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Changes in Sea Level

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Executive Summary

This chapter assesses the current state of knowledge of the rate of change of global average and regional sea level in relation to climate change. We focus on the 20th and 21st centuries. However, because of the slow response to past conditions of the oceans and ice sheets and the consequent land movements, we consider changes in sea level prior to the historical record, and we also look over a thousand years into the future.

Past changes in sea level
From recent analyses, our conclusions are as follows:

• Since the Last Glacial Maximum about 20,000 years ago, sea level has risen by over 120 m at locations far from present and former ice sheets, as a result of loss of mass from these ice sheets. There was a rapid rise between 15,000 and 6,000 years ago at an average rate of 10 mm/yr.

• Based on geological data, global average sea level may have risen at an average rate of about 0.5 mm/yr over the last 6,000 years and at an average rate of 0.1 to 0.2 mm/yr over the last 3,000 years.

• Vertical land movements are still occurring today as a result of these large transfers of mass from the ice sheets to the ocean.

• During the last 6,000 years, global average sea level variations on time-scales of a few hundred years and longer are likely to have been less than 0.3 to 0.5 m.

• Based on tide gauge data, the rate of global average sea level rise during the 20th century is in the range 1.0 to 2.0 mm/yr, with a central value of 1.5 mm/yr (as with other ranges of uncertainty, it is not implied that the central value is the best estimate).

• Based on the few very long tide gauge records, the average rate of sea level rise has been larger during the 20th century than the 19th century.

• No significant acceleration in the rate of sea level rise during the 20th century has been detected.

• There is decadal variability in extreme sea levels but no evidence of widespread increases in extremes other than that associated with a change in the mean.

Factors affecting present day sea level change
Global average sea level is affected by many factors. Our assessment of the most important is as follows.

• Ocean thermal expansion leads to an increase in ocean volume at constant mass. Observational estimates of about 1 mm/yr over recent decades are similar to values of 0.7 to 1.1 mm/yr obtained from Atmosphere-Ocean General Circulation Models (AOGCMs) over a comparable period. Averaged over the 20th century, AOGCM simulations result in rates of thermal expansion of 0.3 to 0.7 mm/yr.

• The mass of the ocean, and thus sea level, changes as water is exchanged with glaciers and ice caps. Observational and modelling studies of glaciers and ice caps indicate a contribution to sea level rise of 0.2 to 0.4 mm/yr averaged over the 20th century.

• Climate changes during the 20th century are estimated from modelling studies to have led to contributions of between –0.2 and 0.0 mm/yr from Antarctica (the results of increasing precipitation) and 0.0 to 0.1 mm/yr from Greenland (from changes in both precipitation and runoff).

• Greenland and Antarctica have contributed 0.0 to 0.5 mm/yr over the 20th century as a result of long-term adjustment to past climate changes.

• Changes in terrestrial storage of water over the period 1910 to 1990 are estimated to have contributed from –1.1 to +0.4 mm/yr of sea level rise.

The sum of these components indicates a rate of eustatic sea level rise (corresponding to a change in ocean volume) from 1910 to 1990 ranging from –0.8 to 2.2 mm/yr, with a central value of 0.7 mm/yr. The upper bound is close to the observational upper bound (2.0 mm/yr), but the central value is less than the observational lower bound (1.0 mm/yr), i.e., the sum of components is biased low compared to the observational estimates. The sum of components indicates an acceleration of only 0.2 mm/yr/century, with a range from –1.1 to +0.7 mm/yr/century, consistent with observational finding of no acceleration in sea level rise during the 20th century. The estimated rate of sea level rise from anthropogenic climate change from 1910 to 1990 (from modelling studies of thermal expansion, glaciers and ice sheets) ranges from 0.3 to 0.8 mm/yr. It is very likely that 20th century warming has contributed significantly to the observed sea level rise, through thermal expansion of sea water and widespread loss of land ice.

Projected sea level changes from 1990 to 2100
Projections of components contributing to sea level change from 1990 to 2100 (this period is chosen for consistency with the IPCC Second Assessment Report), using a range of AOGCMs following the IS92a scenario (including the direct effect of sulphate aerosol emissions) give:

• thermal expansion of 0.11 to 0.43 m, accelerating through the 21st century;
• a glacier contribution of 0.01 to 0.23 m;
• a Greenland contribution of –0.02 to 0.09 m;
• an Antarctic contribution of –0.17 to 0.02 m.

Including thawing of permafrost, deposition of sediment, and the ongoing contributions from ice sheets as a result of climate change since the Last Glacial Maximum, we obtain a range of

global-average sea level rise from 0.11 to 0.77 m. This range reflects systematic uncertainties in modelling.

For the 35 SRES scenarios, we project a sea level rise of 0.09 to 0.88 m for 1990 to 2100, with a central value of 0.48 m. The central value gives an average rate of 2.2 to 4.4 times the rate over the 20th century. If terrestrial storage continued at its present rates, the projections could be changed by –0.21 to +0.11 m. For an average AOGCM, the SRES scenarios give results which differ by 0.02 m or less for the first half of the 21st century. By 2100, they vary over a range amounting to about 50% of the central value. Beyond the 21st century, sea level rise will depend strongly on the emissions scenario.

The West Antarctic ice sheet (WAIS) has attracted special attention because it contains enough ice to raise sea level by 6 m and because of suggestions that instabilities associated with its being grounded below sea level may result in rapid ice discharge when the surrounding ice shelves are weakened. The range of projections given above makes no allowance for ice-dynamic instability of the WAIS. It is now widely agreed that major loss of grounded ice and accelerated sea level rise are very unlikely during the 21st century.

Our confidence in the regional distribution of sea level change from AOGCMs is low because there is little similarity between models. However, models agree on the qualitative conclusion that the range of regional variation is substantial compared with the global average sea level rise. Nearly all models project greater than average rise in the Arctic Ocean and less than average rise in the Southern Ocean.

Land movements, both isostatic and tectonic, will continue through the 21st century at rates which are unaffected by climate change. It can be expected that by 2100 many regions currently experiencing relative sea level fall will instead have a rising relative sea level.

Extreme high water levels will occur with increasing frequency (i.e. with reducing return period) as a result of mean sea level rise. Their frequency may be further increased if storms become more frequent or severe as a result of climate change.

Longer term changes

If greenhouse gas concentrations were stabilised, sea level would nonetheless continue to rise for hundreds of years. After 500 years, sea level rise from thermal expansion may have reached only half of its eventual level, which models suggest may lie within ranges of 0.5 to 2.0 m and 1 to 4 m for CO_2 levels of twice and four times pre-industrial, respectively.

Glacier retreat will continue and the loss of a substantial fraction of the total glacier mass is likely. Areas that are currently marginally glaciated are most likely to become ice-free.

Ice sheets will continue to react to climate change during the next several thousand years even if the climate is stabilised. Models project that a local annual-average warming of larger than 3°C sustained for millennia would lead to virtually a complete melting of the Greenland ice sheet. For a warming over Greenland of 5.5°C, consistent with mid-range stabilisation scenarios, the Greenland ice sheet contributes about 3 m in 1,000 years. For a warming of 8°C, the contribution is about 6 m, the ice sheet being largely eliminated. For smaller warmings, the decay of the ice sheet would be substantially slower.

Current ice dynamic models project that the WAIS will contribute no more than 3 mm/yr to sea level rise over the next thousand years, even if significant changes were to occur in the ice shelves. However, we note that its dynamics are still inadequately understood to make firm projections, especially on the longer time-scales.

Apart from the possibility of an internal ice dynamic instability, surface melting will affect the long-term viability of the Antarctic ice sheet. For warmings of more than 10°C, simple runoff models predict that a zone of net mass loss would develop on the ice sheet surface. Irreversible disintegration of the WAIS would result because the WAIS cannot retreat to higher ground once its margins are subjected to surface melting and begin to recede. Such a disintegration would take at least a few millennia. Thresholds for total disintegration of the East Antarctic Ice Sheet by surface melting involve warmings above 20°C, a situation that has not occurred for at least 15 million years and which is far more than predicted by any scenario of climate change currently under consideration.

11.1 Introduction

Sea level change is an important consequence of climate change, both for societies and for the environment. In this chapter, we deal with the measurement and physical causes of sea level change, and with predictions for global-average and regional changes over the next century and further into the future. We reach qualitatively similar conclusions to those of Warrick *et al*. (1996) in the IPCC WGI Second Assessment Report (IPCC, 1996) (hereafter SAR). However, improved measurements and advances in modelling have given more detailed information and greater confidence in several areas. The impacts of sea level change on the populations and eco-systems of coastal zones are discussed in the IPCC WGII TAR (IPCC, 2001).

The level of the sea varies as a result of processes operating on a great range of time-scales, from seconds to millions of years. Our concern in this report is with climate-related processes that have an effect on the time-scale of decades to centuries. In order to establish whether there is a significant anthropogenic influence on sea level, the longer-term and non-climate-related processes have to be evaluated as well.

"Mean sea level" at the coast is defined as the height of the sea with respect to a local land benchmark, averaged over a period of time, such as a month or a year, long enough that fluctuations caused by waves and tides are largely removed. Changes in mean sea level as measured by coastal tide gauges are called "relative sea level changes", because they can come about either by movement of the land on which the tide gauge is situated or by changes in the height of the adjacent sea surface (both considered with respect to the centre of the Earth as a fixed reference). These two terms can have similar rates (several mm/yr) on time-scales greater than decades. To infer sea level changes arising from changes in the ocean, the movement of the land needs to be subtracted from the records of tide gauges and geological indicators of past sea level. Widespread land movements are caused by the isostatic adjust-ment resulting from the slow viscous response of the mantle to the melting of large ice sheets and the addition of their mass to the oceans since the end of the most recent glacial period ("Ice Age") (Section 11.2.4.1). Tectonic land movements, both rapid displacements (earthquakes) and slow movements (associated with mantle convection and sediment transport), can also have an important effect on local sea level (Section 11.2.6).

We estimate that global average eustatic sea level change over the last hundred years is within the range 0.10 to 0.20 m (Section 11.3.2). ("Eustatic" change is that which is caused by an alteration to the volume of water in the world ocean.) These values are somewhat higher than the sum of the predictions of the contributions to sea level rise (Section 11.4). The discrep-ancy reflects the imperfect state of current scientific knowledge. In an attempt to quantify the processes and their associated rates of sea level change, we have critically evaluated the error estimates (Box 11.1). However, the uncertainties remain substantial, although some have narrowed since the SAR on account of improved observations and modelling.

Box 11.1: Accuracy

For indicating the uncertainty of data (measurements or model results), two options have been used in this chapter.

1. For data fulfilling the usual statistical requirements, the uncertainty is indicated as $\pm$ 1 standard deviation ($\pm 1\sigma$).
2. For limited data sets or model results, the full range is shown by quoting either all available data or the two extremes. In these cases, outliers may be included in the data set and the use of an arithmetic mean or central value might be misleading.

To combine uncertainties when adding quantities, we used the following procedures:
- Following the usual practice for independent uncertainties, the variances were added (i.e. the standard deviations were combined in quadrature).
- Ranges were combined by adding their extreme values, because in these cases the true value is very likely to lie within the overall range.
- To combine a standard deviation with a range, the standard deviation was first used to derive a range by taking $\pm$ 2 standard deviations about the mean, and then the ranges were combined.

Eustatic sea level change results from changes to the density or to the total mass of water. Both of these relate to climate. Density is reduced by thermal expansion occurring as the ocean warms. Observational estimates of interior temperature changes in the ocean reported by Warrick *et al*. (1996) were limited, and estimates of thermal expansion were made from simple ocean models. Since the SAR, more observational analyses have been made and estimates from several Atmosphere-Ocean General Circulation Models (AOGCMs) have become available (Section 11.2.1). Thermal expansion is expected to contribute the largest component to sea level rise over the next hundred years (Section 11.5.1.1). Because of the large heat capacity of the ocean, thermal expansion would continue for many centuries after climate had been stabilised (Section 11.5.4.1).

Exchanges with water stored on land will alter the mass of the ocean. (Note that sea level would be unaffected by the melting of sea ice, whose weight is already supported by the ocean.) Groundwater extraction and impounding water in reservoirs result in a direct influence on sea level (Section 11.2.5). Climate change is projected to reduce the amount of water frozen in glaciers and ice caps (Sections 11.2.2, 11.5.1.1) because of increased melting and evaporation. Greater melting and evaporation on the Greenland and Antarctic ice sheets (Sections 11.2.3, 11.5.1.1) is also projected, but might be outweighed by increased precipita-tion. Increased discharge of ice from the ice sheets into the ocean is also possible. The ice sheets react to climate change by adjusting their shape and size on time-scales of up to millennia, so they could still be gaining or losing mass as a result of climate variations over a history extending as far back as the last glacial period, and they would continue to change for thousands of years after climate had been stabilised (Section 11.5.4.3).

Sea level change is not expected to be geographically uniform (Section 11.5.2), so information about its distribution is needed to inform assessments of the impacts on coastal regions. Since the SAR, such information has been calculated from several AOGCMs. The pattern depends on ocean surface fluxes, interior conditions and circulation. The most serious impacts are caused not only by changes in mean sea level but by changes to extreme sea levels (Section 11.5.3.2), especially storm surges and exceptionally high waves, which are forced by meteorological conditions. Climate-related changes in these therefore also have to be considered.

11.2 Factors Contributing to Sea Level Change

11.2.1 Ocean Processes

The pattern of sea level in ocean basins is maintained by atmospheric pressure and air-sea fluxes of momentum (surface wind stress), heat and fresh water (precipitation, evaporation, and fresh-water runoff from the land). The ocean is strongly density stratified with motion preferentially along density surfaces (e.g. Ledwell et al., 1993, 1998). This allows properties of water masses, set by interaction with the atmosphere or sea ice, to be carried thousands of kilometres into the ocean interior and thus provides a pathway for warming of surface waters to enter the ocean interior.

As the ocean warms, the density decreases and thus even at constant mass the volume of the ocean increases. This thermal expansion (or steric sea level rise) occurs at all ocean temperatures and is one of the major contributors to sea level changes during the 20th and 21st centuries. Water at higher temperature or under greater pressure (i.e., at greater depth) expands more for a given heat input, so the global average expansion is affected by the distribution of heat within the ocean. Salinity changes within the ocean also have a significant impact on the local density and thus local sea level, but have little effect on global average sea level change.

The rate of climate change depends strongly on the rate at which heat is removed from the ocean surface layers into the ocean interior; if heat is taken up more readily, climate change is retarded but sea level rises more rapidly. Climate change simulation requires a model which represents the sequestration of heat in the ocean and the evolution of temperature as a function of depth.

The large heat capacity of the ocean means that there will be considerable delay before the full effects of surface warming are felt throughout the depth of the ocean. As a result, the ocean will not be in equilibrium and global average sea level will continue to rise for centuries after atmospheric greenhouse gas concentrations have stabilised.

The geographical distribution of sea level change is principally determined by alterations to the ocean density structure, with consequent effects on ocean circulation, caused by the modified surface momentum, heat and water fluxes. Hsieh and Bryan (1996) have demonstrated how the first signals of sea level rise are propagated rapidly from a source region (for instance, a region of heat input) but that full adjustment takes place more slowly. As a result, the geographical distribution of sea level change may take many decades to centuries to arrive at its final state.

11.2.1.1 Observational estimates of ocean warming and ocean thermal expansion

Previous IPCC sea level change assessments (Warrick and Oerlemans, 1990; Warrick et al., 1996) noted that there were a number of time-series which indicate warming of the ocean and a resultant thermal expansion (i.e. a steric sea level rise) but there was limited geographical coverage. Comparison of recent ocean temperature data sets (particularly those collected during the World Ocean Circulation Experiment) with historical data is beginning to reveal large-scale changes in the ocean interior. (Section 2.2.2.5 includes additional material on ocean warming, including studies for which there are no estimates of ocean thermal expansion.) However, the absence of comprehensive long ocean time-series data makes detection of trends difficult and prone to contamination by decadal and interannual variability. While there has been some work on interannual variability in the North Atlantic (e.g. Levitus, 1989a,b, 1990) and North Pacific (e.g. Yasuda and Hanawa, 1997; Zhang and Levitus, 1997), few studies have focused on long-term trends.

The most convincing evidence of ocean warming is for the North Atlantic. An almost constant rate of interior warming, with implied steric sea level rise, is found over 73 years at Ocean Station S (south-east of Bermuda). Comparisons of trans-ocean sections show that these changes are widespread (Table 11.1). On decadal time-scales, variations in surface steric height from station S compare well with sea level at Bermuda (Roemmich, 1990) and appear to be driven by changes in the wind stress curl (Sturges and Hong, 1995; Sturges et al., 1998). Variability in the western North Atlantic (Curry et al., 1998) is related to changes in convective activity in the Labrador Sea (Dickson et al., 1996). Over the 20 years up to the early 1990s there has been a cooling of the Labrador Sea Water (as in the Irminger Sea, Read and Gould, 1992), and more recently in the western North Atlantic (Koltermann et al., 1999). For the South Atlantic, changes are more uncertain, particularly those early in the 20th century.

A warming of the Atlantic layer in the Arctic Ocean is deduced by comparison of modern oceanographic sections collected on board ice-breakers (e.g., Quadfasel et al., 1991; Carmack et al., 1997; Swift et al., 1997) and submarines (e.g. Morison et al., 1998; Steele and Boyd, 1998) with Russian Arctic Ocean atlases compiled from decades of earlier data (Treshnikov, 1977; Gorshkov, 1983). It is not yet clear whether these changes result from a climate trend or, as argued by Grotefendt et al. (1998), from decadal variability. The published studies do not report estimates of steric sea level changes; we note that a warming of 1°C over the central 200 m of the Atlantic layer would result in a local rise of steric sea level of 10 to 20 mm.

Observations from the Pacific and Indian Oceans cover a relatively short period, so any changes seen may be a result of decadal variability. Wong (1999), Wong et al. (1999), Bindoff and McDougall (1994) and Johnson and Orsi (1997) studied changes in the South Pacific. Bindoff and McDougall (2000) studied changes in the southern Indian Ocean. These authors

Table 11.1: *Summary of observations of interior ocean temperature changes and steric sea level rise during the 20th century.*

Reference	Dates of data	Location, section or region	Depth range (m)	Temperature change (°C/century)	Steric rise (mm/yr) (and heat uptake)	
North Atlantic Ocean						
Read and Gould (1992)	1962–1991	55°N, 40°–10°W	50–3000	–0.3		
Joyce and Robbins (1996)	1922–1995	Ocean Station S 32.17°N, 64.50°W	1500–2500	0.5	0.9 (0.7 W/m^{-2})	
Joyce *et al*. (1999)	1958, 1985, 1997	20°N–35°N 52°W and 66°W		0.57	1.0	
Parrilla *et al*. (1994), Bryden *et al*. (1996)	1957, 1981, 1992	24°N	800–2500	Peak of 1 at 1100 m	0.9 (1 W/m^{-2})	
Roemmich and Wunsch (1984)	1959, 1981	36°N	700–3000	Peak of 0.8 at 1500 m	0.9	
Arhan *et al*. (1998)	1957, 1993	8°N	1000–2500	Peak of 0.45 at 1700 m	0.6	
Antonov (1993)	1957–1981	45°N–70°N	0–500	Cooling		
			800–2500	0.4		
South Atlantic Ocean						
Dickson *et al*. (2001), Arbic and Owens (2001)	1926, 1957	8°S, 33.5°W–12.5°W	1000–2000 (Steric expansion for 100 m to bottom is shown in the right-hand half of the last column)	0.30	–0.1	0.0
	1926, 1957	8°S, 12°W–10.5°E		0.23	0.2	0.2
	1983, 1994	11°S, 34°W–13°W		0.30	1.1	4.4
	1983, 1994	11°S, 12.5°W–12°E		0.08	0.3	2.2
	1926, 1957	16°S, 37°W–14°W		0.10	–0.8	–2.5
	1926, 1957	16°S, 13.5°W–10.5°E		0.05	–0.2	–0.7
	1958, 1983	24°S, 40.5°W–14°W		0.41	0.1	1.0
	1958, 1983	24°S, 13.5°W–12.5°E		0.46	0.6	1.0
	1925, 1959	32°S, 48.5°W–14°W		0.13	–0.4	–0.2
Arctic Ocean						
See text			200–1500	Peak of >1 at 300 m		
North Pacific Ocean						
Thomson and Tabata (1989)	1956-1986	Ocean station Papa 50°N, 145°W			1.1	
Roemmich (1992)	1950-1991	32°N (off the coast of California)	0–300		0.9 ± 0.2	
Wong (1999), Wong *et al*. (1999, 2001)	1970s, 1990s	3.5°S–60°N			1.4	
		31.5°S–60°N			0.85	
Antonov (1993)	1957-1981	North of 30°N	0–500	Cooling		
South Pacific Ocean						
Holbrook and Bindoff (1997)	1955-1988	S. Tasman Sea	0–100		0.3	
Ridgway and Godfrey (1996), Holbrook and Bindoff (1997)	1955, mid-1970s	Coral and Tasman Seas	0–100	Warming		
	Since mid-1970s		0–450	Cooling		
Bindoff and Church (1992)	1967, 1989-1990	Australia-170°E 43°S			0.9	
		28°S			1.4	
Shaffer *et al*. (2000)	1967-1995	Eastern S Pacific 43°S			0.5	
		28°S			1.1	
Indian Ocean						
Bindoff and McDougall (1999)	1959-1966, 1987	30°S –35°S	0–900		1.6	
Atlantic, Pacific and Indian Oceans						
Levitus *et al*. (2000), Antonov *et al*. (2000)	1955-1995	Global average	0–300	0.7	(0.3 Wm^{-2})	
			0–3000		0.55 mm/yr (0.5 Wm^{-2})	

found changes in temperature and salinity in the upper hundreds of metres of the ocean which are consistent with a model of surface warming and freshening in the formation regions of the water masses and their subsequent subduction into the upper ocean. Such basin-scale changes are not merely a result of vertical thermocline heave, as might result from variability in surface winds.

In the only global analysis to date, Levitus *et al.* (2000) finds the ocean has stored 20×10^{22} J of heat between 1955 and 1995 (an average of 0.5 Wm^{-2}), with over half of this occurring in the upper 300 m for a rate of warming of 0.7°C/century. The steric sea level rise equivalent is 0.55 mm/yr, with maxima in the sub-tropical gyre of the North Atlantic and the tropical eastern Pacific.

In summary, while the evidence is still incomplete, there are widespread indications of thermal expansion, particularly in the sub-tropical gyres, of the order 1 mm/yr (Table 11.1). The evidence is most convincing for the North Atlantic but it also extends into the Pacific and Indian Oceans. The only area where cooling has been observed is in the sub-polar gyre of the North Atlantic and perhaps the North Pacific sub-polar gyre.

11.2.1.2 Models of thermal expansion
A variety of ocean models have been employed for estimates of ocean thermal expansion. The simplest and most frequently quoted is the one-dimensional (depth) upwelling-diffusion (UD) model (Hoffert *et al.*, 1980; Wigley and Raper, 1987, 1992, 1993; Schlesinger and Jiang, 1990; Raper *et al.*, 1996), which represents the variation of temperature with depth. Kattenberg *et al.* (1996) demonstrated that results from the GFDL AOGCM could be reproduced by the UD model of Raper *et al.* (1996). Using this model, the best estimate of thermal expansion from 1880 to 1990 was 43 mm (with a range of 31 to 57 mm) (Warrick *et al.*, 1996). Raper and Cubasch (1996) and Raper *et al.* (2001) discuss ways in which the UD model requires modification to reproduce the results of other AOGCMs. The latter work shows that a UD model of the type used in the SAR may be inadequate to represent heat uptake into the deep ocean on the time-scale of centuries. De Wolde *et al.* (1995, 1997) developed a two dimensional (latitude-depth, zonally averaged) ocean model, with similar physics to the UD model. Their best estimate of ocean thermal expansion in a model forced by observed sea surface temperatures over the last 100 years was 35 mm (with a range of 22 to 51 mm). Church *et al.* (1991) developed a subduction model in which heat is carried into the ocean interior through an advective process, which they argued better represented the oceans with movement of water along density surfaces and little vertical mixing. Jackett *et al.* (2000) developed this model further and tuned it
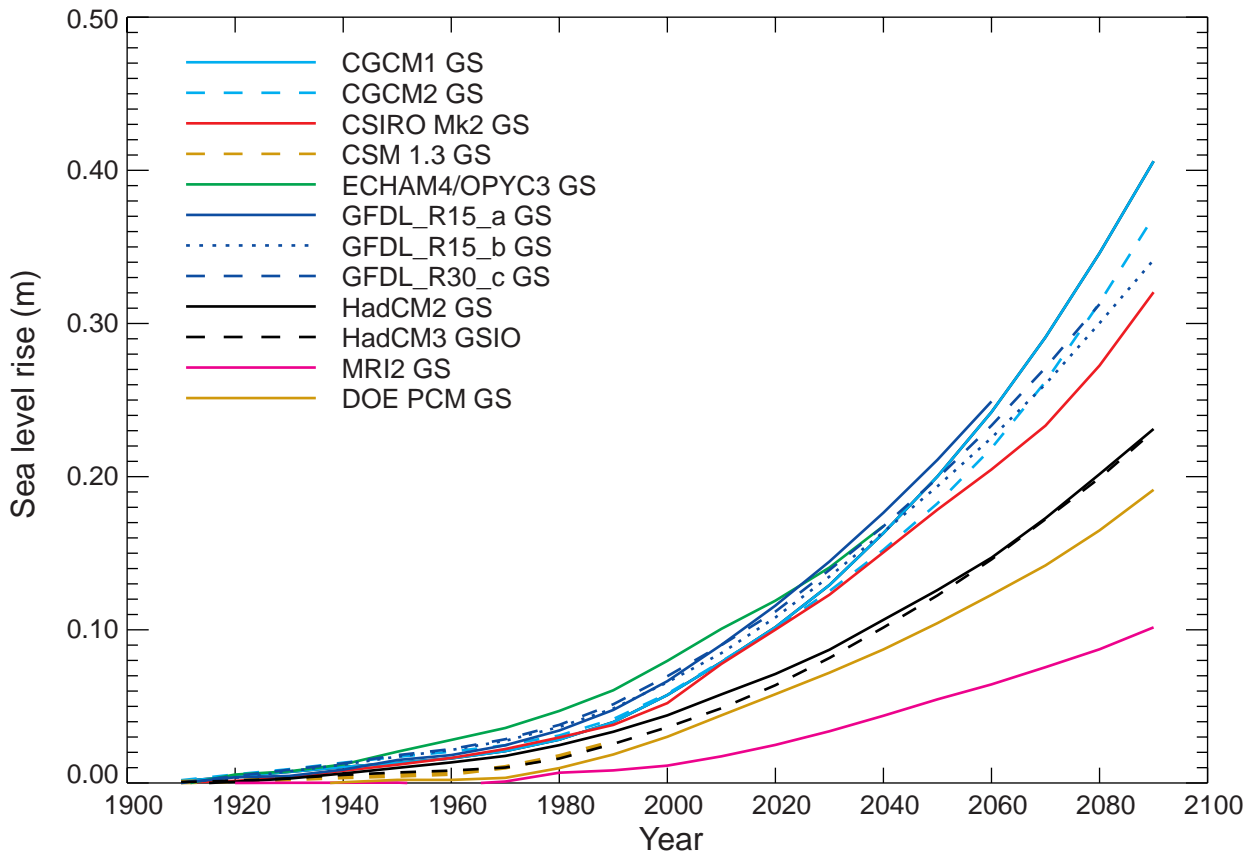

Figure 11.1: Global average sea level changes from thermal expansion simulated in AOGCM experiments with historical concentrations of greenhouse gases in the 20th century, then following the IS92a scenario for the 21st century, including the direct effect of sulphate aerosols. See Tables 8.1 and 9.1 for further details of models and experiments.

by comparison with an AOGCM, obtaining an estimate of 50 mm of thermal expansion over the last 100 years.

The advantage of these simple models is that they require less computing power than AOGCMs and so the sensitivity of results to a range of uncertainties can easily be examined. However, the simplifications imply that important processes controlling the penetration of heat from the surface into the ocean interior are not reproduced and they cannot provide information on the regional distribution of sea level rise. The most satisfactory way of estimating ocean thermal expansion is through the use of AOGCMs (Chapter 8, Section 8.3) (Gregory, 1993; Cubasch *et al.*, 1994; Bryan, 1996; Jackett *et al.*, 2000; Russell *et al.*, 2000; Gregory and Lowe, 2000). Improvements over the last decade relate particularly to the representation of the effect on mixing by processes which operate on scales too small to be resolved in global models, but which may have an important influence on heat uptake (see Section 8.5.2.2.4). The geographical distribution of sea level change due to density and circulation changes can be obtained from AOGCM results (various methods are used; see Gregory *et al.*, 2001). The ability of AOGCMs to simulate decadal variability in the ocean interior has not yet been demonstrated adequately, partly because of the scarcity of observations of decadal variability in the ocean for testing these models. This is not only an issue of evaluation of model performance; it is also relevant for deciding whether observed trends in

sea level and interior ocean temperatures represent a change which is significantly larger than the natural internal variability of the climate system.

A number of model simulations of the 20th century (Table 9.1) have recently been completed using realistic greenhouse gas and aerosol forcings. Results for global average thermal expansion over periods during the 20th century are given in Figure 11.1 and Table 11.2. They suggest that over the last hundred years the average rate of sea level rise due to thermal expansion was of the order of 0.3 to 0.7 mm/yr, a range which encompasses the simple model estimates, rising to 0.6 to 1.1 mm/yr in recent decades, similar to the observational estimates (Section 11.2.1.1).

11.2.2 Glaciers and Ice Caps

> **Box 11.2:** Mass balance terms for glaciers, ice caps and ice sheets
>
> A glacier, ice cap or ice sheet gains mass by accumulation of snow (snowfall and deposition by wind-drift), which is gradually transformed to ice, and loses mass (ablation) mainly by melting at the surface or base with subsequent runoff or evaporation of the melt water. Some melt water may refreeze within the snow instead of being lost, and some snow may sublimate or be blown off the surface. Ice may also be removed by discharge into a floating ice shelf or glacier tongue, from which it is lost by basal melting and calving of icebergs. Net accumulation occurs at higher altitude, net ablation at lower altitude; to compensate for net accumulation and ablation, ice flows downhill by internal deformation of the ice and sliding and bed deformation at the base. The rate at which this occurs is mainly controlled by the surface slope, the ice thickness, the effective ice viscosity, and basal thermal and physical conditions. The mass balance for an individual body of ice is usually expressed as the rate of change of the equivalent volume of liquid water, in m^3/yr; the mass balance is zero for a steady state. Mass balances are computed for both the whole year and individual seasons; the winter mass balance mostly measures accumulation, the summer, surface melting. The specific mass balance is the mass balance averaged over the surface area, in m/yr. A mass balance sensitivity is the derivative of the specific mass balance with respect to a climate parameter which affects it. For instance, a mass balance sensitivity to temperature is in m/yr/°C.

Table 11.2: Rate and acceleration of global-average sea level rise due to thermal expansion during the 20th century from AOGCM experiments with historical concentrations of greenhouse gases, including the direct effect of sulphate aerosols. See Tables 8.1 and 9.1 for further details of models and experiments. The rates are means over the periods indicated, while a quadratic fit is used to obtain the acceleration, assumed constant. Under this assumption, the rates apply to the midpoints (1950 and 1975) of the periods. Since the midpoints are 25 years apart, the difference between the rates is 25 times the acceleration. This relation is not exact because of interannual variability and non-constant acceleration.

	Rate of sea level rise (mm/yr)		Acceleration (mm/yr/century)
	1910 [a] to 1990 [b]	1960 to 1990 [b]	1910 [a] to 1990 [b]
CGCM1 GS	0.48	0.79	0.7 ± 0.2
CGCM2 GS	0.50	0.71	0.5 ± 0.3
CSIRO Mk2 GS	0.47	0.72	1.1 ± 0.2
CSM 1.3 GS	0.34	0.70	1.2 ± 0.3
ECHAM4/OPYC3 GS	0.75	1.09	1.0 ± 0.5
GFDL_R15_a GS	0.59	0.97	1.4 ± 0.4
GFDL_R15_b GS	0.60	0.88	1.1 ± 0.3
GFDL_R30_c GS	0.64	0.97	1.2 ± 0.3
HadCM2 GS	0.42	0.60	0.8 ± 0.2
HadCM3 GSIO	0.32	0.64	1.3 ± 0.4
DOE PCM GS	0.25	0.63	0.8 ± 0.4

[a] The choice of 1910 (rather than 1900) is made to accommodate the start date of some of the model integrations.

[b] The choice of 1990 (rather than 2000) is made because observational estimates referred to here do not generally include much data from the 1990s.

11.2.2.1 Mass balance studies

The water contained in glaciers and ice caps (excluding the ice sheets of Antarctica and Greenland) is equivalent to about 0.5 m of global sea level (Table 11.3). Glaciers and ice caps are rather sensitive to climate change; rapid changes in their mass are possible, and are capable of producing an important contribution to the rate of sea level rise. To evaluate this contribution, we need to know the rate of change of total glacier mass. Unfortunately sufficient measurements exist to determine the mass balance (see Box 11.2 for definition) for only a small minority of the world's 10^5 glaciers.

Table 11.3: *Some physical characteristics of ice on Earth.*

	Glaciers	Ice caps	Glaciers and ice caps [a]	Greenland ice sheet [b]	Antarctic ice sheet [b]
Number	>160 000	70			
Area (10^6 km^2)	0.43	0.24	0.68	1.71	12.37
Volume (10^6 km^3)	0.08	0.10	0.18 ± 0.04	2.85	25.71
Sea-level rise equivalent [d]	0.24	0.27	0.50 ± 0.10	7.2 [c]	61.1 [c]
Accumulation (sea-level equivalent, mm/yr) [d]			1.9 ± 0.3	1.4 ± 0.1	5.1 ± 0.2

Data sources: Meier and Bahr (1996), Warrick *et al.* (1996), Reeh *et al.* (1999), Huybrechts *et al.* (2000), Tables 11.5 and 11.6.

[a] Including glaciers and ice caps on the margins of Greenland and the Antarctic Peninsula, which have a total area of 0.14×10^6 km^2 (Weideck and Morris, 1996). The total area of glaciers and ice-caps outside Greenland and Antarctica is 0.54×10^6 km^2 (Dyurgerov and Meier, 1997a). The glaciers and ice caps of Greenland and Antarctica are included again in the next two columns.

[b] Grounded ice only, including glaciers and small ice caps.

[c] For the ice sheets, sea level rise equivalent is calculated with allowance for isostatic rebound and sea water replacing grounded ice, and this therefore is less than the sea level equivalent of the ice volume.

[d] Assuming an oceanic area of 3.62×10^8 km^2.

A possible approximate approach to this problem is to group glaciers into climatic regions, assuming glaciers in the same region to have a similar specific mass balance. With this method, we need to know only the specific mass balance for a typical glacier in each region (Kuhn *et al.*, 1999) and the total glacier area of the region. Multiplying these together gives the rate of change of glacier mass in the region. We then sum over all regions.

In the past decade, estimates of the regional totals of area and volume have been improved by the application of high resolution remote sensing and, to a lesser extent, by radio-echo-sounding. New glacier inventories have been published for central Asia and the former Soviet Union (Dolgushin and Osipova, 1989; Liu *et al.*, 1992; Kuzmichenok, 1993; Shi *et al.*, 1994; Liu and Xie, 2000; Qin *et al.*, 2000), New Zealand (Chinn, 1991), India (Kaul, 1999) South America (Casassa, 1995; Hastenrath and Ames, 1995; Skvarca *et al.*, Aniya *et al*, 1997; Kaser, 1999; 1995; Kaser *et al.*, 1996; Rott *et al.*, 1998), and new estimates made for glaciers in Antarctica and Greenland apart from the ice sheets (Weidick and Morris, 1996).

By contrast, specific mass balance is poorly known. Continuous mass balance records longer than 20 years exist for about forty glaciers worldwide, and about 100 have records of more than five years (Dyurgerov and Meier, 1997a). Very few have both winter and summer balances; these data are critical to relating glacier change to climatic elements (Dyurgerov and Meier, 1999). Although mass balance is being monitored on several dozen glaciers worldwide, these are mostly small (<20 km^2) and not representative of the size class that contains the majority of the mass (>100 km^2). The geographical coverage is also seriously deficient; in particular, we are lacking information on the most important maritime glacier areas. Specific mass balance exhibits wide variation geographically and over time (Figure 11.2). While glaciers in most parts of the world have had negative mass balance in the

past 20 years, glaciers in New Zealand (Chinn, 1999; Lamont *et al.*, 1999) and southern Scandinavia (Tvede and Laumann, 1997) have been advancing, presumably following changes in the regional climate.

Estimates of the historical global glacier contribution to sea level rise are shown in Table 11.4. Dyurgerov and Meier (1997a) obtained their estimate by dividing a large sample of measured glaciers into seven major regions and finding the mass balance for each region, including the glaciers around the ice sheets. Their area-weighted average for 1961 to 1990 was equivalent to 0.25 ± 0.10 mm/yr of sea level rise. Cogley and Adams (1998) estimated a lower rate for 1961 to 1990. However, their results may be not be representative of the global average because they do not make a correction for the regional biases in the sample of well investigated glaciers (Oerlemans, 1999). When evaluating data based on observed mass balance, one should note a worldwide glacier retreat
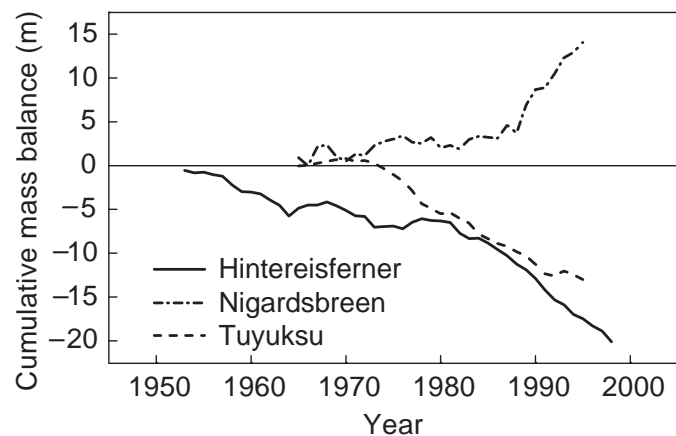

Figure 11.2: Cumulative mass balance for 1952-1998 for three glaciers in different climatic regimes: Hintereisferner (Austrian Alps), Nigardsbreen (Norway), Tuyuksu (Tien Shan, Kazakhstan).

Table 11.4: *Estimates of historical contribution of glaciers to global average sea level rise.*

Reference	Period	Rate of sea-level rise (mm/yr)	Remarks
Meier (1984)	1900 to 1961	0.46 ± 0.26	
Trupin *et al.* (1992)	1965 to 1984	0.18	
Meier (1993)	1900 to 1961	0.40	
Zuo and Oerlemans (1997), Oerlemans (1999)	1865 to 1990	0.22 ± 0.07[a]	Observed temperature changes with mass balance sensitivities estimated from precipitation in 100 regions
	1961 to 1990	0.3[a]	
Dyurgerov and Meier (1997b)	1961 to 1990	0.25 ± 0.10	Area-weighted mean of observed mass balance for seven regions
Dowdeswell *et al.* (1997)	1945 to 1995 approx.	0.13	Observed mass balance, Arctic only
Gregory and Oerlemans (1998)	1860 to 1990	0.15[a]	General Circulation Model (GCM) temperature changes with mass balance sensitivities from Zuo and Oerlemans (1997)
	1960 to 1990	0.26[a]	

[a] These papers give the change in sea level over the period indicated, from which we have calculated the rate of sea level rise.

following the high stand of the middle 19th century and subsequent small regional readvances around 1920 and 1980.

11.2.2.2 Sensitivity to temperature change

A method of dealing with the lack of mass balance measurements is to estimate the changes in mass balance as a function of climate, using mass balance sensitivities (see Box 11.2 for definition) and observed or modelled climate change for glacier covered regions. Mass-balance modelling of all glaciers individually is not practical because no detailed description exists for the great majority of them, and because local climate data are not available; even regional climate models do not have sufficient resolution, while downscaling methods cannot generally be used because local climate measurements have not been made (see Section 10.7). A number of authors have estimated past glacier net mass loss using past temperature change with present day glacier covered areas and mass balance sensitivities (Table 11.4). In this report, we project future mass balance changes using regional mass balance sensitivities which take account of regional and seasonal climatic information, instead of using the heuristic model of Wigley and Raper (1995) employed by Warrick *et al.* (1996).

Meier (1984) intuitively scaled specific mass balance according to mass balance amplitude (half the difference between winter and summer specific mass balance). Braithwaite and Zhang (1999) demonstrated a dependence of mass balance sensitivity on mass balance amplitude. Oerlemans and Fortuin (1992) derived an empirical relationship between the mass balance sensitivity of a glacier to temperature change and the local average precipitation, which is the principal factor determining its mass turnover rate. Zuo and Oerlemans (1997) extended this idea by distinguishing the effects of temperature changes in summer and outside summer; the former have a stronger influence on mass loss, in general. They made a calculation of glacier net mass loss since 1865. For 1961 to 1990, they obtained a rate of 0.3 mm/yr of sea level rise (i.e., a total of 8 mm, Oerlemans, 1999), very similar to the result of Dyurgerov and

Meier (1997b). Gregory and Oerlemans (1998) applied local seasonal temperature changes over 1860 to 1990 calculated by the HadCM2 AOGCM forced by changing greenhouse gases and aerosols (HadCM2 GS in Table 9.1) to the glacier model of Zuo and Oerlemans.

Zuo and Oerlemans (1997), Gregory and Oerlemans (1998) and Van de Wal and Wild (2001) all stress that the global average glacier mass balance depends markedly on the regional and seasonal distribution of temperature change. For instance, Gregory and Oerlemans (1998) find that projected future glacier net mass loss is 20% greater if local seasonal variation is neglected, and 20% less if regional variation is not included. The first difference arises because annual average temperature change is greater than summer temperature change at high latitudes, but the mass balance sensitivity is greater to summer change. The second is because the global average temperature change is less than the change at high latitudes, where most glaciers are found (Section 9.3.2).

Both the observations of mass balance and the estimates based on temperature changes (Table 11.4) indicate a reduction of mass of glaciers and ice caps in the recent past, giving a contribution to global-average sea level of 0.2 to 0.4 mm/yr over the last hundred years.

11.2.2.3 Sensitivity to precipitation change

Precipitation and accumulation changes also influence glacier mass balance, and may sometimes be dominant (e.g. Raper *et al.*, 1996). Generally, glaciers in maritime climates are more sensitive to winter accumulation than to summer conditions (Kuhn, 1984). AOGCM experiments suggest that global-average annual mean precipitation will increase on average by 1 to 3%/°C under the enhanced greenhouse effect (Figure 9.18). Glacier mass balance modelling indicates that to compensate for the increased ablation from a temperature rise of 1°C a precipitation increase of 20% (Oerlemans, 1981) or 35% (Raper *et al.*, 2000) would be required. Van de Wal and Wild (2001) find that the effect of

precipitation changes on calculated global-average glacier mass changes in the 21st century is only 5% of the temperature effect. Such results suggest that the evolution of the global glacier mass is controlled principally by temperature changes rather than precipitation changes. Precipitation changes might be significant in particular localities, especially where precipitation is affected by atmospheric circulation changes, as seems recently to have been the case with southern Scandinavian glaciers (Oerlemans, 1999).

11.2.2.4 Evolution of area

The above calculations all neglect the change of area that will accompany loss of volume. Hence they are inaccurate because reduction of area will restrict the rate of melting. A detailed computation of transient response with dynamic adjustment to decreasing glacier sizes is not feasible at present, since the required information is not available for most glaciers. Oerlemans *et al.* (1998) undertook such detailed modelling of twelve individual glaciers and ice caps with an assumed rate of temperature change for the next hundred years. They found that neglecting the contraction of glacier area could lead to an overestimate of net mass loss of about 25% by 2100.

Dynamic adjustment of glaciers to a new climate occurs over tens to hundreds of years (Jóhannesson *et al.*, 1989), the time-scale being proportional to the mean glacier thickness divided by the specific mass balance at the terminus. Since both quantities are related to the size of the glacier, the time-scale is not necessarily longer for larger glaciers (Raper *et al.*, 1996; Bahr *et al.*, 1998), but it tends to be longer for glaciers in continental climates with low mass turnover (Jóhannesson *et al.*, 1989; Raper *et al.*, 2000).

Meier and Bahr (1996) and Bahr *et al.* (1997), following previous workers, proposed that for a glacier or an ice sheet in a steady state there may exist scaling relationships of the form $V \propto A^c$ between the volume V and area A, where c is a constant. Such relationships seem well supported by the increasing sample of glacier volumes measured by radio-echo-sounding and other techniques, despite the fact that climate change may be occurring on time-scales similar to those of dynamic adjustment. If one assumes that the volume-area relationship always holds, one can use it to deduce the area as the volume decreases. This idea can be extended to a glacier covered region if one knows the distribution of total glacier area among individual glaciers, which can be estimated using empirical functions (Meier and Bahr, 1996; Bahr, 1997). Using these methods, Van de Wal and Wild (2001) found that contraction of area reduces the estimated glacier net mass loss over the next 70 years by 15 to 20% (see also Section 11.5.1.1).

11.2.3 Greenland and Antarctic Ice Sheets

Together, the present Greenland and Antarctic ice sheets contain enough water to raise sea level by almost 70 m (Table 11.3), so that only a small fractional change in their volume would have a significant effect. The average annual solid precipitation falling onto the ice sheets is equivalent to 6.5 mm of sea level, this input being approximately balanced by loss from melting and iceberg calving. The balance of these processes is not the same for the two ice sheets, on account of their different climatic regimes. Antarctic temperatures are so low that there is virtually no surface runoff; the ice sheet mainly loses mass by ice discharge into floating ice shelves, which experience melting and freezing at their underside and eventually break up to form icebergs. On the other hand, summer temperatures on the Greenland ice sheet are high enough to cause widespread melting, which accounts for about half of the ice loss, the remainder being discharged as icebergs or into small ice-shelves.

Changes in ice discharge generally involve response times of the order of 10^2 to 10^4 years. The time-scales are determined by isostasy, the ratio of ice thickness to yearly mass turnover, processes affecting ice viscosity, and physical and thermal processes at the bed. Hence it is likely that the ice sheets are still adjusting to their past history, in particular the transition to interglacial conditions. Their future contribution to sea level change therefore has a component resulting from past climate changes as well as one relating to present and future climate changes.

For the 21st century, we expect that surface mass balance changes will dominate the volume response of both ice sheets. A key question is whether ice-dynamical mechanisms could operate which would enhance ice discharge sufficiently to have an appreciable additional effect on sea level rise.

11.2.3.1 Mass balance studies

Traditionally, the state of balance of the polar ice sheets has been assessed by estimating the individual mass balance terms, and making the budget. Only the mass balance of the ice sheet resting on bedrock (the grounded ice sheet) needs to be considered, because changes in the ice shelves do not affect sea level as they are already afloat. Recent mass balance estimates for Greenland and Antarctica are shown in Tables 11.5 and 11.6. Most progress since the SAR has been made in the assessment of accumulation, where the major obstacle is poor coverage by *in situ* measurements. New methods have made use of atmospheric moisture convergence analysis based on meteorological data, remotely sensed brightness temperatures of dry snow, and GCMs (see references in the tables). Recent accumulation estimates display a tendency for convergence towards a common value, suggesting a remaining error of less than 10% for both ice sheets.

For Greenland (Table 11.5), runoff is an important term but net ablation has only been measured directly at a few locations and therefore has to be calculated from models, which have considerable sensitivity to the surface elevation data set and the parameters of the melt and refreezing methods used (Reeh and Starzer, 1996; Van de Wal, 1996; Van de Wal and Ekholm, 1996; Janssens and Huybrechts, 2000). Summing best estimates of the various mass balance components for Greenland gives a balance of $-8.5 \pm 10.2\%$ of the input, or $+0.12 \pm 0.15$ mm/yr of global sea level change, not significantly different from zero.

During the last five years, some mass balance estimates have been made for individual Greenland sectors. A detailed comparison of the ice flux across the 2,000 m contour with total accumulation revealed most of the accumulation zone to be near to equilibrium, albeit with somewhat larger positive and negative local imbalances (Thomas *et al.*, 1998, 2000). These results are

Table 11.5: *Current state of balance of the Greenland ice sheet (10^{12} kg/yr).*

Source and remarks	A Accumulation	B Runoff	C Net accumulation	D Iceberg production	E Bottom melting	F Balance
Benson (1962)	500	272	228	215		+13
Bauer (1968)	500	330	170	280		−110
Weidick (1984)	500	295	205	205		± 0
Ohmura and Reeh (1991): New accumulation map	535					
Huybrechts *et al*. (1991): Degree-day model on 20 km grid	539	256	283			
Robasky and Bromwich (1994): Atmospheric moisture budget analysis from radiosonde data, 1963-1989	545					
Giovinetto and Zwally (1995a): Passive microwave data of dry snow	461[a]					
Van de Wal (1996): Energy-Balance model on 20 km grid	539	316	223			
Jung-Rothenhäusler (1998): Updated accumulation map	510					
Reeh *et al*. (1999)	547	276	271	239	32	± 0
Ohmura *et al*. (1999): Updated accumulation map with GCM data; runoff from ablation-summer temperature parametrization	516	347	169			
Janssens and Huybrechts (2000): recalibrated degree-day model on 5 km grid; updated precipitation and surface elevation maps	542	281	261			
Zwally and Giovinetto (2000): Updated calculation on 50 km grid			216[b]			
Mean and standard devation	520 ± 26	297 ± 32	225 ± 41	235 ± 33	32 ± 3[c]	−44 ± 53[d]

[a] Normalised to ice sheet area of $1.676{\times}10^{6}$ km^2 (Ohmura and Reeh, 1991).

[b] Difference between net accumulation above the equilibrium line and net ablation below the equilibrium line.

[c] Melting below the fringing ice shelves in north and northeast Greenland (Rignot *et al.*, 1997).

[d] Including the ice shelves, but nearly identical to the grounded ice sheet balance because the absolute magnitudes of the other ice-shelf balance terms (accumulation, runoff, ice-dynamic imbalance) are very small compared to those of the ice sheet *(F=A–B–D–E).*

Table 11.6: *Current state of balance of the Antarctic ice sheet (10^{12} kg/yr).*

Source and remarks	A Accumulation over grounded ice	B Accumulation over all ice sheet	C Ice shelf melting	D Runoff	E Iceberg production	F Flux across grounding line
Kotlyakov *et al*. (1978)		2000	320	60	2400	
Budd and Smith (1985)	1800	2000			1800	1620
Jacobs *et al*. (1992). Ice shelf melting from observations of melt water outflow, glaciological field studies and ocean modelling.	1528	2144	544	53	2016	
Giovinetto and Zwally (1995a). Passive microwave data of dry snow.	1752[a]	2279[a]				
Budd *et al*. (1995). Atmospheric moisture budget analysis from GASP data, 1989 to 1992.		2190[b]				
Jacobs *et al*. (1996). Updated ice-shelf melting assessment.			756			
Bromwich *et al*. (1998). Atmospheric moisture budget analysis from ECMWF reanalysis and evaporation/ sublimation forecasts, 1985 to 1993.		2190[b]				
Turner *et al*. (1999). Atmospheric moisture budget analysis from ECMWF reanalysis, 1979 to 1993.		2106				
Vaughan *et al*. (1999). 1800 *in situ* measurements interpolated using passive microwave control field.	1811	2288				
Huybrechts *et al*. (2000). Updated accumulation map.	1924	2344				
Giovinetto and Zwally (2000). Updated map on 50 km grid.	1883[c]	2326[c]				
Mean and standard deviation.	1843 ± 76[d]	2246 ± 86[d]	540 ± 218	10 ± 10[e]	2072 ± 304	

[a] Normalised to include the Antarctic Peninsula.

[b] Specific net accumulation multiplied by total area of $13.95{\times}10^{6}$ km^2 (Fox and Cooper, 1994).

[c] Normalised to include the Antarctic Peninsula, and without applying a combined deflation and ablation adjustment.

[d] Mean and standard deviation based only on accumulation studies published since 1995.

[e] Estimate by the authors.

The mass balance of the ice sheet including ice shelves can be estimated as $B–C–D–E=−376 ± 384{\times}10^{12}$ kg/yr, which is $−16.7 ± 17.1\%$ of the total input *B*.

Assuming the ice shelves are in balance (and noting that the runoff derives from the grounded ice, not the ice shelves) would imply that $0=F+(B −A)−C−E$, in which case the flux across the grounding line would be $F=A−B+C+E =2209 ± 391{\times}10^{12}$ kg/yr.

likely to be only little influenced by short-term variations, because in the ice sheet interior, quantities that determine ice flow show little variation on a century time-scale. Recent studies have suggested a loss of mass in the ablation zone (Rignot *et al.*, 1997; Ohmura *et al.*, 1999), and have brought to light the important role played by bottom melting below floating glaciers (Reeh *et al.*, 1997, 1999; Rignot *et al.*, 1997); neglect of this term led to erroneous results in earlier analyses.

For Antarctica (Table 11.6), the ice discharge dominates the uncertainty in the mass balance of the grounded ice sheet, because of the difficulty of determining the position and thickness of ice at the grounding line and the need for assumptions about the vertical distribution of velocity. The figure of Budd and Smith (1985) of $1,620 \times 10^{12}$ kg/yr is the only available estimate. Comparing this with an average value of recent accumulation estimates for the grounded ice sheet would suggest a positive mass balance of around +10% of the total input, equivalent to −0.5 mm/yr of sea level. Alternatively, the flux across the grounding line can be obtained by assuming the ice shelves to be in balance and using estimates of the calving rate (production of icebergs), the rate of melting on the (submerged) underside of the ice shelves, and accumulation on the ice shelves. This results in a flux of $2,209 \pm 391 \times 10^{12}$ kg/yr across the grounding line and a mass balance for the grounded ice equivalent to $+1.04 \pm 1.06$ mm/yr of sea level (Table 11.6). However, the ice shelves may not be in balance, so that the error estimate probably understates the true uncertainty.

11.2.3.2 Direct monitoring of surface elevation changes
Provided that changes in ice and snow density and bedrock elevation are small or can be determined, elevation changes can be used to estimate changes of mass of the ice sheets. Using satellite altimetry, Davis *et al.* (1998) reported a small average thickening between 1978 and 1988 of 15 ± 20 mm/yr of the Greenland ice sheet above 2,000 m at latitudes up to 72°N. Krabill *et al.* (1999) observed a similar pattern above 2,000 m from 1993 to 1998 using satellite referenced, repeat aircraft laser altimetry. Together, these results indicate that this area of the Greenland ice sheet has been nearly in balance for two decades, in agreement with the mass budget studies mentioned above (Thomas *et al.*, 2000). Krabill *et al.* (1999) observed markedly different behaviour at lower altitudes, with thinning rates in excess of 2 m/yr in the south and east, which they attributed in part to excess flow, although a series of warmer-than-average summers may also have had an influence. In a recent update, Krabill *et al.* (2000) find the total ice sheet balance to be $−46 \times 10^{12}$ kg/yr or 0.13 mm/yr of sea level rise between 1993 and 1999, but could not provide an error bar. Incidentally, this value is very close to the century time-scale imbalance derived from the mass budget studies (Table 11.5), although the time periods are different and the laser altimetry results do not allow us to distinguish between accumulation, ablation, and discharge.

Small changes of ± 11 mm/yr were reported by Lingle and Covey (1998) in a region of East Antarctica between 20° and 160°E for the period 1978 to 1988. Wingham *et al.* (1998) examined the Antarctic ice sheet north of 82°S from 1992 to 1996, excluding the marginal zone. They observed no change in

East Antarctica to within ± 5 mm/yr, but reported a negative trend in West Antarctica of $−53 \pm 9$ mm/yr, largely located in the Pine Island and Thwaites Glacier basins. They estimated a century-scale mass imbalance of $−6\% \pm 8\%$ of accumulation for 63% of the Antarctic ice sheet, concluding that the thinning in West Antarctica is likely to result from a recent accumulation deficit. However, the measurements of Rignot (1998a), showing a 1.2 ± 0.3 km/yr retreat of the grounding line of Pine Island Glacier between 1992 and 1996, suggest an ice-dynamic explanation for the observed thinning. Altimetry records are at present too short to confidently distinguish between a short-term surface mass-balance variation and the longer-term ice-sheet dynamic imbalance. Van der Veen and Bolzan (1999) suggest that at least five years of data are needed on the central Greenland ice sheet.

11.2.3.3 Numerical modelling
Modelling of the past history of the ice sheets and their underlying beds over a glacial cycle is a way to obtain an estimate of the present ice-dynamic evolution unaffected by short-term (annual to decadal) mass-balance effects. The simulation requires time-dependent boundary conditions (surface mass balance, surface temperature, and sea level, the latter being needed to model grounding-line changes). Current glaciological models employ grids of 20 to 40 km horizontal spacing with 10 to 30 vertical layers and include ice shelves, basal sliding and bedrock adjustment.

Huybrechts and De Wolde (1999) and Huybrechts and Le Meur (1999) carried out long integrations over two glacial cycles using 3-D models of Greenland and Antarctica, with forcing derived from the Vostok (Antarctica) and Greenland Ice Core Project (GRIP) ice cores. The retreat history of the ice sheet along a transect in central west Greenland in particular was found to be in good agreement with a succession of dated moraines (Van Tatenhove *et al.*, 1995), but similar validation elsewhere is limited by the availability of well-dated material. Similar experiments were conducted as part of the European Ice Sheet Modelling Initiative (EISMINT) intercomparison exercise (Huybrechts *et al.*, 1998). These model simulations suggest that the average Greenland contribution to global sea level rise has been between −0.1 and 0.0 mm/yr in the last 500 years, while the Antarctic contribution has been positive. Four different Antarctic models yield a sea level contribution of between +0.1 and +0.5 mm/yr averaged over the last 500 years, mainly due to incomplete grounding-line retreat of the West Antarctic ice sheet (WAIS) since the Last Glacial Maximum (LGM) (Figure 11.3). However, substantial uncertainties remain, especially for the WAIS, where small phase shifts in the input sea level time-series and inadequate representation of ice-stream dynamics may have a significant effect on the model outcome. Glacio-isostatic modelling of the solid earth beneath the Antarctic ice sheet with prescribed ice sheet evolution (James and Ivins, 1998) gave similar uplift rates as those presented in Huybrechts and Le Meur (1999), indicating that the underlying ice sheet scenarios and bedrock models were similar, but observations are lacking to validate the generated uplift rates. By contrast, Budd *et al.* (1998) find that Antarctic ice volume is currently increasing at a rate of about 0.08 mm/yr of sea level lowering because in their modelling the Antarctic ice sheet was actually smaller during the LGM than today (for which there
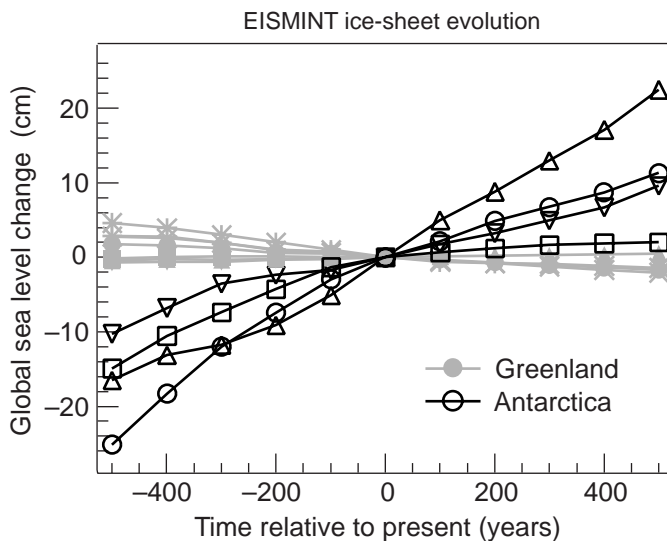
EISMINT ice-sheet evolution

Figure 11.3: Modelled evolution of ice sheet volume (represented as sea level equivalent) centred at the present time resulting from ongoing adjustment to climate changes over the last glacial cycle. Data are from all Antarctic and Greenland models that participated in the EISMINT intercomparison exercise (From Huybrechts *et al.*, 1998).

is, however, little independent evidence) and the effect of the higher accumulation rates during the Holocene dominates over the effects of grounding line changes.

Model simulations of this kind have not included the possible effects of changes in climate during the 20th century. The simulations described later (Section 11.5.1.1), in which an ice sheet model is integrated using changes in temperature and precipitation derived from AOGCM simulations, suggest that anthropogenic climate change could have produced an additional contribution of between –0.2 to 0.0 mm/yr of sea level from increased accumulation in Antarctica over the last 100 years, and between 0.0 and 0.1 mm/yr from Greenland, from both increased accumulation and ablation. The model results for Greenland exhibit substantial interannual variability. Furthermore, because of rising temperatures during the 20th century, the contribution for recent decades is larger than the average for the century. These points must be borne in mind when comparing with results of the direct observation methods for short periods in recent decades (Sections 11.2.3.1 and 11.2.3.2). Note also that the observational results include the ongoing response to past climate change as well as the effect of 20th century climate change.

11.2.3.4 Sensitivity to climatic change

The sensitivity of the ice sheet's surface mass balance has been studied with multiple regression analyses, simple meteorological models, and GCMs (Table 11.7). Most progress since the SAR has been made with several coupled AOGCMs, especially in the "time-slice" mode in which a high-resolution AGCM (Atmospheric General Circulation Model) is driven by output from a low-resolution transient AOGCM experiment for a limited duration of time. Model resolution of typically 100 km allows for a more realistic topography crucial to better

resolving temperature gradients and orographic forcing of precipitation along the steep margins of the polar ice sheets. Even then, GCMs do not yet perform well in reproducing melting directly from the surface energy fluxes. The ablation zone around the Greenland ice sheet is mostly narrower than 100 km, and the important role played by topography therefore requires the use of downscaling techniques to transfer information to local and even finer grids (Glover, 1999). An additional complication is that not all melt water runs off to the ocean and can be partly retained on or in the ice sheet (Pfeffer *et al.*, 1991; Janssens and Huybrechts, 2000).

For Greenland, estimates of the sensitivity to a 1°C local warming over the ice sheet are close to 0.3 mm/yr (with a total range of +0.1 to +0.4 mm/yr) of global sea level equivalent. This range mainly reflects differences in the predicted precipitation changes and the yearly distribution of the temperature increase, which is predicted to be larger in winter than in summer in the GCMs, but is assumed uniform in the studies of Van de Wal (1996) and Janssens and Huybrechts (2000). Another difference amongst the GCM results concerns the time window over which the sensitivities are assessed. The CSIRO9/T63 sensivities are estimated from high-resolution runs forced with observed SSTs for the recent past (Smith *et al.*, 1998; Smith, 1999), whereas the ECHAM data are given as specific mass balance changes for doubled minus present atmospheric CO_2. Thompson and Pollard (1997) report similar results to the ECHAM studies but the corresponding sensitivity value could not be calculated because the associated temperature information is not provided. Some palaeoclimatic data from central Greenland ice cores indicate that variations in precipitation during the Holocene are related to changes in atmospheric circulation rather than directly to local temperature (Kapsner *et al.*, 1995; Cuffey and Clow, 1997), such that precipitation might not increase with temperature (in contrast with Clausen *et al.*, 1988). For glacial-interglacial transitions, the ice cores do exhibit a strong positive correlation between temperature and precipitation (Dansgaard *et al.*, 1993; Dahl-Jensen *et al.*, 1993; Kapsner *et al.*, 1995; Cuffey and Marshall, 2000), as simulated by AOGCMs for anthropogenic warming. Although other changes took place at the glacial-interglacial transition, this large climate shift could be argued to be a better analogue for anthropogenic climate change than the smaller fluctuations of the Holocene. To allow for changes in circulation patterns and associated temperature and precipitation patterns, we have used time-dependent AOGCM experiments to calculate the Greenland contribution (Section 11.5.1).

For Antarctica, mass-balance sensitivities for a 1°C local warming are close to –0.4 mm/yr (with one outlier of –0.8 mm/yr) of global sea level equivalent. A common feature of all methods is the insignificant role of melting, even for summer temperature increases of a few degrees, so that only accumulation changes need to be considered. The sensitivity for the case that the change in accumulation is set proportional to the relative change in saturation vapour pressure is at the lower end of the sensitivity range, suggesting that in a warmer climate changes in atmospheric circulation and increased moisture advection can become equally important, in particular close to the ice sheet margin (Bromwich, 1995; Steig, 1997). Both ECHAM3 and

Table 11.7: *Mass balance sensitivity of the Greenland and Antarctic ice sheets to a 1°C local climatic warming.*

Source	dB/dT (mm/yr/°C)	Method
Greenland ice sheet		
Van de Wal (1996)	+0.31[a]	Energy balance model calculation on 20 km grid
Ohmura *et al.* (1996)	+0.41[c] [+0.04][cd]	ECHAM3/T106 time slice [$2 \times CO_2 - 1 \times CO_2$]
Smith (1999)	[−0.306][d]	CSIRO9/T63 GCM forced with SSTs 1950-1999
Janssens and Huybrechts (2000)	+0.35[a] [+0.26][b]	Recalibrated degree-day model on 5 km grid with new precipitation and surface elevation maps
Wild and Ohmura (2000)	+0.09[c] [−0.13][cd]	ECHAM4-OPYC3/T106 GCM time slice [$2 \times CO_2 - 1 \times CO_2$]
Antarctic ice sheet		
Huybrechts and Oerlemans (1990)	−0.36	Change in accumulation proportional to saturation vapour pressure
Giovinetto and Zwally (1995b)	−0.80[e]	Multiple regression of accumulation to sea-ice extent and temperature
Ohmura *et al.* (1996)	−0.41[c]	ECHAM3/T106 time slice [$2 \times CO_2 - 1 \times CO_2$]
Smith *et al.* (1998)	−0.40	CSIRO9/T63 GCM forced with SSTs 1950-1999
Wild and Ohmura (2000)	−0.48[c]	ECHAM4-OPYC3/T106 time slice [$2 \times CO_2 - 1 \times CO_2$]

dB/dT Mass balance sensitivity to local temperature change expressed as sea level equivalent. Note that this is not a sensitivity to global average temperature change.

[a] Constant precipitation.

[b] Including 5% increase in precipitation.

[c] Estimated from published data and the original time slice results.

[d] Accumulation changes only.

[e] Assuming sea-ice edge retreat of 150 km per °C.

ECHAM4/OPYC3 give a similar specific balance change over the ice sheet for doubled versus present atmospheric CO_2 to that found by Thompson and Pollard (1997).

In summary, the static sensitivity values suggest a larger role for Antarctica than for Greenland for an identical local temperature increase, meaning that the polar ice sheets combined would produce a sea level lowering, but the spread of the individual estimates includes the possibility that both ice sheets may also balance one another for doubled atmospheric CO_2 conditions (Ohmura *et al.*, 1996; Thompson and Pollard, 1997). For CO_2 increasing according to the IS92a scenario (without aerosol), studies by Van de Wal and Oerlemans (1997) and Huybrechts and De Wolde (1999) calculated sea level contributions for 1990 to 2100 of +80 to +100 mm from the Greenland ice sheet and about −80 mm from the Antarctic ice sheet. On this hundred year time-scale, ice-dynamics on the Greenland ice sheet was found to counteract the mass-balance-only effect by between 10 and 20%. Changes in both the area-elevation distribution and iceberg discharge played a role, although the physics controlling the latter are poorly known and therefore not well represented in the models. Because of its longer response time-scales, the Antarctic ice sheet hardly exhibits any dynamic response on a century time-scale, except when melting rates below the ice shelves were prescribed to rise by in excess of 1 m/yr (O'Farrell *et al.*, 1997; Warner and Budd, 1998; Huybrechts and De Wolde, 1999; see also Section 11.5.4.3).

11.2.4 Interaction of Ice Sheets, Sea Level and the Solid Earth

11.2.4.1 Eustasy, isostasy and glacial-interglacial cycles

On time-scales of 10^3 to 10^5 years, the most important processes affecting sea level are those associated with the growth and decay of the ice sheets through glacial-interglacial cycles. These contributions include the effect of changes in ocean volume and the response of the earth to these changes. The latter are the glacio-hydro-isostatic effects: the vertical land movements induced by varying surface loads of ice and water and by the concomitant redistribution of mass within the Earth and oceans. While major melting of the ice sheets ceased by about 6,000 years ago, the isostatic movements remain and will continue into the future because the Earth's viscous response to the load has a time-constant of thousands of years. Observational evidence indicates a complex spatial and temporal pattern of the resulting isostatic sea level change for the past 20,000 years. As the geological record is incomplete for most parts of the world, models (constrained by the reliable sea level observations) are required to describe and predict the vertical land movements and changes in ocean area and volume. Relative sea level changes caused by lithospheric processes, associated for example with tectonics and mantle convection, are discussed in Section 11.2.6.

Figure 11.4 illustrates global-average sea level change over the last 140,000 years. This is a composite record based on oxygen isotope data from Shackleton (1987) and Linsley
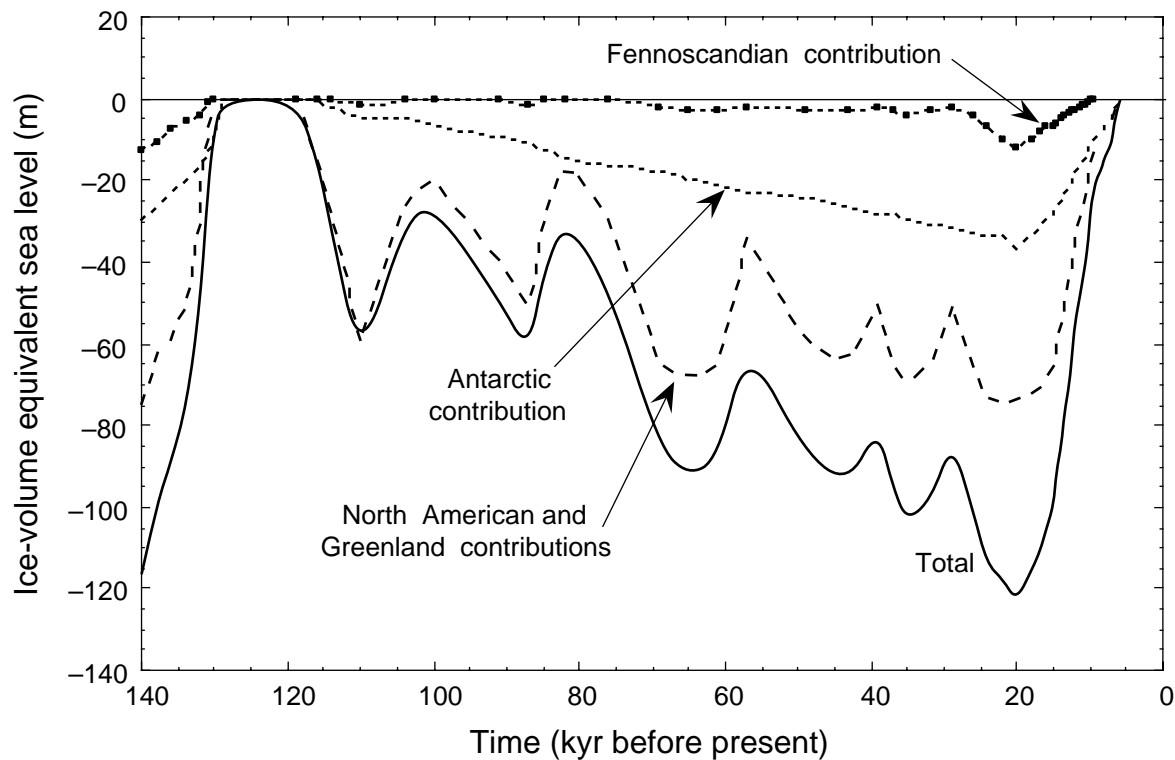
Figure 11.4: Estimates of global sea level change over the last 140,000 years (continuous line) and contributions to this change from the major ice sheets: (i) North America, including Laurentia, Cordilleran ice, and Greenland, (ii) Northern Europe (Fennoscandia), including the Barents region, (iii) Antarctica. (From Lambeck, 1999.)

(1996), constrained by the Huon terrace age-height relationships of Chappell *et al.* (1996a), the estimate of the LGM sea level by Yokoyama *et al.* (2000), the late-glacial eustatic sea level function of Fleming *et al.* (1998), and the timing of the Last Interglacial by Stirling *et al.* (1998). These fluctuations demonstrate the occurrence of sea level oscillations during a glacial-interglacial cycle that exceed 100 m in magnitude at average rates of up to 10 mm/year and more during periods of decay of the ice sheets and sometimes reaching rates as high as 40 mm/year (Bard *et al.*, 1996) for periods of very rapid ice sheet decay. Current best estimates indicate that the total LGM land-based ice volume exceeded present ice volume by 50 to 53×10^6 km^3 (Yokoyama *et al.*, 2000).

Local sea level changes can depart significantly from this average signal because of the isostatic effects. Figure 11.5 illustrates typical observational results for sea level change since the LGM in regions with no significant land movements other than of a glacio-hydro-isostatic nature. Also shown are model predictions for these localities, illustrating the importance of the isostatic effects. Geophysical models of these isostatic effects are well developed (see reviews by Lambeck and Johnston, 1998; Peltier, 1998). Recent modelling advances have been the development of high-resolution models of the spatial variability of the change including the detailed description of ice loads and of the melt-water load distribution (Mitrovica and Peltier, 1991; Johnston, 1993) and the examination of different assumptions about the physics of the earth (Peltier and Jiang, 1996; Johnston *et al.*, 1997; Kaufmann and Wolf, 1999; Tromp and Mitrovica, 1999).

Information about the changes in ice sheets come from field observations, glaciological modelling, and from the sea level observations themselves. Much of the emphasis of recent work on glacial rebound has focused on improved calculations of ice sheet parameters from sea level data (Peltier, 1998; Johnston and Lambeck, 2000; see also Section 11.3.1) but discrepancies between glaciologically-based ice sheet models and models inferred from rebound data remain, particularly for the time of, and before, the LGM. The majority of ice at this time was contained in the ice sheets of Laurentia and Fennoscandia but their combined estimated volume inferred from the rebound data for these regions (e.g., Nakada and Lambeck, 1988; Tushingham and Peltier, 1991, 1992; Lambeck *et al.*, 1998) is less than the total volume required to explain the sea level change of about 120 to 125 m recorded at low latitude sites (Fairbanks, 1989; Yokoyama *et al.*, 2000). It is currently uncertain how the remainder of the ice was distributed. For instance, estimates of the contribution of Antarctic ice to sea level rise since the time of the LGM range from as much as 37 m (Nakada and Lambeck, 1988) to 6 to 13 m (Bentley, 1999; Huybrechts and Le Meur, 1999). Rebound evidence from the coast of Antarctica indicates that ice volumes have changed substantially since the LGM (Zwartz *et al.*, 1997; Bentley, 1999) but these observations, mostly extending back only to 8,000 years ago, do not provide good constraints on the LGM volumes. New evidence from exposure age dating of moraines and rock surfaces is beginning to provide new constraints on ice thickness in Antarctica (e.g., Stone *et al.*, 1998) but the evidence is not yet sufficient to constrain past volumes of the entire ice sheet.

Figure 11.5: Examples of observed relative sea level change (with error bars, right-hand side) and model predictions for four different locations. The model predictions (left-hand side) are for the glacio-hydro-eustatic contributions to the total change (solid line, right hand side). (a) Angermann River, Sweden, near the centre of the former ice sheet over Scandinavia. The principal contribution to the sea level change is the crustal rebound from the ice unloading (curve marked ice, left-hand side) and from the change in ocean volume due to the melting of all Late Pleistocene ice sheets (curve marked esl). The combined predicted effect, including a small water loading term (not shown), is shown by the solid line (right-hand side), together with the observed values. (b) A location near Stirling, Scotland. Here the ice and esl contributions are of comparable magnitude but opposite sign (left-hand side) such that the rate of change of the total contribution changes sign (right-hand side). This result is typical for locations near former ice margins or from near the centres of small ice sheets. (c) The south of England where the isostatic contributions from the water (curve marked water) and ice loads are of similar amplitude but opposite sign. The dominant contribution to sea level change is now the eustatic contribution. This behaviour is characteristic of localities that lie well beyond the ice margins where a peripheral bulge created by the ice load is subsiding as mantle material flows towards the region formerly beneath the ice. (d) A location in Australia where the melt-water load is the dominant cause of isostatic adjustment. Here sea level has been falling for the past 6,000 years. This result is characteristic of continental margin sites far from the former areas of glaciation. (From Lambeck, 1996.)
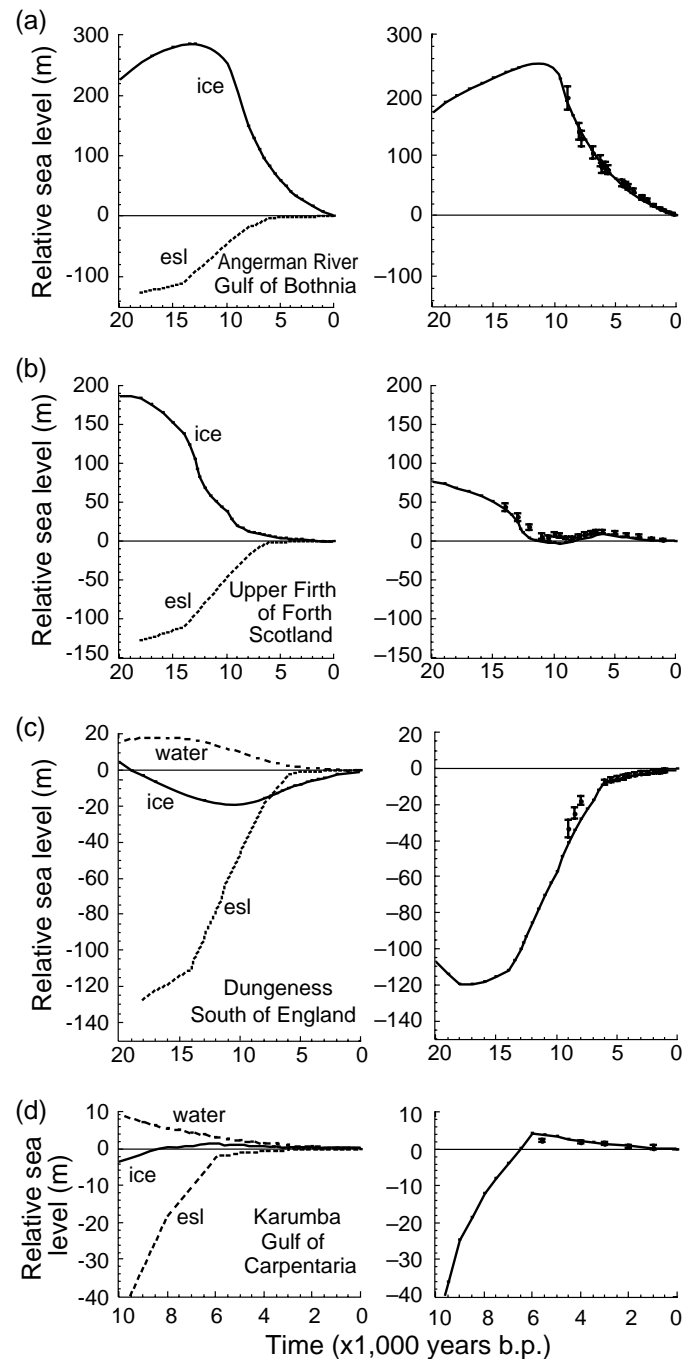
11.2.4.2 Earth rotation constraints on recent sea level rise

Changes in the Earth's ice sheets introduce a time-dependency in the Earth's inertia tensor whose consequences are observed both in the planet's rotation (as an acceleration in its rotation rate and as a shift in the position of the rotation axis) and in the acceleration of the rotation of satellite orbits about the Earth (Wu and Peltier, 1984; Lambeck, 1988). Model estimates of these changes are functions of mass shifts within and on the Earth and are dependent, therefore, on the past ice sheet geometries, on the Earth's rheology, and on the recent past and present rates of melting of the residual ice sheets. Other geophysical processes also contribute to the time-dependence

of the rotational and dynamical parameters (e.g. Steinberger and O'Connell, 1997). Hence, unique estimates of recent melting cannot be inferred from the observations.

Some constraints on the present rates of change of the ice sheets have, nevertheless, been obtained, in particular through a combination of the rotational observations with geological and tide gauge estimates of sea level change (Wahr *et al.*, 1993; Mitrovica and Milne, 1998; Peltier, 1998; Johnston and Lambeck, 1999). Results obtained so far are preliminary because observational records of the change in satellite orbits are relatively short (Nerem and Klosko, 1996; Cheng *et al.*, 1997) but they will become important as the length of the record increases. Peltier

(1998) has argued that if the polar ice sheets contributed, for example, 0.5 mm/yr to the global sea level rise, then the rotational constraints would require that most of this melting derived from Greenland. Johnston and Lambeck (1999) concluded that a solution consistent with geological evidence, including constraints on sea level for the past 6,000 years (Section 11.3.1), is for a non-steric sea level rise (i.e., not resulting from ocean density changes) of 1.0 ± 0.5 mm/yr for the past 100 years, with 5 to 30% originating from Greenland melting. However, all such estimates are based on a number of still uncertain assumptions such that the inferences are more indicative of the potential of the methodology than of actual quantitative conclusions.

11.2.5 Surface and Ground Water Storage and Permafrost

An important contribution to present day sea level rise could result from changes in the amount of water stored in the ground, on the surface in lakes and reservoirs, and by modifications to surface characteristics affecting runoff or evapotranspiration rates. Changing practices in the use of land and water could make these terms larger in future. However, very little quantitative information is available. For some of the components of the terrestrial water budget, Gornitz *et al.* (1997), updated by Gornitz (2000), give net results which differ substantially from those of Sahagian (2000) and Vörösmarty and Sahagian (2000), and also from those of Sahagian *et al.* (1994) used by Warrick *et al.* (1996). The largest positive contribution to sea level probably comes from ground water mining, which means the extraction of ground water from storage in aquifers in excess of the rate of natural recharge. Gornitz *et al.* (1997) estimate that ground water is mined at a rate that has been increasing in time, currently equivalent to 0.2 to 1.0 mm/yr of sea level, but they assume that much of this infiltrates back into aquifers so the contribution to sea level rise is only 0.1 to 0.4 mm/yr. Sahagian (2000) considers fewer aquifers; consequently he obtains a smaller total of 0.17 mm/yr from mining, but assumes that all of this water eventually reaches the ocean through the atmosphere or runoff. If Sahagian's assumption were applied to the inventory of Gornitz *et al.* it would imply a sea level contribution of 0.2 to 1.0 mm/yr.

Volumes of many of the world's large lakes have been reduced in recent decades through increased irrigation and other water use. Sahagian *et al.* (1994) and Sahagian (2000) estimate that the reduced volumes of the Caspian and Aral Seas (and associated ground water) contribute 0.03 and 0.18 mm/year to sea level rise, on the assumption that the extracted water reaches the world ocean by evapotranspiration. Recent *in situ* records and satellite altimetry data indicate that substantial fluctuations in the level of the Caspian Sea can occur on decadal time-scales (Cazenave *et al.*, 1997) which suggests that short records may not give a good indication of the long-term average. The reduction of lake volumes in China may contribute a further 0.005 mm/yr (Shi and Zhou, 1992). Assuming there are no other large sources, we take 0.2 mm/yr as the upper limit of the present contribution to sea level from lakes. Gornitz *et al.* (1997) do not include a term from lake volume changes, because they assume the water extracted for irrigation largely enters the ground water rather than the world ocean, so we take zero as the lower limit.

Gornitz *et al.* (1997) estimate there is 13.6 mm of sea level equivalent impounded in reservoirs. Most of this capacity was created, at roughly a constant rate, from 1950 to 1990. This rate of storage represents a reduction in sea level of 0.34 mm/yr. They assume that annually 5 ± 0.5% of the water impounded seeps into deep aquifers, giving a 1990 rate of seepage of 0.61 to 0.75 mm/yr, and a total volume of 15 mm sea level equivalent. We consider that this represents an upper bound, because it is likely that the rate of seepage from any reservoir will decrease with time as the surrounding water table rises, as assumed by Sahagian (2000). On the basis of a typical porosity and area affected, he estimates that the volume trapped as ground water surrounding reservoirs is 1.2 times the volume impounded in reservoirs. His estimate of the storage in reservoirs is 14 to 28 mm sea level equivalent; hence the ground water storage is an additional 17 to 34 mm sea level equivalent. Lack of global inventories means that these estimates of storage may well be too small because of the many small reservoirs not taken into account (rice paddies, ponds, etc., provided they impound water above the water table) (Vörösmarty and Sahagian, 2000). The total stored could be up to 50% larger (Sahagian, 2000).

Gornitz *et al.* (1997) estimate that evapotranspiration of water from irrigated land leads to an increase in atmospheric water content and hence a fall in sea level of 0.14 to 0.15 mm/yr. We consider this to be an overestimate, because it implies a 20th century increase in global tropospheric water content which substantially exceeds observations (Section 2.5.3.2). They further suggest that irrigation water derived from surface sources may infiltrate into aquifers, removing 0.40 to 0.48 mm/yr of sea level equivalent, based on the same assumption as for seepage from reservoirs. Urbanisation leads to reduced infiltration and increased surface runoff, which Gornitz *et al.* (1997) estimate may contribute 0.35 to 0.41 mm/yr of sea level rise. We consider these two terms to be upper bounds because, as with infiltration from reservoirs, a new steady state will be achieved after a period of years, with no further change in storage.

Estimates of the water contributed by deforestation are 0.1 mm/yr (Gornitz *et al.*, 1997) and 0.14 mm/yr (Sahagian, 2000) of sea level rise. Water released by oxidation of fossil fuels, vegetation and wetlands loss is negligible (Gornitz *et al.*, 1997).

Gornitz *et al.* (1997) estimate the total contribution to the 1990 rate of sea level change as −1.2 to −0.5 mm/yr. Integrating their estimates over 1910 to 1990 gives between −32 and −11 mm of sea level rise. In contrast, the estimate of Vörösmarty and Sahagian (2000) for the rate of sea level change from terrestrial storage is 0.06 mm/yr, equivalent to 5.4 mm over 80 years. The estimate of Sahagian *et al.* (1994), quoted by Warrick *et al.* (1996), was 12 mm during the 20th century. These discrepancies emphasise again the unsatisfactory knowledge of these contributions to sea level change.

Table 11.8 shows the ranges we have adopted for the various terms, based on the foregoing discussion. We integrate these terms over 1910 to 1990. (We use the time profiles of Gornitz *et al.* (1997) except that the infiltration from reservoirs is based on the approach of Sahagian (2000), and the rate of

Table 11.8: *Estimates of terrestrial storage terms. The values given are those of Gornitz et al. (1997) and Sahagian (2000). The estimates used in this report are the maximum and minimum values from these two studies. The average rates over the period 1910 to 1990 are obtained by integrating the decadal averages using the time history of contributions estimated by Gornitz et al. (1997).*

| | Rate of sea level rise for 1990 (mm/yr) | | | | | | Average rate 1910 to 1990 (mm/yr) | |
| | Gornitz *et al.* (1997) | | Sahagian (2000) | | This assessment | | | |
	min	max	min	max	min	max	min	max
Groundwater mining	+0.1	+0.4	+0.17		+0.1	+1.0	0.0	0.5
Lakes	0.0		+0.2		0.0	+0.2	0.0	0.1
Impoundment in reservoirs	–0.38	–0.30	–0.70	–0.35	–0.7	–0.3	–0.4	–0.2
Infiltration from reservoirs	–0.75	–0.61	–0.84	–0.42	–0.8	–0.4	–0.5	–0.2
Evapotranspiration	–0.15	–0.14	0.0		–0.1	0.0	–0.1	0.0
Infiltration from irrigation	–0.48	–0.40	0.0		–0.5	0.0	–0.2	0.0
Runoff from urbanisation	+0.35	+0.41	0.0		0.0	+0.4	0.0	0.1
Deforestation	+0.1		+0.14		+0.1	0.14	0.1	0.1
Total					–1.9	+1.0	–1.1	0.4

withdrawal from lakes is assumed constant over the last five decades.) This gives a range of –83 to +30 mm of sea level equivalent, or –1.1 to +0.4 mm/yr averaged over the period. However note that the rate of each of the terms increases during the 20th century.

This discussion suggests three important conclusions: (i) the effect of changes in terrestrial water storage on sea level may be considerable; (ii) the net effect on sea level could be of either sign, and (iii) the rate has increased over the last few decades (in the assessment of Gornitz *et al.* (1997) from near zero at the start of the century to –0.8 mm/yr in 1990).

Estimates of ice volume in northern hemisphere permafrost range from 1.1 to 3.7×10^{13} m^3 (Zhang *et al.*, 1999), equivalent to 0.03 to 0.10 m of global-average sea level. It occupies 25% of land area in the northern hemisphere. The major effects of global warming in presently unglaciated cold regions will be changes in the area of permafrost and a thickening of the active layer (the layer of seasonally thawed ground above permafrost). Both of these factors result in conversion of ground ice to liquid water, and hence in principle could contribute to the sea level change. Anisimov and Nelson (1997) estimated that a 10 to 20% reduction of area could occur by 2050 under a moderate climate-change scenario. In the absence of information about the vertical distribution of the ice, we make the assumption that the volume change is proportional to the area change. By 2100, the upper limit for the conversion of permafrost to soil water is thus about 50% of the total, or 50 mm sea level equivalent.

A thickening active layer will result in additional water storage capacity in the soil and thawing of ground ice will not necessarily make water available for runoff. What water is released could be mainly captured in ponds, thermokarst lakes, and marshes, rather than running off. Since the soil moisture in permafrost regions in the warm period is already very high, evaporation would not necessarily increase. We know of no quantitative estimates for these storage terms. We assume that the fraction which runs off lies within 0 and 50% of the available water. Hence we estimate the contribution of

permafrost to sea level 1990 to 2100 as 0 to 25 mm (0 to 0.23 mm/yr). For the 20th century, during which the temperature change has been about five times less than assumed by Anisimov and Nelson for the next hundred years, our estimate is 0 to 5 mm (0 to 0.05 mm/yr).

11.2.6 Tectonic Land Movements

We define tectonic land movement as that part of the vertical displacement of the crust that is of non-glacio-hydro-isostatic origin. It includes rapid displacements associated with earthquake events and also slow movements within (e.g., mantle convection) and on (e.g., sediment transport) the Earth. Large parts of the earth are subject to active tectonics which continue to shape the planet's surface. Where the tectonics occur in coastal areas, one of its consequences is the changing relationship between the land and sea surfaces as shorelines retreat or advance in response to the vertical land movements. Examples include the Huon Peninsula of Papua New Guinea (Chappell *et al.*, 1996b), parts of the Mediterranean (e.g. Pirazzoli *et al.*, 1994; Antonioli and Oliverio, 1996), Japan (Ota *et al.*, 1992) and New Zealand (Ota *et al.*, 1995). The Huon Peninsula provides a particularly good example (Figure 11.6) with 125,000 year old coral terraces at up to 400 m above present sea level. The intermediate terraces illustrated in Figure 11.6 formed at times when the tectonic uplift rates and sea level rise were about equal. Detailed analyses of these reef sequences have indicated that long-term average uplift rates vary between about 2 and 4 mm/yr, but that large episodic (and unpredictable) displacements of 1 m or more occur at repeat times of about 1,000 years (Chappell *et al.*, 1996b). Comparable average rates and episodic displacements have been inferred from Greek shoreline evidence (Stiros *et al.*, 1994). With major tectonic activity occurring at the plate boundaries, which in many instances are also continental or island margins, many of the world's tide gauge records are likely to contain both tectonic and eustatic signals. One value of the geological data is that it permits evaluations to be made of tectonic stability of the tide gauge locality.

Figure 11.6: The raised 125,000 year old coral terraces of the Huon Peninsula of Papua New Guinea up to 400 m above present sea level (Chappell *et al.*, 1996b).

Over very long time-scales (greater than 10^6 years), mantle dynamic processes lead to changes in the shape and volume of the ocean basins, while deposition of sediment reduces basin volume. These affect sea level but at very low rates (less than 0.01 mm/yr and 0.05 mm/yr, respectively; e.g., Open University, 1989; Harrison, 1990).

Coastal subsidence in river delta regions can be an important contributing factor to sea level change, with a typical magnitude of 10 mm/yr, although the phenomenon will usually be of a local character. Regions of documented subsidence include the Mediterranean deltas (Stanley, 1997), the Mississippi delta (Day *et al.*, 1993) and the Asian deltas. In the South China Sea, for example, the LGM shoreline is reported to occur at a depth of about 165 m below present level (Wang *et al.*, 1990), suggesting that some 40 m of subsidence may have occurred in 20,000 years at an average rate of about 2 mm/yr. Changes in relative sea level also arise through accretion and erosion along the coast; again, such effects may be locally significant.

11.2.7 Atmospheric Pressure

Through the inverse barometer effect, a local increase in surface air pressure over the ocean produces a depression in the sea surface of 1 cm per hPa (1 hPa = 1 mbar). Since water is practically incompressible, this cannot lead to a global-average sea level rise, but a long-term trend in surface air pressure patterns could influence observed local sea level trends. This has been investigated using two data sets: (i) monthly mean values of surface air pressure on a 10°×5° grid for the period 1873 to 1995 for the Northern Hemisphere north of 15°N obtained from the University of East Anglia Climatic Research Unit, and (ii) monthly mean values on a global 5°×5° grid for the period 1871 to 1994 obtained from the UK Met Office (see Basnett and Parker, 1997, for a discussion of the various data sets). The two data sets present similar spatial pattens of trends for their geographical and temporal overlaps. Both yield small trends of the order 0.02 hPa/yr; values of –0.03 hPa/yr occur in limited

regions of the high Arctic and equatorial Pacific. As found by Woodworth (1987), trends are only of the order of 0.01 hPa/yr in northern Europe, where most of the longest historical tide gauges are located. We conclude that long-term sea level trends could have been modified to the extent of ± 0.2 mm/yr, considerably less than the average eustatic rate of rise. Over a shorter period larger trends can be found. For example, Schönwiese *et al.* (1994) and Schönwiese and Rapp (1997) report changes in surface pressure for the period 1960 to 1990 that could have modified sea level trends in the Mediterranean and around Scandinavia by –0.05 and +0.04 mm/yr respectively.

11.3 Past Sea Level Changes

11.3.1 Global Average Sea Level over the Last 6,000 Years

The geological evidence for the past 10,000 to 20,000 years indicates that major temporal and spatial variation occurs in relative sea level change (e.g., Pirazzoli, 1991) on time-scales of the order of a few thousand years (Figure 11.5). The change observed at locations near the former centres of glaciation is primarily the result of the glacio-isostatic effect, whereas the change observed at tectonically stable localities far from the former ice sheets approximate the global average sea level change (for geologically recent times this is primarily eustatic change relating to changes in land-based ice volume). Glacio-hydro-isostatic effects (the Earth's response to the past changes in ice and water loads) remain important and result in a spatial variability in sea level over the past 6,000 years for localities far from the former ice margins. Analysis of data from such sites indicate that the ocean volume may have increased to add 2.5 to 3.5 m to global average sea level over the past 6,000 years (e.g., Fleming *et al.*, 1998), with a decreasing contribution in the last few thousand years. If this occurred uniformly over the past 6,000 years it would raise sea level by 0.4 to 0.6 mm/year. However, a few high resolution sea level records from the French Mediterranean coast indicate that much of this increase occurred between about 6,000 and 3,000 years ago and that the rate over the past 3,000 years was only about 0.1 to 0.2 mm/yr (Lambeck and Bard, 2000). These inferences do not constrain the source of the added water but likely sources are the Antarctic and Greenland ice sheets with possible contributions from glaciers and thermal expansion.

In these analyses of Late Holocene observations, the relative sea level change is attributed to both a contribution from any change in ocean volume and a contribution from the glacio-hydro-isostatic effect, where the former is a function of time only and the latter is a function of both time and position. It is possible to use the record of sea level changes to estimate parameters for a model of isostatic rebound. In doing this, the spatial variability of sea level change determines the mantle rheology, whereas the time dependence determines any correction that may be required to the assumed history of volume change. Solutions from different geographic regions may lead to variations in the rheology due to lateral variations in mantle temperature, for example, but the eustatic term should be the same, within observational and model uncertainties, in each case (Nakada and Lambeck, 1988). If it is assumed that no eustatic change has occurred in the past 6,000

years or so, but in fact eustatic change actually has occurred, the solution for Earth-model parameters will require a somewhat stiffer mantle than a solution in which eustatic change is included. The two solutions may, however, be equally satisfactory for interpolating between observations. For example, both approaches lead to mid-Holocene highstands at island and continental margin sites far from the former ice sheets of amplitudes 1 to 3 m. The occurrence of such sea level maxima places a upper limit on the magnitude of glacial melt in recent millennia (e.g., Peltier, 2000), but it would be inconsistent to combine estimates of ongoing glacial melt with results of calculations of isostatic rebound in which the rheological parameters have been inferred assuming there is no ongoing melt.

The geological indicators of past sea level are usually not sufficiently precise to enable fluctuations of sub-metre amplitude to be observed. In some circumstances high quality records do exist. These are from tectonically stable areas where the tidal range is small and has remained little changed through time, where no barriers or other shoreline features formed to change the local conditions, and where there are biological indicators that bear a precise and consistent relationship to sea level. Such areas include the micro-atoll coral formations of Queensland, Australia (Chappell, 1982; Woodroffe and McLean, 1990); the coralline algae and gastropod vermetid data of the Mediterranean (Laborel *et al.*, 1994; Morhange *et*

al., 1996), and the fresh-to-marine transitions in the Baltic Sea (Eronen *et al.*, 1995; Hyvarinen, 1999). These results all indicate that for the past 3,000 to 5,000 years oscillations in global sea level on time-scales of 100 to 1,000 years are unlikely to have exceeded 0.3 to 0.5 m. Archaeological evidence for this interval places similar constraints on sea level oscillations (Flemming and Webb, 1986). Some detailed local studies have indicated that fluctuations of the order of 1 m can occur (e.g., Van de Plassche *et al.*, 1998) but no globally consistent pattern has yet emerged, suggesting that these may be local rather than global variations.

Estimates of current ice sheet mass balance (Section 11.2.3.1) have improved since the SAR. However, these results indicate only that the ice sheets are not far from balance. Earth rotational constraints (Section 11.2.4.2) and ice sheet altimetry (Section 11.2.3.2) offer the prospect of resolving the ice sheet mass balance in the future, but at present the most accurate estimates of the long-term imbalance (period of several hundred years) follows from the comparison of the geological sea level data with the ice sheet modelling results (Section 11.2.3.3). The above geological estimates of the recent sea level rates may include a component from thermal expansion and glacier mass changes which, from the long-term tempera- ture record in Chapter 2 (Section 2.3.2), could contribute to a sea level lowering by as much as 0.1 mm/yr. These results

Table 11.9: *Recent estimates of sea level rise from tide gauges. The standard error for these estimates is also given along with the method used to correct for vertical land movement (VLM).*

	Region	**VLM** [a]	**Rate ± s.e.** [b] **(mm/yr)**
Gornitz and Lebedeff (1987)	Global	Geological	1.2 ± 0.3
Peltier and Tushingham (1989, 1991)	Global	ICE-3G/M1	2.4 ± 0.9 [c]
Trupin and Wahr (1990)	Global	ICE-3G/M1	1.7 ± 0.13
Nakiboglu and Lambeck (1991)	Global	Spatial decomposition	1.2 ± 0.4
Douglas (1991)	Global	ICE-3G/M1	1.8 ± 0.1
Shennan and Woodworth (1992)	NW Europe	Geological	1.0 ± 0.15
Gornitz (1995) [d]	N America E Coast	Geological	1.5 ± 0.7 [c]
Mitrovica and Davis (1995), Davis and Mitrovica (1996)	Global far field (far from former ice sheets)	PGR Model	1.4 ± 0.4 [c]
Davis and Mitrovica (1996)	N America E Coast	PGR Model	1.5 ± 0.3 [c]
Peltier (1996)	N America E Coast	ICE-4G/M2	1.9 ± 0.6 [c]
Peltier and Jiang (1997)	N America E Coast	Geological	2.0 ± 0.6 [c]
Peltier and Jiang (1997)	Global	ICE-4G/M2	1.8 ± 0.6 [c]
Douglas (1997) [d]	Global	ICE-3G/M1	1.8 ± 0.1
Lambeck *et al.* (1998)	Fennoscandia	PGR Model	1.1 ± 0.2
Woodworth *et al.* (1999)	British Isles	Geological	1.0

[a] This column shows the method used to correct for vertical land motion. ICE-3G/M1 is the Post Glacial Rebound (PGR) model of Tushingham and Peltier (1991). ICE-4G/M2 is a more recent PGR model based on the deglaciation history of Peltier (1994) and the mantle viscosity model of Peltier and Jiang (1996). Nakiboglu and Lambeck (1991) performed a spherical harmonic decomposition of the tide-gauge trends and took the the zero-degree term as the global-average rate. They indicated that a PGR signal would make little contribution to this term. The use of geological data is discussed in the text.

[b] The uncertainty is the standard error of the estimate of the global average rate.

[c] This uncertainty is the standard deviation of the rates at individual sites.

[d] See references in these papers for estimates of sea level rise for various other regions.

suggest that the combined long-term ice sheet imbalance lies within the range 0.1 to 0.3 mm/yr. Results from ice sheet models for the last 500 years indicate an ongoing adjustment to the glacial-interglacial transition of Greenland and Antarctica together of 0.0 to 0.5 mm/yr. These ranges are consistent. We therefore take the ongoing contribution of the ice sheets to sea level rise in the 20th and 21st centuries in response to earlier climate change as 0.0 to 0.5 mm/yr. This is additional to the effect of 20th century and future climate change.

11.3.2 Mean Sea Level Changes over the Past 100 to 200 Years

11.3.2.1 Mean sea level trends

The primary source of information on secular trends in global sea level during the past century is the tide gauge data set of the Permanent Service for Mean Sea Level (PSMSL) (Spencer and Woodworth, 1993). The tide gauge measurement is of the level of the sea surface relative to that of the land upon which the gauge is located and contains information on both the displacement of the land and on changes in ocean volume (eustatic changes). The land displacement may be of two types: that caused by active tectonics and that caused by glacial rebound. Corrections for these effects are required if the change in ocean volume is to be extracted from the tide gauge record. Both corrections are imperfectly known and are based on sea level observations themselves, usually from long geological records. Different strategies have been developed for dealing with these corrections but differences remain that are not inconsequential (see Table 11.9).

The sea level records contain significant interannual and decadal variability and long records are required in order to estimate reliable secular rates that will be representative of the last century. In addition, sea level change is spatially variable because of land movements and of changes in the ocean circulation. Therefore, a good geographic distribution of observations is required. Neither requirement is satisfied with the current tide gauge network and different strategies have been developed to take these differences into consideration. Warrick *et al.* (1996), Douglas (1995) and Smith *et al.* (2000) give recent reviews of the subject, including discussions of the Northern Hemisphere geographical bias in the historical data set.

In the absence of independent measurements of vertical land movements by advanced geodetic techniques (Section 11.6.1), corrections for movements are based on either geological data or geophysical modelling. The former method uses geological evidence from locations adjacent to the gauges to estimate the long-term relative sea level change which is assumed to be caused primarily by land movements, from whatever cause. This is subtracted from the gauge records to estimate the eustatic change for the past century. However, this procedure may underestimate the real current eustatic change because the observed geological data may themselves contain a long-term component of eustatic sea level rise (Section 11.3.1). The latter method, glacial rebound modelling, is also constrained by geological observations to estimate earth response functions or ice load parameters, which may therefore themselves contain a component of long-term eustatic sea level

change unless this component is specifically solved for (Section 11.3.1).

A further underestimate of the rate of sea level rise from the geological approach, compared to that from glacial rebound models, will pertain in forebulge areas, and especially the North American east coast, where the linear extrapolation of geological data could result in an underestimate of the corrected rate of sea level change for the past century typically by 0.3 mm/yr because the glacial rebound signal is diminishing with time (Peltier, 2000). However, in areas remote from the former ice sheets this bias will be considerably smaller.

Also, in adding recent mass into the oceans, most studies have assumed that it is distributed uniformly and have neglected the Earth's elastic and gravitational response to the changed water loading (analogous to glacio-hydro-isostatic effect). This will have the effect of reducing the observed rise at continental margin sites from ongoing mass contributions by as much as 30% (cf. Nakiboglu and Lambeck, 1991).

Table 11.9 summarises estimates of the corrected sea level trends for the past century. Estimates cover a wide range as a result of different assumptions and methods for calculating the rate of vertical land movement, of different selections of gauge records for analysis, and of different requirements for minimum record length.

There have been several more studies since the SAR of trends observed in particular regions. Woodworth *et al.* (1999) provided a partial update to Shennan and Woodworth (1992), suggesting that sea level change in the North Sea region has been about 1 mm/yr during the past century. Lambeck *et al.* (1998) combined coastal tide gauge data from Fennoscandinavia together with lake level records and postglacial rebound models to estimate an average regional rise for the past century of 1.1 ± 0.2 mm/yr. Studies of the North American east coast have been particularly concerned with the spatial dependence of trends associated with the Laurentian forebulge. Peltier (1996) concluded a current rate of order 1.9 ± 0.6 mm/yr, larger than the 1.5 mm/yr obtained by Gornitz (1995), who used the geological data approach, and Mitrovica and Davis (1995), who employed Post Glacial Rebound (PGR) modelling. Note that the observations of thermal expansion (Section 11.2.1.1) indicate a higher rate of sea level rise over recent decades in the sub-tropical gyres of the North Atlantic (i.e., off the North American east coast) than the higher latitude sub-polar gyre. Thus the differences between three lower European values compared with the higher North American values may reflect a real regional difference (with spatial variations in regional sea level change being perhaps several tenths of a millimetre per year – see also Section 11.5.2). In China, relative sea level is rising at about 2 mm/yr in the south but less than 0.5 mm/yr in the north (National Bureau of Marine Management, 1992), with an estimated average of the whole coastline of 1.6 mm/yr (Zhen and Wu, 1993) and with attempts to remove the spatially dependent component of vertical land movement yielding an average of 2.0 mm/yr (Shi, 1996). The two longest records from Australia (both in excess of 80 years in length and not included in Douglas, 1997) are from Sydney and Fremantle, on opposite sides of the continent. They show observed rates of relative sea level rise of 0.86 ± 0.12 mm/yr
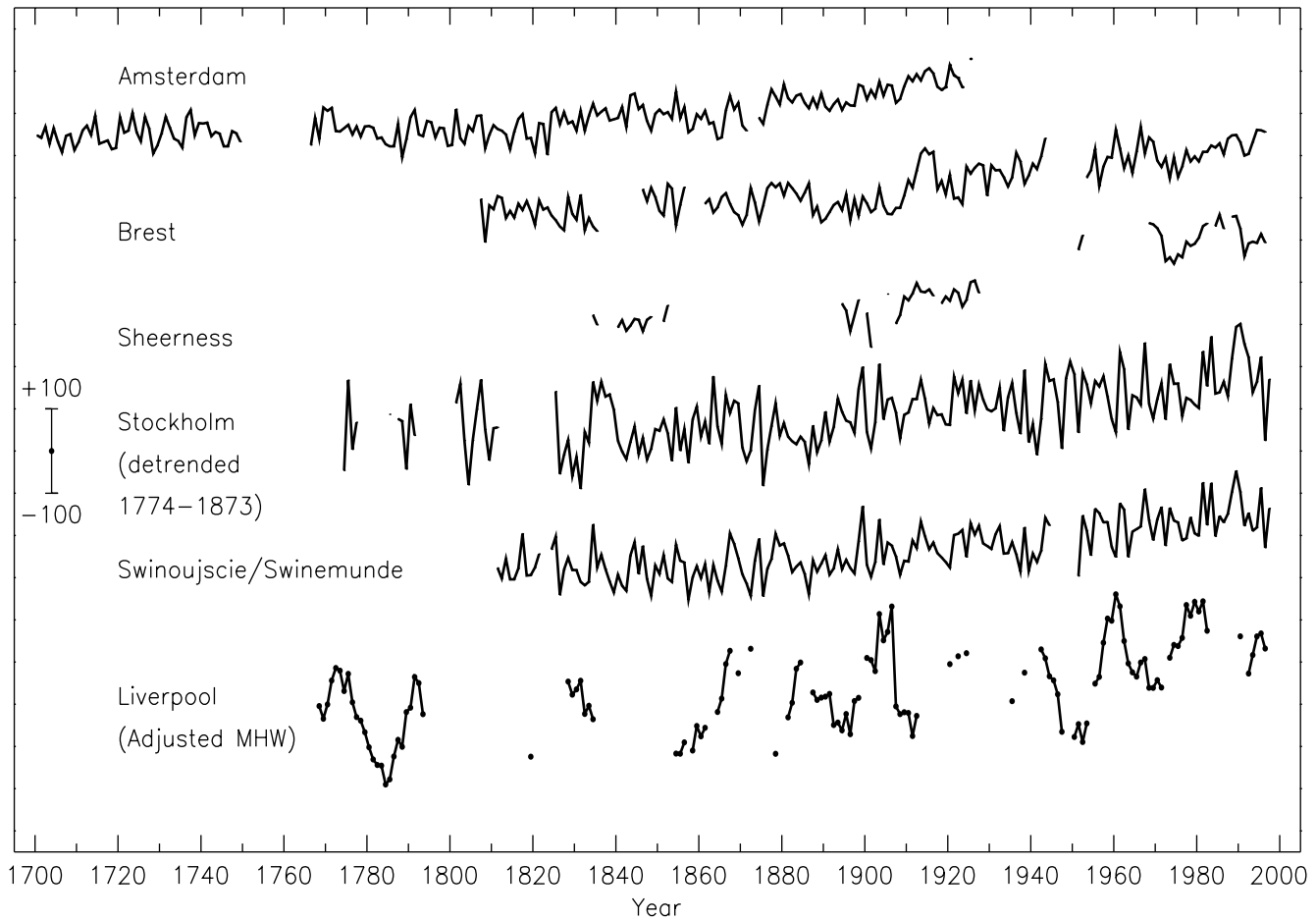
Figure 11.7: Time-series of relative sea level for the past 300 years from Northern Europe: Amsterdam, Netherlands; Brest, France; Sheerness, UK; Stockholm, Sweden (detrended over the period 1774 to 1873 to remove to first order the contribution of postglacial rebound); Swinoujscie, Poland (formerly Swinemunde, Germany); and Liverpool, UK. Data for the latter are of "Adjusted Mean High Water" rather than Mean Sea Level and include a nodal (18.6 year) term. The scale bar indicates ±100 mm. (Adapted from Woodworth, 1999a.)

and 1.38 ± 0.18 mm/yr over the periods 1915 to 1998 and 1897 to 1998 (Mitchell *et al.*, 2000), corresponding to approximately 1.26 mm/yr and 1.73 mm/yr after glacial rebound correction using the Peltier ICE-4G/M2 model, or 1.07 mm/yr and 1.55 mm/yr using the corrections of Lambeck and Nakada (1990).

There have been only two analyses of global sea level change based on the PSMSL data set published since the SAR. Douglas (1997) provided an update to Douglas (1991) and applied the PGR model of Tushingham and Peltier (1991) to a selected set of twenty-four long tide gauge records, grouped into nine geographical areas, with minimum record length 60 years and average length 83 years. However, the only Southern Hemisphere sites included in this solution were from Argentina and New Zealand. The overall global average of 1.8 ± 0.1 mm/yr agreed with the 1991 analysis, with considerable consistency between area-average trends. The standard error of the global rate was derived from the standard deviation of regional trends, assuming that temporal and spatial variability is uncorrelated between regions. Peltier and Jiang (1997) used essentially the same set of stations as Douglas and a new model for postglacial rebound.

From Table 11.9 one can see that there are six global estimates determined with the use of PGR corrections derived from global models of isostatic adjustment, spanning a range from 1.4 mm/yr (Mitrovica and Davis, 1995; Davis and Mitrovica, 1996) to 2.4 mm/yr (Peltier and Tushingham, 1989, 1991). We consider that these five are consistent within the systematic uncertainty of the PGR models, which may have a range of uncertainty of 0.5 mm/yr depending on earth structure parametrization employed (Mitrovica and Davis, 1995). The average rate of the five estimates is 1.8 mm/yr. There are two other global analyses, of Gornitz and Lebedeff (1987) and Nakiboglu and Lambeck (1991), which yield estimates of 1.2 mm/yr, lower than the first group. Because of the issues raised above with regard to the geological data method for land movement correction, the value of Gornitz and Lebedeff may be underestimated by up to a few tenths of a millimetre per year, although such considerations do not affect the method of Nakiboglu and Lambeck. The differences between the former five and latter two analyses reflect the analysis methods, in particular the differences in corrections for land movements and in selections of tide gauges used, including the effect of any spatial variation in thermal expansion. However, all the discrepancies which could

arise as a consequence of different analysis methods remain to be more thoroughly investigated. On the basis of the published literature, we therefore cannot rule out an average rate of sea level rise of as little as 1.0 mm/yr during the 20th century. For the upper bound, we adopt a limit of 2.0 mm/yr, which includes all recent global estimates with some allowance for systematic uncertainty. As with other ranges (see Box 11.1), we do not imply that the central value is the best estimate.

11.3.2.2 Long-term mean sea level accelerations

Comparison of the rate of sea level rise over the last 100 years (1.0 to 2.0 mm/yr) with the geological rate over the last two millennia (0.1 to 0.2 mm/yr; Section 11.3.1) implies a comparatively recent acceleration in the rate of sea level rise. The few very long tide gauge records are especially important in the search for "accelerations" in sea level rise. Using four of the longest (about two centuries) records from north-west Europe (Amsterdam, Brest, Sheerness, Stockholm), Woodworth (1990) found long-term accelerations of 0.4 to 0.9 mm/yr/century (Figure 11.7). Woodworth (1999a) found an acceleration of order 0.3 mm/yr/century in the very long quasi-mean sea level (or 'Adjusted Mean High Water') record from Liverpool. From these records, one can infer that the onset of the acceleration occurred during the 19th century, a suggestion consistent with separate analysis of the long Stockholm record (Ekman, 1988, 1999; see also Mörner, 1973). It is also consistent with some geological evidence from north-west Europe (e.g., Allen and Rae, 1988). In North America, the longest records are from Key West, Florida, which commenced in 1846 and which suggest an acceleration of order 0.4 mm/year/century (Maul and Martin, 1993), and from New York which commenced in 1856 and which has a similar acceleration. Coastal evolution evidence from parts of eastern North America suggest an increased rate of rise between one and two centuries before the 20th century (Kearney and Stevenson, 1991; Varekamp *et al.*, 1992; Kearney, 1996; Van de Plassche *et al.*, 1998; Varekamp and Thomas, 1998; Shaw and Ceman, 2000).

There is no evidence for any acceleration of sea level rise in data from the 20th century data alone (Woodworth, 1990; Gornitz and Solow, 1991; Douglas, 1992). Mediterranean records show decelerations, and even decreases in sea level in the latter part of the 20th century, which may be caused by increases in the density of Mediterranean Deep Water and air pressure changes connected to the North Atlantic Oscillation (NAO) (Tsimplis and Baker, 2000), suggesting the Mediterranean might not be the best area for monitoring secular trends. Models of ocean thermal expansion indicate an acceleration through the 20th century but when the model is subsampled at the locations of the tide gauges no significant acceleration can be detected because of the greater level of variability (Gregory *et al.*, 2001). Thus the absence of an acceleration in the observations is not necessarily inconsistent with the model results.

11.3.2.3 Mean sea level change from satellite altimeter observations

In contrast to the sparse network of coastal and mid-ocean island tide gauges, measurements of sea level from space by satellite
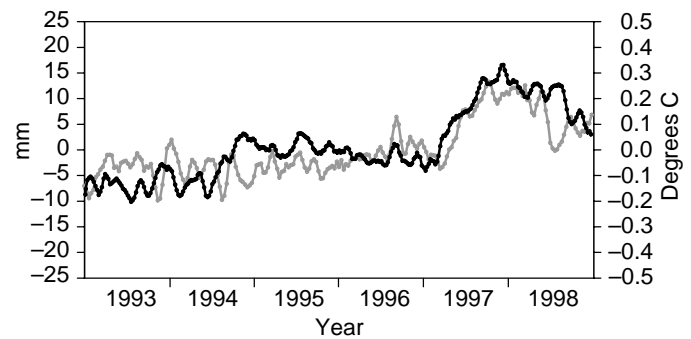

Figure 11.8: Global mean sea level variations (light line) computed from the TOPEX/POSEIDON satellite altimeter data compared with the global averaged sea surface temperature variations (dark line) for 1993 to 1998 (Cazenave *et al.*, 1998, updated). The seasonal components have been removed from both time-series.

radar altimetry provide near global and homogenous coverage of the world's oceans, thereby allowing the determination of regional sea level change. Satellite altimeters also measure sea level with respect to the centre of the earth. While the results must be corrected for isostatic adjustment (Peltier, 1998), satellite altimetry avoids other vertical land movements (tectonic motions, subsidence) that affect local determinations of sea level trends measured by tide gauges. However, achieving the required sub-millimetre accuracy is demanding and requires satellite orbit information, geophysical and environmental corrections and altimeter range measurements of the highest accuracy. It also requires continuous satellite operations over many years and careful control of biases.

To date, the TOPEX/POSEIDON satellite-altimeter mission, with its (near) global coverage from 66°N to 66°S (almost all of the ice-free oceans) from late 1992 to the present, has proved to be of most value to direct estimates of sea level change. The current accuracy of TOPEX/POSEIDON data allows global average sea level to be estimated to a precision of several millimetres every 10 days, with the absolute accuracy limited by systematic errors.

Careful comparison of TOPEX/POSEIDON data with tide gauge data reveals a difference in the rate of change of local sea level of -2.3 ± 1.2 mm/yr (Mitchum, 1998) or -2 ± 1.5 mm/yr (Cazenave *et al.*, 1999). This discrepancy is caused by a combination of instrumental drift, especially in the TOPEX Microwave Radiometer (TMR) (Haines and Bar-Sever, 1998), and vertical land motions which have not been allowed for in the tide gauge data. The most recent estimates of global average sea level rise from the six years of TOPEX/POSEIDON data (using corrections from tide gauge comparisons) are 2.1 ± 1.2 mm/yr (Nerem *et al.*, 1997), 1.4 ± 0.2 mm/yr (Cazenave *et al.*, 1998; Figure 11.8), 3.1 ± 1.3 mm/yr (Nerem, 1999) and 2.5 ± 1.3 mm/yr (Nerem, 1999), of which the last assumes that all instrumental drift can be attributed to the TMR. When Cazenave *et al.* allow for the TMR drift, they compute a sea level rise of 2.6 mm/yr. Their uncertainty of

± 0.2 mm/yr does not include allowance for uncertainty in instrumental drift, but only reflects the variations in measured global sea level. Such variations correlate with global average sea surface temperature, perhaps indicating the importance of steric effects through ocean heat storage. Cazenave *et al.* (1998) and Nerem *et al.* (1999) argue that ENSO events cause a rise and a subsequent fall in global averaged sea level of about 20 mm (Figure 11.8). These findings indicate that the major 1997/98 El Niño-Southern Oscillation (ENSO) event could bias the above estimates of sea level rise and also indicate the difficulty of separating long-term trends from climatic variability.

After upgrading many of the geophysical corrections on the original European Remote Sensing (ERS) data stream, Cazenave *et al.* (1998) find little evidence of sea level rise over the period April 1992 to May 1996. However, over the time span of overlap between the ERS-1 and TOPEX/POSEIDON data, similar rates of sea level change (about 0.5 mm/yr) are calculated. For the period April 1992 to April 1995, Anzenhofer and Gruber (1998) find a sea level rise of 2.2 ± 1.6 mm/yr.

In summary, analysis of TOPEX/POSEIDON data suggest a rate of sea level rise during the 1990s greater than the mean rate of rise for much of the 20th century. It is not yet clear whether this is the result of a recent acceleration, of systematic differences between the two measurement techniques, or of the shortness of the record (6 years).

11.3.3 Changes in Extreme Sea Levels: Storm Surges and Waves

Lack of adequate data sets means we can not ascertain whether there have been changes in the magnitude and/or frequency of storm surges (aperiodic changes associated with major meteorological disturbances resulting in sea level changes of up to several metres and lasting a few hours to days) in many regions of the world. Zhang *et al.* (1997, 2000) performed a comprehensive analysis of hourly tide gauge data from the east coast of North America, and concluded that there had been no discernible secular trend in storm activity or severity during the past century. European analyses include that of Woodworth (1999b), who found no significant increase in extreme high water level distributions from Liverpool from 1768 to 1993 to those from later epochs, other than what can be explained in terms of changes in local tidal amplitudes, mean sea level and vertical land movement. Vassie (reported in Pugh and Maul, 1999) and Bijl *et al.* (1999) concluded that there was no discernible trend over the last century in the statistics of non-tidal sea level variability around the UK and the eastern North Sea (Denmark, Germany and the Netherlands), above the considerable natural sea level variability on decadal time-scales. In South America, D'Onofrio *et al.* (1999) observed a trend of extreme levels at Buenos Aires of 2.8 mm/yr over 1905 to 1993. On the basis of available statistics, the South American result is consistent with the local mean sea level trend.

Variations in surge statistics can also be inferred from analysis of meteorological data. Kass *et al.* (1996) and the WASA Group (1998) showed that there are no significant overall trends in windiness and cyclonic activity over the North Atlantic

and north-west Europe during the past century, although major variations on decadal times-scales exist. An increase in storminess in the north-east Atlantic in the last few decades (Schmith *et al.*, 1998) and a recent trend towards higher storm surge levels on the German and Danish coasts (Langenberg *et al.*, 1999) is consistent with natural variability evident over the last 150 years. Pirazzoli (2000) detected evidence for a slight decrease in the main factors contributing to surge development on the French Atlantic coast in the last 50 years. Correlation between the frequency of Atlantic storms and ENSO was demonstrated by Van der Vink *et al.* (1998).

Increases in wave heights of approximately 2 to 3 m over the period 1962 to 1985 off Land's End, south-west England (Carter and Draper, 1988), increases in wave height over a neighbouring area at about 2%/yr since 1950 (Bacon and Carter, 1991, 1993) and wave height variations simulated by the Wave Action Model (WAM) (Günther *et al.*, 1998) are all consistent with decadal variations over most of the north-east Atlantic and North Sea. This variability could be related to the NAO (Chapter 2, Section 2.6.5).

11.4 Can 20th Century Sea Level Changes be Explained?

In order to have confidence in our ability to predict future changes in sea level, we need to confirm that the relevant processes (Section 11.2) have been correctly identified and evaluated. We attempt this by seeing how well we can account for the current rate of change (Section 11.3). We note that:

- some processes affecting sea level have long (centuries and longer) time-scales, so that current sea level change is also related to past climate change,

- some relevant processes are not determined solely by climate,

- fairly long records (at least 50 years according to Douglas, 1992) are needed to detect a significant trend in local sea level, because of the influence of natural variability in the climate system, and

- the network of tide gauges with records of this length gives only a limited coverage of the world's continental coastline and almost no coverage of the mid-ocean.

The estimated contributions from the various components of sea level rise during the 20th century (Table 11.10, Figure 11.9) were constructed using the results from Section 11.2. The sum of these contributions for the 20th century ranges from –0.8 mm/yr to 2.2 mm/yr, with a central value of 0.7 mm/yr. The upper bound is close to the observational upper bound (2.0 mm/yr), but the central value is less than the observational lower bound (1.0 mm/yr), and the lower bound is negative i.e. the sum of components is biased low compared to the observational estimates. Nonetheless, the range is narrower than the range given by Warrick *et al.* (1996), as a result of greater constraints on all the contributions, with the exception of the terrestrial storage terms. In particular, the long-term contribution from the

Table 11.10: *Estimated rates of sea level rise components from observations and models (mm/yr) averaged over the period 1910 to 1990. (Note that the model uncertainties may be underestimates because of possible systematic errors in the models.) The 20th century terms for Greenland and Antarctica are derived from ice sheet models because observations cannot distinguish between 20th century and long-term effects. See Section 11.2.3.3.*

	Minimum (mm/yr)	Central value (mm/yr)	Maximum (mm/yr)
Thermal expansion	0.3	0.5	0.7
Glaciers and ice caps	0.2	0.3	0.4
Greenland – 20th century effects	0.0	0.05	0.1
Antarctica – 20th century effects	− 0.2	− 0.1	0.0
Ice sheets – adjustment since LGM	0.0	0.25	0.5
Permafrost	0.00	0.025	0.05
Sediment deposition	0.00	0.025	0.05
Terrestrial storage (not directly from climate change)	− 1.1	− 0.35	0.4
Total	− 0.8	0.7	2.2
Estimated from observations	1.0	1.5	2.0

ice sheets has been narrowed substantially from those given in Warrick *et al.* (1996) by the use of additional constraints (geological data and models of the ice sheets) (Section 11.3.1).

The reason for the remaining discrepancy is not clear. However, the largest uncertainty (by a factor of more than two) is in the terrestrial storage terms. Several of the components of the terrestrial storage term are poorly determined and the quoted limits require several of the contributions simultaneously to lie at the extremes of their ranges. This coincidence is improbable unless the systematic errors affecting the estimates are correlated. Furthermore, while coupled models have improved considerably in recent years, and there is general agreement between the observed and modelled thermal expansion contribution, the models' ability to quantitatively simulate decadal changes in ocean temperatures and thus thermal expansion has not been adequately tested. Given the poor global coverage of high quality tide gauge records and the uncertainty in the corrections for land motions, the observationally based rate of sea level rise this century should also be questioned.

In the models, at least a third of 20th century anthropogenic eustatic sea level rise is caused by thermal expansion, which has a geographically non-uniform signal in sea level change. AOGCMs do not agree in detail about the patterns of geographical variation (see Section 11.5.2). They all give a geographical spread of 20th century trends at individual grid points which is characterised by a standard deviation of 0.2 to 0.5 mm/yr (Gregory *et al.*, 2001). This spread is a result of a combination of spatial non-uniformity of trends and the uncertainty in local trend estimates arising from temporal variability. As yet no published study has revealed a stable pattern of observed non-uniform sea level change. Such a pattern would provide a critical test of models. If there is significant non-uniformity, a trend from a single location would be an inaccurate estimate of the global average. For example, Douglas (1997) averaged nine regions and found a standard deviation of about 0.3 mm/yr (quoted by Douglas as a standard error), similar to the range expected from AOGCMs.

A common perception is that the rate of sea level rise should have accelerated during the latter half of the 20th century. The tide gauge data for the 20th century show no significant acceleration (e.g., Douglas, 1992). We have obtained estimates based on AOGCMs for the terms directly related to anthropogenic climate change in the 20th century, i.e., thermal expansion (Section 11.2.1.2), ice sheets (Section 11.2.3.3), glaciers and ice caps (Section 11.5.1.1) (Figure 11.10a). The estimated rate of sea level rise from anthropogenic climate change ranges from 0.3 to 0.8 mm/yr (Figure 11.10b). These terms do show an acceleration through the 20th century (Figure 11.10a,b). If the terrestrial storage terms have a negative sum (Section 11.2.5), they may offset some of the acceleration in recent decades. The total computed rise (Figure 11.10c) indicates an acceleration of only 0.2 mm/yr/century, with a range from −1.1 to +0.7 mm/yr/century,
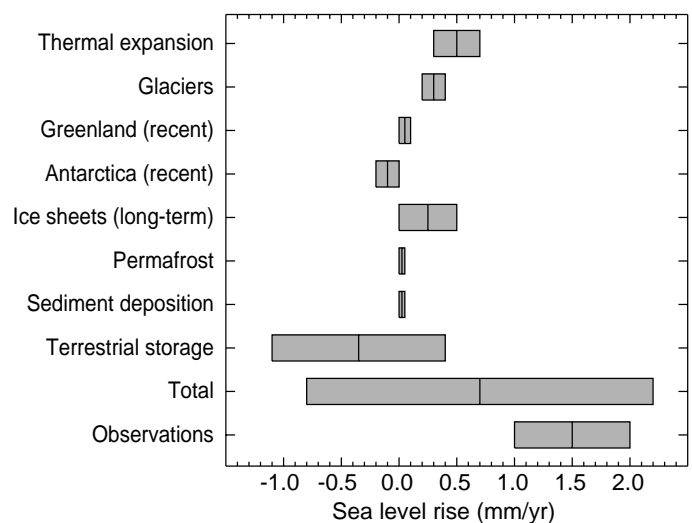

Figure 11.9: Ranges of uncertainty for the average rate of sea level rise from 1910 to 1990 and the estimated contributions from different processes.
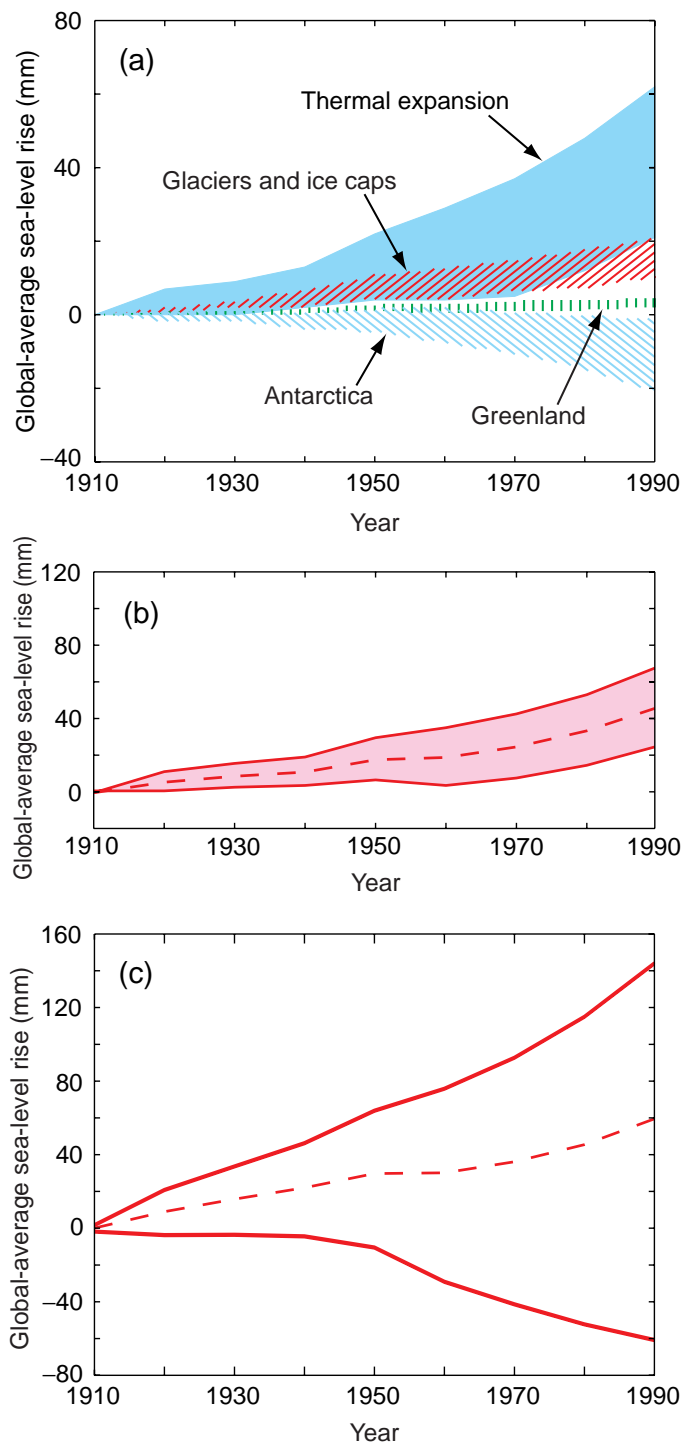
Figure 11.10: Estimated sea level rise from 1910 to 1990. (a) The thermal expansion, glacier and ice cap, Greenland and Antarctic contributions resulting from climate change in the 20th century calculated from a range of AOGCMs. Note that uncertainties in land ice calculations have not been included. (b) The mid-range and upper and lower bounds for the computed response of sea level to climate change (the sum of the terms in (a) plus the contribution from permafrost). These curves represent our estimate of the impact of anthropogenic climate change on sea level during the 20th century. (c) The mid-range and upper and lower bounds for the computed sea level change (the sum of all terms in (a) with the addition of changes in permafrost, the effect of sediment deposition, the long-term adjustment of the ice-sheets to past climate change and the terrestrial storage terms).

consistent with observational finding of no acceleration in sea level rise during the 20th century (Section 11.3.2.2). The sum of terms not related to recent climate change is −1.1 to +0.9 mm/yr (i.e., excluding thermal expansion, glaciers and ice caps, and changes in the ice sheets due to 20th century climate change). This range is less than the observational lower bound of sea level rise. Hence it is very likely that these terms alone are an insufficient explanation, implying that 20th century climate change has made a contribution to 20th century sea level rise.

Recent studies (see Sections 2.3.3, 2.3.4) suggest that the 19th century was unusually cold on the global average, and that an increase in solar output may have had a moderate influence on warming in the early 20th century (Section 12.4.3.3). This warming might have produced some thermal expansion and could have been responsible for the onset of glacier recession in the early 20th century (e.g., Dowdeswell *et al.*, 1997), thus providing a possible explanation of an acceleration in sea level rise commencing before major industrialisation.

11.5 Future Sea Level Changes

11.5.1 Global Average Sea Level Change 1990 to 2100

Warrick *et al.* (1996) made projections of thermal expansion and of loss of mass from glaciers and ice-sheets for the 21st century for the IS92 scenarios using two alternative simple climate models. Since the SAR, time-dependent experiments have been run with several AOGCMs (Chapter 9.1.2, Table 9.1) following the IS92a scenario (Leggett *et al.*, 1992) for future concentrations of greenhouse gases, including the direct effect of sulphate aerosols. In Section 11.5.1.1, we use the AOGCM IS92a results to derive estimates of thermal expansion and land ice melt, employing methods from the literature as described in Section 11.2, and we add contributions from thawing of permafrost, sediment deposition, and the continuing adjustment of the ice sheets to climate changes since the LGM. The choice of scenario is not the principal consideration; the main point is that the AOGCMs all follow the same scenario, so the range of results reflects the systematic uncertainty inherent in the modelling of sea level changes. The use of IS92a also facilitates comparison with the result of Warrick *et al.* (1996).

To quantify the uncertainty resulting from the uncertainty in future emissions, and to obtain results consistent with the global-average temperature change projections of Section 9.3.3, in Section 11.5.1.2 we derive projections for thermal expansion and land ice melt for the scenarios of the IPCC Special Report on Emissions Scenarios (SRES) (Nakićenović *et al.*, 2000) (see also Box 9.1 in Chapter 9, Section 9.1). The results are given as sea level change relative to 1990 in order to facilitate comparison with previous IPCC reports, which used 1990 as their base date.

11.5.1.1 Projections for a single scenario based on a range of AOGCMs
Thermal expansion
Over the hundred years 1990 to 2090, the AOGCM experiments for IS92a including sulphate aerosols (GS experiments – see Chapter 9, Table 9.1) show global-average sea level rise from

Table 11.11: *Sea level rise from thermal expansion from AOGCM experiments following the IS92a scenario for the 21st century, including the direct effect of sulphate aerosols. See Chapter 8, Table 8.1 and Chapter 9, Table 9.1 for further details of models and experiments. See Table 11.2 for thermal expansion from AOGCM experiments for the 20th century.*

Experiment	ΔT_g (°C)	Sea level rise (m)	
	1990 to 2090[a]	1990 to 2040	1990 to 2090[a]
CGCM1 GS	3.7	0.12	0.37
CGCM2 GS	3.6	0.11	0.33
CSIRO Mk2 GS	2.7	0.11	0.28
ECHAM4/OPYC3 GS[b]	—	0.11	—
GFDL_R15_a GS[c]	—	0.13	—
GFDL_R15_b GS	3.2	0.12	0.29
GFDL_R30_c GS	2.8	0.12	0.31
HadCM2 GS	2.5	0.07	0.20
HadCM3 GSIO	2.8	0.07	0.20
MRI2 GS	1.5	0.04	0.09
DOE PCM GS	1.9	0.07	0.17

[a] An end date of 2090, rather than 2100, is chosen to match the last available date in some of the experiments.

[b] This experiment ends at 2050.

[c] This experiment ends at 2065.

ΔT_g Global average surface air temperature change.

thermal expansion in the range 0.09 to 0.37 m (Figure 11.1, Table 11.11). There is an acceleration through the 21st century; expansion for 2040 to 2090 is greater than for 1990 to 2040 by a factor of 1.4 to 2.1. Since the models experience the same forcing, the differences in the thermal expansion derive from differences in the physical behaviour of the models. Broadly speaking, the range of results reflects the systematic uncertainty of modelling in three factors: the size of the surface warming, the effectiveness of heat uptake by the ocean for a given warming (Gregory and Mitchell, 1997) and the expansion resulting from a given heat uptake (Russell *et al.*, 2000). The separation of the first two factors parallels the distinction made in Section 9.3.4.2 between the effects of climate feedback and heat uptake on the rate of climate change. Since models differ in regard to the second and third factors, experiments with a similar temperature change do not necessarily have a similar thermal expansion, as the results demonstrate.

Glaciers and ice caps

To make projections for future loss of mass from glaciers and ice caps, we have applied the methods of Gregory and Oerlemans (1998) and Van de Wal and Wild (2001) (Sections 11.2.2.2, 11.2.2.4) to the seasonally and geographically dependent temperature changes given by a range of AOGCM IS92a experiments including sulphate aerosols (Table 11.12). We adjust the results to be consistent with the assumption that the climate of 1865 to 1895 was 0.15 K warmer than the steady state for glaciers, following Zuo and Oerlemans (1997) (see also Section 11.4). Precipitation changes are not included, as they are not expected to have a strong influence on the global average (Section 11.2.2.3).

Table 11.12: *Calculations of glacier melt from AOGCM experiments following the IS92a scenario for the 21st century, including the direct effect of sulphate aerosols. See Tables 8.1 and 9.1 for further details of models and experiments.*

Experiment	B (mm/yr) 1990	ΔT_g (°C) 1990 to 2090	Sea level rise (m) 1990 to 2090		$\partial B/\partial T_g$ (mm/yr/°C)
			Constant area	Changing area	
CGCM1 GS	0.43	3.7	0.15	0.11	0.65
CSIRO Mk2 GS	0.52	2.7	0.15	0.11	0.73
CSM 1.3 GS	0.45	1.8	0.10	0.07	0.61
ECHAM4/OPYC3 GS[a]	0.56	–	–	–	0.64
GFDL_R15_a GS[b]	0.42	–	–	–	0.58
GFDL_R15_b GS	0.44	3.2	0.13	0.09	0.54
GFDL_R30_c GS	0.33	2.8	0.12	0.08	0.53
HadCM2 GS[c]	0.44	2.5	0.11	0.08	0.61
HadCM3 GSIO	0.31	2.8	0.11	0.08	0.62
MRI2 GS	0.22	1.5	0.06	0.05	0.60
DOE PCM GS	0.42	1.9	0.09	0.06	0.59

[a] This experiment ends at 2050.

[b] This experiment ends at 2065.

[c] Similar results for constant area were obtained for an ensemble of HadCM2 GS experiments by Gregory and Oerlemans (1998).

B Global glacier mass balance for constant glacier area, expressed as sea level equivalent.

$\partial B/\partial T_g$ Sensitivity of global glacier mass balance for constant glacier area, expressed as sea level equivalent, to global average surface air temperature change.

ΔT_g Global average surface air temperature change.

Table 11.13: *Calculations of ice sheet mass changes using temperature and precipitation changes from AOGCM experiments following the IS92a scenario for the 21st century, including the direct effect of sulphate aerosols, to derive boundary conditions for an ice sheet model. See Tables 8.1 and 9.1 for further details of models and experiments.*

Experiment	Greenland					Antarctica			
	Sea level rise (m) 1990 to 2090	Sensitivity (mm/yr/ °C)		$\Delta T/\Delta T_g$	1/P dP/dT (%/°C)	Sea level rise (m) 1990 to 2090	Sensitivity (mm/yr/ °C)		$\Delta T/\Delta T_g$
		dB/dT_g	dB/dT				dB/dT_g	dB/dT	
CGCM1 GS	0.03	0.13	0.10	1.3	2.7	− 0.02	− 0.12	− 0.11	1.1
CSIRO Mk2 GS	0.02	0.16	0.08	2.0	5.9	− 0.07	− 0.37	− 0.33	1.1
CSM 1.3 GS	0.02	0.15	0.05	3.1	7.8	− 0.04	− 0.31	− 0.27	1.1
ECHAM4/OPYC3 GS [a]	–	0.03	0.03	1.2	6.5	–	− 0.48	− 0.32	1.5
GFDL_R15_a GS [b]	–	0.12	0.06	1.9	4.1	–	− 0.18	− 0.22	0.8
HadCM2 GS	0.02	0.10	0.07	1.4	4.0	− 0.04	− 0.21	− 0.17	1.2
HadCM3 GSIO	0.02	0.09	0.06	1.4	4.5	− 0.07	− 0.35	− 0.28	1.3
MRI2 GS	0.01	0.08	0.05	1.6	4.4	− 0.01	− 0.14	− 0.12	1.2
DOE PCM GS	0.02	0.14	0.06	2.2	5.6	− 0.07	− 0.48	− 0.30	1.6

[a] This experiment ends at 2050.

[b] This experiment ends at 2065.

dB/dT_g Ice-sheet mass balance sensitivity to global-average surface air temperature change, expressed as sea level equivalent.

dB/dT Ice-sheet mass balance sensitivity to surface air temperature change averaged over the ice sheet, expressed as sea level equivalent.

$\Delta T/\Delta T_g$ Slope of the regression of surface air temperature change averaged over the ice sheet against global-average change.

1/P dP/dT Fractional change in ice-sheet average precipitation as a function of temperature change.

For constant glacier area, from the AOGCM IS92a experiments including sulphate aerosol, predicted sea level rise from glacier melt over the hundred years 1990 to 2090 lies in the range 0.06 to 0.15 m. The variation is due to three factors. First, the global average temperature change varies between models. A larger temperature rise tends to give more melting, but they are not linearly related, since the total melt depends on the time-integrated temperature change. Second, the global mass balance sensitivity to temperature change varies among AOGCMs because of their different seasonal and regional distribution of temperature change. Third, the glaciers are already adjusting to climate change during the 20th century, and any such imbalance will persist during the 21st century, in addition to the further imbalance due to future climate change. The global average temperature change and glacier mass balance sensitivity may not be independent factors, since both are affected by regional climate feedbacks. The sensitivity and the present imbalance are related factors, because a larger sensitivity implies a greater present imbalance.

With glacier area contracting as the volume reduces, the estimated sea level rise contribution is in the range 0.05 to 0.11 m, about 25% less than if constant area is assumed, similar to the findings of Oerlemans *et al.* (1998) and Van de Wal and Wild (2001). The time-dependence of glacier area means the results can no longer be represented by a global glacier mass balance sensitivity.

Glaciers and ice caps on the margins of the Greenland and Antarctic ice sheets are omitted from these calculations, because they are included in the ice sheet projections below. These ice masses have a large area (Table 11.3), but experience little ablation on account of being in very cold climates. Van de Wal

and Wild (2001) find that the Greenland marginal glaciers contribute an additional 6% to glacier melt in a scenario of CO_2 doubling over 70 years. Similar calculations using the AOGCM IS92a results give a maximum contribution of 14 mm for 1990 to 2100. For the Antarctic marginal glaciers, the ambient temperatures are too low for there to be any significant surface runoff. Increasing temperatures will increase the runoff and enlarge the area experiencing ablation, but their contribution is very likely to remain small. For instance, Drewry and Morris (1992) calculate a contribution of 0.012 mm/yr/°C to the global glacier mass balance sensitivity from the glacier area of 20,000 km^2 which currently experiences some melting on the Antarctic Peninsula.

Lack of information concerning glacier areas and precipitation over glaciers, together with uncertainty over the projected changes in glacier area, lead to uncertainty in the results. This is assessed as ± 40%, matching the uncertainty of the observed mass balance estimate of Dyurgerov and Meier (1997b).

Greenland and Antarctic ice sheets
To make projections of Greenland and Antarctic ice sheet mass changes consistent with the IS92a AOGCM experiments including sulphate aerosols, we have integrated the ice-sheet model of Huybrechts and De Wolde (1999) using boundary conditions of temperature and precipitation derived by perturbing present day climatology according to the geographically and seasonally dependent pattern changes predicted by the T106 ECHAM4 model (Wild and Ohmura, 2000) for a doubling of CO_2. To generate time-dependent boundary conditions, these patterns were scaled with the area average changes over the ice sheets as a function of time for each AOGCM experiment using a method similar to that described by Huybrechts *et al.* (1999).

Table 11.14: *Sea level rise 1990 to 2100 due to climate change derived from AOGCM experiments following the IS92a scenario, including the direct effect of sulphate aerosols. See Tables 8.1 and 9.1 for further details of models and experiments. Results were extrapolated to 2100 for experiments ending at earlier dates. The uncertainties shown in the land ice terms are those discussed in this section. For comparison the projection of Warrick et al. (1996) (in the SAR) is also included. Note that the minimum of the sum of the components is not identical with the sum of the minima because the smallest values of the components do not all come from the same AOGCM, and because for each model the land ice uncertainties have been combined in quadrature; similarly for the maxima, which also include non-zero contributions from smaller terms.*

Experiment		Sea level rise (m) 1990 to 2100									
		Expansion		Glaciers		Greenland		Antarctica [a]		Sum [b]	
		min	max	min	max	min	max	min	max	min	max
CGCM1 GS		0.43		0.03	0.23	0.00	0.07	–0.07	0.02	0.45	0.77
CSIRO Mk2 GS		0.33		0.02	0.22	–0.01	0.08	–0.12	–0.04	0.29	0.60
ECHAM4/OPYC3 GS		0.30		0.02	0.18	–0.02	0.03	–0.17	–0.06	0.19	0.48
GFDL_R15_a GS		0.38		0.02	0.19	–0.01	0.09	–0.09	–0.01	0.37	0.67
HadCM2 GS		0.23		0.02	0.17	–0.01	0.05	–0.09	0.00	0.21	0.48
HadCM3 GSIO		0.24		0.02	0.18	0.00	0.05	–0.13	–0.03	0.18	0.46
MRI2 GS		0.11		0.01	0.11	0.00	0.03	–0.04	0.00	0.11	0.31
DOE PCM GS		0.19		0.01	0.13	–0.01	0.06	–0.13	–0.04	0.12	0.37
Range		0.11	0.43	0.01	0.23	–0.02	0.09	–0.17	0.02	0.11	0.77
Central value		0.27		0.12		+0.04		–0.08		0.44	
SAR 7.5.2.4	Best estimate	0.28		0.16		+0.06		–0.01		0.49	
	Range									0.20	0.86

[a] Note that this range does not allow for uncertainty relating to ice-dynamical changes in the West Antarctic ice sheet. See Section 11.5.4.3 for a full discussion.

[b] Including contributions from permafrost, sedimentation, and adjustment of ice sheets to past climate change.

The marginal glaciers and ice caps on Greenland and Antarctica were included in the ice sheet area. The calculated contributions from these small ice masses have some uncertainty resulting from the limited spatial resolution of the ice sheet model.

For 1990 to 2090 in the AOGCM GS experiments, Greenland contributes 0.01 to 0.03 m and Antarctica –0.07 to –0.01 m to global average sea level (Table 11.13). Note that these sea level contributions result solely from recent and projected future climate change; they do not include the response to past climate change (discussed in Sections 11.2.3.3 and 11.3.1).

Mass balance sensitivities are derived by regressing rate of change of mass against global or local temperature change (note that they include the effect of precipitation changes) (Table 11.13). The Greenland local sensitivities are smaller than some of the values reported previously from other methods (Section 11.2.3.4 and Table 11.7) and by Warrick *et al.* (1996) because of the larger precipitation increases and the seasonality of temperature changes (less increase in summer) predicted by AOGCMs, and the smaller temperature rise in the ablation zone (as compared to the ice-sheet average) projected by the T106 ECHAM4 time slice results. The Antarctic sensitivities are less negative than those in Table 11.7 because the AOGCMs predict smaller precipitation increases.

The use of a range of AOGCMs represents the uncertainty in modelling changing circulation patterns, which lead to both changes in temperature and precipitation, as noted by Kapsner *et al.* (1995) and Cuffey and Clow (1997) from the results from Greenland ice cores. The range of AOGCM thermodynamic and circulation responses gives a range of 4 to 8%/°C for Greenland precipitation increases, generally less than indicated by ice-cores

for the glacial-interglacial transition, but more than for Holocene variability (Section 11.2.3.4). If precipitation did not increase at all with greenhouse warming, Greenland local sensitivities would be larger, by 0.05 to 0.1 mm/yr/°C (see also Table 11.7). Given that all AOGCMs agree on an increase, but differ on the strength of the relationship, we include an uncertainty of ±0.02 mm/yr/°C in Table 11.13 on the Greenland local sensitivities, being the product of the standard deviation of precipitation increase (1.5%/°C) and the current Greenland accumulation (1.4 mm/yr sea level equivalent, Table 11.5).

Estimates of Greenland runoff (Table 11.5) have a standard error of about ± 10%. This reflects uncertainty in the degree-day method (Braithwaite, 1995) and refreezing parametrization (Janssens and Huybrechts, 2000) used to calculate Greenland ablation. Given that a typical size of the sensitivity of ablation to temperature change is 0.3 mm/yr/°C (Table 11.7), we adopt an additional uncertainty of ± 0.03 mm/yr/°C for the local Greenland sensitivities in Table 11.13. We include a separate uncertainty of the same size to reflect the possible sensitivity to use of different high-resolution geographical patterns of temperature and precipitation change (the T106 ECHAM4 pattern was the only one available). As an estimate of the uncertainty related to changes in iceberg discharge and area-elevation distribution, we ascribe an uncertainty of ± 10% to the net mass change, on the basis of the magnitude of the dynamic response for Greenland described in Section 11.2.3.4.

For Antarctica, uncertainty introduced by ablation model parameters need not be considered because melting remains very small for the temperature scenarios considered for the 21st century. Ice-dynamical uncertainties are much more difficult to
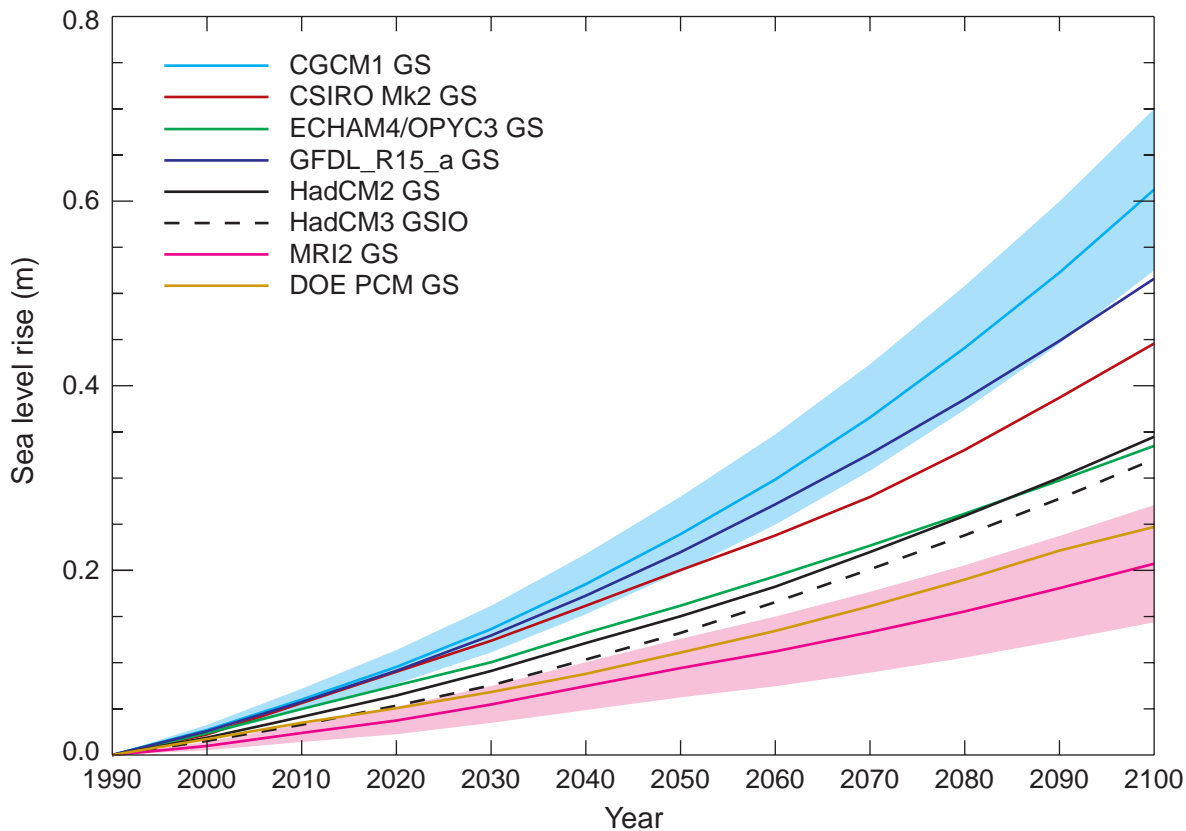
Figure 11.11: Global average sea level rise 1990 to 2100 for the IS92a scenario, including the direct effect of sulphate aerosols. Thermal expansion and land ice changes were calculated from AOGCM experiments, and contributions from changes in permafrost, the effect of sediment deposition and the long-term adjustment of the ice sheets to past climate change were added. For the models that project the largest (CGCM1) and the smallest (MRI2) sea level change, the shaded region shows the bounds of uncertainty associated with land ice changes, permafrost changes and sediment deposition. Uncertainties are not shown for the other models, but can be found in Table 11.14. The outermost limits of the shaded regions indicate our range of uncertainty in projecting sea level change for the IS92a scenario.

determine. See Section 11.5.4.3 for a detailed discussion. We include an uncertainty of 0.08 mm/yr/°C on the local sensitivity, which is its inter-model standard deviation, to reflect the spread of precipitation changes as a function of temperature.

Total

To obtain predictions of global average sea level rise for 1990-2100 for the IS92a scenario with sulphate aerosols, we calculate the sum of the contributions from thermal expansion, glaciers and ice sheets for each AOGCM, and add the 0 to 0.5 mm/yr from the continuing evolution of the ice sheets in response to past climate change (Section 11.2.3.1) and smaller terms from thawing of permafrost (Section 11.2.5) and the effect of sedimentation (Section 11.2.6). The range of our results is 0.11 to 0.77 m (Table 11.14, Figure 11.11), which should be compared with the range of 0.20 to 0.86 m given by Warrick *et al.* (1996) (SAR Section 7.5.2.4, Figure 7.7) for the same scenario. The AOGCMs have a range of effective climate sensitivities from 1.4 to 4.2°C (Table 9.1), similar to the range of 1.5 to 4.5°C used by Warrick *et al.* The AOGCM thermal expansion values are generally larger than those of Warrick *et al.* (SAR Section 7.5.2.4, Figure 7.8), but the other terms are mostly smaller (i.e., more negative in the case of Antarctica).

Warrick *et al.* included a positive term to allow for the possible instability of the WAIS. We have omitted this because it is now widely agreed that major loss of grounded ice and accelerated sea level rise are very unlikely during the 21st century (Section 11.5.4.3). The size of our range is an indication of the systematic uncertainty in modelling radiative forcing, climate and sea level changes. Uncertainties in modelling the carbon cycle and atmospheric chemistry are not covered by this range, because the AOGCMs are all given similar atmospheric concentrations as input.

11.5.1.2 Projections for SRES scenarios

Few AOGCM experiments have been done with any of the SRES emissions scenarios. Therefore to establish the range of sea level rise resulting from the choice of different SRES scenarios, we use results for thermal expansion and global-average temperature change from a simple climate model based on that of Raper *et al.* (1996) and calibrated individually for seven AOGCMs (CSIRO Mk2, CSM 1.3, ECHAM4/OPYC3, GFDL_R15_a, HadCM2, HadCM3, DOE PCM). The calibration is discussed in Chapter 9, Section 9.3.3 and the Appendix to Chapter 9. The AOGCMs used have a range of effective climate sensitivity of 1.7 to 4.2°C (Table 9.1). We
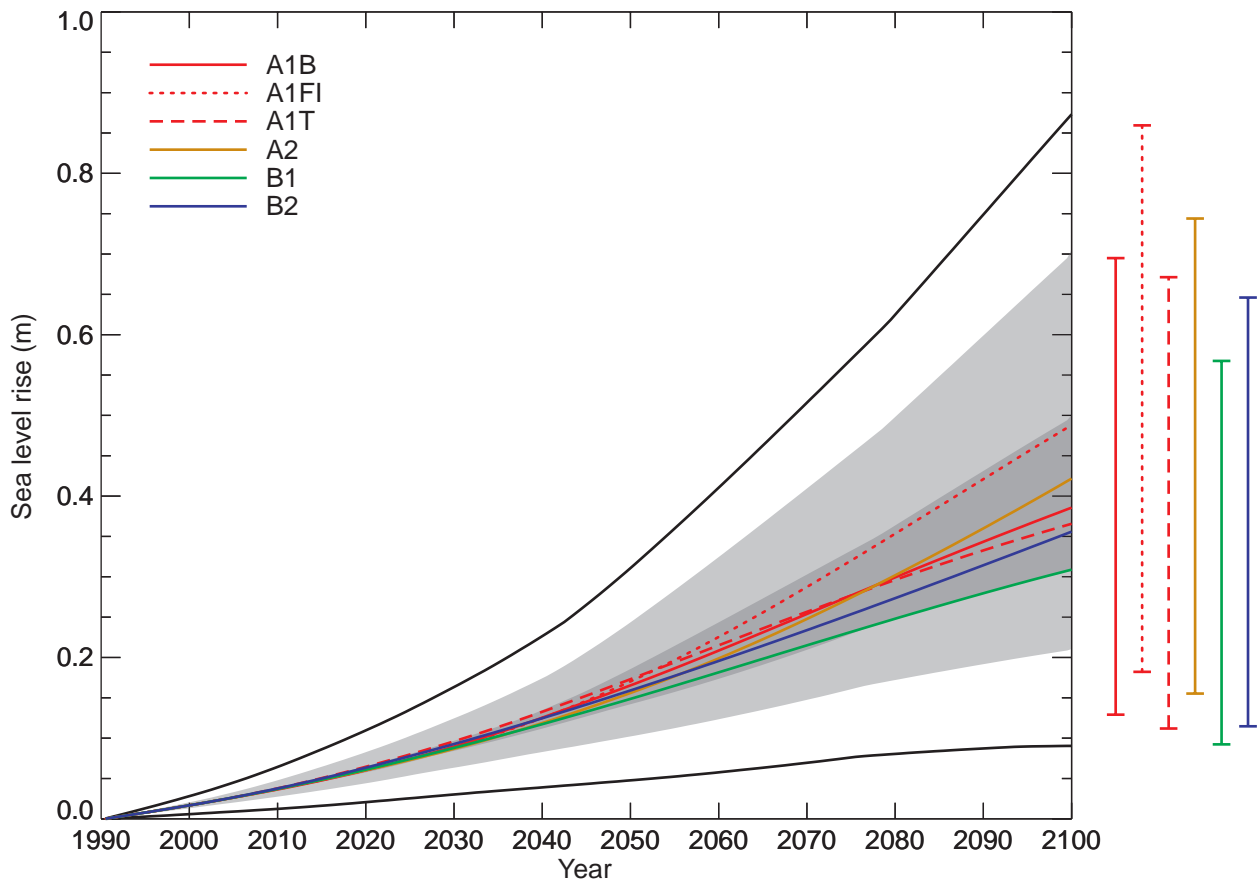
Figure 11.12: Global average sea level rise 1990 to 2100 for the SRES scenarios. Thermal expansion and land ice changes were calculated using a simple climate model calibrated separately for each of seven AOGCMs, and contributions from changes in permafrost, the effect of sediment deposition and the long-term adjustment of the ice sheets to past climate change were added. Each of the six lines appearing in the key is the average of AOGCMs for one of the six illustrative scenarios. The region in dark shading shows the range of the average of AOGCMs for all 35 SRES scenarios. The region in light shading shows the range of all AOGCMs for all 35 scenarios. The region delimited by the outermost lines shows the range of all AOGCMs and scenarios including uncertainty in land-ice changes, permafrost changes and sediment deposition. Note that this range does not allow for uncertainty relating to ice-dynamical changes in the West Antarctic ice sheet. See 11.5.4.3 for a full discussion. The bars show the range in 2100 of all AOGCMs for the six illustrative scenarios.

calculate land-ice changes using the global average temperature change from the simple model and the global average mass balance sensitivities estimated from the AOGCM IS92a experiments in Section 11.5.1.1 (Tables 11.12 and 11.13). We add contributions from the continuing evolution of the ice sheets in response to past climate change, thawing of permafrost, and the effect of sedimentation (the same as in Section 11.5.1.1). The methods used to make the sea level projections are documented in detail in the Appendix to this chapter.

For the complete range of AOGCMs and SRES scenarios and including uncertainties in land-ice changes, permafrost changes and sediment deposition, global average sea level is projected to rise by 0.09 to 0.88 m over 1990 to 2100, with a central value of 0.48 m (Figure 11.12). The central value gives an average rate of 2.2 to 4.4 times the rate over the 20th century.

The corresponding range reported by Warrick *et al.* (1996) (representing scenario uncertainty by using all the IS92 scenarios with time-dependent sulphate aerosol) was 0.13 to 0.94 m, obtained using a simple model with climate sensitivities

of 1.5 to 4.5°C. Their upper bound is larger than ours. Ice sheet mass balance sensitivities derived from AOGCMs (see Section 11.5.1.1) are smaller (less positive or more negative) than those used by Warrick *et al.*, while the method we have employed for calculating glacier mass loss (Sections 11.2.2 and 11.5.1.1) gives a smaller sea level contribution for similar scenarios than the heuristic model of Wigley and Raper (1995) employed by Warrick *et al.*

In addition, Warrick *et al.* included an allowance for ice-dynamical changes in the WAIS. The range we have given does not include such changes. The contribution of the WAIS is potentially important on the longer term, but it is now widely agreed that major loss of grounded ice from the WAIS and consequent accelerated sea-level rise are very unlikely during the 21st century. Allowing for the possible effects of processes not adequately represented in present models, two risk assessment studies involving panels of experts concluded that there was a 5% chance that by 2100 the WAIS could make a substantial contribution to sea level rise, of 0.16 m (Titus and
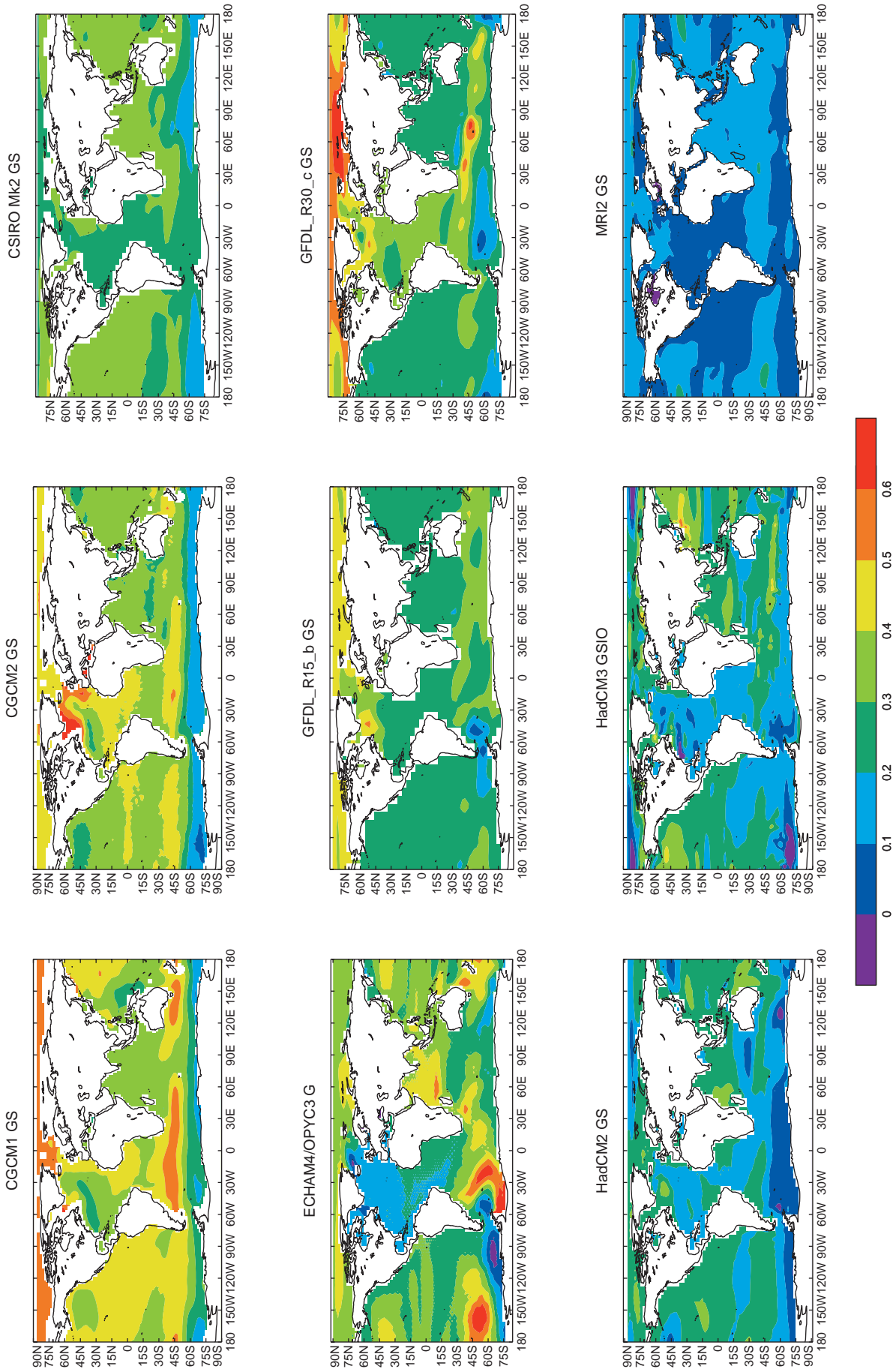
Figure 11.13: Sea level change in metres over the 21st century resulting from thermal expansion and ocean circulation changes calculated from AOGCM experiments following the IS92a scenario and including the direct effect of sulphate aerosol (except that ECHAM4/OPYC3 G is shown instead of GS, because GS ends in 2050). Each field is the difference in sea level change between the last decade of the experiment and the decade 100 years earlier. See Tables 8.1 and 9.1 for further details of models and experiments.

Table 11.15: *Spatial standard deviation, local minimum and local maximum of sea level change during the 21st century due to ocean processes, from AOGCM experiments following the IS92a scenario for greenhouse gases, including the direct effect of sulphate aerosols. See Tables 8.1 and 9.1 for further details of models and experiments. Sea level change was calculated as the difference between the final decade of each experiment and the decade 100 years earlier. Sea level changes due to land ice and water storage are not included.*

Experiment	Std deviation (m)	Divided by global average		
		Std deviation	Minimum	Maximum
CGCM1 GS	0.07	0.19	0.3	1.6
CGCM2 GS	0.08	0.23	0.2	2.2
CSIRO Mk2 GS	0.05	0.15	0.5	1.3
ECHAM4/OPYC3 G[a]	0.10	0.34	−1.2	2.3
GFDL_R15_b GS	0.05	0.18	0.3	1.8
GFDL_R30_c GS	0.07	0.25	0.2	2.5
HadCM2 GS	0.06	0.29	−0.1	1.7
HadCM3 GSIO	0.07	0.32	−0.5	2.2
MRI2 GS	0.04	0.35	−1.2	2.2

[a] This experiment does not include sulphate aerosols. The ECHAM4/OPYC3 experiment including sulphates extends only to 2050.

Narayanan, 1996) or 0.5 m (Vaughan and Spouge, 2001). These studies also noted a 5% chance of WAIS contributing a sea level fall of 0.18 m or 0.4 m respectively. (See Section 11.5.4.3 for a full discussion.)

The range we have given also does not take account of uncertainty in modelling of radiative forcing, the carbon cycle, atmospheric chemistry, or storage of water in the terrestrial environment. The recent publications by Gornitz *et al.* (1997) and Sahagian (2000) indicate that this last term could be significant (Section 11.2.5). Future changes in terrestrial storage depend on societal decisions on the use of ground water, the building of reservoirs and other factors. We are not currently in a position to make projections incorporating future changes in these factors, although we note that the assumptions behind the construction of the SRES scenarios imply increasing water consumption, which may entail both more ground water extraction and more reservoir capacity. Continued anthropogenic water storage on land at its current rate could change the projected sea level rise 1990 to 2100 by between −0.21 and +0.11 m. We emphasise that estimates of the relevant factors are highly uncertain (see Sections 11.2.5 and 11.4).

The evolution of sea level rise for the average of the seven AOGCMs for each of the six illustrative SRES scenarios is shown in Figure 11.12, and the shading shows the range for all 35 SRES scenarios. It is apparent that the variation due to the choice of scenario alone is relatively small over the next few decades. The range spanned by the SRES scenarios by 2040 is only 0.02 m or less. By 2100, the scenario range has increased to 0.18 m, about 50% of the central value. All the AOGCMs have a similar range at 2100 expressed as a fraction of their central value. Of the six illustrative scenarios, A1FI gives the largest sea level rise and B1 the smallest.

The average-AOGCM range for all 35 scenarios (dark shading in Figure 11.12) covers about one third of the all-AOGCM range (light shading). That is, for sea level rise 1990 to 2100, the uncertainty in climate sensitivity and heat uptake, represented by the spread of AOGCMs, is more important than the uncertainty from choice of emissions scenario. This is different for three reasons from the case of global average temperature change (Section 9.3.2.1), where the scenario and modelling uncertainties are comparable. First, the compensation between climate sensitivity and heat uptake does not apply to thermal expansion. Second, models with large climate sensitivity and temperature change consequently have a large land-ice melt contribution to sea level. Third, both thermal expansion and land-ice melt depend on past climate change, being approximately proportional to the time-integral of temperature change; the SRES scenarios differ by less in respect of the time-integral of temperature change over the interval 1990 to 2100 than they do in respect of the temperature change at 2100.

11.5.2 Regional Sea Level Change

The geographical distribution of sea level change caused by ocean processes can be calculated from AOGCM results (see Gregory *et al.*, 2001, for methods). This was not possible with the simple climate models used by Warrick *et al.* (1996). Results for sea level change from ocean processes in the 21st century are shown in Figure 11.13 for AOGCM experiments used in Section 11.5.1.1. Some regions show a sea level rise substantially more than the global average (in many cases of more than twice the average), and others a sea level fall (Table 11.15) (note that these figures do not include sea level rise due to land ice changes). The standard deviation of sea level change is 15 to 35% of the global average sea level rise from thermal expansion.

In each of these experiments, a non-uniform pattern of sea level rise emerges above the background of temporal variability in the latter part of the 21st century. However, the patterns given by the different models (Figure 11.13) are not similar in detail. The largest correlations are between models which are similar in formulation: 0.65 between CGCM1 and CGCM2, 0.63 between GFDL_R15_b and GFDL_R30_c. The largest correlations between models from different centres are 0.60 between CSIRO Mk2 and HadCM2, 0.58 between CGCM2 and GFDL_R30_c. The majority of correlations are less than 0.4, indicating no

significant similarity (Gregory *et al.*, 2001). The disagreement between models is partly a reflection of the differences in ocean model formulation that are also responsible for the spread in the global average heat uptake and thermal expansion (Sections 11.2.1.2, 11.5.1.1). In addition, the models predict different changes in surface windstress, with consequences for changes in ocean circulation and subduction. More detailed analysis is needed to elucidate the reasons for the differences in patterns. The lack of similarity means that our confidence in predictions of local sea level changes is low. However, we can identify a few common features on the regional and basin scale (see also Gregory *et al.*, 2001).

Seven of the nine models in Table 11.14 (also Bryan, 1996; Russell *et al.*, 2000) exhibit a maximum sea level rise in the Arctic Ocean. A possible reason for this is a freshening of the Arctic due to increased river runoff or precipitation over the ocean (Bryan, 1996; Miller and Russell, 2000). The fall in salinity leads to a reduction of density, which requires a compensating sea level rise in order to maintain the pressure gradient at depth.

Seven of the models (also Gregory, 1993; Bryan, 1996) show a minimum of sea level rise in the circumpolar Southern Ocean south of 60°S. This occurs despite the fact that the Southern Ocean is a region of pronounced heat uptake (e.g., Murphy and Mitchell, 1995; Hirst *et al.*, 1996). The low thermal expansion coefficient at the cold temperatures of the high southern latitudes, changes in wind patterns and transport of the heat taken up to lower latitudes are all possible explanations.

Bryan (1996) draws attention to a dipole pattern in sea level change in the north-west Atlantic; there is a reduced rise south of the Gulf Stream extension and enhanced rise to the north, which corresponds to a weakening of the sea surface gradient across the current. This would be consistent with a weakening of the upper

branch of the North Atlantic circulation, which is a response to greenhouse warming observed in many AOGCM experiments (e.g., Manabe and Stouffer, 1993, 1994; Hirst, 1998). This can be seen in all the models considered here except ECHAM4/OPYC3, in which the Atlantic thermohaline circulation does not weaken (Latif and Roeckner, 2000).

Local land movements, both isostatic and tectonic (Sections 11.2.4.1, 11.2.6), will continue in the 21st century at rates which are unaffected by climate change, and should be added to the regional variation described in this section. On account of the increased eustatic rate of rise in the 21st century (Section 11.5.1) it can be expected that by 2100 many regions currently experiencing relative sea level fall owing to isostatic rebound will instead have a rising relative sea level.

All the global models discussed here have a spatial resolution of 1 to 3°. To obtain information about mean sea level changes at higher resolution is currently not practical; a regional model such as that of Kauker (1998) would be needed.

11.5.3 Implications for Coastal Regions

To determine the practical consequences of projections of global sea level rise in particular coastal regions, it is necessary to understand the various components leading to relative sea level changes. These components include local land movements, global eustatic sea level rise, any spatial variability from that global average, local meteorological changes and changes in the frequency of extreme events.

11.5.3.1 Mean sea level

Titus and Narayanan (1996) propose a simple method for computing local projections of mean sea level rise given historical observations at a site and projections of global average sea level (such as Figure 11.12). To allow for local land movements and our current inability to model sea level change accurately (Section 11.4), they propose linearly extrapolating the historical record and adding to this a globally averaged projection. However, they point out that to avoid double counting, it is necessary to correct the global projection for the corresponding modelled trend of sea level rise during the period of the historical observations. Caution is required in applying this method directly to the projections of this chapter for several reasons. First, current model projections indicate substantial spatial variability in sea level rise. This variability has a standard deviation of up to 0.1 m by 2100; some locations experience a sea level rise of more than twice the global-average thermal expansion, while others may have a fall in sea level (Section 11.5.2; Table 11.15). Second, there are uncertainties in the accuracy of the trend from the historical record and in modelling of past sea level changes (Sections 11.3.2.1 and 11.4). Third, as well as changes in mean sea level there may be changes in the local meteorological regime resulting in modified storm surge statistics (Section 11.5.3.2). If the method of Titus and Narayanan is not applied, it is nonetheless important to recognise that in all models and scenarios the rate of local sea level rise in the 21st century is projected to be greater than in the 20th century at the great majority of coastal locations.
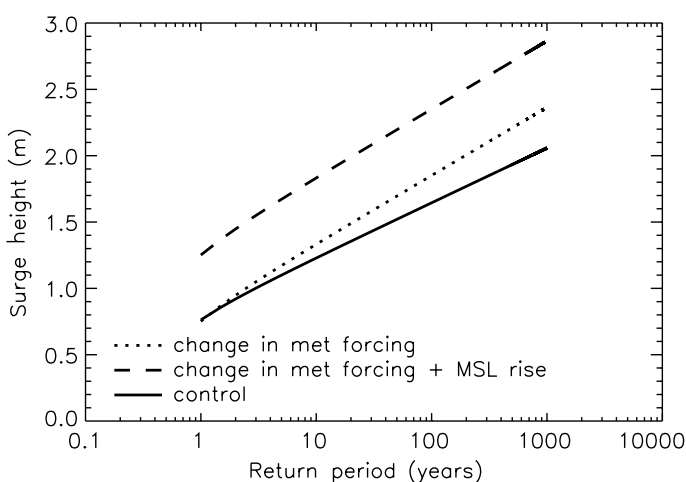
Figure 11.14: Frequency of extreme water level, expressed as return period, from a storm surge model for present day conditions (control) and the projected climate around 2100 for Immingham on the east coast of England, showing changes resulting from mean sea level rise and changes in meteorological forcing. The water level is relative to the sum of present day mean sea level and the tide at the time of the surge. (From Lowe *et al.*, 2001.)

11.5.3.2 Extremes of sea level: storm surges and waves

The probability of flood risk in coastal areas is generally expressed in terms of extreme sea level distributions. Such distributions are usually computed from observed annual maximum sea levels from several decades of tide gauge data, or from numerical models. While such distributions are readily available for many locations, a worldwide set has never been computed to common standards for studies of impacts of global sea level change.

Changes in the highest sea levels at a given locality could result mainly from two effects. First, if mean sea level rises, the present extreme levels will be attained more frequently, all else being equal. This may imply a significant increase in the area threatened with inundation (e.g., Hubbert and McInnes, 1999) and an increased risk within the existing flood plain. The effect can be estimated from a knowledge of the present day frequency of occurrence of extreme levels (e.g., Flather and Khandker, 1993; Lowe *et al.*, 2001; Figure 11.14).

Second, changes in storm surge heights would result from alterations to the occurrence of strong winds and low pressures. At low-latitude locations, such as the Bay of Bengal, northern Australia and the southern USA, tropical cyclones are the primary cause of storm surges. Changes in frequency and intensity of tropical cyclones could result from alterations to sea surface temperature, large-scale atmospheric circulation and the characteristics of ENSO (Pittock *et al.*, 1996) but no consensus has yet emerged (see Box 10.2). In other places, such as southern Australia and north-west Europe, storm surges are associated with mid-latitude low-pressure systems. For instance, Hubbert and McInnes (1999) showed that increasing the wind speeds in historical storm surge events associated with the passage of cold fronts could lead to greater flooding in Port Phillip Bay, Victoria, Australia. Changes in extratropical storms also cannot be predicted with confidence (Section 9.3.6.3).

Several studies have attempted to quantify the consequences of changes in storm climatology for the north-west European continental shelf using regional models of the atmosphere and ocean. Using five-year integrations of the ECHAM T106 model for present and doubled CO_2, Von Storch and Reichardt (1997) and Flather and Smith (1998) did not find any significant changes in extreme events compared with the variability of the control climate (see also WASA Group, 1998). However, Langenberg *et al.* (1999) reported increases of 0.05 to 0.10 m in five-winter-mean high-water levels around all North Sea coasts, judged to be significant compared with observed natural variability. Lowe *et al.* (2001) undertook a similar study using multi-decadal integrations of the Hadley Centre regional climate model for the present climate and the end of the 21st century (Figure 11.14), finding statistically significant changes of up to 0.2 m in five-year extremes in the English Channel. Differences between these various results relate to the length of model integration and to systematic uncertainty in the modelling of both the atmospheric forcing and the ocean response.

Changes in wind forcing could result in changes to wave heights, but with the short integrations available, the WASA Group (Rider *et al.*, 1996) were not able to identify any significant changes for the North Atlantic and North Sea for a doubling of CO_2. Günther *et al.* (1998) noted that changes in future wave climate were similar to patterns of past variation.

11.5.4 Longer Term Changes

Anthropogenic emissions beyond 2100 are very uncertain, and we can only indicate a range of possibilities for sea level change. On the time-scale of centuries, thermal expansion and ice sheet changes are likely to be the most important processes.

11.5.4.1 Thermal expansion

The most important conclusion for thermal expansion is that it would continue to raise sea level for many centuries after stabilisation of greenhouse gas concentrations, so that its eventual contribution would be much larger than at the time of stabilisation.

Table 11.16: *Sea level rise due to thermal expansion in 2×CO₂ and 4×CO₂ experiments. See Chapter 8, Table 8.1 for further details of models.*

	Sea level rise (m) in 2×CO₂ experiment		$\Delta h_2/\Delta h_1$	Final sea level rise (m)	
	at 2×CO₂ (Δh_1)	500 yr later (Δh_2)		2×CO₂ experiment	4×CO₂ experiment
CLIMBER	0.16	0.67	4.2	0.78	1.44
ECHAM3/LSG	0.06	0.57	9.2	1.53[a]	2.56[a]
GFDL_R15_a	0.13	1.10	8.5	1.96	3.46
HadCM2	0.09	0.70	7.8	−	−
BERN2D GM	0.23	1.12	4.9	1.93	3.73
BERN2D HOR	0.22	0.92	4.2	1.28	4.30
UVic GM	0.11	0.44	3.9	0.53	1.24
UVic H	0.13	0.71	5.6	1.19	2.62
UVic HBL	0.10	0.44	4.3	0.65	1.78

[a] Estimated from the ECHAM3/LSG experiments by fitting the time-series with exponential impulse response functions (Voss and Mikolajewicz, 2001).
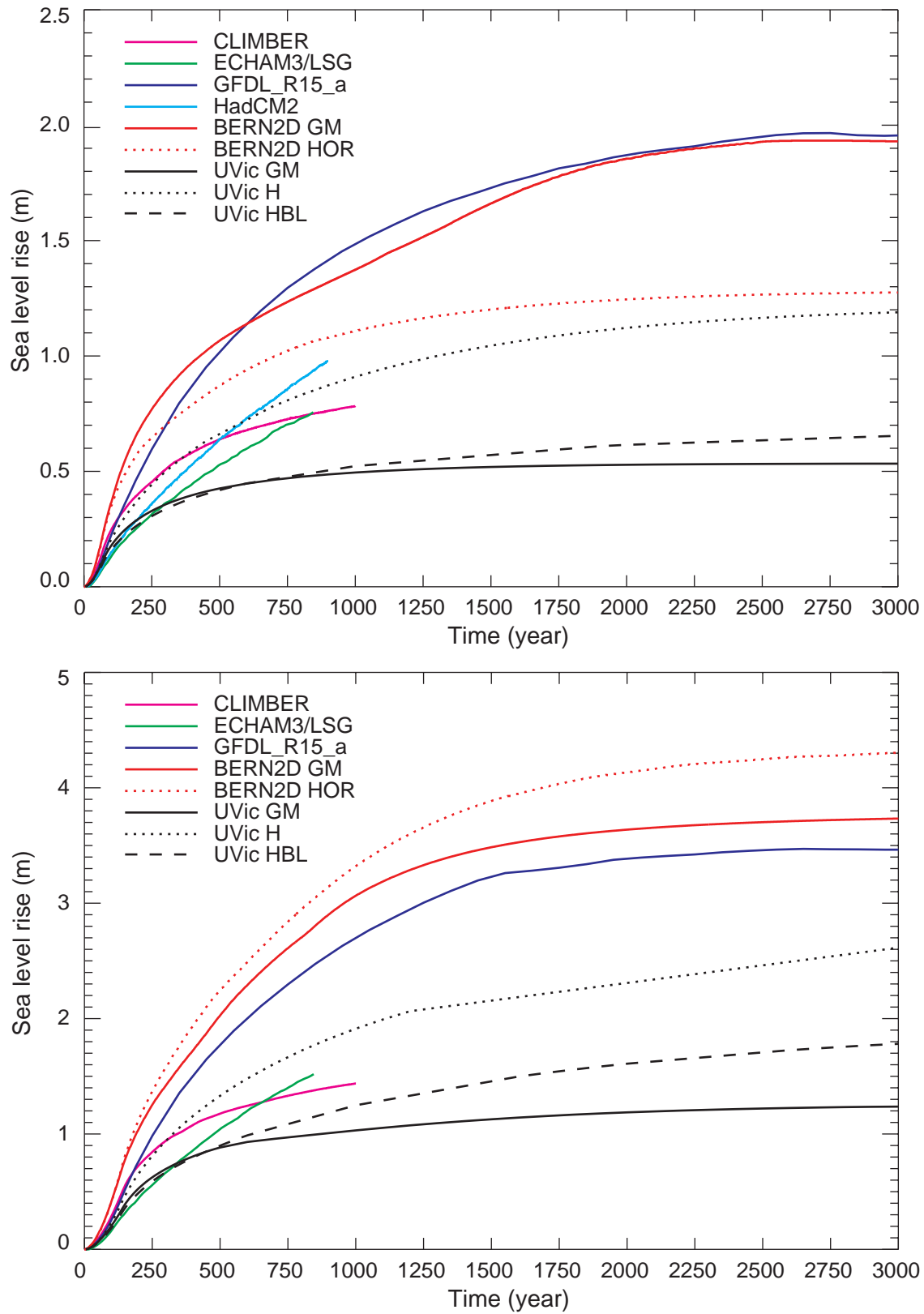
Figure 11.15: Global average sea level rise from thermal expansion in model experiments with CO_2 (a) increasing at 1%/yr for 70 years and then held constant at 2× its initial (preindustrial) concentration; (b) increasing at 1%/yr for 140 years and then held constant at 4× its initial (preindustrial) concentration.

A number of investigations have aimed to quantify this delayed but inevitable consequence of the enhanced greenhouse effect, using a simple scenario in which carbon dioxide concentration increases rapidly (at 1%/yr, not intended as a realistic historical scenario) up to double or four times its initial value (referred to as $2 \times CO_2$ and $4 \times CO_2$), and thereafter remains constant. ($2 \times CO_2$ is about 540 ppm by volume, and $4 \times CO_2$ about 1080 ppm.) Long experiments of this kind have been run with three AOGCMs (Chapter 8, Table 8.1): GFDL (Manabe and Stouffer, 1994; Stouffer and Manabe, 1999), ECHAM3/LSG (Voss and Mikolajewicz, 2001) and HadCM2 (Senior and Mitchell, 2000), but owing to the computational requirement, only one of these (GFDL) has been continued until a steady state is reached. Models of intermediate complexity (Chapter 8, Table 8.1) have also been employed: CLIMBER, BERN2D and UVic. These models have a less detailed representation of some important processes, but are less expensive to run for millennia.

Thermal expansion could be greater in one model than another either because the surface warming is larger, or because the warming penetrates more deeply (Figure 11.15, Table 11.16; the suffixes to the BERN2D and UVic model names indicate versions of the models with different parametrizations of heat transport processes). For instance, UVic H and UVic GM show markedly different expansion, although they have similar surface warming (Weaver and Wiebe, 1999). The $4 \times CO_2$ experiment with each model generally has around twice the expansion of the $2 \times CO_2$ experiment, but the BERN2D HOR $4 \times CO_2$ experiment has more than three times, because the Atlantic thermohaline circulation collapses, permitting greater warming and adding about 0.5 m to thermal expansion (Knutti and Stocker, 2000). (See also Section 9.3.4.4.) In all the models reported here, a
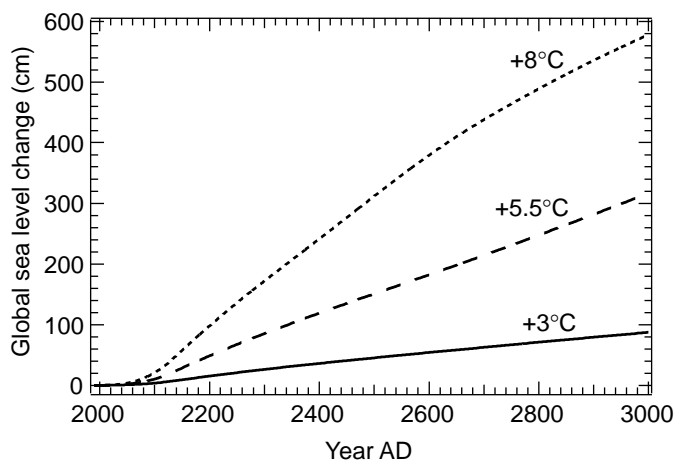
vertical temperature gradient is maintained, although in some cases weakened. If the whole depth of the ocean warmed to match the surface temperature, thermal expansion would be considerably larger.

The long time-scale (of the order of 1,000 years) on which thermal expansion approaches its eventual level is characteristic of the weak diffusion and slow circulation processes that transport heat to the deep ocean. On account of the time-scale, the thermal expansion in the $2 \times CO_2$ experiments after 500 years of constant CO_2 is 4 to 9 times greater than at the time when the concentration stabilises. Even by this time, it may only have reached half of its eventual level, which models suggest may lie within a range of 0.5 to 2.0 m for $2 \times CO_2$ and 1 to 4 m for $4 \times CO_2$. For the first 1,000 years, the $4 \times CO_2$ models give 1 to 3 m.

11.5.4.2 Glaciers and ice caps
Melting of all existing glaciers and ice caps would raise sea level by 0.5 m (Table 11.3). For 1990 to 2100 in IS92a, the projected loss from land-ice outside Greenland and Antarctica is 0.05 to 0.11 m (Table 11.12). Further contraction of glacier area and retreat to high altitude will restrict ablation, so we cannot use the 21st century rates to deduce that there is a time by which all glacier mass will have disappeared. However, the loss of a substantial fraction of the total glacier mass is likely. The viability of any particular glacier or ice cap will depend on whether there remains any part of it, at high altitude, where ablation does not exceed accumulation over the annual cycle. Areas which are currently marginally glaciated are most likely to become ice-free.

11.5.4.3 Greenland and Antarctic ice sheets
Several modelling studies have been conducted for time periods of several centuries to millennia (Van de Wal and Oerlemans, 1997; Warner and Budd, 1998; Huybrechts and De Wolde, 1999; Greve, 2000). A main conclusion is that the ice sheets would continue to react to the imposed climatic change during the next millennium, even if the warming stabilised early in the 22nd century. Whereas Greenland and Antarctica may largely counteract one another for most of the 21st century (Section 11.2.3.4), this situation would no longer hold after that and their combined contribution would be a rise in sea level.

Greenland ice sheet
The Greenland ice sheet is the most vulnerable to climatic warming. As the temperature rises, ablation will increase. For moderate warming, the ice sheet can be retained with reduced extent and modified shape if this results in less ablation and/or a decrease in the rate of ice discharge into the sea, each of which currently account for about half the accumulation (Section 11.2.3). The discharge can be reduced by thinning of the ice sheet near the grounding line. Ablation can be reduced by a change in the area-elevation distribution. However, once ablation has increased enough to equal accumulation, the ice sheet cannot survive, since discharge cannot be less than zero. This situation occurs for an annual-average warming of 2.7°C for the present ice-sheet topography, and for a slightly larger warming for a retreating ice sheet (Huybrechts *et al.*, 1991; see also Oerlemans,

Figure 11.16: Response of the Greenland Ice Sheet to three climatic warming scenarios during the third millennium expressed in equivalent changes of global sea level. The curve labels refer to the mean annual temperature rise over Greenland by 3000 AD as predicted by a two-dimensional climate and ocean model forced by greenhouse gas concentration rises until 2130 AD and kept constant after that. (From Huybrechts and De Wolde, 1999.) Note that projected temperatures over Greenland are generally greater than globally averaged temperatures (by a factor of 1.2 to 3.1 for the range of AOGCMs used in this chapter). See Table 11.13 and Chapter 9, Fig 9.10c.

1991; Van de Wal and Oerlemans, 1994). Models show that under these circumstances the Greenland ice sheet eventually disappears, except for residual glaciers at high altitudes. By using the AOGCM ratios of the Greenland temperature to the global average (Table 11.13) with the results of the calibrated simple model (Section 11.5.1.2 and Chapter 9, Section 9.3.3) we project increases in Greenland temperatures by 2100 of more than 2.7°C for nearly all combinations of SRES scenarios and AOGCMs. The maximum by 2100 is 9°C.

Huybrechts and De Wolde (1999) (Figure 11.16) (see also Letreguilly *et al.*, 1991) find the Greenland ice sheet contributes about 3 m of sea level rise equivalent over a thousand years under their mid-range scenario, in which the Greenland temperature change passes through 4°C in 2100 before stabilising at 5.5°C after 2130. Taking into account the high-latitude amplification of warming, this temperature change is consistent with mid-range stabilisation scenarios (Chapter 9, Section 9.3.3.1 and Figure 9.17(b)). For a warming of 8°C, they calculate a contribution of about 6 m. Their experiments take into account the effect of concomitant increases in precipitation (which reduces sensitivity) but also of the precipitation fraction falling as rain (which strongly enhances sensitivity for the larger temperature increases). Disregarding the effects of accumulation changes and rainfall, Greve (2000) reports that loss of mass would occur at a rate giving a sea level rise of between 1 mm/yr for a year-round temperature perturbation of 3°C to as much as 7 mm/yr for a sustained warming of 12°C, the latter being an extreme scenario in which the ice sheet would be largely eliminated within 1,000 years.

West Antarctic ice sheet

The WAIS contains enough ice to raise sea level by 6 m. It has received particular attention because it has been the most dynamic part of the Antarctic ice sheet in the recent geological past, and because most of it is grounded below sea level – a situation that according to models proposed in the 1970s could lead to flow instabilities and rapid ice discharge into the ocean when the surrounding ice shelves are weakened (Thomas, 1973; Weertman, 1974; Thomas *et al.*, 1979). Geological evidence suggests that WAIS may have been smaller than today at least once during the last million years (Scherer *et al.*, 1998). The potential of WAIS to collapse in response to future climate change is still a subject of debate and controversy.

The discharge of the WAIS is dominated by fast-flowing ice streams, dynamically constrained at four boundaries: the transition zone where grounded ice joins the floating ice shelf (Van der Veen, 1985; Herterich, 1987), the interface of ice with bedrock that is lubricated by sediment and water (Blankenship *et al.*, 1986; Anandakrishnan *et al.*, 1998; Bell *et al.*, 1998), the shear zone where fast-moving ice meets relatively static ice at the transverse margins of ice streams (Echelmeyer *et al.*, 1994; Jacobson and Raymond, 1998), and the ice-stream onset regions where slowly flowing inland ice accelerates into the ice streams. Mechanisms have been proposed for dynamic changes at each of these boundaries.

Early studies emphasised the role of the ice shelf boundary in ice discharge by introducing the concept of a "back-stress"

believed to buttress the grounded ice sheet and prevent it from collapsing. Recent work, however, both modelling and measurement, places greater emphasis on the other ice stream boundaries. Force balance studies on Ice Stream B show no evidence of stresses generated by the Ross ice shelf, and mechanical control emanates almost entirely from the lateral margins (Whillans and Van der Veen, 1997). If confirmed for the other Siple Coast ice streams, this suggests ice stream flow fields to be little influenced by conditions in the ice shelf, similar to the situation elsewhere at the Antarctic margin (Mayer and Huybrechts, 1999).

Nonetheless, there is a considerable body of evidence for ice stream variability, and the above analyses may not apply to dynamic situations involving large thinning or grounding line change. Ice Stream C largely stopped about a century ago (Retzlaff and Bentley, 1993) and Ice Stream B decelerated by 20% within a decade (Stephenson and Bindschadler, 1988). The mechanisms for these oscillations are not well understood and have been ascribed to processes such as basal water diversion to a neighbouring ice stream (Anandakrishnan and Alley, 1997) or thermomechanical interactions between competing catchment areas (Payne, 1998). Despite ice stream variability in the latter model on the millenial time-scale, the overall volume of the WAIS hardly changed, supporting the suggestion that ice streams may act to remove the imbalance of individual drainage basins and to stabilise rather than destabilise WAIS (Hindmarsh, 1993).

The WAIS was much larger during the LGM and has probably lost up to two thirds of its volume since then (Bindschadler, 1998). The largest losses have been from grounding line retreat below the present WAIS ice shelves (Ross and Filchner-Ronne), most likely as a gradual response to rising sea levels subsequent to melting of the Northern Hemisphere ice sheets (Ingolfsson *et al.*, 1998). Local imbalances, both positive and negative, are presently occurring (Shabtaie and Bentley, 1987; Whillans and Bindschadler, 1988; Bindschadler and Vornberger, 1998; Hamilton *et al.*, 1998; Rignot, 1998a,b; Wingham, *et al.*, 1998), but there is no conclusive observational evidence (from monitoring of surface elevation, see Section 11.2.3.2) that WAIS overall is making a significant contribution to global average sea level change (Bentley, 1997, 1998a,b; Bindschadler, 1998; Oppenheimer, 1998; Wingham *et al.* 1998). Conway *et al.* (1999) suggest that grounding line retreat since the LGM may still be ongoing, giving an average rate of recession corresponding to a rate of sea level rise of 0.9 mm/yr (Bindschadler, 1998). If projected into the future, this would imply disappearance of WAIS in 4,000 to 7,000 years (Bindschadler, 1998). However, geological estimates of ocean volume increase over the last 6,000 years place an upper limit of about half this amount on global sea level rise (Section 11.3.1).

Recent spectacular break-ups of the Larsen ice shelves in the Antarctic Peninsula (Vaughan and Doake, 1996; Doake *et al.*, 1998) demonstrate the existence of an abrupt thermal limit on ice shelf viability associated with regional atmospheric warming (Skvarca *et al.*, 1998). However, the WAIS ice shelves are not immediately threatened by this mechanism, which would require a further warming of 10°C before the −5°C mean annual isotherm reached their ice fronts (Vaughan and Doake, 1996). Although

atmospheric warming would increase the rate of deformation of the ice, causing the ice shelf to thin, response time-scales are of the order of several hundred years (Rommelaere and MacAyeal, 1997; Huybrechts and de Wolde, 1999).

In view of these considerations, it is now widely agreed that major loss of grounded ice, and accelerated sea level rise, is very unlikely during the 21st century. An interdisciplinary panel of international experts applying the techniques of risk assessment to the future evolution of WAIS concluded that there is a 98% chance that WAIS will not collapse in the next 100 years, defined as a change that contributes at least 10 mm/yr to global sea level change (Vaughan and Spouge, 2001). The probability of a contribution to sea level (exceeding 0.5 m) by the year 2100 was 5%. These results are broadly consistent with an earlier assessment by Titus and Narayanan (1996) based on a US-only panel, who found a 5% chance of a 0.16 m contribution and 1% chance of a 0.3 m contribution to sea level rise from WAIS by 2100. We note that Vaughan and Spouge also report a probability of 5% for WAIS giving a sea level fall exceeding 0.4 m within the same time frame, while Titus and Narayanan give 0.18 m.

Nonetheless, on a longer time-scale, changes in ice dynamics could result in significantly increased outflow of ice into the ice shelves and a grounding line retreat. Large-scale models show both of these phenomena to be sensitive to basal melting below the ice shelves (Warner and Budd, 1998; Huybrechts and De Wolde, 1999). Model studies do not agree on the sensitivity of the basal melting to an oceanic warming: for instance, one shows a quadrupling of the basal melting rate below the Amery ice shelf in East Antarctica for an adjacent sea warming of 1°C (Williams *et al.*, 1998), while another claims that warmer sea temperatures would reduce melting rates below the Ronne-Filchner ice shelves through alteration to sea-ice formation and the thermohaline circulation (Nicholls, 1997). Changes in open ocean circulation may also play a role. Warner and Budd (1998) suggest that even for moderate climatic warmings of a few degrees, a large increase in bottom melting of 5 m/yr becomes the dominant factor in the longer-term response of the Antarctic ice sheet. In their model, this causes the demise of WAIS ice shelves in a few hundred years and would float a large part of the WAIS (and marine portions of East Antarctica) after 1,000 years. Predicted rates of sea level rise are between 1.5 and 3.0 mm/yr depending on whether accumulation rates increase together with the warming. Allowing for runoff in addition to increased accumulation, Huybrechts and De Wolde (1999) find a maximum Antarctic contribution to global sea level rise of 2.5 mm/yr for an extreme scenario involving a warming of 8°C and a bottom melting rate of 10 m/yr. These figures are upper limits based on results currently available from numerical models, which do not resolve ice streams explicitly and which may not adequately predict the effect of ice shelf thinning on grounding line retreat owing to physical uncertainties.

Based on a wide-ranging review, Oppenheimer (1998) argues that WAIS could disintegrate within five to seven centuries following a warming of only a few degrees. Such a collapse implies a rate of sea level rise of 10 mm/yr and an average speed-up of the total outflow by at least a factor of 10 (Bentley, 1997, 1998a,b). However, the majority opinion of a recent expert panel reported by Vaughan and Spouge (2001) is that such outflow

rates are not attainable. It is, therefore, also plausible that WAIS may not make a significant contribution to sea level rise over time-scales less than a millennium. Vaughan and Spouge (2001) attribute a 50% probability to the latter scenario, but retained an equally large probability that the sea level rise will be larger than 2 mm/yr after 1,000 years, emphasising the inadequacy of our current understanding of the dynamics of WAIS, especially for predictions on the longer time-scales.

Independent of bottom melting below the ice shelves and the possibility of an ice-dynamic instability, surface melting sets an upper temperature limit on the viability of the Antarctic ice sheet, because runoff would eventually become the dominant wastage mechanism (as would be the case for Greenland in a climate several degrees warmer than today). For warmings of more than 10°C, simple runoff models predict that an ablation zone would develop around the Antarctic coast, making the mass balance at sea level sufficiently negative that the grounded ice would no longer be able to feed an ice shelf. Also the WAIS ice shelves would disintegrate to near to their inland limits as summer temperatures rise above the thermal limit of ice shelf viability believed to be responsible for the recent collapse of ice shelves at the northern tip of the Antarctic Peninsula. Disintegration of WAIS would in that case result, because the WAIS cannot retreat to higher ground once its margins are subjected to surface melting and begin to recede (Huybrechts, 1994). Depending on the strength of the warming, such a disintegration would take at least a few millennia.

East Antarctic ice sheet

Thresholds for disintegration of the East Antarctic ice sheet by surface melting involve warmings above 20°C, a situation that has not occurred for at least the last 15 million years (Barker *et al.*, 1999), and which is far more than thought possible under any scenario of climatic change currently under consideration. In that case, the ice sheet would decay over a period of at least 10,000 years. However, the recent inference of complex flow patterns in the interior of the East Antarctic ice sheet demonstrates the existence of ice-streaming features penetrating far inland, which may be indicative of a more dynamic regime than believed so far (Bamber *et al.*, 2000; Huybrechts *et al.*, 2000).

11.6 Reducing the Uncertainties in Future Estimates of Sea Level Change

It is valuable to note that the reduction in the uncertainty of estimation of the long-term ice sheet imbalance reported in Sections 11.3.1 and 11.4 came from indirect constraints and the synthesis of information of different types. Such syntheses offer promise for further progress.

11.6.1 Observations of Current Rates of Global-averaged and Regional Sea Level Change

Sections 11.3.2.1 and 11.4 reveal significant uncertainty in the analysis of 20th century sea level change. Also, we have little knowledge of the regional pattern of sea level change. Observational determination of such a pattern would be a

powerful test of the coupled models required for projections of globally averaged and regional sea level rise. Requirements for reducing uncertainties include:

- A global tide gauge network (the 'GLOSS Core Network') (IOC, 1997) for measuring relative change.
- A programme of measurements of vertical land movements at gauge sites by means of GPS (Global Positioning System), DORIS Beacons and/or absolute gravity meters (Neilan *et al.*, 1998).
- Improved models of postglacial rebound.
- A reanalysis of the historical record, including allowing for the impact of variable atmospheric forcing.
- A subset of mostly island tide gauge stations devoted to ongoing calibration of altimetric sea level measurements (Mitchum, 1998).
- An ongoing high-quality (TOPEX/POSEIDON class) satellite radar altimeter mission (Koblinsky *et al.*, 1992) and careful control of biases within a mission and between missions.
- Space gravity missions to estimate the absolute sea surface topography (Balmino *et al.*, 1996, 1999) and its temporal changes, to separate thermal expansion from an increase in ocean mass from melting of glaciers and ice sheets (NASA, 1996, NRC, 1997) and changes in terrestrial storage.

For assessment of possible changes to the severity of storm surges, analyses of historical storm surge data in conjunction with meteorological analyses are needed for the world's coastlines, including especially vulnerable regions.

11.6.2   Ocean Processes

Requirements for improved projections of ocean thermal expansion include:

- Global estimates of ocean thermal expansion through analysis of the historical data archive of ocean observations and a programme of new observations, including profiling floats measuring temperature and salinity and limited sets of full-depth repeat oceanographic sections and time-series stations.
- Testing of the ability of AOGCMs to reproduce the observed three-dimensional and time-varying patterns of ocean thermal expansion.
- An active programme of ocean and atmosphere model improvement, with a particular focus on the representation of processes which transport heat into and within the interior of the ocean.

11.6.3   Glaciers and Ice Caps

Requirements for improved projections of glacier contributions include (see also Haeberli *et al.*, 1998):

- A strategy of worldwide glacier monitoring, including the application of remote sensing techniques (laser altimetry, aerial photography, high-resolution satellite visible and infrared imagery e.g. from ASTER and Landsat).
- A limited number of detailed and long-term mass balance measurements on glaciers in different climatic regions of the world, with an emphasis on winter and summer balances in order to provide a more direct link with meteorological observations.

- Development of energy balance and dynamical models for more detailed quantitative analysis of glacier geometry changes with respect to mass balance and climate change.
- Glacier inventory data to determine the distribution of glacier parameters such as area and area-altitude relations, so that mass balance, glacier dynamics and runoff/sea level rise models can be more realistically framed.

11.6.4   Greenland and Antarctic Ice Sheets

- Continued observations with satellite altimeters, including the upcoming satellite laser altimeter on ICESat and the radar interferometer on CRYOSAT. Measurements should be continued for at least 15 years (with intercalibration between missions) to establish the climate sensitivities of the mass balance and decadal-scale trends.
- Satellite radar altimetry and synthetic aperture radar interferometry (ERS-1, ERS-2 and Radarsat) for detailed topography, changes in ice sheet volume and surface velocity of the ice sheets (Mohr *et al.*, 1998; Joughin *et al.*, 1999), as well as short-term variability in their flow (Joughin *et al.*, 1996a,b) and grounding line position (Rignot, 1998a,b,c).
- Determination of the Earth's time-variant gravity field by the Gravity Recovery and Climate Experiment (GRACE) satellite flown concurrently with ICESat to provide an additional constraint on the contemporary mass imbalances (Bentley and Wahr, 1998). This could provide estimates of sea level change to an accuracy of ± 0.35 mm/yr.
- Geological observations of sea level change during recent millennia combined with improved postglacial rebound models and palaeoclimatic and palaeoglaciological studies to learn what changes have occurred in the past.
- Further analysis of Earth rotational parameters in combination with sea level measurements.
- Improved estimates of surface mass balance (including its spatial and temporal variability) from *in situ* observations, accumulation rates inferred from atmospheric moisture budgets and improved estimates of the rate of iceberg calving and the melt-water flux.
- Improved calculation of the surface mass balance within ice sheet models or by atmospheric models, with attention to modelling of changes in sea-ice concentration because of the consequent effect on moisture transports and accumulation.
- Improved understanding and modelling of the dynamics of ice sheets, ice streams and ice shelves (requiring combined studies using glaciological, oceanographic and satellite observations), including the physics of iceberg calving.

11.6.5   Surface and Ground Water Storage

Surface and ground water storage changes are thought to be having a significant impact on sea level, but their contribution is very uncertain (Table 11.10, Figure 11.9), and could be either positive or negative. They may become more important in the future, as a result of changes related not only to climate, but also to societal decisions that are beyond the scope of this scientific assessment. There are several general issues in climate-related aspects:

• A more thorough investigative search of historical records could provide addition information on ground water mining, and storage in reservoirs.
• Accurate satellite measurements of variations in the Earth's gravity (Herring, 1998) to detect changes in land water storage due to water-table variations and impoundments.
• A better understanding of seepage losses beneath reservoirs and in irrigation is required.
• A unified systems approach is needed to trace the path of water more accurately through the atmosphere, hydrologic, and biosphere sub-systems, and to account for various feedbacks (including the use of GCMs and improved hydrologic models).
• Satellite remote sensing offers useful technology for monitoring the global hydrologic budget. A cumulative volume estimate for the many small reservoirs might be possible using high-resolution radar data, targeted ground studies and a classification of land use classes from satellite data and also of changes in deforestation and other land-use transformations (Koster *et al.*, 1999).

11.6.6   Summary

Sea level change involves many components of the climate system and thus requires a broad range of research activities. A more detailed discussion of the requirements is given in the report of the recent IGBP/GAIM Workshop on sea level changes (Sahagian and Zerbini, 1999). We recognise that it is important to assign probabilities to projections, but this requires a more critical and quantitative assessment of model uncertainties than is possible at present.

Appendix 11.1: Methods for projections of global-average sea level rise

This Appendix describes the methods used in this report to make sea level rise projections for the SRES scenarios for the 21st century. The results are discussed in Section 11.5.1.2 and shown in Figure 11.12 and Appendix II.

Global-average sea-level rise $\Delta h(t)$ is a function of time t and is expressed relative to the level in 1990. It comprises several components, which are all zero at 1990:

$$\Delta h(t) = X(t) + g(t) + G(t) + A(t) + I(t) + p(t) + s(t)$$

The components are sea-level rise due to:

X	thermal expansion.
g	loss of mass of glaciers and ice caps.
G	loss of mass of the Greenland ice sheet due to projected and recent climate change.
A	loss of mass of the Antarctic ice sheet due to projected and recent climate change.
I	loss of mass of the Greenland and Antarctic ice sheets due to the ongoing adjustment to past climate change.
p	runoff from thawing of permafrost.
s	deposition of sediment on the ocean floor.

The components X, g, G and A are estimated for each of 35 SRES scenarios using the projections of an upwelling-diffusion energy-balance (UD/EB) model calibrated separately for each of seven AOGCMs (Appendix 9.1).

Thermal expansion X is obtained directly from the thermal expansion $X_m(t)$ projected by the UD/EB model:

$$X(t) = X_m(t) - X_m(1990)$$

No uncertainty is included in this term, because the uncertainty is sufficiently represented by the use of a range of AOGCMs. The term g from glaciers and ice caps is estimated using the global average temperature change $T_m(t)$ projected by the UD/EB model. First, we obtain the loss of mass g_u with respect to the glacier steady state without taking contraction of glacier area into account.

$$g_u(t) = g_{1990} + \int_{1990}^{t} (T_{1990} + \Delta T_b + T_m(t') - T_m(1990)) \frac{\partial B_g}{\partial T_g} \, dt'$$

where g_{1990} is the sea-level rise from glaciers and ice caps up to 1990 calculated from AOGCM results without contraction of glacier area, T_{1990} is the AOGCM global average temperature change at 1990 with respect to the climate of the late 19th century, $\Delta T_b = 0.15$ K the difference in the global average temperature between the late 19th century and the glacier steady state (see 11.5.1.1) and $\partial B_g/\partial T_g$ is the sensivity of global glacier mass balance for constant glacier area to global-average temperature change, expressed as sea level equivalent (from Table 11.11). Second, we estimate the loss of mass g_s with respect to the glacier steady state taking into account contraction of glacier area. This is done by using an empirical relationship between the loss of mass for changing and for constant area. The relationship was obtained by a quadratic fit to the AOGCM IS92a results of Section 11.5.1.1.

$$g_s(t) = 0.934 g_u(t) - 1.165 g_u^2(t)$$

for g_u and g_s in metres. Third, we calculate the change since 1990.

$$g(t) = g_s(t) - g_s(1990)$$

The uncertainty $\delta g(t)$ on this term is calculated assuming an uncertainty of ±40% (standard deviation) in the mass balance sensitivities, as discussed in Section 11.5.1.1.

$$\delta g(t) = 0.40 g(t)$$

The term G from the Greenland ice sheet is calculated according to

$$G(t) = \int_{1990}^{t} (T_{1990} + T_m(t') - T_m(1990)) \frac{dB_G}{dT_g} \, dt'$$

where dB_G/dT_g is the sensitivity of the Greenland mass balance to global-average temperature change, expressed as sea level equivalent (from Table 11.12). The uncertainty on this term comprises two components, as discussed in Section 11.5.1.1. The first uncertainty is a mass balance uncertainty

$$\delta G_1(t) = \int_{1990}^{t} (T_{1990} + T_m(t') - T_m(1990)) \frac{\Delta T_G}{\Delta T_g} \delta m_G \, dt'$$

where $\delta m_G = 0.05$ mm/yr/°C and $\Delta T_G/\Delta T_g$ is the ratio of Greenland average temperature change to global average temperature change (from Table 11.12). The first uncertainty is the combination in quadrature of 0.03 mm/yr/°C from ablation parametrization,

0.03 mm/yr/°C from high-resolution patterns, and 0.02 mm/yr/°C from precipitation changes, as discussed in section 11.5.1.1. The second uncertainty is an ice-dynamic uncertainty.

$$\delta G_2(t) = 0.1G(t)$$

The term A from the Antarctic ice sheet is calculated according to

$$A(t) = \int_{1990}^{t} (T_{1990} + T_m(t') - T_m(1990)) \frac{dB_A}{dT_g} \, dt'$$

where dB_A/dT_g is the sensitivity of the Antarctic mass balance to global-average temperature change, expressed as sea level equivalent (from Table 11.12). Ice-dynamical uncertainty for the Antarctic is not included and is discussed in Section 11.5.4.3. There is no uncertainty for ablation. Precipitation change uncertainty is calculated as discussed in Section 11.5.1.1 according to

$$\delta A(t) = \int_{1990}^{t} (T_{1990} + T_m(t') - T_m(1990)) \frac{\Delta T_A}{\Delta T_g} \, \delta m_A \, dt'$$

where $\delta m_A = 0.08$ mm/yr/°C and $\Delta T_A/\Delta T_g$ is the ratio of Antarctic average temperature change to global average temperature change (from Table 11.12).

The uncertainties on the above terms are combined in quadrature:

$$\delta h_v = \sqrt{(\delta g)^2 + (\delta G_1)^2 + (\delta G_1)^2 + (\delta A)^2}$$

The remaining terms are calculated assuming they contribute to sea-level rise at a constant rate, independent of AOGCM and scenario, thus:

$$I(t) = \int_{1990}^{t} \frac{dI}{dt'} \, dt' \qquad p(t) = \int_{1990}^{t} \frac{dp}{dt'} \, dt' \qquad s(t) = \int_{1990}^{t} \frac{ds}{dt'} \, dt'$$

The rates each have a range of uncertainty. For dI/dt, this is 0.0 to 0.5 mm/yr (Section 11.3.1, Table 11.9), for dp/dt 0 to 0.23 mm/yr (the upper bound is more precisely 25 mm divided by 110 years, section 11.2.5), for ds/dt 0 to 0.05 mm/yr (Section 11.2.6, Table 11.9). The central rates are 0.25, 0.11 and 0.025 mm/yr for the three terms. We denote I calculated at the minimum rate by I_{min} and at the maximum rate by I_{max}; similarly for p and s. The minimum projected sea-level rise $\Delta h_{min}(t)$ for a given AOGCM and SRES scenario is given by

$$\Delta h_{min}(t) = X(t) + g(t) + G(t) + A(t) - 2\delta h_v(t) + I_{min}(t) + P_{min}(t) + s_{min}(t)$$

and the maximum is

$$\Delta h_{max}(t) = X(t) + g(t) + G(t) + A(t) + 2\delta h_v(t) + I_{max}(t) + P_{max}(t) + s_{max}(t)$$

In these formulae, δh_v has been doubled to convert from an uncertainty to a range, following Box 11.1.

Table 11.17: *Parameters used in sea-level projections to simulate AOGCM results.*

AOGCM	T_{1990} (°C)	g_{1990} (m)	$\partial B_g/\partial T_g$ (mm/yr°C)	dB_G/dT_g (mm/yr/°C)	dB_A/dT_g (mm/yr/°C)	$\Delta T_G/\Delta T_g$	$\Delta T_A/\Delta T_g$
CSIRO Mk2	0.593	0.022	0.733	0.157	− 0.373	2.042	1.120
CSM 1.3	0.567	0.021	0.608	0.146	− 0.305	3.147	1.143
ECHAM4/OPYC3	0.780	0.027	0.637	0.029	− 0.478	1.153	1.484
GFDL_R15_a	0.635	0.015	0.576	0.121	− 0.177	1.879	0.799
HadCM2	0.603	0.027	0.613	0.096	− 0.214	1.441	1.239
HadCM3	0.562	0.021	0.622	0.085	− 0.354	1.443	1.288
DOE PCM	0.510	0.017	0.587	0.136	− 0.484	2.165	1.618

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