpler models, however, is that they do not realistically represent the processes involved in the penetration and distribution of heat from the surface to deeper layers of the ocean.

With two-dimensional, zonally averaged dynamical models coupled to an energy balance climate model (as developed by Harvey, 1992), the upwelling velocities can
be related to the vertical diffusion coefficient. With such a model, Harvey (1994) has demonstrated that the 1-D upwelling diffusion models have a significantly faster surface transient response than the 2-D model, due to transient weakening of thermohaline overturning in the 2-D model which damps surface temperature warming. This difference is most pronounced if the upwelling rate in the upwelling diffusion models is constant (Raper et al., 1996).

Subduction models, while still relatively simple, introduce heat into the ocean by advection along isopycnal (constant density) surfaces. Such a model has been developed by Church et al. (1991). In this model, the estimates of the magnitude of global thermal expansion are independent of assumptions regarding the magnitudes of eddy diffusivity. Furthermore, since this model is three-dimensional, it provides information on the regional differences in sea level change adjustments. Church et al. (1991) demonstrate a small dependence of the mean sea level rise on the spatial distribution of warming, for a given global mean temperature rise. With this model, Church et al. (1991) obtain a sea level rise due to thermal expansion of about 6.9 cm over the last one hundred years. The major disadvantage of this model is that it does not predict surface temperature changes, and it also ignores deep water formation and associated changes in convective heat fluxes.

A more realistic way to simulate ocean warming and thermal expansion is to use an OGCM. One example is the study of Mikołajewicz et al. (1990), who used the Large-Scale Geostrophic (LSG) OGCM of the Max-Planck-Institute (MPI). The model was forced with time-dependent surface temperature changes caused by progressively increasing atmospheric CO₂, calculated separately from an average of the equilibrium results from several GCMs for doubled CO₂.

However, in concept, the most realistic simulations involve the use of fully coupled AOGCMs. In such experiments, changes in the three-dimensional ocean and atmosphere feed back on each other as the modelled climate evolves. Although such models are very demanding on computer resources, transient experiments are increasingly being performed with coupled models (see Chapter 6). Sea level changes have been analysed from some of them. Among these are the experiments of the Geophysical Fluid Dynamics Laboratory, Princeton (see review from Gates et al., 1992), of the MPI, Hamburg (Cubasch et al., 1992, Cubasch et al., 1994a), and of the UK Meteorological Office Hadley Centre, Bracknell (Gregory, 1993: Mitchell et al., 1995). A recent simulation using the UKMO coupled model (Mitchell et al., 1995) with historical forcing changes gives a 1880 to 1990 sea level rise due to thermal expansion of 3.6 cm – in accord with simpler models.

There is no guarantee, however, that coarse resolution AOGCMs adequately represent the processes governing water mass formation and thus the penetration of surface heat into the ocean interior. Furthermore, these models also suffer from important unsolved problems with which thermal expansion calculations are particularly sensitive. First, there is the "cold start" problem. Neglecting past changes in greenhouse gas forcing before running scenarios of future forcing changes in AOGCMs leads to significant underestimation of both future warming and sea level rise due to thermal expansion of sea water. For example, Cubasch et al. (1994a) showed that starting the MPI coupled model from 1935 instead of 1985, but using the same future forcing scenario beyond 1985, leads to estimates of 7 cm instead of 4 cm for the period 1985–2050. The GFDL model shows a much shorter time constant (about 10 years) compared to the MPI model (30 years), and different versions of the UKMO model show significant differences. A method to correct simulations for this "cold start" effect has been proposed by Hasselmann et al. (1993). Nonetheless, the issue has not been fully resolved, and the magnitude and relevance of this error for the surface air temperature and, especially, for sea level remains controversial.

Second, there is the problem of decadal variability and predictability. The initial conditions for AOGCMs are never fully known. This alone would be sufficient to prevent a prediction of the state of the ocean over several decades. To estimate the error that stems from this uncertainty, Cubasch et al. (1994a) carried out four 50-year integrations with the MPI coupled model. In all the experiments the greenhouse gas forcing started with conditions in 1986 and then followed the IPCC (1990) Scenario A. The only differences were in the initial conditions of the atmosphere and ocean. The decadal means of global warming and sea level rise showed large differences. The global mean sea level rise due to thermal expansion was 4.2 cm in the last years of integration, but in many regions of the world ocean, the local mean response had about the same size as two standard deviations of the decadal means between the runs. These problems, and others like the consequences of differences in spin-up techniques, obviously limit the reliability of the conclusions presently resulting from these coupled ocean-atmosphere model transient experiments.

Comparison of thermal expansion resulting from these various model experiments is difficult because of
differences in experimental design, in model assumptions, particularly those concerning emission scenarios and climate sensitivity, and in the complexity of processes incorporated in different models. Carefully designed intermodel comparisons have not yet been carried out.

In summary, oceanic expansion is almost certainly one consequence of global warming. Although observational data are still too sparse and regionally specific to allow global estimates of thermal expansion, analyses of recent data sets in relation to previous hydrographic surveys are beginning to indicate large-scale, coherent changes in subsurface ocean temperatures. Thermal expansion can also be predicted from a range of models, from simple to very complex. Based on model experiments, the estimated rise in sea level due to thermal expansion over the last 100 years lies within the range 2–7 cm.

7.3.2 Glaciers and Ice Caps
7.3.2.1 Processes causing change in glaciers and ice caps
The amount of land ice on Earth has fluctuated as the climate has changed. This section is concerned with the numerous mountain glaciers, ice fields and ice caps of the world, exclusive of the Greenland and Antarctic ice sheets (Figure 7.2), and their past and future contributions to changes in sea level. The potential contributions of glaciers and ice caps are not insignificant because the rates of ice accumulation and loss are more intense than those on the huge ice sheets.

The change in mass of these glaciers can be expressed as the sum of the mass balance at the surface (the difference between accumulation and loss, which can be positive or negative), plus the meltwater that is refrozen internally (internal accumulation), minus any iceberg calving flux. The change in mass of ice on land results in a change in sea level of opposite sign. The complexity stems from the fact that the surface mass balance, internal accumulation, iceberg calving flux, as well as the area and shape of the ice masses, are functions of time. The surface balance can be measured in a straightforward way, although it is a labour-intensive process (Østrem and Brugman, 1991). It can also be inferred, using hydrometeorological or climatological models, as the difference between snow accumulation (sometimes equated to winter precipitation) and snow/ice melt (often modelled using temperature or runoff as proxy); both terms are required for accurate modelling, including their seasonal differences.

Internal accumulation, on the other hand, is more difficult to measure. This term is often ignored in the modelling of glacier mass balances using climatological data, with the result that the calculated loss of water to the sea may be overestimated. Simple models to estimate the amount of internal accumulation have been developed for mountain glaciers (Trabant and Mayo, 1985) and for Arctic glaciers including Greenland (e.g., Pfeffer et al., 1991; Reeh, 1991). The rate of iceberg calving can be measured (e.g., Brown et al., 1982) but cannot be modelled with confidence.

Figure 7.2: Global distribution of glaciers (solid dots) and ice sheets (pebble pattern). Note: a single dot may represent one or many individual glaciers, and many glaciers around the margin of the Antarctic ice sheet are omitted due to lack of information.
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Changes in mass balance and/or iceberg calving produce changes in the area and surface profile of a glacier. Thus, models of long-term (more than a few years) changes in glaciers with respect to climate require incorporation of the flow processes – a dynamics model. Although these models are in common use and are of some numerical complexity, the basal sliding process for temperate glaciers is still very uncertain. A strong interaction between iceberg calving and glacier dynamics exists (Meier, 1994). The distribution of glacier mean thicknesses must also be considered because many small glaciers will likely disappear during the next century.

Internal accumulation and iceberg calving can be safely ignored in many regions, such as in much of Europe. In this case, conventional surface observations, or modelled relations, of snow accumulation and melt rates yield sufficient information for predicting volume changes as a result of climate change. In some areas of significant glacier cover, such as in the Arctic, all terms (and their time-dependencies) must be taken into consideration. This complexity, together with the paucity of observational data in many glaciated regions, makes a global synthesis of the present or future role of glaciers in sea level change a daunting task.

7.3.2.2 Changes in the last hundred years

Thinning of glaciers since the mid-19th century has been obvious and pervasive in many parts of the world (for example, see Figure 7.3). This long-term ice loss has considerable spatial and short-term variability. For instance, negative mass balances and rapid thinning have been observed in this century in the Alps (Haebeler and Hoelzle, 1995) and in south-central Alaska but not in the Canadian Arctic (Fisher and Koerner, 1994). Significant positive spatial correlations in annual mass balances are found between some glaciated regions, such as Kamchatka and the Northern Rockies (Trupin et al., 1992), while negative correlations are found between other regions, such as Washington State and Alaska (Walters and Meier, 1989).

Table 7.1 shows improved estimates of the volume of glaciers and ice caps of the world, obtained by using data from the World Glacier Inventory (IAHS/ICSI/UNEP/UNESCO, 1989). This publication presents the numbers of glaciers increasing by size categories from less than 0.03 to greater than 1024 km$^2$ for about 45 well-inventoried regions. Unfortunately, this includes only a small fraction of the world’s glacier cover. For the purposes of the present report, these data were extended to a global estimate of ice cover according to size category

![Figure 7.3: Cumulative mass balances, in metres of water equivalent, for the glaciers Hintereisferner (Austria), Rhône (Switzerland), Sarennes (France), South Cascade (United States), Storbreen (Norway), and Storglaciären (Sweden). These are among the few glaciers with long observational time series that have been extended using well calibrated hydrometeorological models. All values are relative to 1890.]

by: (1) selecting 12 regions with very complete inventories that were typical of the world’s glaciated regions; (2) adding an estimate for coastal Alaska and adjacent Canada using U. S. Geological Survey data on 321 large glaciers to extend the inventory, and extending these data by power-law scaling to smaller glacier sizes; (3) assembling numbers of glaciers in each size class and total area for the 13 regions; (4) assigning each of the 31 regions defined by Meier (1984) to one or more of these typical inventory regions; (5) weighting these regions according to the 31-region areas, (6) summing the above. Using a power law regression of known glacier area/volume relationships (e.g., Macheret et al., 1988; Chen and Ohmura, 1990; Meier, 1993), the mean thicknesses for glaciers in each size category were estimated. On this basis, the total volume of glaciers and ice caps is estimated to be about 180,000 km$^3$, and the mean thicknesses range from 7 m to 655 m. Much of the glacier area and volume is comprised of glaciers in the 100 to 1000 km$^2$ size classes.
Table 7.1: Some physical characteristics of ice on Earth. Accuracy is better than 10% unless indicated otherwise

<table>
<thead>
<tr>
<th></th>
<th>Antarctic Ice Sheet</th>
<th>Greenland Ice Sheet</th>
<th>Glaciers and ice caps</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Area (10^6 km²)</strong></td>
<td>12.1</td>
<td>1.71</td>
<td>0.68</td>
</tr>
<tr>
<td><strong>Volume (10^6 km³)</strong></td>
<td>29^†</td>
<td>2.95</td>
<td>0.18 ± 0.04</td>
</tr>
<tr>
<td><strong>Volume (sea level equivalent, m)</strong></td>
<td>73^†</td>
<td>7.4</td>
<td>0.5 ± 0.10</td>
</tr>
<tr>
<td><strong>Accumulation (10^{12} kg/yr)</strong></td>
<td>1660</td>
<td>553</td>
<td>670 ± 100</td>
</tr>
<tr>
<td><strong>Runoff (10^{12} kg/yr)</strong></td>
<td>53 ± 30</td>
<td>237</td>
<td>690 ± 100*</td>
</tr>
<tr>
<td><strong>Iceberg discharge (10^{12} kg/a)</strong></td>
<td>2016</td>
<td>316</td>
<td>50 ± 30</td>
</tr>
</tbody>
</table>

(from ice shelves)


† Total volume, including the amount currently below sea level (1.9 × 10^6 km³, or 5 m sea level rise equivalent).

* Includes runoff into closed basins in Central Asia (60 × 10^{12} kg/yr).

Table 7.2: Estimates of global glacier mass balances during this century.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Time period</th>
<th>No of observed glaciers^1 (regions)^2</th>
<th>Global area^3 (10^4 km²)</th>
<th>Models: balance dynamic</th>
<th>Sea level change (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Meier (1984)</td>
<td>1900–61</td>
<td>25 (31)</td>
<td>540</td>
<td>a, f</td>
<td>0.46</td>
</tr>
<tr>
<td>Trupin et al. (1992)</td>
<td>1965–84</td>
<td>85 (31)</td>
<td>540</td>
<td>b, g</td>
<td>0.18</td>
</tr>
<tr>
<td>Meier (1993)</td>
<td>1900–61</td>
<td>– (31)</td>
<td>640</td>
<td>c, f</td>
<td>0.40</td>
</tr>
<tr>
<td>Dyurgerov (1994)</td>
<td>1960–93</td>
<td>23–76 (8)</td>
<td>589^4</td>
<td>d, g</td>
<td>0.35^5</td>
</tr>
<tr>
<td></td>
<td>1985–93</td>
<td>“</td>
<td>“</td>
<td>“</td>
<td>0.60^6</td>
</tr>
</tbody>
</table>

Balance Models:

a) Balance correlated with annual amplitude for measured glaciers in 13 regions, then extended globally based on measured or climatically estimated annual amplitudes.

b) Observed balance data correlated to regional precipitation, runoff data; extended globally using climatological data.

c) As in a), internal accumulation estimated.

d) Statistical analysis of all observed balance data.

Dynamics Models:

f) Balance adjusted by estimate of mean areal shrinkage (10%).

g) None (no dynamic response considered).

Notes

† Number of observed glaciers used in correlations.

* Number of regions into which data were aggregated.

@ Assumed total area of glaciers for calculating global contribution.

§ Does not include contribution of glaciers in drainage basins with internal runoff (52 × 10^3 km²).

Many attempts have been made to quantify the global rate of change of the world’s glacier cover, beginning with the seminal work of Thorarinsson (1940). Recent analyses are listed in Table 7.2. The principal difficulty is the lack of observational data in many areas and some of the analyses have extended the data using climate information. Virtually no glaciers are being regularly measured in the mountains bordering the Gulf of Alaska or in Patagonia, two areas thought to contribute significantly to present-day sea level rise (Meier, 1985). The global number of glaciers on which the mass balance was directly measured was six or less through 1956, then rose to 50–60 during the period 1967 to 1989, coinciding with the new programmes of the International Hydrological Decade/Programme and the establishment of the World Glacier Monitoring Service; the number has fallen slightly since 1989 (Dyurgerov and Meier, 1994).

Clearly, insufficient data exist for a long-term global synthesis without extending the data using climate information. Simplistic models have been used to take changes in glacier dynamics into account. Table 7.2 lists examples of several recent compilations with brief sketches of the approaches used. It is obvious that further refinement is needed. Table 7.2 shows that the global contribution of glacier wastage to current sea level rise in this century is uncertain by a factor of about two. There are many reasons for this uncertainty, including: (1) different time periods used in analysis; (2) differences in the total glacier area, due to incomplete data from many of the highly glacierised regions of the world, and inconsistent consideration of small glaciers in Greenland and in Antarctica which are not part of the ice sheets; (3) incomplete climatic data from many parts of the world because of lack of weather stations near the glaciers; (4) crude approximations to dynamic feedbacks, instead of a real dynamics model; (5) general neglect of refreezing of meltwater in cold firm and of iceberg calving.

The results listed in Table 7.2 and Figure 7.4 suggest that the sea level rise contribution due to changes in glacier volumes averaged about 0.35 mm/yr between 1890 and 1990, and 0.60 mm/yr between 1985 and 1993. The uncertainty of these values is difficult to assess; the spread of the individual estimates suggests uncertainties of at least ±0.1 and ±0.05 mm/yr, respectively. None of the studies noted above used an interactive model of the balance-dynamics feedback. Simple models exist (e.g., Joughensson et al., 1989), but are difficult to apply over the vast spectrum of glaciers: large and small, temperate and cold, etc. One recent model of time-changes in glacier profiles, using typical or averaged values of ice thickness and a parameter specifying the localisation of thickness changes near the terminus, produces an estimate of sea level rise due to thinning of glaciers over the last 100 years of 0.5 mm/yr (Schwiter and Raymond, 1993).

7.3.2.3 Sensitivity to climate changes

The contribution of glacier wastage to sea level rise in the future could be calculated for any climate scenario, assuming knowledge of the dependence of the mass balance, internal accumulation and iceberg calving on the meteorological environment, and how the ensuing changes feed back to the glacier area and volume. Most of this knowledge does not currently exist. Alternatively, the contribution can be estimated using simple models developed by comparing past changes in glacier volume with observed changes in air temperature, or by calculating the present-day relation of mass balance to temperature and precipitation – the "sensitivity" (e.g., Oerlemans and Fortuin, 1992).

The term "sensitivity" can be defined in several ways. The expression often used to estimate sea level change from glacier volume change is the ratio of the rate of change of glacier volume to a small change in a climatic
parameter (such as temperature). Here, this ratio is termed the static sensitivity of glaciers to climate change. The static sensitivity ignores changes in the configuration of the glacier. The dynamic response of the glacier and the lowering of its surface due to increased melting will change the effect of a non-zero mass balance over timescales ranging from decades to centuries. Most glaciers in the world have not been in equilibrium with their climatic environment, at least not since the Little Ice Age, so there is a problem in computing the sensitivity value in a non-equilibrium situation with an unknown or ill-defined initial boundary condition.

Nonetheless, estimates of the static sensitivity have been obtained by different authors using different methods. The methods include: (1) hydrometeorological modelling using precipitation and temperature (e.g., Martin, 1978; Kuhn et al., 1979; Tangborn, 1980; Chen and Funk, 1990); (2) degree-day modelling (e.g., Laumann and Reeh, 1993; Johannesson et al., 1995; Wigley et al., 1996); (3) mass balance or volume change observations related to an assumed linear change in global air temperature (e.g., Meier, 1993); (4) energy-balance modelling (Oerlemans and Fortuin, 1992). The resulting values range from 0.58 to 2.6 mm/yr/°C, with much of the spread related to assumptions about the initial mass balance condition (Wigley et al., 1996).

7.3.3 The Greenland and Antarctic Ice Sheets

7.3.3.1 Processes causing change in the ice sheets

Most of the non-oceanic water on Earth resides in the two great ice sheets (Table 7.1), and most of their volume lies on land above sea level. Thus, loss of only a small fraction of this volume could have a significant effect on sea level.

In Antarctica, recent break-ups of the Larsen and Wordie Ice Shelves in the Antarctic Peninsula and discharges of enormous icebergs from the Flicher and Ross Ice Shelves, and the discovery of major recent changes in certain Antarctic ice streams, have focused public attention on the possibility of “collapse” of this ice reservoir within the next century, with potential impacts on sea level. Changes in floating ice shelves, of course, cannot affect sea level directly. Nevertheless, ice shelves are part of a complex, coupled ice flow system involving the inland ice and relatively fast-moving ice streams that discharge ice from land to sea, and changes in the rate of discharge can affect sea level. Whether such dynamic processes can be affected by climate changes on the time-scale of a century is a key issue.

In this sense, the Greenland ice sheet, which has no floating ice shelves of consequence, is different from the Antarctic ice sheet. In Antarctica, temperatures are so low that comparatively little surface melting occurs and the ice loss is mainly by iceberg calving, the rates of which are determined by dynamic processes involving long response times. In Greenland, ice loss from surface melting and runoff is of the same order of magnitude as loss from iceberg calving (Table 7.1). Thus, climate change in Greenland could be expected to have immediate effects on the surface mass balance of the ice sheet through melting and runoff as well as through accumulation.

On both ice sheets, the residence times for particles of ice range from the order of $10^5$ yr or more for ice near an ice divide, to $10^2$ yr or less for ice near the equilibrium line which separates the area of annual mass gain (accumulation) from the area of annual mass loss (melting and/or calving). These long residence times and their variations over the ice sheets further complicate the modelling of the response of ice sheets to climate change.

7.3.3.2 Current state of balance

Current changes of the ice sheets can be measured using surface mass balance observations or by geodetic (volume-change) methods. However, the ice sheets respond to processes at all time-scales, ranging from the last glacial-interglacial transition to decade-scale fluctuations in the temperature and precipitation fields. As most of our observations extend over a few decades only, this immediately poses a problem: how can we decide from observations whether a small change in ice sheet configuration is a response to a short-term climatic fluctuation or an ongoing process of slow adjustment to changes that happened a long time ago? This question requires numerical modelling studies of ice sheet dynamics.

Mass balance studies

One approach to estimating the current state of balance is to collect all available data on specific balance (net gain or loss of ice at the surface) and on iceberg production from ice shelves or outlet glaciers, and make the sum. When the interest is in the effect on sea level, the mass flux across grounding lines (the boundary between grounded and floating ice) should be considered rather than iceberg calving. With the data currently available, this procedure leads to very uncertain estimates.

The vastness of the ice sheets is the major obstacle in mass balance studies. In principle, the net mass balance of an ice sheet can be determined by summing the balance observations at the surface and the loss of ice by calving. However, measurements of the surface balance still give a poor coverage of both the Greenland and Antarctic ice
sheets, and estimating calving rates is an even more uncertain exercise. Mass balance estimates for Antarctica and Greenland are shown in Tables 7.3 and 7.4.

For the Antarctic ice sheet, two studies attempted to compare accumulation with mass flux across grounding lines. Budd and Smith (1985) used existing ice velocity measurements to estimate outflow from the grounded ice. They concluded that "...the total influx over the Antarctic ice sheet of about 2000 km³/yr (= 1800 × 10¹² kg/yr) is probably nearly balanced by the outflow with a discrepancy most likely in the range of 0 to +20%". Bentley and Giovinetto (1991) made an assessment of the imbalance for drainage basins for which a reasonable amount of data was available. Most basins seem to have a positive balance. Extrapolating the results to the entire grounded ice sheet in three different ways yields an imbalance between +2 and +25% of the total input. These studies thus suggest that grounded ice volume is increasing, but the error bars are very large.

For the Greenland ice sheet, early estimates (Bauer, 1968) suggested a negative balance. More recently, several mass balance studies of a more local nature have been carried out. Kostecka and Whillans (1988) compared mass balance with ice velocity measurements along two transects (International Glaciological expedition to Greenland [EGIG] traverse [71°N] and the Ohio State University traverse [65°N]). Results suggest no significant changes in ice thickness, with perhaps a very slight thickening at the Ohio State University traverse (0.06 ± 0.08 m/yr). For the Dye 3 station (65°N, 43.5°W) Reeh and Gudmundsson (1986) obtain a change in ice thickness of 0.03 ± 0.06 m/yr, i.e., not significantly different from zero.

From south-west Greenland there is a large amount of information on fluctuations of outlet glaciers (Weidick,
There has been a general retreat of outlet glaciers since the end of the 19th century. This retreat has slowed down in recent decades (a significant number of outlet glaciers are now advancing). It is not exactly clear how such fluctuations relate to the total volume of the ice sheet, but it appears that the Greenland ice sheet may have contributed significantly to sea level rise during the first part of this century and that its present contribution may be close to zero.

**Geodetic methods**

The most direct method for determining the current state of balance is to measure continuously and very accurately the surface elevation of the ice sheets. Assuming that grounding lines can be located, and changes in density and bed topography are sufficiently small or can be determined otherwise, a trend in the amount of grounded ice mass can be detected.

Ground-based levelling studies along transects on the ice sheets have been very limited. The EGRG line in Greenland, however, has been studied in some detail (Mälzer and Seckel, 1975; Seckel, 1977; Kock, 1993; Moeller, 1994). There appears to be a general thickening from 1959 to 1968 followed by a thinning to 1992, with an overall change of 0.1 m/yr or less. It is not known how representative this finding is of the whole Greenland ice sheet.

Recent attempts to measure the surface elevation include the use of satellite radar altimetry. The results are still controversial. Zwally et al. (1989) used satellite radar altimetry to estimate the change in surface elevation of the Greenland ice sheet south of 72°N (excluding the margins). For the period 1978–1985, they found an ice thickening rate of 0.23 ± 0.04 m/yr which implied a 25% to 45% positive balance for that period. Douglas et al. (1990) criticised this result on the basis of inadequate calibration of satellite orbits, but Zwally et al. (1990) obtained a similar result using orbits that were consistently calculated and adjusted to a common ocean reference. Lingle et al. (1991) used this method in the ablation zone, but due to errors in the altimetry on these sloping and undulating margins, the thickness-change results are not significantly different from zero. Ice sheet modelling by Huybrechts (1994) suggests an average thickening south of 72°N, in agreement with Zwally's results but somewhat smaller in magnitude.

**Numerical modelling**

A different approach simulates the evolution of the ice sheets using numerical models that include some of the relevant physical processes, attempting to constrain resulting

**Figure 7.5:** Long-term model simulations of the changes in ice volume of the Antarctic and Greenland ice sheets due to temperature and sea level changes (courtesy of Ph. Huybrechts).

Ice sheet histories as much as possible by (palaeo) field evidence of both a geological and glaciological (ice cores) nature. The integrations should be performed over at least one glacial cycle to remove transient effects. One of the problems of such an approach is how to specify the forcing history (i.e., how to formulate the mass balance per unit area, which is the driving force, as a function of time and space).

Using a modelling approach, Huybrechts (1990, 1994) has carried out long integrations with a fairly comprehensive model of the Antarctic and Greenland ice sheets. This model solves the coupled mechanic and thermodynamic equations and includes ice shelves. The forcing follows a specified temperature and a sea level history. Simple parametrisations are used for the generation of the mass balance fields from a uniform temperature signal, which is taken from the Vostok ice core, and there is no direct effect of insolation variations on the mass balance. From the model, the Antarctic ice sheet shows a very strong response to the glacial-interglacial transition, as shown in Figure 7.5. This is mainly a response to the rise in sea level at that time and involves a substantial shrinkage of the West Antarctic ice sheet. The Greenland ice sheet shows a much weaker response, except in the beginning of the integration (which must be considered as transient effects). The Greenland ice sheet responds solely to the temperature signal (sea level has very little effect). With respect to current mass balance, these model simulations suggest that the contribution of the Greenland ice sheet to sea level rise may be close to zero, while that of the Antarctic ice sheet may be positive.

**Summary**

The paucity of relevant data does not allow a meaningful judgement of the current state of balance of the Greenland
and Antarctic ice sheets. Different workers claim changes with even different sign, up to (and perhaps exceeding) 25% of the annual mass turnover (even more for south Greenland). A major problem is the use of data on the decadal time-scale to infer long-term changes. At present it can be concluded that an imbalance of up to 25% cannot be detected in a definite way by current methods/data. In terms of sea level change, a ±25% imbalance implies:

Antarctica ±1.4 mm/yr
Greenland ±0.4 mm/yr

7.3.3.3 Sensitivity to Climate Change
Tools to study the sensitivity of ice sheet mass balance comprise regression analyses, simple meteorological models (e.g., energy-balance modelling of a melting glacier surface), and GCMs. The sensitivities of Antarctica and Greenland to a 1°C warming based on these various methods are shown in Tables 7.5 and 7.6.

Ultimately, GCMs should be the best tools to study the mass balance of ice sheets, because in principle they include most relevant processes. Nevertheless, published climate change studies using GCMs rarely report or discuss changes in the water mass balance at the surface of ice sheets in spite of the crucial importance of such information for predictions of sea level change. This is an indication that confidence in such quantification is still low. Partly because ice accumulation at the surface of ice sheets is very slow, small absolute errors (and regional-scale precipitation patterns from control runs of GCMs are often significantly different from observed climate) come out as large relative errors. Even GCMs which fare well on average are not entirely reliable for polar climate studies (Genthon, 1994). Nevertheless, there have been significant improvements in recent years, mainly associated with increased resolution (Simmonds, 1990; Budd and Simmonds, 1991; Genthon, 1994). It is expected that more credible results from AOGCMs will be available in the near future.

For Greenland, a number of estimates of the sensitivity of the ice sheet to climate change have been made with simple meteorological models (degree-day models and energy balance models). The most studied quantity is the change in mean specific (surface) balance for a uniform 1°C increase in air temperature (Table 7.5). The earlier studies extrapolated calculations for a particular site to the entire ice sheet. In the later studies where calculations were made on a 20 km grid, the sensitivity values are somewhat

Table 7.5: Sensitivity of the Greenland ice sheet to 1°C climatic warming.

<table>
<thead>
<tr>
<th>Source</th>
<th>Sensitivity (mm/yr in equivalent sea level change)</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ambach and Kuhn (1989)</td>
<td>+0.31</td>
<td>Based on analysis of EGHG data; energy balance considerations at equilibrium line extrapolated to entire ice sheet.</td>
</tr>
<tr>
<td></td>
<td>[+0.24]*</td>
<td></td>
</tr>
<tr>
<td>Bindschadler (1985)</td>
<td>+0.57 to +0.77</td>
<td>Based on simple flowline model extrapolated to entire ice sheet and including estimated change in iceberg calving.</td>
</tr>
<tr>
<td>Braithwaite and Olesen (1990)</td>
<td>+0.36 to +0.48</td>
<td>Extrapolation of energy balance calculation for South-West Greenland.</td>
</tr>
<tr>
<td>Oerlemans et al. (1991)</td>
<td>+0.37</td>
<td>Energy balance model; ice sheet divided in four &quot;climatic sectors&quot;.</td>
</tr>
<tr>
<td></td>
<td>[+0.28]*</td>
<td></td>
</tr>
<tr>
<td>Huybrechts et al. (1991)</td>
<td>+0.30</td>
<td>Degree-day model on 20 km grid.</td>
</tr>
<tr>
<td></td>
<td>[+0.22]*</td>
<td></td>
</tr>
<tr>
<td>van de Wal and Oerlemans (1994)</td>
<td>+0.30</td>
<td>Energy balance model on 20 km grid.</td>
</tr>
<tr>
<td></td>
<td>[+0.21]*</td>
<td></td>
</tr>
</tbody>
</table>

* Includes 5% increase in accumulation.
Table 7.6: Sensitivity of the Antarctic ice sheet to 1°C climatic warming (mm/yr in equivalent sea level change).

<table>
<thead>
<tr>
<th>Source/Method</th>
<th>Sensitivity</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Muszynski and Birchfield (1985)</td>
<td>-0.38</td>
<td>Regression on 208 data points</td>
</tr>
<tr>
<td>Fortuin and Oerlemans (1990)</td>
<td>-0.20</td>
<td>Grounded ice only</td>
</tr>
<tr>
<td></td>
<td></td>
<td>regression on 486 data points</td>
</tr>
<tr>
<td>Fortuin and Oerlemans (1992)</td>
<td>-0.27</td>
<td>2-dimensional atmospheric model: +1°C warming at coast and associated greenhouse forcing over ice sheet</td>
</tr>
<tr>
<td>Change in accumulation assumed</td>
<td>-0.34</td>
<td>20 km grid over grounded ice</td>
</tr>
<tr>
<td>proportional to saturation vapour pressure</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Temperature/sea ice/mass balance coupling</td>
<td>-0.7</td>
<td>Sensitivity of accumulation to sea ice extent (Giovinetto and Zwally, 1995) coupled with sensitivity of sea ice extent to air temperature (Parkinson and Bindschadler, 1984)</td>
</tr>
</tbody>
</table>

less. Nonetheless, the differences between the various studies are not very large. This must be due partly to the fact that empirical constants/parameters used in the different methods come mostly from the same data sets. The various approaches have some weak points in common as well (treatment of meltwater infiltration and refreezing, and of albedo). Also, accumulation rates are treated in the same way in all studies: either fixed, or changed in proportion to the saturation vapour pressure of the overlying air. However, there is little physical reason to believe that the precipitation pattern over Greenland is invariant for climate change. In fact, some studies using current information (Bromwich et al., 1993; Bromwich, 1995) or results from ice cores (Kapsner et al., 1995) suggest that there is little relation between temperature and precipitation changes over Greenland during the Holocene owing to the dominant effect of changes in atmospheric circulation.

Iceberg calving is a major component in the mass balances of both ice sheets. Reeh (1994) has estimated that the iceberg discharge from Greenland is $319 \times 10^{12}$ kg/yr, compared with a discharge of $239 \times 10^{12}$ kg/yr due to melting and meltwater runoff. The calving data are still approximate, because many major calving glaciers, especially in East Greenland, have not been measured, and because the seasonality of calving is not well defined even for the measured glaciers. Possible changes in the iceberg calving regimes due to climate change are not well understood, because a physical relation between calving rates and climate, geometry and ice flow has not yet been defined (Reeh, 1994). Much of the discharge from the Antarctic ice sheet is by calving of icebergs, but the various estimates of this flux differ. The most serious problem is that iceberg breakoff may be very episodic; in some areas the mean time between major calving events may be >100 years, so that historical data are of limited use. Also, the method of measuring the discharge using observations of current iceberg distributions is limited by uncertainty regarding mean lifetimes of icebergs of differing sizes, a critical relation needed for computing fluxes.

For the Antarctic ice sheet, possible changes in accumulation rates associated with eventual warming have been studied using regression techniques (e.g., Muszynski and Birchfield, 1985; Fortuin and Oerlemans, 1990). The problem here is the dependence within the set of predictors that are used in a multiple regression. Broadly speaking, when going from the margin of the ice sheet to the interior, the accumulation rate changes by an order of magnitude, typically from 0.4 m/yr to 0.04 m/yr (expressed as water-equivalent). This is partly due to lower air temperature, but also to other factors (e.g., distance to moisture source, topographic effects). So the sensitivity of the accumulation rate to temperature (identified as the regression coefficient for temperature) depends on the choice of the other predictors. This explains the large difference in the results of Muszynski and Birchfield (1985) and Fortuin and Oerlemans (1990); see Table 7.6. In addition, accumulation rates on the Antarctic ice sheet may be sensitive to changes in sea ice extent and concentration (Giovinetto and Zwally,
An extensive sensitivity study with a two-dimensional (vertical plane) meteorological model was conducted by Fortuin and Oerlemans (1992). In this model, the zonal mean distribution of temperature and precipitation is calculated with a four-layer dynamic model, which includes a radiation scheme. For a uniform warming, this model predicts a higher accumulation rate and a stronger evaporation in the coastal regions than in the interior. The net effect over the grounded ice is an increase in the mass balance corresponding to a \(-0.27 \text{ mm/yr} \) sea level change for a \(1^\circ\text{C} \) warming. This value is slightly less than the \(-0.34 \text{ mm/yr} \) found if the change in accumulation is set proportional to the relative change in saturation vapour pressure.

In some parts of the Antarctic Peninsula, surface temperatures are sufficiently high that melting occurs. Drewry and Morris (1992) estimated the potential contribution to sea level due to increased runoff from that part of the peninsula where melting may occur (about 20,000 km\(^2\)). They suggest a sensitivity of 0.012 mm of sea level rise/\(^\circ\text{C}\), which is small compared to the sensitivity for the whole of Antarctica as listed in Table 7.6.

In summary, based on Tables 7.5 and 7.6, the value of the static sensitivity, in terms of equivalent sea level change, is estimated here as 0.30 mm/yr/\(^\circ\text{C}\) for the Greenland ice sheet, and \(-0.30 \text{ mm/yr} \) for the Antarctic ice sheet. The spread of the individual estimates suggests uncertainties of at least \(\pm0.15 \text{ mm/yr} \) for both Greenland and Antarctica (excluding the possibility of collapse of the West Antarctic ice sheet – see Section 7.5.5).

### 7.3.4 Surface Water and Ground Water Storage

Changes in the terrestrial liquid-water budget can affect sea level. Shiklomanov (1993) points out that human activity affects the hydrologic cycle through water diversions, transformations of stream networks, and changes of the surface characteristics of drainage basins, as well as causing climatic changes on regional or global scales which affect water transfers between land and sea. Both direct anthropogenic changes, as well as natural and anthropogenic effects on the climate, are involved, and it is often difficult to separate the two. Some estimates of the contributions from the major components of the liquid-water budget are as follows:

- **Ground water depletion.** Ground water pumped at a rate in excess of recharge may add to sea level. Much of this water is used for irrigation and a major fraction is transpired or evaporated to the atmosphere or contributes to runoff, eventually reaching the sea. Estimates of the current rate of ground water depletion range from 0.07 to 0.38 mm/yr in sea level equivalent (Klige, 1982; Meier, 1983; Sahagian et al., 1994).

- **Surface reservoir and lake storage.** The filling of surface water reservoirs transfers water from sea to land, tending to lower sea level. Lakes depend on the balance of precipitation and evaporation, and also respond to changes in upstream runoff which may be affected by human action. Some of the world’s large lakes are currently rising; others are falling, due in part to upstream irrigation and evaporation. The current effect of these non-synchronous changes on global sea level is probably small. Estimates of the current rate of increase in water storage in artificial reservoirs range widely, from about \(-0.09 \) to \(-0.54 \text{ mm/yr} \) sea level equivalent (e.g., Golubev, 1983; Chao, 1988, 1994; Shiklomanov, 1993; Gornitz et al., 1994; Rodenburg, 1994; Sahagian et al., 1994).

- **Deforestation.** Gornitz et al. (1994) estimate that the combustion and oxidation of forests transfer water to the sea at a rate of about 0.03 mm/yr in sea level equivalent, but that decreased runoff (Henderson-Sellers et al., 1993) increases water on land at a rate of 0.15 mm/yr. Sahagian et al. (1994), on the other hand, suggest that tropical forest loss and desertification are currently contributing to sea level rise by 0.15 mm/yr, due to the loss of biomass water, soil moisture and water vapour in the atmosphere. The net effect is thus uncertain.

- **Loss of wetlands.** Sahagian et al. (1994) estimate a sea level rise component of at least 0.01 mm/yr due to the loss of wetlands.

- **Other changes.** Ice-rich permafrost may contain up to twice as much water (frozen) as the same soil in a thawed state. Lachenbruch and Marshall (1986) estimate a secular thawing of permafrost in Arctic Alaska equivalent to about 10 mm of ice melt/yr. If half of the pore water runs off on thawing, and if half of the global permafrost area is similarly ice-rich and degrading (certainly an upper limit), the contribution of permafrost thawing to sea level rise could be up to 0.1 mm/yr. Gornitz et al. (1994) have estimated the following additional effects: the water pumped for irrigation that is returned by infiltration to soil...
moisture; the water added to the atmosphere as water vapour in irrigated areas; the water that seeps into the ground each year below reservoirs; and the increased water vapour content of the atmosphere in the vicinity of the reservoirs. These effects may be appreciable, causing a decrease in the overall rate of sea level rise. However, these results are conjectural, and a more accurate budget calculation will require comprehensive global hydrological modelling of the type now under development for use in general atmospheric circulation models.

The above estimates of the terrestrial liquid-water components affecting sea level vary considerably, and the spread is so large that the sum could be either positive or negative. Earlier estimates (e.g., Klige, 1982; Meier, 1983; Robin, 1986) and more recent work (Chao, 1994; Rodenburg, 1994) suggest that the sum is, in fact, close to zero. Our best estimate at this time is that the current contribution to sea level rise due to these hydrologic factors is between −0.40 and +0.75 mm/yr, with a mean estimate of about +0.1 mm/yr. Over the last 100 years, these factors may have contributed about 0.5 cm to sea level rise, although the uncertainties are very large indeed.

Few authors have ventured an estimate for future years. Sahagian et al. (1994) point out that the rate of dam building has slowed markedly, but that depletion of ground water reservoirs is likely to increase with growing demands for water, apart from any effects of climate change. They estimate that the contribution of anthropogenic changes in land hydrology to sea level rise will be 2.6 cm in the next 50 years if the present rates are maintained. This estimate, largely based on the expected continued depletion of ground water reserves, is reasonable, but a future decline in dam building is not supported by the analysis of Shiklomanov (1993). Unfortunately, the contribution to sea level from changes in hydrologic practices due to climate trends has not yet been fully analysed.

### 7.4 Can Sea Level Changes During the Last 100 Years be Explained?

A critical issue is whether the rise in sea level observed over the last 100 years can be explained. A synthesis of the model and observational data pertaining to the factors discussed in Section 7.3 is presented in Table 7.7. In general, there is broad agreement that both thermal expansion and glaciers have contributed to the observed sea level rise, but there are very large uncertainties regarding the role of the ice sheets and other hydrologic factors.

<table>
<thead>
<tr>
<th>Component contributions</th>
<th>Low</th>
<th>Middle</th>
<th>High</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thermal expansion</td>
<td>2</td>
<td>4</td>
<td>7</td>
</tr>
<tr>
<td>Glaciers/small ice caps</td>
<td>2</td>
<td>3.5</td>
<td>5</td>
</tr>
<tr>
<td>Greenland ice sheet</td>
<td>−4</td>
<td>0</td>
<td>4</td>
</tr>
<tr>
<td>Antarctic ice sheet</td>
<td>−14</td>
<td>0</td>
<td>14</td>
</tr>
<tr>
<td>Surface water and ground water storage</td>
<td>−5</td>
<td>0.5</td>
<td>7</td>
</tr>
<tr>
<td>TOTAL</td>
<td>−19</td>
<td>8</td>
<td>37</td>
</tr>
<tr>
<td>OBSERVED</td>
<td>10</td>
<td>18</td>
<td>25</td>
</tr>
</tbody>
</table>

For thermal expansion, the range 2–7 cm reported in Table 7.7 derives from model simulations, especially those carried out for this report and other recent model results (see Section 7.5). The various studies that have addressed this issue give answers that are of the same direction and similar magnitude. Observational data related to thermal expansion are presently too sparse to make global-scale estimates. Overall, it is likely that some oceanic thermal expansion would have occurred given the global mean warming of 0.3 to 0.6 °C observed over the same time period.

It is clear that many of the world’s glaciers have retreated over the last 100 years. However, continuous, long-term measurements of the mass balances of glaciers and ice caps are very limited. Based on a combination of observations and simple models (see Section 7.3.2.2), it is concluded that glaciers and ice caps may have accounted for 2–5 cm of the observed sea level rise.

With respect to the Greenland and Antarctic ice sheets, there is simply insufficient evidence, either from models or data, to say whether the average mass balances have been positive or negative. Thus, the “zero” entries in Table 7.7 should be interpreted as a reflection of the current poor state of knowledge, rather than as an estimate of the current state of balance. As mentioned previously (Section 7.3.3.2), an imbalance of up to 25% of the annual mass turnover cannot be ruled out by existing data and methods, giving the wide range of uncertainty indicated in Table 7.7. However, a large positive mass balance of both ice sheets would seem unlikely, as this would have led to a substantial sea level lowering and would therefore be highly inconsistent with the observed sea level rise.

The current estimates of changes in surface water and ground water storage are very uncertain and speculative.
There is no compelling recent evidence to alter the conclusion of IPCC (1990) that the most likely net contribution during the last 100 years has been near zero or perhaps slightly positive, with an uncertainty of about ±6 cm.

In total, based on models and observations, the combined range of uncertainty regarding the contributions of thermal expansion, glaciers, ice sheets and land water storage to past sea level change is about −19 cm to +37 cm—a very wide band of uncertainty which easily embraces the observed sea level rise (10–25 cm). The major source of uncertainty relates to the current mass balance of the polar ice sheets. Although the apparent discrepancy between the middle values of the “total” and “observed” estimates in Table 7.7 might suggest a net positive contribution from the ice sheets, the role of the other factors cannot be ruled out within the overall uncertainties. This problem in reconciling the past change, especially in relation to climate-related factors, emphasises the uncertainties in projections of the future, as noted below.

### 7.5 How Might Sea Level Change in the Future?

#### 7.5.1 Recent Projections

In 1990, the IPCC concluded that for Scenario A, or the “Business-as-Usual” scenario, sea level would rise 66 cm by the year 2100 (Warrick and Oerlemans, 1990). Since then, additional estimates of future sea level rise have been made, as shown in Table 7.8. Although these more recent estimates seem lower than that of IPCC (1990), direct comparisons cannot easily be made due to differences in assumptions regarding factors such as emission scenarios, gas concentration changes, radiative forcing changes, the climate sensitivity and initial conditions.

For example, the estimate of Church et al. (1991) of 35 cm by 2050 was based on a rate of global warming which was twice that projected by IPCC (1990). The projection of Wigley and Raper (1993) of 46 cm by 2100 was based on a lower emission scenario (IPCC (1990)-B). The Wigley and Raper (1992) estimate of 48 cm by the year 2100 was based on the revisions of gas cycle models and lower

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**Table 7.8: Recent estimates of future global sea level rise (cm) (updated from Warrick and Oerlemans, 1990). It should be noted that the estimates of sea level change are not strictly comparable, as they also reflect the combination of the authors’ different estimates of future emissions, radiative forcing changes and model parameters, which are difficult to disentangle.**

<table>
<thead>
<tr>
<th>Source (emission scenario)</th>
<th>Thermal Expansion</th>
<th>Glaciers and ice caps</th>
<th>Greenland ice sheet</th>
<th>Antarctic ice sheet</th>
<th>Best Estimate</th>
<th>Total rise(^a)</th>
<th>Range(^b)</th>
<th>To (Year)</th>
</tr>
</thead>
<tbody>
<tr>
<td>IPCC90-A</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(Warrick &amp; Oerlemans, 1990)</td>
<td>43</td>
<td>18</td>
<td>10</td>
<td>-5</td>
<td>66</td>
<td>31 to 110</td>
<td>2100</td>
<td></td>
</tr>
<tr>
<td>Church et al. (1991)(^c)</td>
<td>25</td>
<td>[........]</td>
<td>10 (all ice)</td>
<td>[........]</td>
<td>35</td>
<td>15 to 70</td>
<td>2050</td>
<td></td>
</tr>
<tr>
<td>Wigley &amp; Raper (1992)(^d)</td>
<td>[........]</td>
<td>[.........]</td>
<td>[.........]</td>
<td></td>
<td>48</td>
<td>15 to 90</td>
<td>2100</td>
<td></td>
</tr>
<tr>
<td>Wigley &amp; Raper (1993)</td>
<td>25</td>
<td>[........]</td>
<td>[........]</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Titus &amp; Narayanan (1995)(^e)</td>
<td>21</td>
<td>9</td>
<td>5</td>
<td>-1</td>
<td>46(^e)</td>
<td>3 to 124(^f)</td>
<td>2100</td>
<td></td>
</tr>
<tr>
<td>IPCC projections, this report(^i)</td>
<td>21</td>
<td>12</td>
<td>7</td>
<td>-7</td>
<td>27</td>
<td></td>
<td>2100</td>
<td></td>
</tr>
<tr>
<td>This report (Section 7.5.3.2)(^j)</td>
<td>15</td>
<td>12</td>
<td>7</td>
<td>-7</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

\(^a\) In most cases from 1990.

\(^b\) No confidence intervals given unless indicated otherwise.

\(^c\) Assumes rapid warming of 3°C by 2050 for best case.

\(^d\) For IPCC emission scenario IS92a.

\(^e\) For IPCC (1990) Scenario B, best estimate model parameters.

\(^f\) For IPCC (1990) Scenarios A and C, with high and low model parameters, respectively.

\(^g\) Incorporates subjective probability distributions for model parameter values based on expert opinion.

\(^h\) Represents 90% confidence interval.

\(^i\) For the IPCC IS92a forcing scenario, using a climate sensitivity of 2.5°C for the mid projection and 1.5°C and 4.5°C for the low and high projections, respectively. Also see Raper et al., 1996.

\(^j\) For IPCC IS92a forcing scenario, with a constant 2.2°C climate sensitivity (no range provided).

\(^k\) Also see Raper et al., 1996.
radiative forcing changes implied by IPCC (1992). The probabilistic estimates of sea level rise provided by Titus and Narayanan (1995, 1996) were derived from subjective estimates of probability distributions of model parameter values made by a panel of expert reviewers; the result was a median estimate of 34 cm by the year 2100. Despite the differences in methods and assumptions, all of these recent “best estimates” of future sea level rise still fall within a range of 3–6 cm/decade.

7.5.2 Revised IPCC Projections

In this section, we offer revised future sea level projections that are consistent with the gas concentration, radiative forcing and climate changes discussed in earlier chapters. The individual contributions from oceanic thermal expansion, glaciers and ice caps, and the ice sheets of Greenland and Antarctica are calculated separately and summed to give the total projected sea level rise to the year 2100.

7.5.2.1 Methods and assumptions

Projections of changes in sea level are made using a simple global climate model, a global glacier melt model, and sensitivity values relating temperature change to ice sheet mass balances. The modelling approaches are similar to those used in IPCC (1990) (Warrick and Oerlemans, 1990), but the models themselves have been substantially revised and updated as discussed below. As in Chapter 6, there are two sets of projections for each greenhouse gas emission scenario based on two alternative views of future aerosol concentrations. In the first, aerosol concentrations change in response to the changing emissions of their precursors assumed in the IS92 scenarios. In the second, future aerosol concentrations are held constant at 1990 levels for the purpose of sensitivity analyses. The latter set is included because of the large uncertainties in future aerosol forcing, related both to uncertainties in future emission changes and their consequent effects on radiative forcing change. Projections corresponding to "aerosols constant at 1990 levels" provide an estimate of the sea level response to a situation where global emissions of aerosol precursors remain similar to 1990 levels.

The projections of oceanic thermal expansion are made using an upwelling diffusion-energy balance climate model, that of Wigley and Raper (1987, 1992, 1993). Since changes in surface air temperature and oceanic thermal expansion are interactive (in general, they tend to be inversely related – the greater the thermal expansion, the less the surface warming for a positive change in radiative forcing), it is appropriate for the sake of consistency that this is the same model used to project temperature changes in Chapter 6. The model has been recently updated and revised (Raper et al., 1996) to incorporate different land/ocean climate sensitivities and temperature-dependent upwelling rates to simulate a slow-down of the thermohaline circulation with global warming, effects that are suggested by recent coupled ocean-atmosphere model experiments (Cubasch et al., 1992; Manabe and Stouffer, 1994; Murphy, 1995). As demonstrated in Chapter 6, the revised model is able to emulate well the thermal expansion and temperature predictions of more complicated coupled ocean-atmosphere models (see Sections 6.3.1–6.3.3 for a description of model parameters, assumptions and results). The predicted temperature changes from this model were used to force the glacier and ice sheet models.

Changes in the volume of glaciers and ice caps were estimated with a revised version of the global glacier model used in IPCC (1990) (Wigley and Raper, 1995). In the revised model, the driving force for melting (or change in ice volume) is the difference between the ice volume and the “equilibrium” value of the ice volume, the latter itself being a function of temperature change. The model has a small number of parameters and is tuned against observationally based estimates of the glacier volume change over the period 1900 to 1961. The model incorporates a range of characteristic response times due to glacier dynamics; this is necessary because projections are needed for the next 100 years, which is of the same order of magnitude as real glacier response times. The model is regionally disaggregated to take into account the variations in the altitudinal range of the world’s glaciers and their response times, so that as the more “sensitive” glaciers begin to disappear during model simulations, the average characteristic response time and other model parameters values are altered, giving a non-linear response of glacier melt to temperature change. At the initial year of the simulation (1880), it is assumed that the world’s glaciers are in steady-state with respect to the prevailing climate and that they contain a volume equal to 30 cm in sea level equivalent.

For the Greenland and Antarctic ice sheets, the simplifying assumption made for the present set of model projections is that the dynamic response can be ignored on the time-scale of decades to a century (in Section 7.5.3 this assumption is relaxed). Accordingly, as for the IPCC (1990) assessment, the changes in the surface mass balance of the ice sheets are represented by static sensitivity values (in terms of sea level equivalent) as discussed above. For Greenland the sensitivity value is $0.30 \pm 0.15 \text{ mm/yr/°C}$,
and it is assumed that the Greenland ice sheet was in equilibrium for the initial year (1880) of the simulation. For Antarctica, the sensitivity value is $-0.20 \pm 0.25 \text{ mm/yr}^\circ\text{C}$ (including a term for the possible instability of the West Antarctic ice sheet), and it is assumed that the mass balance of the Antarctic ice sheet was negative in 1880 in accordance with glacial-interglacial model simulations (Section 7.3.3.2). (An imbalance of $0.1 \pm 0.5 \text{ mm/yr}$ in sea level equivalent has been assumed as a baseline trend that is extrapolated over the simulation period).

Possible changes in surface and ground water storage are not taken into account, for three reasons: (1) the available data are insufficient for meaningful extrapolation; (2) existing studies (e.g., Sahagian et al., 1994) suggest that the future contribution would, in the worst case, be rather small; (3) such changes are not, for the most part, caused directly by climate change.

For all the sea level projections, the climate model was run from 1765 to 2100. Up to 1990, the same historical radiative forcing changes (including aerosol effects) were used; thereafter, the various IS92 emission scenarios were applied. The glacier and ice sheet models, which were all run from 1880, were forced with the model-derived global mean temperature changes (except in the case of Greenland, for which the temperature changes were scaled by a factor of 1.5 in accordance with AOGCM results). Projections were made for the following sets of scenarios (see Chapter 6 for an elaboration of methods and assumptions regarding scenarios):

- **IS92a–f**, using "best-estimate" model parameters, both including and excluding the effects of changes in aerosol concentrations after 1990;
- **IS92a**: low, mid and high projections, both including and excluding the effects of changes in aerosol concentrations after 1990;
- extreme range of projections, based on the highest and lowest forcing scenarios and with model parameters chosen to maximise or minimise sea level changes.

### 7.5.2.2 Modelled past changes

For the period 1880 to 1990, the calculated change in sea level change ranges from $-1 \text{ cm}$ to $17 \text{ cm}$, with a middle estimate of $7 \text{ cm}$ (of which more than half is due to thermal expansion, followed by glaciers). This is low in comparison to the observed range based on tide gauge records ($10-25 \text{ cm}$; see Section 7.2.1) because it only takes into account an estimate of the anthropogenic component of past climate forcing, giving a lower temperature change than that observed (about $0.29^\circ\text{C}$ as compared to $0.45^\circ\text{C}$; see Section 6.3). Similarly, for the projections of future sea level rise only the estimated anthropogenic climate forcing is taken into account; no other assumed climate- or non-climate-related (or "unexplained") component of the observed past sea level rise is extrapolated into the future projections. For this reason, both the past and future projections are likely to be underestimated.

![Figure 7.6: Projections of global sea level rise over the period 1990 to 2100 for Scenarios IS92a–f, using best-estimate model parameters, including the effects of changing aerosol concentrations after 1990 (a) and, to indicate the sensitivity of projections to aerosol effects, for aerosol amounts constant at 1990 levels (b).](image-url)
7.5.2.3 Scenarios IS92a-f
The 1990–2100 changes in sea level for scenarios IS92a-f, with "best-estimate" model parameters, are shown in Figure 7.6(a). The range of the projections shown is determined by the range of future emissions under the IS92 scenarios and does not include the additional uncertainties in model parameter values. For the first decades of the projection period, the choice of emission scenario has little effect on the rate of sea level rise; even by the year 2050, the range of projected sea level rise is still relatively small, 18–21 cm. This is a consequence of lags in the climate system, caused primarily by the thermal inertia of the ocean, and of the continuing response of the ocean, climate and ice masses to past changes in radiative forcing and temperature. In the short term (i.e., several decades), future sea level rise is largely determined by past emissions of greenhouse gases. During the second half of the next century, however, the curves diverge noticeably. By the year 2100, the uncertainty introduced by the emission scenarios gives a range of sea level rise of 38–55 cm. The effect of holding aerosol amounts constant at 1990 levels is to increase the range of projected sea level rise based on the six IS92 scenarios, as shown in Figure 7.6(b).

7.5.2.4 Scenario IS92a
The 1990–2100 rise in sea level for IS92a, with high, middle and low projections based on model uncertainties, is shown in Figure 7.7 Taking into account future changes in aerosol amounts under the IS92a Scenario (Figure 7.7, solid curves), sea level is projected to rise by 20 cm by the year 2050, within a range of uncertainty of 7–39 cm. By the year 2100, sea level is estimated to rise by 49 cm, with a range of uncertainty of 20–86 cm.

Figure 7.7: High, middle and low projections of global sea level rise over the period 1990 to 2100 for Scenario IS92a, for aerosol amounts constant at 1990 levels (dashed curves) and for changing aerosol amounts after 1990 (solid curves).

7.5.2.5 Extreme range
The extreme projections of sea level rise, taking into account uncertainties in both model parameters and future radiative forcing changes, are shown in Figure 7.9. There is

Figure 7.8: The projected individual contributions to global sea level change, 1990 to 2100, for Scenario IS92a (including the effects of changes in aerosol amounts beyond 1990).

For the middle projection under IS92a, more than half of the rise by the end of the next century is due to oceanic thermal expansion alone. This is followed by the contribution from glaciers and ice caps and from the Greenland ice sheet. The Antarctic ice sheet actually causes a slight decrease in sea level due to increased accumulation as a result of atmospheric warming (as shown in Figure 7.8).

To indicate the sensitivity of the projections to assumptions concerning aerosols, Figure 7.7 (dashed curves) shows results with aerosol concentrations held constant at 1990 levels: the IS92a projections at the year 2100 are about 10% higher than the projections that include aerosol changes. The inclusion of aerosol effects tends to lower the estimated changes in radiative forcing, compared to those due to greenhouse gases alone, both for the past and the future (see Chapters 2 and 6). This directly lowers the heating and consequent thermal expansion of the oceans, as well as reducing the surface temperature changes that drive changes in the glaciers and ice sheets.

The estimates of sea level rise presented here are lower than those given by IPCC (1990). For example, the "best estimate" value for IS92a by the year 2100 is 49 cm, compared with 66 cm for the corresponding case in IPCC (1990). This change is due primarily to the lower temperature projection (see Section 6.3.3), but it also reflects the compensating effects of a slow-down in the thermohaline circulation (which was not considered in 1990 and which leads to an increase in thermal expansion) and the changes made to the glacier model.
an order of magnitude difference between the highest and lowest projections. The lowest projection shows sea level rising at an average rate of about 1 mm/yr over the next century, a rate comparable to that which has occurred over the last 100 years. The highest projection indicates an average rate of about 9–10 mm/yr, a rate which, on a global scale, is probably unprecedented over at least the last several thousand years. Although this range should be considered extreme, no attempt has been made to quantify the confidence interval.

7.5.2.6 Summary
The "best estimate" for IS92a is that sea level will rise by 49 cm by the year 2100, with a range of uncertainty of 20–86 cm. These projections of future sea level rise are lower than those presented in IPCC (1990). The differences are due primarily to the lower temperature projections, the inclusion of a slow-down of the thermohaline circulation, and changes to the glacier model. However, the basic understanding of climate-sea level relationships has not changed fundamentally since IPCC (1990). Thus, if future temperature change is higher than expected, sea level rise will also be higher.

Large uncertainties remain. In the particular set of models used above, these uncertainties derive mainly from uncertainties in radiative forcing and model parameters affecting temperature change, especially the value of the climate sensitivity (see Chapter 6). Relatively speaking, uncertainties in future greenhouse gas emissions have comparatively little effect on the projected sea level rise, particularly over the first half of the next century, due largely to lags in the climate system. Nonetheless, combining the various sources of uncertainty, the extreme range of sea level projections is very large — an order of magnitude difference between the highest and lowest.

7.5.3 Possible Inter-Model Differences
In Section 7.5.2 above, a single set of integrated climate and ice-melt models was chosen to maximise consistency with the various chapters of this report. Consistency was achieved in three ways. First, the same simple climate model that was used to estimate oceanic thermal expansion was also used to estimate global-mean temperature change, thus ensuring consistency between Chapters 6 and 7. Second, this climate model was demonstrated to emulate both the temperature and thermal expansion predictions of certain AOGCMs, selected during the IPCC process as the standard for gauging the acceptability of a simple climate model as the means for examining the effects of various IPCC emission scenarios. Third, the land-ocean temperature differences, the changes in thermohaline circulation, and the magnitude of temperature changes in polar regions — as well as the global temperature and thermal expansion predictions — produced by the AOGCMs, were explicitly taken into account in using this set of models. Largely because of this consistency, these model results are promulgated as the IPCC sea level rise projections for the purposes of the present IPCC Assessment.

Using a single set of models, however, ignores the differences and uncertainties that might arise from alternative models. For this reason, in this section we present and compare the results of another set of climate, glacier and ice sheet models. This alternative set of models incorporates several recent advancements in modelling and is a credible complement to those models of Section 7.5.2. The models were forced by the identical set of IS92 radiative forcing scenarios for purposes of comparison. However, these models do not necessarily meet all the same requirements of consistency with the other chapters, and therefore were not put forward to the Summary for Policymakers as the IPCC sea-level projections. Nonetheless, the results are to be considered internally consistent, plausible and "state of the art". In this regard, the results highlight the uncertainties that could arise from various modelling approaches and emphasise the need for systematic inter-model comparisons as a post-IPCC activity in order to improve future projections, as discussed below.
7.5.3.1 Methods and assumptions
For predicting temperature changes and thermal expansion, a simple two-dimensional energy-balance climate model having latitudinal and seasonal resolution was used (Bintanja, 1995; Bintanja and Oerlemans, 1995; de Wolde et al., 1995). This zonal-mean model has a climate sensitivity of approximately 2.2°C, and so would be expected to give lower sea-level changes than the best estimates described above (which use a sensitivity of 2.5°C). The model was first calibrated against the seasonal cycle of present-day observations of surface air temperature, ocean temperature (Levitus, 1982), and snow and sea-ice cover. Radiative forcing values for 1765, referenced to 1990 (see Section 6.3.2), were then used to obtain the initial state, after which the model was integrated over the period 1765–2100. Using comparable forcings, the thermal expansion results of this model have been compared to several coupled GCM results and found to be in reasonable agreement. The latitudinally and seasonally varying changes in the surface air temperature from the model were used to determine the sea level contributions from the glacier and ice sheet models.

For glaciers and ice caps, a range of sensitivity to climate change was used (from Oerlemans and van de Fortuin, 1992). The value of the sensitivity varies latitudinally, depending on the present-day precipitation rate, since glaciers in wetter regions are more sensitive than those in drier regions. These sensitivity values are time-independent and do not take into account the dynamic response of glaciers in a changing climate. Since the dynamic response may be very different for individual glaciers, it is uncertain how such an averaged dynamic behaviour should be included. However, it is assumed that on the time-scales considered here, the warming associated with the lowering of the ice surface and the decrease of the area due to the retreat of the glacier terminus will have a counterbalancing effect on the sensitivity values. The model calculations start in 1990. On the assumption that most glaciers are not in equilibrium with the present climate, a constant trend of 0.5 mm/yr is included (consistent with observations and the global glacier model results in Section 7.5.2).

Dynamic ice sheet models were used to estimate the sea level contributions of Greenland and Antarctica. These models take into account the effects of a changing climate on the dynamic responses of ice sheets; these flow responses were not explicitly included in the static sensitivity values used in Section 7.5.2. The two-dimensional time-dependent Greenland ice sheet model (after van de Wal and Oerlemans (1994), modified for dynamics in accordance with Mahaffy (1976) and Cadee (1992)) has a horizontal resolution of 20 × 20 km and is driven by atmospheric temperature changes, which change the surface mass balance. Ablation is calculated with an energy-balance model (van de Wal and Oerlemans, 1994), while the accumulation rate is kept constant at its present-day value (as described by Ohmura and Reeh, 1991). The initial state of the fully coupled model is its present-day equilibrium state.

The Antarctic ice sheet model (Huybrechts, 1990) is a three-dimensional, thermomechanic model coupled to a mass balance model that is driven by temperature changes interpolated onto a 40 km grid. Since accumulation appears to be strongly related to temperature in Antarctica, the accumulation rate is perturbed in proportion to the saturation water vapour pressure as temperature changes. The initial state of the ice sheet was obtained by integrating the ice sheet model over several glacial cycles. Although the model shows a continuing, long-term negative mass balance (refer to Figure 7.5), no trend is included for future projections because of the large uncertainties in the current state of balance due to the paucity of observations; the projections are calculated as the difference between the perturbed run and the reference run.

For both the Antarctic and Greenland ice sheet models as well as the glacier model, simulations begin in 1990, with temperature perturbations referenced to 1990. This is because of the non-linear response of the models and the tuning to present-day climate.

7.5.3.2 Model results and comparisons
In general, the projections of future sea-level rise from this set of models are substantially lower than the revised IPCC projections presented in Section 7.5.2 for the identical set of IS92 forcing scenarios. To illustrate, Figure 7.10 shows the highest and lowest projections of sea level rise for the period 1990–2100, along with the “best estimates” for IS92a. For IS92a, sea level is estimated to rise by 27 cm by the year 2100 (34 cm for the sensitivity case with constant aerosols), about 45% lower than the corresponding projection in Section 7.5.2. Possible explanations for the differences are given below. It should also be noted that, unlike Section 7.5.2, the value of the climate sensitivity (2.2°C) did not vary for the high and low projections, resulting in a smaller range of estimates than would otherwise have been the case.

The individual contributions to the total projected sea level change for IS92a (including aerosols) are shown in Figure 7.11 (compare to Figure 7.8). The relative contributions shown in Figure 7.11 are similar to those presented in Section 7.5.2: most of the future rise is caused
by oceanic thermal expansion and increased melting of glaciers and ice caps, with a positive contribution from Greenland and a negative contribution from Antarctica (from increased accumulation). However, there are some large apparent differences regarding the absolute contributions between the two sets of model results.

For thermal expansion, the contribution to sea-level rise by 2100 is 15 cm, about half of that obtained by the simple climate model used in Section 7.5.2. This lower thermal expansion, however, can be largely explained by two factors. First, the climate sensitivity of the climate model is lower (2.2°C, as compared to 2.5°C used for the best-estimate projections in Section 7.5.2). Second, the thermohaline circulation is represented differently by the two models. For the revised IPCC projections in Section 7.5.2, the simple climate model simulates the slow-down of the thermohaline circulation with global warming, in accordance with most recent coupled GCM results (see Section 6.3), which allows greater warming at depth and larger thermal expansion. In the zonal-mean model used here, the thermohaline circulation was kept constant, with comparatively less thermal expansion. Other possible explanations for the differences in thermal explanation have yet to be examined fully.

For the Antarctic ice sheet, the dynamic model gives relatively much larger negative contributions to sea level by the year 2100 than the constant sensitivity values used in Section 7.5.2 (~7 cm as compared to about ~1 cm). For both the dynamic Antarctic and Greenland models used here, however, the portion of sea level change attributed to dynamic changes was found to be minor on this time-scale and does not appear to account for much of the inter-model differences. Rather, the inter-model differences are due largely to three other factors. First, the Antarctic temperature forcings were very different. For the revised IPCC projections, the global-mean temperature change was used (consistent with recent AOGCM results which show little enhancement of temperature changes for Antarctica). In contrast, the dynamic ice sheet model used here was forced with enhanced polar temperature changes obtained by the two-dimensional climate model. Second, the ice-sheet sensitivity value used for the revised IPCC projections was considerably lower, since it included a term for the possible instability of the West Antarctic ice sheet. Third, the revised IPCC projections included an extrapolation of an assumed 1880 negative mass balance (a relatively small effect) throughout the simulated period. Together, these factors result in a substantial difference between the two sets of model results for Antarctica.

For the Greenland ice sheet and glaciers, the sea level projections obtained by the two sets of models are in closer agreement, perhaps by coincidence, despite the different modelling approaches that were followed.

In summary, the uncertainties in sea level projections attributed to inter-model differences may be significant. Most of the differences between the model results discussed in this section can probably be attributed to assumptions about the climate sensitivity, changes in the thermohaline...
circulation, and the spatial patterns of temperature changes. However, no comprehensive model intercomparisons of the different component models used for sea level projections have been carried out. Such comparisons are required to identify the directions for improved models and to narrow the uncertainties in future projections.

### 7.5.4 Possible Longer-term (>100 years) Changes

In IPCC (1990), it was shown that on the decadal timescale, sea level could be expected to continue to rise throughout the next century even if greenhouse forcing were stabilised by 2030 (Warrick and Oerlemans, 1990). Since IPCC (1990) several model experiments have been carried out, using both complex and simple models, that reveal the longer-term, multi-century implications of greenhouse-gas forcing on sea level.

One such transient experiment involved the use of an AOGCM in which the CO$_2$ concentration was increased by 1% per year until it doubled, after 70 years (e.g., Manabe and Stouffer, 1993, 1994). Although the concentration was stabilised after doubling, sea level continued to rise from thermal expansion alone. At the end of five hundred years, sea level had risen one metre due to thermal expansion and was still rising, even though temperature changes had largely been stabilised.

![Figure 7.12: Long-term (1990 to 2500) projections of global sea level rise for stabilisation of CO$_2$ concentration at 450 ppmv (S450) and 650 ppmv (S650). Dots denote the dates of CO$_2$ stabilisation (from Raper et al., 1996). Calculations assume the "observed" history of forcing to 1990, including aerosol effects (see Figure 6.18) and then CO$_2$ concentration increases only beyond 1990.](image)

Similar results have also been derived using a simpler climate model (Raper et al., 1996; Wigley, 1995). To illustrate, Figure 7.12 (reproduced from Raper et al., 1996) shows the long-term effects on sea level of stabilising the atmospheric concentrations of CO$_2$ at 450 ppmv in 2100 and 650 ppmv in 2200 (Scenarios S450 and S650, respectively — see Chapter 2), for the high, middle and low sets of climate and ice-melt model parameters. The changes shown are those arising from CO$_2$ alone. In all but the lowest projections, sea level continues to rise at a scarcely unabated rate for many centuries after concentration stabilisation. Figure 7.13 provides another illustration of the long-term effects of anthropogenic forcing on sea level. In this scenario, anthropogenic emissions of CO$_2$, CH$_4$, N$_2$O, the halocarbons and SO$_x$ (an important precursor of aerosols) follow IS92a to 2100 and then are assumed to decline linearly to zero over 2100 to 2200. In this scenario, total radiative forcing peaks around the year 2160, but sea level is still rising by 2500, at which time it has reached 150 cm.

![Figure 7.13: Long-term (1990 to 2500) projection of global sea level rise under an extended emission scenario comprising IS92a Scenario to 2100, with a linear decrease in greenhouse gas emissions to zero by the year 2200.](image)

In both the complex coupled atmosphere-ocean models and the simpler climate models, most of the residual sea level rise is due to the thermal inertia of the oceans and continued thermal expansion. In addition, in the case where simple ice sheet models are included, the assumed large response times involving ice dynamics result in continuing effects on ice volumes after temperature changes have
largely ceased. Overall, these results reinforce the conclusions of IPCC (1990) that a long-term "sea level rise commitment" must accompany greenhouse-gas-induced warming. Thus, even if greenhouse gas concentrations were stabilised, sea level would continue to rise for many centuries because of the large inertia in the ocean-ice-atmosphere climate system.

7.5.5 Possible Instabilities of the West Antarctic Ice Sheet

The West Antarctic Ice Sheet (WAIS) is a marine ice sheet – it rests on a bed well below sea level. It has long been argued (Weertman, 1974) that the WAIS may be inherently unstable because the interior, grounded ice (inland ice) cannot respond fast enough to changes in thickness of the floating portions at their junction, the grounding line. It has also been argued (Thomas, 1973, 1985) that the large abutting ice shelves create "back pressure" which prevents the collapse of the inland ice, such that ice shelf thinning or break-up could cause the grounded ice to "surge" – another critical element contributing to marine ice sheet instability.

These notions are changing. It is now known that the activity of the WAIS is dominated by fast-flowing, wet-based ice streams whose characteristics blend gradually into those of the floating ice shelves and whose response times to changes in the grounding line appear to be very rapid (Alley and Whillans, 1991). However, the effects of these dynamic ice streams on the stability of the WAIS is very much in dispute. In the view of some glaciologists, the ability of ice streams to transport ice rapidly from the interior to the ocean, on a time-scale of the order of 100 years, indicates an enhanced capability for a drastically accelerated discharge. A contrary view is that the short response time of ice streams removes the flux imbalance at the grounding line so that the purported instability may not exist.

Recent theoretical work is equivocal. Several recent treatments support the idea that the transition zone between the grounded and floating ice does not act as a source of instability (Van der Veen, 1985; Herterich, 1987; Barcillon and MacAyeal, 1993, Lestemant, 1994). On the other hand, there is also support for the idea of instability, which may include the concept of ice shelf buttressing at the grounding line (NASA, 1991). A recent theoretical development that suggests dramatic instability of a marine ice sheet is the so-called "binge-purge" cycle put forward to explain the massive outpourings of icebergs (Heinrich events) from Northern Hemisphere ice sheets during the last ice age (Alley and MacAyeal, 1994; MacAyeal, 1994). A model study of the WAIS over the last million years that incorporated ice streams and their slippery beds suggests that the ice sheet did collapse in the past but that the outflow rates were only a few times faster than at present (MacAyeal, 1992).

Recent observational work does not present a clear answer either. On the one hand, there is evidence suggesting unstable behaviour of the WAIS: Ice Stream B is currently flowing too rapidly for a steady-state; the current growth of the Crary Ice Rise is affecting the regional velocity field and perhaps reducing the discharge of Ice Stream B; and Ice Stream C has stagnated in the last 100–150 years (Retzlaff and Bentley, 1993). Furthermore, there is geologic evidence that this ice sheet has been largely or completely absent at some time after its initial formation (Scherrer, 1991; Burkle, 1993), which suggests transient behaviour in this part of the Ross Ice Shelf system. On the other hand, there is evidence that does not support the notion of WAIS instability: the steady flow for the last 1500 years (except for one pulse a few hundred years ago) as suggested by flow tracers in the Ross Ice Shelf; the current growth, rather than collapse, of the glaciers feeding Pine Island Bay (which lost its ice shelf in the recent geologic past); and the lack of evidence of drastic change in the height or flow of the WAIS at Byrd Station in the last 30,000 years (Whillans, 1976).

Given our present knowledge, it is clear that while the ice sheet has had a very dynamic history, estimating the likelihood of a collapse during the next century is not yet possible. If collapse occurs, it will probably be due more to climate changes of the last 10,000 years rather than to greenhouse-induced warming. Nonetheless, such a collapse, once initiated, would be irreversible. Our ignorance of the specific circumstances under which West Antarctica might collapse limits the ability to quantify the risk of such an event occurring, either in total or in part, in the next 100 to 1000 years.

7.6 Spatial and Temporal Variability

7.6.1 Geologic and Geophysical Effects

The only globally coherent geological contribution to long-term sea level change about which we possess detailed understanding due to a detailed theory of the process is post-glacial rebound (Peltier and Tushingham, 1989; Lambeck, 1990). This is the process by which the solid Earth and the ocean have continued to adjust to the effects of deglaciation throughout the Holocene period (last 10,000 years). Sea level changes due to longer time-scale geological processes (e.g., sea floor spreading) are sufficiently small to be of little interest to this report (e.g.,
see Harrison, 1989; Meier, 1990). Most Holocene geological sea level data have been assimilated into, or used to verify, geodynamic models of the Earth (e.g., Tushingham and Peltier, 1991; Peltier, 1994). These models attempt to achieve a consistency between the geological sea level measurements, the history of glaciation, and the physics of the solid Earth, including the resulting changes in the gravitational field of the Earth. Mantle viscosity is determined from a best fit to the data. The models result in estimates of relative sea level at any point on the Earth’s surface (including the interior of continents where for “relative sea level” one infers geoid height relative to the land surface). A recent example is shown in Figure 7.14. Such models are the only ones which can be employed on a global basis to estimate the rate one would expect sea level to be changing at the present time at each location due to the continuing response of the solid Earth to deglaciation.

There is some debate at present concerning the precise form of the Earth’s viscosity profile. Some authors believe that the geological sea level data constrain the viscosity to be approximately uniform, while others think either that there is a large increase in viscosity below a 660 km discontinuity, or that the present data do not constrain the profile well enough to be useful. It is not clear at the present time how the uncertainty in the viscosity profile propagates into expected rates of vertical land movement. However, it is to be expected that more sophisticated post-glacial rebound models will be developed as this debate is resolved. For example, the removal of post-glacial rebound-related land movements from tide gauge records by means of Peltier’s ICE-4G model (Peltier, 1994) has not been investigated.

Superimposed upon post-glacial rebound are a variety of local and regional isostatic and tectonic effects, many of which cannot be modelled accurately (for a review, see

Figure 7.14: The present day rate of relative sea level change (in mm/yr) based on the topographically and gravitationally self-consistent theory of glacial isostatic adjustment of Peltier (1994). The calculation employed the ICE-4G model of the last deglaciation event of the current ice age and an internal radial viscoelastic structure comprising a 120 km thick lithosphere and an upper mantle of viscosity $2 \times 10^{21}$ Pa s. This model somewhat overpredicts the ongoing rate of sea level rise due to post-glacial forebulge collapse along the east coast of the USA.
Emery and Aubrey, 1991). This is most obvious in the Mediterranean where the available archaeological sea level data primarily reflect local tectonic land movements (Flemming, 1969, 1978, 1993; Flemming and Webb, 1986) and where the Peltier post-glacial rebound models do not reproduce relative land-sea movements at all. Most large river deltas exhibit submergence associated with sedimentary isostasy which is clearly identified in tide gauge data. Very local effects, which have no relation to regional geology, can take place within the harbour of the tide gauge itself. However, differences between tide gauge sea level records can be used to provide maps of relative land movements on a regional basis, and such maps are usually consistent with previous geological knowledge of the area (Emery and Aubrey, 1991).

Studies of the vertical crustal motions on different time-scales can help elucidate patterns of tectonic deformation affecting tide gauge data. For example, along convergent plate boundaries, the uplift recorded by tide gauges or geodetic levelling measurements generally represent interseismic deformation that typically spans over 90% of the earthquake cycle (Bilham, 1991). However, the deformation during, and also shortly before or after, a major earthquake may exceed the average interseismic rate by an order of magnitude. Deformation associated with major earthquakes may temporarily cause displacements resulting in sea level changes comparable to the global sea level change (Bilham and Barrientos, 1991). These recent rates of crustal motion may differ substantially from longer-term geological trends derived from raised marine terraces (see, for example, Kelsey et al., 1994; Mitchell et al., 1994). Analyses of the changes in uplift rates on different time-scales may contribute to improved modelling of the earthquake cycle and extraction of the sea level signal from the tidal and geodetic data in those regions.

Anthropogenic effects (e.g., extraction of water from aquifers) can also result in considerable rates of subsidence (e.g., 10 cm/yr at Bangkok, Thailand). Ground water extraction has been an important factor in Venice sinking (Frassetto, 1991), while large recent rates of sea level rise at Manila, Philippines, has been blamed on coastal reclamation (Spencer and Woodworth, 1993).

It is important to realise that all sea level measurements, whether tide gauge, archaeological or geological, are measures of the level of the ocean relative to a land datum. The exceptions are those from space (e.g., from radar altimetry or via the Global Positioning System) for which the datum is in effect the computed satellite orbit. Such ground based measurements will always, therefore, contain some kind of inherent land-ocean level ambiguity. However, with the advent of GPS and other forms of advanced geodetic measurement, it is now possible to measure vertical land movements independently of sea level changes (Section 7.7.1).

A good example of alternative methods of monitoring vertical land movements is provided by the recent breakthrough in development of absolute gravimeters (Carter et al., 1994). Measurements are now repeatable to the 1 or 2 microgal level (1 gal = 1 cm/s$^2$). For instance, a joint USA-Canada effort to produce the first reliable gravimetric measurement of glacial rebound shows at Churchill, Canada, that gravity has been decreasing at 1.6 microgal/yr since 1987. Theoretical models which account for viscous rebound of crust and mantle beneath Canada predict a linear relationship between gravity fall and crustal rise of about 0.15 microgal/mm. These gravity observations therefore indicate crustal uplift to be 11 mm/yr. At the same location, tide gauge records show that the Hadon Bay water level has been falling at an average rate of 11 mm/yr for the past 50 years.

Even if land levels were to be measured to good precision, there would still in principle be corresponding changes in the geoid to take into account. For example, Wagner and McAdoo (1986) presented maps of secular trends in geoid heights to be expected from post-glacial rebound. However, these are approximately an order of magnitude less than the corresponding changes in vertical land movements, and in studies of sea level changes over century time-scales or less, the geoid is considered in most applications to be time independent. Repeated space gravity missions, together with long-term laser tracking of dedicated geodetic satellites, should ideally be mounted to monitor such changes, while they can continue to be studied within geodynamic models.

7.6.2 Dynamic Effects

The variability of sea level on interannual to interdecadal time-scales is a major complication in determining reliable sea level trends and accelerations (Douglas, 1992). It is also clear that the global average sea level change is not a good indicator of local changes at any particular place.

In large part, these interannual to interdecadal fluctuations result from the fact that sea level topography, referenced to the geoid, is closely related to the dynamics and thermodynamics of the ocean. If the ocean were homogeneous and at rest, with a uniform atmospheric pressure field above it and no wind, the sea surface would correspond to the geoid (i.e., an equipotential of the gravity field). However, it does not; it differs from the geoid by ± one metre. Ocean and atmosphere are non-homogeneous
and continuously moving within a variety of time and space scales, under gravitational forcing (for tides) and thermal forcing from the Sun (including variable wind stress and heat and fresh water exchanges at the sea surface).

In analysing trends, high-frequency ocean signals, swells, tides and surges are generally easily removed from sea level records by filtering techniques, as illustrated, among others, by Chelton and Enfield (1986) and Sturges (1987). But the analysis of the low-frequency residuals remains extremely difficult because these data have red spectra (the spectral energy keeps rising at low frequency), and the variability of the lowest frequencies that can generally be resolved with the longer available sea level records is of the same amplitude as (or larger than) the rising sea level signal. It is thus very important to understand the physical causes of these long-period events in order to be able to correct the data and improve the signal-to-noise ratio. Unfortunately, the forcing mechanisms are not well-known, especially at the decadal and interdecadal time-scale. The processes involve natural oscillations of the ocean-atmosphere system, which result in perturbations of the three dimensional state of the ocean, in terms of the dynamics and thermodynamics and the feedback on the state of the atmosphere.

At the short-term interannual frequencies, one major event is the El Niño-Southern Oscillation phenomenon in the Pacific Ocean. Associated sea level oscillations have been clearly observed by in situ tide gauges and satellite altimetry, typically with eastward propagating equatorial Kelvin waves and westward reflected or locally-forced Rossby waves (Wyrtki, 1979, 1985; Miller and Cheney, 1990). Monthly maps of sea level anomalies are now produced routinely to document and follow this interannual variability of the Pacific Equatorial Ocean.

Along single coastlines, the efficiency of wave propagation processes is a major candidate for explaining the often observed coherency of the long-period sea level variabilities, as studied by Enfield and Allen (1980) and Chelton and Davis (1982) for the Pacific coast of the USA. At basin scale, long-period baroclinic Rossby wave propagation has been demonstrated to possibly lead to coherent sea level signals; Sturges (1987) thus explained the 5–8 years period coherency in the long records available for San Francisco and Honolulu, with amplitudes of 5–15 cm at these long periods.

At the interannual to decadal periods, sea level fluctuations must often be driven primarily by atmospheric forcing, wind and pressure. Maul and Hanson (1991) have observed a significant coherency, with peaks at 4 and 13 years, between the sea level variability of the tidal records along the Atlantic coast of the USA and the long-period, basin-scale variations in North Atlantic atmospheric surface pressure known as the North Atlantic Oscillation (Rogers, 1984). Additionally, Sturges and Hong (1995) have demonstrated that, at Bermuda, sea level and thermocline variability, estimated from a simple ocean model forced by the wind, is in remarkably good agreement with observations at long periods.

However, sea surface temperature variations, and hence buoyancy forcing by the atmosphere, are coherent with changes in the wind (Kushnir, 1994). Thus, thermodynamic processes also have to be considered in interpreting long-term sea level variations. Decadal to interdecadal changes in temperature and salinity of the three-dimensional structure of the ocean have been widely reported in many places (Gordon et al., 1992). Taking the North Atlantic as an example, one can refer to the Great Salinity Anomaly traced by Dickson et al. (1988) around the North Atlantic subpolar gyre from the mid-1960s to the late 1970s, and to the cooling at intermediate depth in the North Atlantic subpolar gyre observed by Lazier and Gerschey (1991), Read and Gould (1992) and Koltmann and Sy (1994), due to drastic changes in the production rates of Labrador Sea Water (LSW) and its property characteristics. This freshening is also present in the 24.5°N sections of hydrographic data, between 400 and 500 m (Lavin, 1993), due to some amount of cooler LSW circulating in the subtropical gyre. The complex picture emerging from these observations is related to the North Atlantic Deep Water (NADW) production, which propels the global planetary Thermohaline Conveyor Belt circulation (Broecker et al., 1985). Changes in the production rate of NADW must change the northward transport of upper ocean warm water, and it could lead to an enhancement of sea rise in the mid-latitudes of the North Atlantic (Mikolajewicz et al., 1990).

Large-scale ocean circulation has a direct signature on the sea surface topography, through geostrophic balance. Sea surface slope variabilities are thus thought to be good indicators of the large-scale ocean transport variabilities. Maud et al. (1990) have shown that it is true for the Gulf Stream for the semi-annual and annual band. However, as noticed by Sturges and Hong (1995) this has not been fully demonstrated for interannual and lower frequencies from the analysis of long coastal sea level records, until recently the only available source of data (see also Whitworth and Peterson, 1985, for the Antarctic Circumpolar Current). Hence, there is interest in long-term, deep-sea bottom pressure measurements (as used in WOCE) and satellite altimetry for studying this relation between ocean slope and ocean transport variabilities.
As noted by Gates et al. (1992), there are now a sufficient number of AOGCM integrations over 50–100 year and longer periods to provide preliminary information on the simulation of atmospheric and oceanic decadal variability. Over these last years, a great deal of interest has focused on the behaviour of the thermohaline circulation of the North Atlantic, the formation of NADW, the Conveyor Belt Circulation, and the possible existence of multiple equilibrium states for the global ocean circulation. The effect of these processes on the interdecadal variabilities of sea level has been noted above. Another mode of decadal climate variability, in the North Pacific Ocean, has been found in AOGCMs of the MPI, attributed to a cycle involving air-sea interactions between the sub-tropical gyre circulation and the Aleutian low pressure system (Latif and Barnett, 1994). Besides its effect on sea surface temperature, it has an effect on sea level of 3.4 cm in decadal variability.

Coupled model simulations of climate change under increasing CO₂ have also shown that the regional differences in sea level change are larger than the globally averaged change (Mikolajewicz et al., 1990; Gregory, 1993; Cubasch et al., 1994b). This spatial variation comes mostly from the geographical distribution of surface temperature changes, but ocean dynamics do modify it, especially in areas where temperature changes occur to considerable depth. Different climate models give different patterns of local response, although there are some common features, such as a strong relative rise off the east coast of North America, and a marked relative fall north of the Ross Sea. An example from Gregory (1993) is shown in Figure 7.15. However, the non-equilibrium response in sea level of the world ocean to warming is, as yet, very poorly understood, and it is not clear that the present generation of AOGCMs can resolve such a process correctly.

Since the launch of the TOPEX/POSEIDON altimeter satellite in 1992, it has been possible to construct similar global maps of sea level change from direct measurements (Cheney et al., 1994). As in the model simulations, regional variations observed by the altimeter are quite large, with amplitudes of the order of 5 cm on monthly time-scales, as compared to the global average increase of 3 mm/yr derived from these data. The combination of altimetry with sub-surface data and winds will provide a way of interpreting these low-frequency sea level phenomena.

7.6.3 Trends in Extremes

The statistical treatment of data on extreme sea levels has progressed considerably over the past few years. For example, Tawn and Vassie (1991) developed a "revised joint probability method" applicable to all types of tidal regime, while Coles and Tawn (1990) employed

![Figure 7.15: Relative sea level change from a transient AOGCM experiment over the years 66–75 (the decade centred on the time of CO₂ doubling). Relative sea level change here is calculated as the difference between the anomaly and control minus the change in global average sea level of 10.2 cm (source: Gregory, 1993).](image)
multivariate extremes and the spatial properties of extreme processes to improve estimations at nearby sites with little or no data. Extreme levels, computed by a variety of old and new methods, are available for much of the world coastline for input to coastal impact studies (e.g., see de Mesquita and Franca, 1990) and Simon (1994), for coastal areas in Brazil and France, respectively.

The determination of a trend in extreme values is a difficult procedure since, by definition, the data become increasingly rare as the extremes are approached and estimates of statistics in the tails have increasingly wide confidence limits. Secular trends in annual maxima have tended to be studied primarily in north-west Europe where long records exist. For example, the spatial distribution of secular trends in annual maxima around the UK appears to be similar on average to those derived from tide gauge mean sea level and geological sea level data sets (Dixon and Tawn, 1992). Extreme levels, and their apparent temporal variation, have been studied at a number of “cities on water”, such as Venice (Frassetto, 1991; Pirazzoli, 1991; Rusconi, 1993).

The continental shelf of north-west Europe is one example of an area where trends in storm surge activity have been studied. From analyses of storm surge frequency around the coast of the UK over the past 70 years, and at the Hook of Holland (Hoozemans and Wiersma, 1992) and the German coast over the past few centuries (Rodhe, 1980), there is no evidence of a long-term trend in storm surge activity, although Rodhe did point to a possible low-frequency (approximately 80-year) periodicity in flooding. On the other hand, evidence exists for a trend over the same period in regional winds and air pressures that would result in enhanced German Bight mean sea level (and the surge activity that it reflects) based on gridded Norwegian meteorological data for the shelf applied to a numerical model (Tsimplis et al., 1994). Meanwhile, Führbötter and Toppe (1991) observed tide gauge data in the German Bight directly and also indicate an increase of storm surge frequency in the area, at least since 1960 and possibly before. Local wind data for the Bight itself appear to show no obvious trend (Hoozemans and Wiersma, 1992) from 1950 to 1980, and in spite of large interdecadal variation, very little at all over century time-scales (Schmidt, 1991; Schmidt and von Storch, 1993). Clearly, findings will depend on the periods analysed and data spans. However, the difficulty in drawing conclusions relevant to climatic change studies from this relatively well-instrumented part of the world cautions against drawing conclusions for areas with considerably less data.

Numerical tidal models are frequently employed, with reductions in ocean depth, for the study of palaeotides (Scott and Greenberg, 1983) and the historical development of bedforms (Proctor and Carter, 1989; Austin, 1991). Conversely, they can be run with increased depth to predict changes in tidal amplitude that might accompany potential future sea level rise (Rijkswaterstaat, 1986). The effect of changes in depth (a surrogate for sea level rise) on storm surges can also be investigated. For example, Flather and Khandker (1993) described how tides and surges might be modified in the Bay of Bengal if sea level is modified, assuming that the character of the prevailing meteorology does not also change. In most cases, there will be little change in tide and surge (of order of a few centimetres per metre of sea level rise, unless in locally resonant situations).

7.7 Major Uncertainties and How to Reduce Them

7.7.1 Monitoring of Sea Level Change

For a future global sea level monitoring system, it is essential that information is integrated from many sources, with global and regional products (useful to scientists and non-scientists alike) derived from the blended data sets (Eden, 1990). Improvements in such a monitoring system would include five elements.

1. It should include a global sea level monitoring system based on a network of modern tide gauges. The Global Sea Level Observing System (GLOSS) of the Intergovernmental Oceanographic Commission (IOC) is a co-ordinated project for the monitoring of long-term global sea level change and is intended to serve the various purposes of oceanographic and climate change research into the next century (IOC, 1990; Woodworth, 1991). GLOSS consists of a network of approximately 300 tide gauges worldwide, of which over two-thirds are now operational. Technical developments in recent years have seen many of the traditional float and stilling well tide gauges replaced by modern systems based on pneumatic and acoustic principles (Spencer, 1992). Many of these have satellite or telephone data transmission equipment, enabling real-time data access and fault checking. Bottom pressure recorders (Spencer and Vassie, 1985), inverted echo sounders (Wimbush, 1990) and thermistor chain moorings for dynamic height (McPhaden, 1993) now provide quasi-sea level measurements in several areas of the deep ocean, which will provide information on the ocean circulation.
Changes in Sea Level

(2) An improved geodetic network is required which enables many of the sea level measurements to be placed in a stable global co-ordinate system with sub-centimetric accuracy, and which provides a decoupling of land and ocean level signals in the tide gauge records. Remarkable progress has been made in the past five years toward achieving the required network, through the development of the International Terrestrial Reference Frame (ITRF) by the International Earth Rotation Service (IERS). The ITRF is based on Very Long Baseline Interferometry (VLBI), Satellite Laser Ranging (SLR) and Global Positioning System (GPS) observations (Carter et al., 1989; Bilham, 1991; Ashkenazi et al., 1993; Baker, 1993; Carter, 1994). At many locations measurements of land movements near to the tide gauge sites by means of GPS will be supplemented by absolute gravity recording, with a change of gravity of 1 microgal (approximately the current accuracy of the technique) corresponding to a change of land level of 5 mm (Carter, 1994). VLBI, GPS and SLR enable precise time-series of polar motion and changes in rotation rate to be compiled, which may be pertinent to the problems addressed in this chapter.

(3) The monitoring system should include near-global observations of the ocean and ice sheets by means of satellite altimetry. Radar altimetry has become a major tool for studying sea level changes over most of the world ocean. Although a considerable amount of altimeter data have been acquired previously from, for example, the USA Navy Geosat satellite and the European ERS-1 mission, the launch of TOPEX/POSEIDON in 1992 marked the start of a new era of precise sea level measurements from space. TOPEX/POSEIDON is providing sea level data with a single-point precision of about 5 cm (Fu et al., 1994), enabling global mean sea level to be measured every 10 days with a precision of a few mm. Indeed, results from the first 2 years of the mission suggest an apparent sea level rise of approximately 3 mm/y (Wagner et al., 1994; Nerem, 1995). Given the number of potential altimeter errors at the millimetre level, it is premature to attach major significance to this result, but if measurements of this quality can be calibrated and maintained throughout a long-term program of multiple altimeter missions and integrated with the long-term GLOSS gauges, global sea level monitoring will be placed on a much firmer basis (Koblinsky et al., 1992). Meanwhile, ice sheet topographies from ERS-1 and other near-polar orbiting radar altimetric satellites should result in complementary data sets of ice balance. When operational, laser altimetry over ice should be more precise than, and should complement, the radar altimeter data. However, the full potential of satellite altimetry to measure changes in ice mass will not be realised until truly global coverage is available.

Altimetry is a very important tool now available for oceanographic research, providing new insights into ocean tides, the ocean mesoscale, and basin-scale changes, in addition to global sea level change. There are corresponding advances in geophysics through the acquisition of detailed maps of the mean sea surface. In order to compute fields of absolute ocean currents for input to climate models with predictive sea level capability, precise altimetry of the mean sea surface must be complemented by detailed knowledge of the geoid by means of space gravity missions (Koblinsky et al., 1992).

(4) Measurements are required of the temperature and salinity fields of the global ocean. Variations in long distance acoustic pulse travel times can provide for monitoring of average basin temperatures (Munk, 1989).

(5) For a better understanding of the ocean response to climate variability and future sea level rise, there is a need for further developments of coupled atmosphere, ocean and ice models which explicitly include sea level predictions. The continued growth in computer power will strongly support this effort, as it allows the complexity of the models to increase, thus improving their performance and realism. The scientific programmes WOCE and CLIVAR (Climate Variability programme) and the development of GOOS will provide the necessary observations to feed these models. Data assimilation techniques will thus be of primary importance in combining such data sets with models for their optimal use in understanding the climatic role of the ocean and predicting the evolution of sea level.

7.7.2 Oceanic Thermal Expansion
A strategy for long-term observation of the three-dimensional state of the ocean needs to be defined. This is the second goal of WOCE and will be developed under new programmes, like the oceanic component of CLIVAR and GOOS.
Observations derived from three-dimensional monitoring of the ocean will then have to be combined with models in order to make predictions. It can be anticipated that the on-going progress in ocean modelling (Semtner, 1994), and in data assimilation techniques within these models, will allow the improvement of our understanding of the interannual to interdecadal variabilities of the thermodynamic state of the ocean and the thermal contribution to changes in sea level, both temporally and spatially. Thus, continued support of both monitoring programmes and dynamic ocean modelling are necessary to enable global and regional predictions of future sea level changes.

7.7.3 Glaciers and Ice Caps

There remain large uncertainties regarding the future change in global glacier volume as a consequence of global warming. There are four major gaps in knowledge that need to be filled in order to better estimate the contributions of glaciers to sea level change.

1. Probably the most critical need is the development of simple, yet realistic, models which include the processes linking meteorology to mass balance to dynamic response, and which include the resulting feedbacks (Meier, 1965). The importance of incorporating dynamic feedback is obvious when considering that in the next 50–100 years the estimated glacier volume change is expected to be a significant fraction of the total present volume, resulting in a major change in glacier area. Thus, the contribution of glaciers to sea level rise will reach a maximum and then decline despite continuing warming (Kuhn, 1985). The meteorological calculations can be accomplished by simple hydrometeorological (HM) models in which accumulation and energy balance are approximated by functions of seasonal precipitation and air temperature (e.g., Liestol, 1967; Martin, 1978; Tangborn, 1980; Reeh, 1991), perhaps parameterized for mountain glaciers as functions of altitude (e.g., Kuhn, 1981; Oerlemans and Hoogendoorn, 1989; Laumann and Reeh, 1993). The dynamic processes can be modelled using conventional glacier flow models (e.g., Bindschadler, 1982; Stroeven et al., 1989) or by simpler models that address the response times (Johannesson et al., 1989; Schwitter and Raymond, 1993).

2. Model studies need to be extended to ice masses that are very different from the small mountain glaciers of the Alps and Scandinavia in order to encompass the broad spectrum of glacier behaviour. Current models are based on well-studied reference glaciers, practically all of which are small, mid-latitude glaciers (Kuhn, 1993). However, in terms of potential sea level rise, these glaciers are relatively insignificant. The glaciers that are especially critical with respect to sea level rise include the following: the large valley and piedmont glaciers of south-east Alaska; the Patagonian Ice Caps; and the monsoon-nourished glaciers of central Asia. The ice caps of the Arctic are also important and have regimes significantly different from those commonly modelled.

3. There is a need to further quantify the process of internal accumulation (refreezing of meltwater) for incorporation into mass balance models. Internal accumulation (Trabant and Mayo, 1985; Pfeffer et al., 1991; Reeh, 1991) is pervasive at high latitudes and at high altitudes, yet is frequently neglected in glacier-meteorological studies.

4. Better understanding is required of iceberg calving flux and its interaction with ice flow dynamics. Iceberg calving is a major component of the mass balance of some large glaciers in the Arctic, Alaska, and at high southern latitudes. However, the iceberg calving flux interacts with ice flow dynamics in ways that are not well understood (Meier, 1994). This problem needs further investigation and incorporation into models.

7.7.4 Greenland and Antarctic Ice Sheets

Of all the terms that enter the sea level change equation, the largest uncertainties pertain to the Earth’s major ice sheets. Relatively small changes in these ice sheets could have major effects on global sea level, yet we are not even certain of the sign of their present contribution. Obviously, better determination of the present ice sheet mass balances are needed, and additional historical (palaeoclimatic and palaeoglaciologic) studies are required in order to learn how changes have occurred in the past. The results of such studies need to be incorporated into improved modelling schemes in order to anticipate how the ice sheets will change in response to future climate perturbations.

Field (surface-based) observations of mass balances – the accumulation of snow and superimposed ice and the discharge of icebergs and meltwater – need to be continued and extended. Special attention also needs to be given to detecting change in ice dynamics, including the flow of ice streams and iceberg-calving rates. Additional observations
of atmospheric water-vapour flux divergence would be useful in order to better define the processes leading to possible changes in snow accumulation. New ice cores in Greenland and high time-resolution ice cores from Antarctica will be needed to show the spatial and temporal changes in atmospheric circulation that occurred with the abrupt climate changes in the past, in order to test models of climate change in the high latitudes.

Special attention needs to be given to current changes in ice shelves, such as the break-up of a large part of the Larsen Ice Shelf on the Antarctic Peninsula in 1995, and to dynamic interactions between ice sheets, ice streams and ice shelves. While ice shelf break-up does not contribute directly to sea level rise, changes in the interconnected grounded ice will affect sea level in ways that cannot be predicted at the present. Improved understanding will require combined studies using glaciological, oceanographic and satellite observations. Because sea ice extent affects water vapour transport in the Antarctic, better definitions of variations in sea ice extent and of the distribution of ice thickness are needed.

The current mass balances of the Antarctic and Greenland Ice Sheets need to be determined and future changes monitored. Perhaps the most powerful tool for doing this in the near future is satellite altimetry. A laser altimeter is urgently needed on a polar-orbiting satellite, especially to detect changes in the low-accumulation regions and critical ice sheet margins where the existing radar altimetry is not adequate. These measurements should commence as soon as possible in order to provide a baseline for detecting greenhouse-induced changes in the future. The observed seasonal and interannual variations in surface elevation will provide information on precipitation variations, which can be used in energy-balance models and to test atmospheric general circulation model results. Accurate mapping of surface elevation changes by laser altimetry will also permit monitoring of ice sheet stability. Early warning of ice sheet collapse would thus be obtained before major ice shelf break-up or massive iceberg discharges were observed. Satellite synthetic-aperture radar interferometry may allow monitoring of ice stream velocities and grounding zone locations (Goldstein et al., 1993; Herzfeld, 1994).

Improved modelling of the ice sheets is vitally needed. This should involve developing and coupling more realistic models of both atmospheric circulation and ice dynamics. Major uncertainties are associated with estimating future snowfall on the major ice sheets. Most existing models relate accumulation changes to changes in air temperature, which assumes that warmer air delivers more snow and that atmospheric circulation will not change. Palaeoclimatic data show that storm strengths and trajectories have changed in the past and have greatly affected accumulation. This raises the possibility that future circulation changes will occur and will also affect precipitation. It is important that atmospheric and snow-surface processes in ice sheet regions be understood well enough so that model-based predictions of snow accumulation can be made directly, rather than using predictions of temperature and assumed temperature sensitivity to estimate precipitation. Palaeoclimatic data from high-resolution ice cores are needed to test these models.

Improvement is also needed in the ice dynamics models. Short-term, transient behaviour has been observed in major ice streams. The fact that these changes exist means that the physics incorporated in contemporary ice sheet models is not completely realistic. In particular, treatment of basal sliding and/or deforming basal till, together with the longitudinal stresses involved, needs to be incorporated. Another key aspect, currently not well understood and therefore impossible to model with confidence, is the rate of iceberg calving. These are the key elements in defining the interactions between ice sheets, ice streams and ice shelves. These, together with changes of snow accumulation, are the most uncertain aspects of predicting changes in the ice sheets over the next decades to centuries, and need major improvement.

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