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Observations: Oceanic Climate Change and Sea Level

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This chapter should be cited as:

Bindoff, N.L., J. Willebrand, V. Artale, A. Cazenave, J. Gregory, S. Gulev, K. Hanawa, C. Le Quéré, S. Levitus, Y. Nojiri, C.K. Shum, L.D. Talley and A. Unnikrishnan, 2007: Observations: Oceanic Climate Change and Sea Level. In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

5.4.5 Biological Changes Relevant to Ocean Biogeochemistry

Changes in biological activity are an important part of the carbon cycle but are difficult to quantify at the global scale. Marine export production (the fraction of primary production that is not respired at the ocean surface and thus sinks to depth) is the biological process that has the largest influence on element cycles. There are no global observations on changes in export production or respiration. However, estimates of changes in primary production provide partial information. A reduction in global oceanic primary production by about 6% between the early 1980s and the late 1990s was estimated based on the comparison of chlorophyll data from two satellites (Gregg et al., 2003). The errors in this estimate are potentially large because it is based on the comparison of data from two different sensors. Nevertheless, a change in biological fluxes of this order of magnitude is plausible considering that biological production is controlled primarily by nutrient input from intermediate waters, and that a decrease in intermediate water renewal has been observed during that period as indicated by the decrease in O_2 . Shifts and trends in plankton biomass have been observed for instance in the North Atlantic (Beaugrand and Reid, 2003), the North Pacific (Karl, 1999; Chavez et al., 2003) and in the Southern Indian Ocean (Hirawake et al., 2005), but the spatial and temporal coverage is limited. The potential impacts of changes in marine ecosystems or dissolved organic matter on climate are discussed in Section 7.3.4, and the impact of climate on marine ecosystems in Chapter 4 of the Working Group II contribution to the IPCC Fourth Assessment Report.

5.4.6 Consistency with Physical Changes

It is clearly established that climate variability affects the oceanic content of natural and anthropogenic DIC and the air-sea flux of CO_2 , although the amplitude and physical processes responsible for the changes are less well known. Variability in the marine carbon cycle has been observed in response to physical changes associated with the dominant modes of climate variability such as El Niño events and the PDO (Feely et al., 1999; Takahashi et al., 2006), and the NAO (Bates et al., 2002; Johnson and Gruber, 2007). The regional patterns of anthropogenic CO_2 storage are consistent with those of CFCs and with changes in heat content. The observed trends in CO_2 , DIC, pH and carbonate species can be primarily explained by the response of the ocean to the increase in atmospheric CO_2 .

Large-scale changes in the O_2 content of the thermocline have been observed between the 1970s and the late 1990s. These changes are everywhere consistent with the local changes in ocean ventilation as identified either by changes in density gradients or by changes in apparent CFC ages. Nevertheless, an influence of changes in marine biology cannot be ruled out. The available data are insufficient to say if the changes in O_2 are caused by natural variability or are trends that are likely to persist in the future, but they do indicate that large-scale changes

in ocean physics influence natural biogeochemical cycles, and thus the cycles of O_2 and CO_2 are likely to undergo changes if ocean circulation changes persist in the future.

5.5 Changes in Sea Level

5.5.1 Introductory Remarks

Present-day sea level change is of considerable interest because of its potential impact on human populations living in coastal regions and on islands. This section focuses on global and regional sea level variations, over time spans ranging from the last decade to the past century; a brief discussion of sea level change in previous centuries is given in Section 5.5.2.4. Changes over previous millennia are discussed in Section 6.4.3.

Processes in several nonlinearly coupled components of the Earth system contribute to sea level change, and understanding these processes is therefore a highly interdisciplinary endeavour. On decadal and longer time scales, global mean sea level change results from two major processes, mostly related to recent climate change, that alter the volume of water in the global ocean: i) thermal expansion (Section 5.5.3), and ii) the exchange of water between oceans and other reservoirs (glaciers and ice caps, ice sheets, other land water reservoirs - including through anthropogenic change in land hydrology, and the atmosphere; Section 5.5.5). All these processes cause geographically non-uniform sea level change (Section 5.5.4) as well as changes in the global mean; some oceanographic factors (e.g., changes in ocean circulation or atmospheric pressure) also affect sea level at the regional scale, while contributing negligibly to changes in the global mean. Vertical land movements such as resulting from glacial isostatic adjustment (GIA), tectonics, subsidence and sedimentation influence local sea level measurements but do not alter ocean water volume; nonetheless, they affect global mean sea level through their alteration of the shape and hence the volume of the ocean basins containing the water.

Measurements of present-day sea level change rely on two different techniques: tide gauges and satellite altimetry (Section 5.5.2). Tide gauges provide sea level variations with respect to the land on which they lie. To extract the signal of sea level change due to ocean water volume and other oceanographic change, land motions need to be removed from the tide gauge measurement. Land motions related to GIA can be simulated in global geodynamic models. The estimation of other land motions is not generally possible unless there are adequate nearby geodetic or geological data, which is usually not the case. However, careful selection of tide gauge sites such that records reflecting major tectonic activity are rejected, and averaging over all selected gauges, results in a small uncertainty for global sea level estimates (Appendix 5.A.4). Sea level change based on satellite altimetry is measured with respect to the Earth's centre of mass, and thus is not distorted by land motions, except for a small component due to large-scale deformation of ocean basins from GIA.

Frequently Asked Question 5.1

Is Sea Level Rising?

Yes, there is strong evidence that global sea level gradually rose in the 20th century and is currently rising at an increased rate, after a period of little change between AD 0 and AD 1900. Sea level is projected to rise at an even greater rate in this century. The two major causes of global sea level rise are thermal expansion of the oceans (water expands as it warms) and the loss of land-based ice due to increased melting.

Global sea level rose by about 120 m during the several millennia that followed the end of the last ice age (approximately 21,000 years ago), and stabilised between 3,000 and 2,000 years ago. Sea level indicators suggest that global sea level did not change significantly from then until the late 19th century. The instrumental record of modern sea level change shows evidence for onset of sea level rise during the 19th century. Estimates for the 20th century show that global average sea level rose at a rate of about 1.7 mm yr⁻¹.

Satellite observations available since the early 1990s provide more accurate sea level data with nearly global coverage. This decade-long satellite altimetry data set shows that since 1993, sea level has been rising at a rate of around 3 mm yr⁻¹, significantly higher than the average during the previous half century. Coastal tide gauge measurements confirm this observation, and indicate that similar rates have occurred in some earlier decades.

In agreement with climate models, satellite data and hydrographic observations show that sea level is not rising uniformly around the world. In some regions, rates are up to several times the global mean rise, while in other regions sea level is falling. Substantial spatial variation in rates of sea level change is also inferred from hydrographic observations. Spatial variability of the rates of sea level rise is mostly due to non-uniform changes in temperature and salinity and related to changes in the ocean circulation.

Near-global ocean temperature data sets made available in recent years allow a direct calculation of thermal expansion. It is believed that on average, over the period from 1961 to 2003, thermal expansion contributed about one-quarter of the observed sea level rise, while melting of land ice accounted for less than half. Thus, the full magnitude of the observed sea level rise during that period was not satisfactorily explained by those data sets, as reported in the IPCC Third Assessment Report.

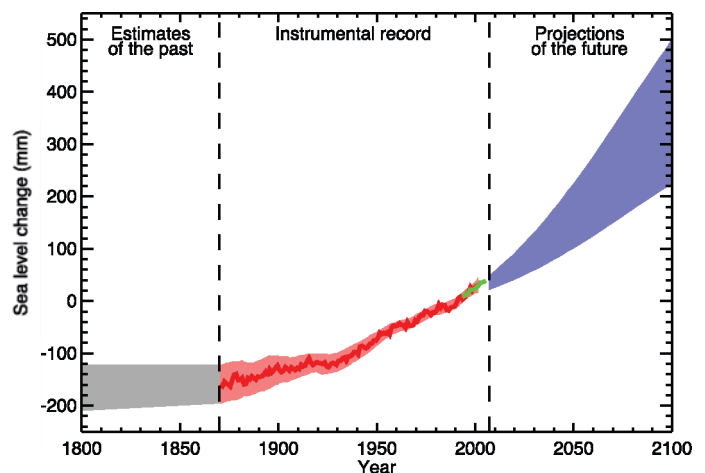
During recent years (1993–2003), for which the observing system is much better, thermal expansion and melting of land ice each account for about half of the observed sea level rise, although there is some uncertainty in the estimates.

The reasonable agreement in recent years between the observed rate of sea level rise and the sum of thermal expansion and loss of land ice suggests an upper limit for the magnitude of change in land-based water storage, which is relatively poorly known. Model results suggest no net trend in the storage of water over land due to climate-driven changes but there are large interannual and decadal fluctuations. However, for the recent period 1993 to 2003,

the small discrepancy between observed sea level rise and the sum of known contributions might be due to unquantified human-induced processes (e.g., groundwater extraction, impoundment in reservoirs, wetland drainage and deforestation).

Global sea level is projected to rise during the 21st century at a greater rate than during 1961 to 2003. Under the IPCC Special Report on Emission Scenarios (SRES) A1B scenario by the mid-2090s, for instance, global sea level reaches 0.22 to 0.44 m above 1990 levels, and is rising at about 4 mm yr⁻¹. As in the past, sea level change in the future will not be geographically uniform, with regional sea level change varying within about ±0.15 m of the mean in a typical model projection. Thermal expansion is projected to contribute more than half of the average rise, but land ice will lose mass increasingly rapidly as the century progresses. An important uncertainty relates to whether discharge of ice from the ice sheets will continue to increase as a consequence of accelerated ice flow, as has been observed in recent years. This would add to the amount of sea level rise, but quantitative projections of how much it would add cannot be made with confidence, owing to limited understanding of the relevant processes.

Figure 1 shows the evolution of global mean sea level in the past and as projected for the 21st century for the SRES A1B scenario.



FAQ 5.1, Figure 1. Time series of global mean sea level (deviation from the 1980–1999 mean) in the past and as projected for the future. For the period before 1870, global measurements of sea level are not available. The grey shading shows the uncertainty in the estimated long-term rate of sea level change (Section 6.4.3). The red line is a reconstruction of global mean sea level from tide gauges (Section 5.5.2.1), and the red shading denotes the range of variations from a smooth curve. The green line shows global mean sea level observed from satellite altimetry. The blue shading represents the range of model projections for the SRES A1B scenario for the 21st century, relative to the 1980 to 1999 mean, and has been calculated independently from the observations. Beyond 2100, the projections are increasingly dependent on the emissions scenario (see Chapter 10 for a discussion of sea level rise projections for other scenarios considered in this report). Over many centuries or millennia, sea level could rise by several metres (Section 10.7.4).

The TAR chapter on sea level change provided estimates of climate and other anthropogenic contributions to 20th-century sea level rise, based mostly on models (Church et al., 2001). The sum of these contributions ranged from -0.8 to 2.2 mm yr^{-1} , with a mean value of 0.7 mm yr^{-1} , and a large part of this uncertainty was due to the lack of information on anthropogenic land water change. For observed 20th-century sea level rise, based on tide gauge records, Church et al. (2001) adopted as a best estimate a value in the range of 1 to 2 mm yr^{-1} , which was more than twice as large as the TAR's estimate of climate-related contributions. It thus appeared that either the processes causing sea level rise had been underestimated or the rate of sea level rise observed with tide gauges was biased towards higher values.

Since the TAR, a number of new results have been published. The global coverage of satellite altimetry since the early 1990s (TOPOgraphy EXperiment (TOPEX)/Poseidon and Jason) has improved the estimate of global sea level rise and has revealed the complex geographical patterns of sea level change in open oceans. Near-global ocean temperature data for the last 50 years have been recently made available, allowing the first observationally based estimate of the thermal expansion contribution to sea level rise in past decades. For recent years, better estimates of the land ice contribution to sea level are available from various observations of glaciers, ice caps and ice sheets.

In this section, we summarise the current knowledge of present-day sea level rise. The observational results are assessed, followed by our current interpretation of these observations in terms of climate change and other processes, and ending with a discussion of the sea level budget (Section 5.5.6).

5.5.2 Observations of Sea Level Changes

5.5.2.1 20th-Century Sea Level Rise from Tide Gauges

Table 11.9 of the TAR listed several estimates for global and regional 20th-century sea level trends based on the Permanent Service for Mean Sea Level (PSMSL) data set (Woodworth and Player, 2003). The concerns about geographical bias in the PSMSL data set remain, with most long sea level records stemming from the NH, and most from continental coastlines rather than ocean interiors. Based on a small number (~ 25) of high-quality tide gauge records from stable land regions, the rate of sea level rise has been estimated as 1.8 mm yr^{-1} for the past 70 years (Douglas, 2001; Peltier, 2001), and Miller and Douglas (2004) find a range of 1.5 to 2.0 mm yr^{-1} for the 20th century from 9 stable tide gauge sites. Holgate and Woodworth (2004) estimated a rate of $1.7 \pm 0.4 \text{ mm yr}^{-1}$ sea level change averaged along the global coastline during the period 1948 to 2002, based on data from 177 stations divided into 13 regions. Church et al. (2004) (discussed further below) determined

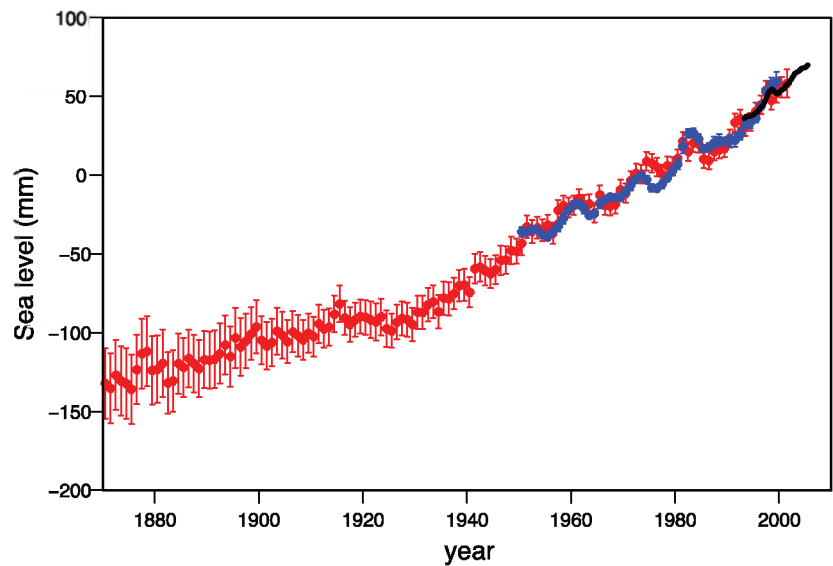


Figure 5.13. Annual averages of the global mean sea level (mm). The red curve shows reconstructed sea level fields since 1870 (updated from Church and White, 2006); the blue curve shows coastal tide gauge measurements since 1950 (from Holgate and Woodworth, 2004) and the black curve is based on satellite altimetry (Leuliette et al., 2004). The red and blue curves are deviations from their averages for 1961 to 1990, and the black curve is the deviation from the average of the red curve for the period 1993 to 2001. Error bars show 90% confidence intervals.

a global rise of $1.8 \pm 0.3 \text{ mm yr}^{-1}$ during 1950 to 2000, and Church and White (2006) determined a change of $1.7 \pm 0.3 \text{ mm yr}^{-1}$ for the 20th century. Changes in global sea level as derived from analyses of tide gauges are displayed in Figure 5.13. Considering the above results, and allowing for the ongoing higher trend in recent years shown by altimetry (see Section 5.5.2.2), we assess the rate for 1961 to 2003 as $1.8 \pm 0.5 \text{ mm yr}^{-1}$ and for the 20th century as $1.7 \pm 0.5 \text{ mm yr}^{-1}$.

While the recently published estimates of sea level rise over the last decades remain within the range of the TAR values (i.e., $1\text{--}2 \text{ mm yr}^{-1}$), there is an increasing opinion that the best estimate lies closer to 2 mm yr^{-1} than to 1 mm yr^{-1} . The lower bound reported in the TAR resulted from local and regional studies; local and regional rates may differ from the global mean, as discussed below (see Section 5.5.2.5).

A critical issue concerns how the records are adjusted for vertical movements of the land upon which the tide gauges are located and of the oceans. Trends in tide gauge records are corrected for GIA using models, but not for other land motions. The GIA correction ranges from about 1 mm yr^{-1} (or more) near to former ice sheets to a few tenths of a millimetre per year in the far field (e.g., Peltier, 2001); the error in tide-gauge based global average sea level change resulting from GIA is assessed as 0.15 mm yr^{-1} . The TAR mentioned the developing geodetic technologies (especially the Global Positioning System; GPS) that hold the promise of measuring rates of vertical land movement at tide gauges, no matter if those movements are due to GIA or to other geological processes. Although there has been some model validation, especially for GIA models, systematic problems with such techniques, including short data spans, have yet to be fully resolved.

5.5.2.2 Sea Level Change during the Last Decade from Satellite Altimetry

Since 1992, global mean sea level can be computed at 10-day intervals by averaging the altimetric measurements from the TOPEX/Poseidon (T/P) and Jason satellites over the area of coverage (66°S to 66°N) (Nerem and Mitchum, 2001). Each 10-day estimate of global mean sea level has an accuracy of approximately 5 mm. Numerous papers on the altimetry results (see Cazenave and Nerem, 2004, for a review) show a current rate of sea level rise of $3.1 \pm 0.7 \text{ mm yr}^{-1}$ over 1993 to 2003 (Cazenave and Nerem, 2004; Leuliette et al., 2004; Figure 5.14). A significant fraction of the 3 mm yr^{-1} rate of change has been shown to arise from changes in the Southern Ocean (Cabanès et al., 2001).

The accuracy needed to compute mean sea level change pushes the altimeter measurement system to its performance limits, and thus care must be taken to ensure that the instrument is precisely calibrated (see Appendix 5.A.4.1). The tide gauge calibration method (Mitchum, 2000) provides diagnoses of problems in the altimeter instrument, the orbits, the measurement corrections and ultimately the final sea level data. Errors in determining the altimeter instrument drift using the tide gauge calibration, currently estimated to be about 0.4 mm yr^{-1} , are almost entirely driven by errors in knowledge of vertical land motion at the gauges (Mitchum, 2000).

Altimetry-based sea level measurements include variations in the global ocean basin volume due to GIA. Averaged over the oceanic regions sampled by the altimeter satellites, this effect yields a value close to -0.3 mm yr^{-1} in sea level (Peltier, 2001), with possible uncertainty of 0.15 mm yr^{-1} . This number is subtracted from altimetry-derived global mean sea level in order to obtain the contribution due to ocean (water) volume change.

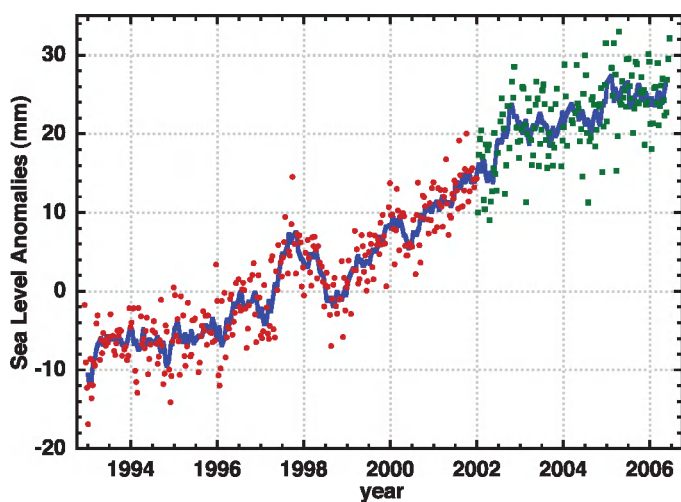


Figure 5.14. Variations in global mean sea level (difference to the mean 1993 to mid-2001) computed from satellite altimetry from January 1993 to October 2005, averaged over 65°S to 65°N . Dots are 10-day estimates (from the TOPEX/Poseidon satellite in red and from the Jason satellite in green). The blue solid curve corresponds to 60-day smoothing. Updated from Cazenave and Nerem (2004) and Leuliette et al. (2004).

Altimetry from T/P allows the mapping of the geographical distribution of sea level change (Figure 5.15a). Although regional variability in coastal sea level change had been reported from tide gauge analyses (e.g., Douglas, 1992; Lambeck, 2002), the global coverage of satellite altimetry provides unambiguous evidence of non-uniform sea level change in open oceans, with some regions exhibiting rates of sea level change about five times the global mean. For the past decade, sea level rise shows the highest magnitude in the western Pacific and eastern Indian oceans, regions that exhibit large interannual variability associated with ENSO. Except for the Gulf Stream region, most of the Atlantic Ocean shows sea level rise during the past decade. Despite the global mean rise, Figure 5.15a shows that sea level has been dropping in some regions (eastern Pacific and western Indian Oceans). These spatial patterns likely reflect decadal fluctuations rather than long-term trends. Empirical Orthogonal Functions (EOF) analyses of altimetry-based sea level maps over 1993 to 2003 show a strong influence of the 1997–1998 El Niño, with the geographical patterns of the dominant mode being very similar to those of the sea level trend map (e.g., Nerem et al., 1999).

5.5.2.3 Reconstructions of Sea Level Change during the Last 50 Years Based on Satellite Altimetry and Tide Gauges

Attempts have been made to reconstruct historical sea level fields by combining the near-global coverage from satellite altimeter data with the longer but spatially sparse tide gauge records (Chambers et al., 2002; Church et al., 2004). These sea level reconstructions use the short altimeter record to determine the principal EOF of sea level variability, and the tide gauge data to estimate the evolution of the amplitude of the EOFs over time. The method assumes that the geographical patterns of decadal sea level trends can be represented by a superposition of the patterns of variability that are manifest in interannual variability. The sea level for the period 1870 to 2000 (Church and White, 2006) shown in Figure 5.13 is based on this approach. As a caveat, note that variability on different time scales may have different characteristic patterns (see Section 5.5.4.1).

The trends in the EOF amplitudes (and the implied global correlations) allow the reconstruction of a spatially variable rate of sea level rise. Figure 5.16a (updated from Church et al., 2004) shows the geographical distribution of linear sea level trends for 1955 to 2003 based on this reconstruction technique. Comparison with the altimetry-based trend map for the shorter period (1993 to 2003) indicates quite different geographical patterns. These differences mainly arise from thermal expansion changes through time (see Section 5.5.3)

Changes in spatial sea level patterns through time may help reconcile apparently inconsistent estimates of regional variations in tide-gauge based sea level rise. For example, the minimum in rise along the northwest Australian coast is consistent with the results of Lambeck (2002) in having smaller rates of sea level rise and indeed sea level fall off north-western Australia over the last few decades. In addition, for the North Atlantic Ocean,

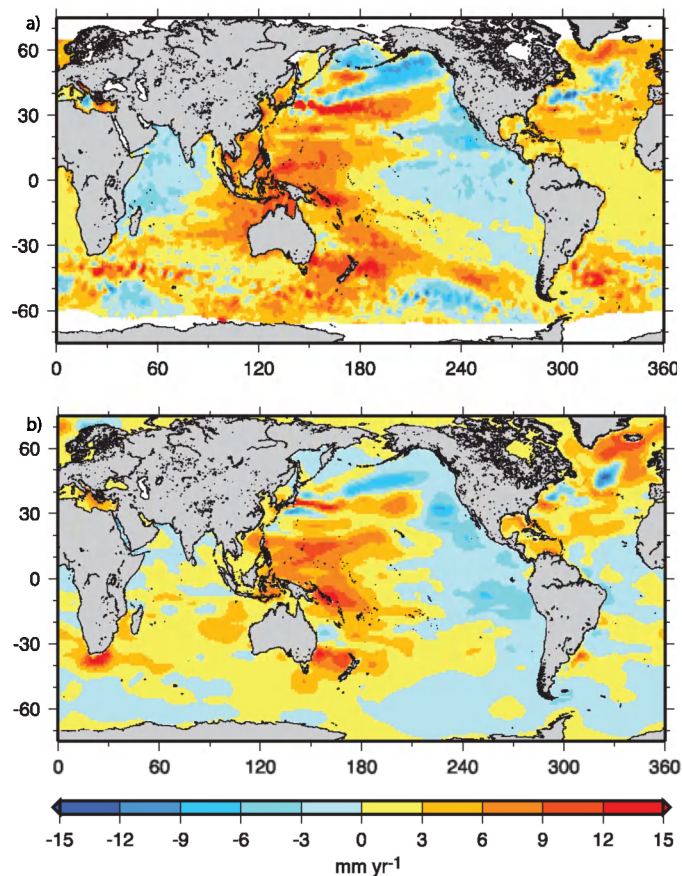


Figure 5.15. (a) Geographic distribution of short-term linear trends in mean sea level (mm yr^{-1}) for 1993 to 2003 based on TOPEX/Poseidon satellite altimetry (updated from Cazenave and Nerem, 2004) and (b) geographic distribution of linear trends in thermal expansion (mm yr^{-1}) for 1993 to 2003 (based on temperature data down to 700 m from Ishii et al., 2006).

the rate of rise reaches a maximum (over 2 mm yr^{-1}) in a band running east-northeast from the US east coast. The trends are lower in the eastern than in the western Atlantic (Lambeck et al., 1998; Woodworth et al., 1999; Mitrovica et al., 2001).

5.5.2.4 Interannual and Decadal Variability and Long-Term Changes in Sea Level

Sea level records contain a considerable amount of interannual and decadal variability, the existence of which is coherent throughout extended parts of the ocean. For example, the global sea level curve in Figure 5.13 shows an approximately 10 mm rise and fall of global mean sea level accompanying the 1997–1998 ENSO event. Over the past few decades, the time series of the first EOF of Church et al. (2004) represents ENSO variability, as shown by a significant (negative) correlation with the Southern Oscillation Index. The signature of the 1997–1998 El Niño is also clear in the altimetric maps of sea level anomalies (see Section 5.5.2.2). Model results suggest that large volcanic eruptions produce interannual to decadal fluctuations in the global mean sea level (see Section 9.5.2).

Holgate and Woodworth (2004) concluded that the 1990s had one of the fastest recorded rates of sea level rise averaged along the global coastline ($\sim 4 \text{ mm yr}^{-1}$), slightly higher than the altimetry-based open ocean sea level rise (3 mm yr^{-1}). However, their analysis also shows that some previous decades had comparably large rates of coastal sea level rise (e.g., around 1980; Figure 5.17). White et al. (2005) confirmed the larger sea level rise during the 1990s around coastlines compared to the open ocean but found that in some previous periods the coastal rate was smaller than the open ocean rate, and concluded that over the last 50 years the coastal and open ocean rates of change were the same on average. The global reconstruction of Church et al. (2004) and Church and White (2006) also exhibits large decadal variability in the rate of global mean sea level rise, and the 1993 to 2003 rate has been exceeded in some previous decades (Figure 5.17). The variability is smaller in the global reconstruction (standard deviation of overlapping 10-year rates is 1.1 mm yr^{-1}) than in the Holgate and Woodworth (2004) coastal time series (standard deviation 1.7 mm yr^{-1}). The rather

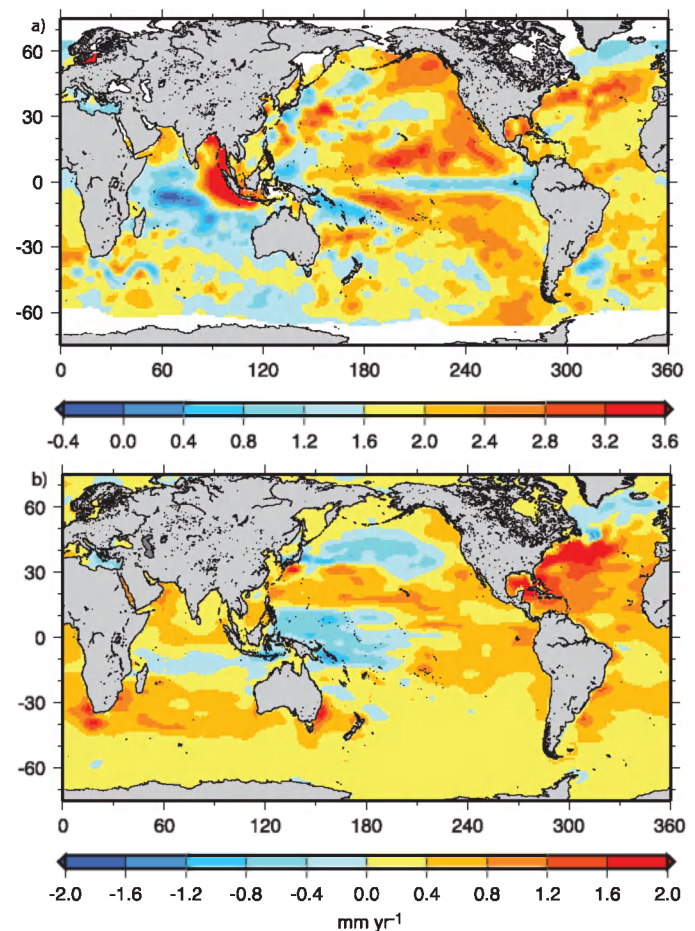


Figure 5.16. (a) Geographic distribution of long-term linear trends in mean sea level (mm yr^{-1}) for 1955 to 2003 based on the past sea level reconstruction with tide gauges and altimetry data (updated from Church et al., 2004) and (b) geographic distribution of linear trends in thermal expansion (mm yr^{-1}) for 1955 to 2003 (based on temperature data down to 700 m from Ishii et al., 2006). Note that colours in (a) denote 1.6 mm yr^{-1} higher values than those in (b).

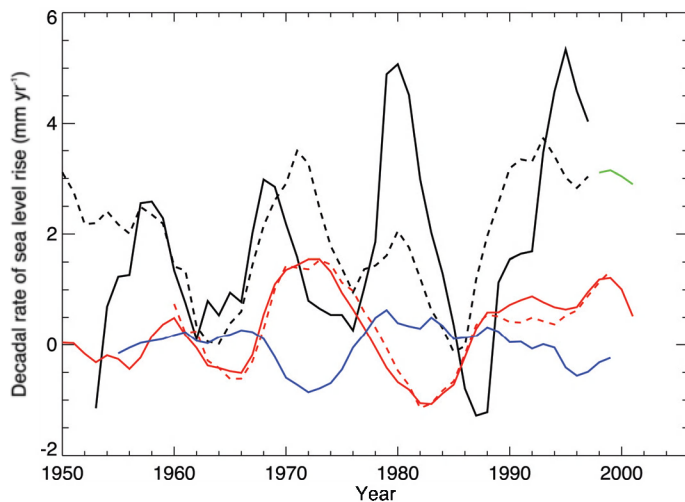


Figure 5.17. Overlapping 10-year rates of global sea level change from tide gauge data sets (Holgate and Woodworth, 2004, in solid black; Church and White, 2006, in dashed black) and satellite altimetry (updated from Cazenave and Nerem, 2004, in green), and contributions to global sea level change from thermal expansion (Ishii et al., 2006, in solid red; Antonov et al., 2005, in dashed red) and climate-driven land water storage (Ngo-Duc et al., 2005, in blue). Each rate is plotted against the middle of its 10-year period.

low temporal correlation ($r = 0.44$) between the two time series suggests that the statistical uncertainty in the linear trends calculated from either data set probably underestimates the systematic uncertainty in the results (Section 5.5.6).

Interannual or longer variability is a major reason why no long-term acceleration of sea level has been identified using 20th-century data alone (Woodworth, 1990; Douglas, 1992). Another possibility is that the sparse tide gauge network may have been inadequate to detect it if present (Gregory et al., 2001). The longest records available from Europe and North America contain accelerations of the order of 0.4 mm yr^{-1} per century between the 19th and 20th century (Ekman, 1988; Woodworth et al., 1999). For the reconstruction shown in Figure 5.13, Church and White (2006) found an acceleration of $1.3 \pm 0.5 \text{ mm yr}^{-1}$ per century over the period 1870 to 2000. These data support an inference that the onset of acceleration occurred during the 19th century (see Section 9.5.2).

Geological observations indicate that during the last 2,000 years (i.e., before the recent rise recorded by tide gauges), sea level change was small, with an average rate of only 0.0 to 0.2 mm yr^{-1} (see Section 6.4.3). The use of proxy sea level data from archaeological sources is well established in the Mediterranean. Oscillations in sea level from 2,000 to 100 yr before present did not exceed $\pm 0.25 \text{ m}$, based on the Roman-Byzantine-Crusader well data (Sivan et al., 2004). Many Roman and Greek constructions are related to the level of the sea. Based on sea level data derived from Roman fish ponds, which are considered to be a particularly reliable source of such information, together with nearby tide gauge records, Lambeck et al. (2004) concluded that the onset of the modern sea level rise occurred between 1850 and 1950. Donnelly et al. (2004) and Gehrels et al. (2004), employing geological data from Connecticut, Maine and Nova Scotia salt-marshes together with

nearby tide gauge records, demonstrated that the sea level rise observed during the 20th century was in excess of that averaged over the previous several centuries.

The joint interpretation of the geological observations, the longest instrumental records and the current rate of sea level rise for the 20th century gives a clear indication that the rate of sea level rise has increased between the mid-19th and the mid-20th centuries.

5.5.2.5 Regional Sea Level Change

Two regions are discussed here to give examples of local variability in sea level: the northeast Atlantic and small Pacific Islands.

Interannual variability in northeast Atlantic sea level records exhibits a clear relationship to the air pressure and wind changes associated with the NAO, with the magnitude and sign of the response depending primarily upon latitude (Andersson, 2002; Wakelin et al., 2003; Woolf et al., 2003). The signal of the NAO can also be observed to some extent in ocean temperature records, suggesting a possible, smaller NAO influence on regional mean sea level via steric (density) changes (Tsimplis et al., 2006). In the Russian Arctic Ocean, sea level time series for recent decades also have pronounced decadal variability that correlates with the NAO index. In this region, wind stress and atmospheric pressure loading contribute nearly half of the observed sea level rise of 1.85 mm yr^{-1} (Proshutinsky et al., 2004).

Small Pacific Islands are the subject of much concern in view of their vulnerability to sea level rise. The Pacific Ocean region is the centre of the strongest interannual variability of the climate system, the coupled ocean-atmosphere ENSO mode. There are only a few Pacific Island sea level records extending back to before 1950. Mitchell et al. (2001) calculated rates of relative sea level rise for the stations in the Pacific region. Using their results (from their Table 1) and focusing on only the island stations with more than 50 years of data (only 4 locations), the average rate of sea level rise (relative to the Earth's crust) is 1.6 mm yr^{-1} . For island stations with record lengths greater than 25 years (22 locations), the average rate of relative sea level rise is 0.7 mm yr^{-1} . However, these data sets contain a large range of rates of relative sea level change, presumably as a result of poorly quantified vertical land motions.

An example of the large interannual variability in sea level is Kwajalein ($8^{\circ}44'N$, $167^{\circ}44'E$) (Marshall Archipelago). As shown in Figure 5.18, the local tide gauge data, the sea level reconstructions of Church et al. (2004) and Church and White (2006) and the shorter satellite altimeter record all agree and indicate that interannual variations associated with ENSO events are greater than 0.2 m . The Kwajalein data also suggest increased variability in sea level after the mid-1970s, consistent with the trend towards more frequent, persistent and intense ENSO events since the mid-1970s (Folland et al., 2001). For the Kwajalein record, the rate of sea level rise, after correction for GIA land motions and isostatic response to atmospheric pressure changes, is $1.9 \pm 0.7 \text{ mm yr}^{-1}$. However,

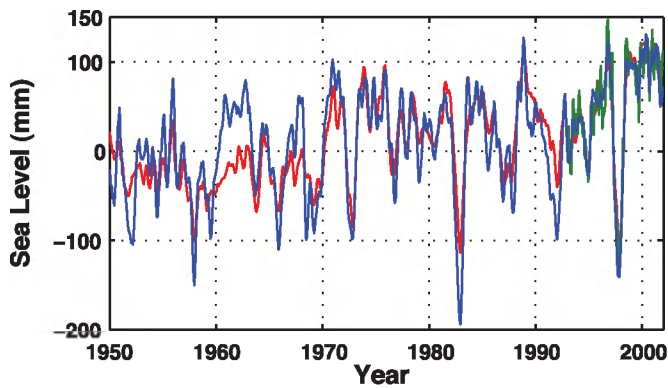


Figure 5.18. Monthly mean sea level curve for 1950 to 2000 at Kwajalein ($8^{\circ}44'N$, $167^{\circ}44'E$). The observed sea level (from tide gauge measurements) is in blue, the reconstructed sea level in red and the satellite altimetry record in green. Annual and semi-annual signals have been removed from each time series and the tide gauge data have been smoothed. The figure was drawn using techniques in Church et al. (2004) and Church and White (2006).

the uncertainties in rates of sea level change increase rapidly with decreasing record length and can be several mm yr^{-1} for decade-long records (depending on the magnitude of the interannual variability). Sea level change on the atolls of Tuvalu (western Pacific) has been the subject of intense interest as a result of their low-lying nature and increasing incidence of flooding. There are two records available at Funafuti, Tuvalu; the first record commences in 1977 and the second (with rigorous datum control) in 1993. After allowing for subsidence affecting the first record, Church et al. (2006) estimate sea level rise at Tuvalu to be $2.0 \pm 1.7 \text{ mm yr}^{-1}$, in agreement with the reconstructed rate of sea level rise.

5.5.2.6 Changes in Extreme Sea Level

Societal impacts of sea level change primarily occur via the extreme levels rather than as a direct consequence of mean sea level changes. Apart from non-climatic events such as tsunamis, extreme sea levels occur mainly in the form of storm surges generated by tropical or extra tropical cyclones. Secular changes and decadal variability in storminess are discussed in Chapter 3. Studies of variations in extreme sea levels during the 20th century based on tide gauge data are fewer than studies of changes in mean sea level for several reasons. A study on changes in extremes, which are caused by changes in mean sea level as well as changes in surges, is more complex than the study of mean sea level changes. Moreover, the hourly sampling interval normally used in tide gauge records is not always sufficient to accurately capture the true extreme. Among the different parameters often used to describe extremes, annual maximum surge is a good indicator of climatic trends. For study of long records extending back to the 19th century or before, annual maximum surge-at-high-water (defined as the maximum of the difference between observed high water and the predicted tide at high water) is a better-suited parameter because during that period high waters and not the full tidal curve were recorded.

Studies of the longest records of extremes are inevitably restricted to a small number of locations. From observed sea level extremes at Liverpool since 1768, Woodworth and Blackman (2002) concluded that the annual maximum surge-at-high-water was larger in the late 18th, late 19th and late 20th centuries than for most of the 20th century, qualitatively consistent with the long-term variability in storminess from meteorological data. From the tide gauge record at Brest from 1860 to 1994, Bouligand and Pirazzoli (1999) found an increasing trend in annual maxima and 99th percentile of surges; however, a decreasing trend was found during the period 1953 to 1994. From non-tidal residuals ('surges') at San Francisco since 1858, Bromirski et al. (2003) concluded that extreme winter residuals have exhibited a significant increasing trend since about 1950, a trend that is attributed to an increase in storminess during this period. Zhang et al. (2000) concluded from records at 10 stations along the east coast of the USA since 1900 that the rise in extreme sea level closely followed the rise in mean sea level. A similar conclusion can be drawn from a recent study of Firing and Merrifield (2004), who found long-term increases in the number and height of daily extremes at Honolulu (interestingly, the highest-ever value being due an anticyclonic oceanic eddy system in 2003), but no evidence for an increase relative to the underlying upward mean sea level trend.

An analysis of 99th percentiles of hourly sea level at 141 stations over the globe for recent decades (Woodworth and Blackman, 2004) showed that there is evidence for an increase in extreme high sea level worldwide since 1975. In many cases, the secular changes in extremes were found to be similar to those in mean sea level. Likewise, interannual variability in extremes was found to be correlated with regional mean sea level, as well as to indices of regional climate patterns.

5.5.3 Ocean Density Changes

Sea level will rise if the ocean warms and fall if it cools, since the density of the water column will change. If the thermal expansivity were constant, global sea level change would parallel the global ocean heat content discussed in Section 5.2. However, since warm water expands more than cold water (with the same input of heat), and water at higher pressure expands more than at lower pressure, the global sea level change depends on the three-dimensional distribution of ocean temperature change.

Analysis of the last half century of temperature observations indicates that the ocean has warmed in all basins (see Section 5.2). The average rate of thermosteric sea level rise caused by heating of the global ocean is estimated to be $0.40 \pm 0.09 \text{ mm yr}^{-1}$ over 1955 to 1995 (Antonov et al., 2005), based on five-year mean temperature data down to 3,000 m. For the 0 to 700 m layer and the 1955 to 2003 period, the averaged thermosteric trend, based on annual mean temperature data from Levitus et al. (2005a), is $0.33 \pm 0.07 \text{ mm yr}^{-1}$ (Antonov et al., 2005). For the same period and depth range, the mean thermosteric rate based on monthly ocean temperature data from Ishii et al. (2006) is $0.36 \pm 0.12 \text{ mm yr}^{-1}$. Figure 5.19

shows the thermosteric sea level curve over 1955 to 2003 for both the Levitus and Ishii data sets. The rate of thermosteric sea level rise is clearly not constant in time and shows considerable fluctuations (Figure 5.17). A rise of more than 20 mm occurred from the late 1960s to the late 1970s (giving peak 10-year rates in the early 1970s) with a smaller drop afterwards. Another large rise began in the 1990s, but after 2003, the steric sea level is decreasing in both estimates (peak rates in the late 1990s). Overlapping 10-year rates from these two estimates have a very high temporal correlation ($r = 0.97$) and the standard deviation of the rates is 0.7 mm yr^{-1} .

The Levitus and Ishii data sets both give $0.32 \pm 0.09 \text{ mm yr}^{-1}$ for the upper 700 m during 1961 to 2003, but the Levitus data set of temperature down to 3,000 m ends in 1998. From the results of Antonov et al. (2005) for thermal expansion, the difference between the trends in the upper 3,000 m and the upper 700 m for 1961 to 1998 is about 0.1 mm yr^{-1} . Assuming that the ocean below 700 m continues to contribute beyond 1998 at a similar rate, with an uncertainty similar to that of the upper-ocean contribution, we assess the thermal expansion of the ocean down to 3,000 m during 1961 to 2003 as $0.42 \pm 0.12 \text{ mm yr}^{-1}$.

For the recent period 1993 to 2003, a value of $1.2 \pm 0.5 \text{ mm yr}^{-1}$ for thermal expansion in the upper 700 m is estimated both by Antonov et al. (2005) and Ishii et al. (2006). Willis et al. (2004) estimate thermal expansion to be $1.6 \pm 0.5 \text{ mm yr}^{-1}$, based on combined *in situ* temperature profiles down to 750 m and satellite measurements of altimetric height. Including the satellite data reduces the error caused by the inadequate sampling of the profile data. Error bars were estimated to be about 2 mm for individual years in the time series, with most of the remaining error due to inadequate profile availability. A close result ($1.8 \pm 0.4 \text{ mm yr}^{-1}$ steric sea level rise for 1993 to 2003) was recently obtained by Lombard et al. (2006), based on a combined analysis of *in situ* hydrographic data and satellite sea surface height and

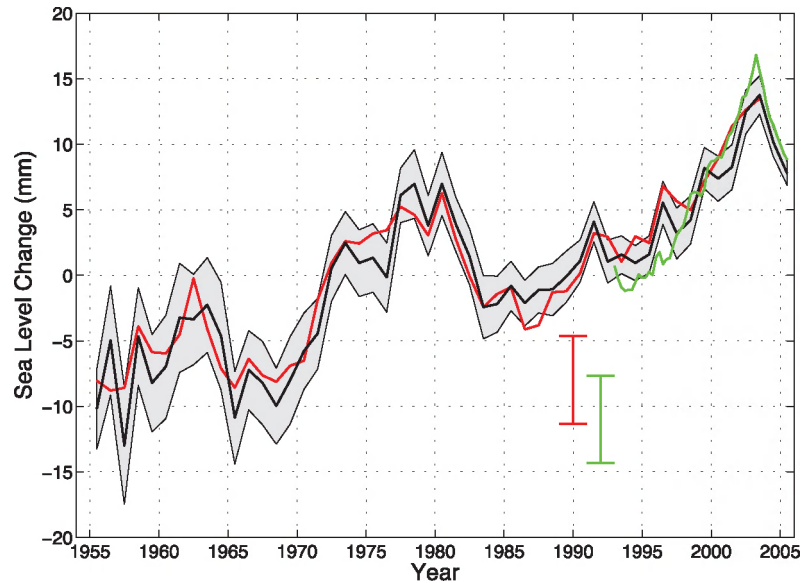


Figure 5.19. Global sea level change due to thermal expansion for 1955 to 2003, based on Levitus et al. (2005a; black line) and Ishii et al. (2006; red line) for the 0 to 700 m layer, and based on Willis et al. (2004; green line) for the upper 750 m. The shaded area and the vertical red and green error bars represent the 90% confidence interval. The black and red curves denote the deviation from their 1961 to 1990 average, the shorter green curve the deviation from the average of the black curve for the period 1993 to 2003.

SST data (Guinehut et al., 2004). It is presently unclear why the latter two estimates are significantly larger than the thermosteric rates based on temperature data alone. It is possible that the *in situ* data underestimate thermal expansion because of poor coverage in Southern Oceans, and it is interesting to note that a model based on assimilation of hydrographic data yields a somewhat higher estimate of 2.3 mm yr^{-1} (Carton et al., 2005). Published estimates of the steric sea level rates for 1955 to 2003 and 1993 to 2003 are shown in Table 5.2.

We assess the thermal expansion of the upper 700 m during 1993 to 2003 as $1.5 \pm 0.5 \text{ mm yr}^{-1}$, and that of the upper 3,000 m as $1.6 \pm 0.5 \text{ mm yr}^{-1}$, allowing for the ocean below 700 m as for the earlier period (see also Section 5.5.6, Table 5.3).

Table 5.2. Recent estimates for steric sea level trends from different studies.

Reference	Steric sea level change with errors (mm yr^{-1})	Period	Depth range (m)	Data Source
Antonov et al. (2005)	0.40 ± 0.09	1955–1998	0–3,000	Levitus et al. (2005b)
Antonov et al. (2005)	0.33 ± 0.07	1955–2003	0–700	Levitus et al. (2005b)
Ishii et al. (2006)	0.36 ± 0.06	1955–2003	0–700	Ishii et al. (2006)
Antonov et al. (2005)	1.2 ± 0.5	1993–2003	0–700	Levitus et al. (2005b)
Ishii et al. (2006)	1.2 ± 0.5	1993–2003	0–700	Ishii et al. (2006)
Willis et al. (2004)	1.6 ± 0.5	1993–2003	0–750	Willis et al. (2004)
Lombard et al. (2006)	1.8 ± 0.4	1993–2003	0–700	Guinehut et al. (2004)

Antonov et al. (2002) attributed about 10% of the global average steric sea level rise during recent decades to halosteric expansion (i.e., the volume increase caused by freshening of the water column). A similar result was obtained by Ishii et al. (2006) who estimated a halosteric contribution to 1955 to 2003 sea level rise of $0.04 \pm 0.02 \text{ mm yr}^{-1}$. While it is of interest to quantify this effect, only about 1% of the halosteric expansion contributes to the global sea level rise budget. This is because the halosteric expansion is nearly compensated by a decrease in volume of the added freshwater when its salinity is raised (by mixing) to the mean ocean value; the compensation would be exact for a linear state equation (Gille, 2004; Lowe and Gregory, 2006). Hence, for global sums of sea level change, halosteric expansion cannot be counted separately from the volume of added land freshwater (which Antonov et al., 2002, also calculate; see Section 5.5.5.1). However, for regional changes in sea level, thermosteric and halosteric contributions can be comparably important (see, e.g., Section 5.5.4.1).

5.5.4 Interpretation of Regional Variations in the Rate of Sea Level Change

Sea level observations show that whatever the time span considered, rates of sea level change display considerable regional variability (see Sections 5.5.2.2 and 5.5.2.3). A number of processes can cause regional sea level variations.

5.5.4.1 Steric Sea Level Changes

Like the sea level trends observed by satellite altimetry (see Section 5.5.2.3), the global distribution of thermosteric sea level trends is not spatially uniform. This is illustrated by Figure 5.15b and Figure 5.16b, which show the geographical distribution of thermosteric sea level trends over two different periods, 1993 to 2003 and 1955 to 2003 respectively (updated from Lombard et al., 2005). Some regions experienced sea level rise while others experienced a fall, often with rates that are several times the global mean. However, the patterns of thermosteric sea level rise over the approximately 50-year period are different from those seen in the 1990s. This occurs because the spatial patterns, like the global average, are also subject to decadal variability. In other words, variability on different time scales may have different characteristic patterns.

An EOF analysis of gridded thermosteric sea level time series since 1955 (updated from Lombard et al., 2005) displays a spatial pattern that is similar to the spatial distribution of thermosteric sea level trends over the same time span (compare Figure 5.20 with Figure 5.16b). In addition, the first principal component is negatively correlated with the Southern Oscillation Index. Thus, it appears that ENSO-related ocean variability accounts for the

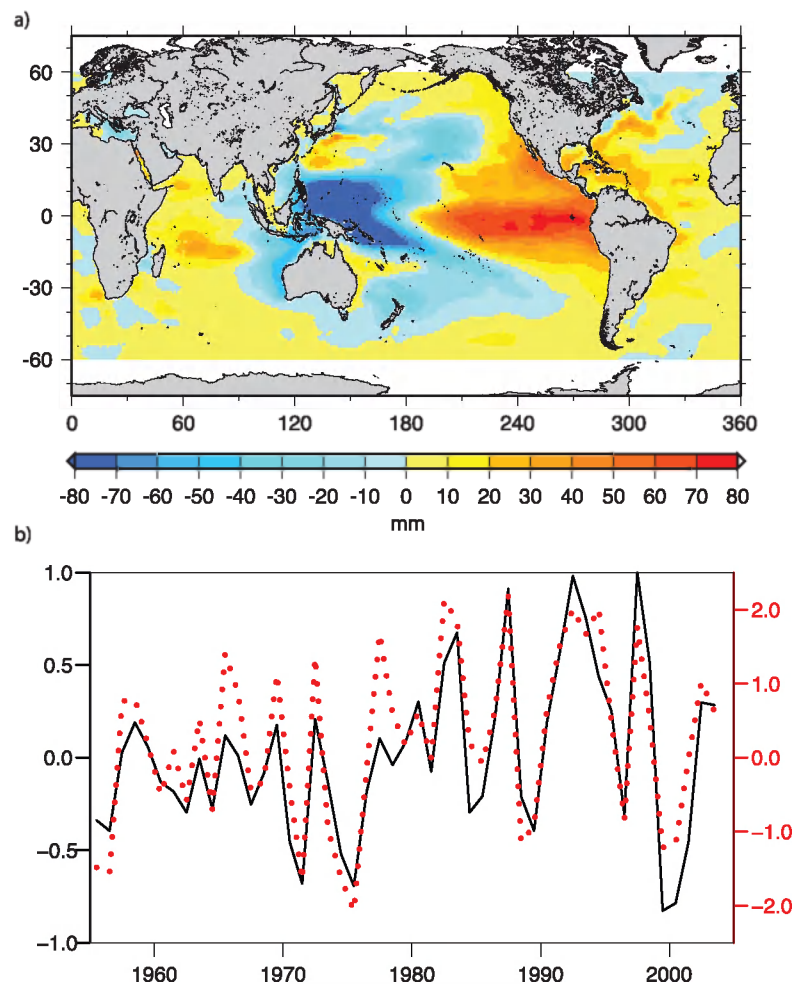


Figure 5.20. (a) First mode of the EOF decomposition of the gridded thermosteric sea level time series of yearly temperature data down to 700 m from Ishii et al. (2006). (b) The normalised principal component (black solid curve) is highly correlated with the negative Southern Oscillation Index (dotted red curve).

largest fraction of variance in spatial patterns of thermosteric sea level. Similarly, decadal thermosteric sea level in the North Pacific and North Atlantic appears strongly influenced by the PDO and NAO respectively.

For the recent years (1993–2003), the geographic distribution of observed sea level trends (Figure 5.15a) shows correlation with the spatial patterns of thermosteric sea level change (Figure 5.15b). This suggests that at least part of the non-uniform pattern of sea level rise observed in the altimeter data over the past decade can be attributed to changes in the ocean's thermal structure, which is itself driven by surface heating effects and ocean circulation. Note that the steric changes due to salinity changes have not been included in these figures due to insufficient salinity data in parts of the World Ocean.

Ocean salinity changes, while unimportant for sea level at the global scale, can have an effect on regional sea level (e.g., Antonov et al., 2002; Ishii et al., 2006; Section 5.5.3). For example, in the subpolar gyre of the North Atlantic, especially in the Labrador Sea, the halosteric contribution nearly

counteracts the thermosteric contribution. This observational result is supported by results from data assimilation into models (e.g., Stammer et al., 2003). Since density changes can result not only from surface buoyancy fluxes but also from the wind, a simple attribution of density changes to buoyancy forcing is not possible.

While much of the non-uniform pattern of sea level change can be attributed to thermosteric volume changes, the difference between observed and thermosteric spatial trends show a high residual signal in a number of regions, especially in the southern oceans. Part of these residuals is likely due to the lack of ocean temperature coverage in remote oceans as well as in deep layers (below 700 m), and to regional salinity change.

5.5.4.2 Ocean Circulation Changes

The highly non-uniform geographical distribution of steric sea level trends is closely connected, through geostrophic balance, with changes in ocean surface circulation. Density and circulation changes result from changes in atmospheric forcing that is primarily by surface wind stress and buoyancy flux (i.e., heat and freshwater fluxes). The wind alone can therefore cause local (but not global) changes in steric sea level. Ocean general circulation models based on the assimilation of ocean data satisfactorily reproduce the spatial structure of sea level trends for the past decade, and show in particular that the tropical Pacific pattern results from decadal fluctuations in the depth of the tropical thermocline and change in equatorial trade winds (Carton et al., 2005; Köhl et al., 2006). The similarity of the patterns of steric and actual sea level change indicates that density changes are the dominant influence. Discrepancies may indicate a significant contribution from changes in the wind-driven barotropic circulation, especially at high latitudes.

5.5.4.3 Surface Atmospheric Pressure Changes

Surface atmospheric pressure also causes regional sea level variations. Over time scales longer than a few days, the ocean adjusts nearly isostatically to changes in atmospheric pressure (inverted barometer effect), that is, for each 1 hPa sea level pressure increase the ocean is depressed by approximately 10 mm, shifting the underlying mass sideways to other regions. For the temporal average, regional changes in sea level caused by atmospheric pressure loading reach about 0.2 m (e.g., between the subtropical Atlantic and the subpolar Atlantic). Such effects are generally corrected for in tide gauge and altimetry-based sea level analyses. The inverted barometer effect has a negligible effect on global mean sea level, because water is nearly incompressible, but is significant when averaged over the area of T/P and Jason-1 altimetry, which does not cover the whole World Ocean (Ponte, 2006). For that reason, the altimetry-based mean sea level curve is corrected for the inverted barometer effect.

5.5.4.4 Solid Earth and Geoid Changes

Geodynamical processes related to the solid Earth's elastic and viscoelastic response to spatially variable ice melt loading (due to the last deglaciation and present-day land ice melt) also cause non-uniform sea level change (e.g., Mitrovica et al., 2001; Peltier, 2001, 2004; Plag, 2006). The solid Earth and oceans continue to respond to the ice and complementary water loads associated with the late Pleistocene and early Holocene glacial cycles through GIA. This process not only drives large crustal uplift near the location of former ice complexes, but also produces a worldwide signature in sea level that results from gravitational, deformational and rotational effects: as the viscous mantle material flows to restore isostasy during and after the last deglaciation, uplift occurs under the former centres of the ice sheets while the surrounding peripheral bulges experience a subsidence. The return of the melt water to the oceans produces an ongoing geoid change resulting in subsidence of the ocean basins and an upward warping of the continents, while the flow of water into the subsiding peripheral bulges contributes a broad scale sea level fall in the far field of the ice complexes. The combined gravitational and deformational effects also perturb the rotation vector of the planet, and this perturbation feeds back into variations in the position of the crust and the geoid (an equipotential surface of the Earth's gravity field that coincides with the mean surface of the oceans). Corrections for GIA effects are made to both tide gauge and altimeter estimates of global sea level change (see Sections 5.5.2.1 and 5.5.2.2).

Self-gravitation and deformation of the Earth's surface in response to the ongoing change in loading by glaciers and ice sheets is another cause of regional sea level variations. Model predictions show quite different patterns of non-uniform sea level change depending on the source of the ice melt (Mitrovica et al., 2001; Plag, 2006), and associated regional sea level variations reach up to a few 0.1 mm yr⁻¹.

5.5.5 Ocean Mass Change

Global mean sea level will rise if water is added to the ocean from other reservoirs in the climate system. Water storage in the atmosphere is equivalent to only about 35 mm of global mean sea level, and the observed atmospheric storage trend of about 0.04 mm yr⁻¹ in recent decades (Section 3.4.2.1) is unimportant compared with changes in ice and water stored on land, described in this subsection. Variations in land water storage result from variations in climatic conditions, direct human intervention in the water cycle and human modification of the land surface.

5.5.5.1 Ocean Mass Change Estimated from Salinity Change

Global salinity changes can be caused by changes in the global sea ice volume (which do not influence sea level) and by ocean mass changes (which do). Thus in principle, global salinity

changes can be used to estimate the global average sea level change due to fresh water input (Antonov et al., 2002; Munk, 2003; Wadhams and Munk, 2004). However, the accuracy of these estimates depends on the accuracy of the estimates for both sea ice volume (Hilmer and Lemke, 2000; Wadhams and Munk, 2004; see also Section 4.4) and global salinity change (Section 5.2.3). We assess that the error in estimates of ocean mass changes derived from salinity changes and sea ice melt is too large to provide useful constraints on the sea level change budget (Section 5.5.6).

5.5.5.2 Land Ice

During the 20th century, glaciers and ice caps have experienced considerable mass losses, with strong retreats in response to global warming after 1970. For 1961 to 2003, their contribution to sea level is assessed as $0.50 \pm 0.18 \text{ mm yr}^{-1}$ and for 1993 to 2003 as $0.77 \pm 0.22 \text{ mm yr}^{-1}$ (see Section 4.5.2).

As discussed in Section 4.6.2.2 and Table 4.6, the Greenland Ice Sheet has also been losing mass in recent years, contributing $0.05 \pm 0.12 \text{ mm yr}^{-1}$ to sea level rise during 1961 to 2003 and $0.21 \pm 0.07 \text{ mm yr}^{-1}$ during 1993 to 2003. Assessments of contributions to sea level rise from the Antarctic Ice Sheet are less certain, especially before the advent of satellite measurements, and are $0.14 \pm 0.41 \text{ mm yr}^{-1}$ for 1961 to 2003 and $0.21 \pm 0.35 \text{ mm yr}^{-1}$ for 1993 to 2003. Geodetic data on Earth rotation and polar wander allow a late-20th century sea level contribution of up to about 1 mm yr^{-1} from land ice (Mitrovica et al., 2006). However, recent estimates of ice sheet mass change exclude the large contribution inferred for Greenland by Mitrovica et al. (2001) from the geographical pattern of sea level change, confirming the lower rates reported above.

5.5.5.3 Climate-Driven Change in Land Water Storage

Continental water storage includes water (both liquid and solid) stored in subsurface saturated (groundwater) and unsaturated (soil water) zones, in the snowpack, and in surface water bodies (lakes, artificial reservoirs, rivers, floodplains and wetlands). Changes in concentrated stores, most notably very large lakes, are relatively well known from direct observation. In contrast, global estimates of changes in distributed surface stores (soil water, groundwater, snowpack and small areas of surface water) rely on computations with detailed hydrological models coupled to global ocean-atmosphere circulation models or forced by observations. Such models estimate the variation in land water storage by solving the water balance equation. The Land Dynamics (LaD) model developed by Milly and Shmakin (2002) provides global 1° by 1° monthly gridded time series of root zone soil water, groundwater and snowpack for the last two decades. With these data, the contributions of time-varying land water storage to sea level rise in response to climate change have been estimated, resulting in a small positive sea level trend of about 0.12 mm yr^{-1} for the last two decades, with larger interannual and decadal fluctuations (Milly et al., 2003).

From a land surface model forced by a global climatic data set based on standard reanalysis products and on observations, land water changes during the past five decades were found to have low-frequency (decadal) variability of about 2 mm in amplitude but no significant trend (Ngo-Duc et al., 2005). These decadal variations are related to groundwater and are caused by precipitation variations. They are strongly negatively correlated with the de-trended thermosteric sea level (Figure 5.17). This suggests that the land water contribution to sea level and thermal expansion partly compensate each other on decadal time scales. However, this conclusion depends on the accuracy of the precipitation in reanalysis products.

5.5.5.4 Anthropogenic Change in Land Water Storage

The amount of anthropogenic change in land water storage systems cannot be estimated with much confidence, as already discussed by Church et al. (2001). A number of factors can contribute to sea level rise. First, natural groundwater systems typically are in a condition of dynamic equilibrium where, over long time periods, recharge and discharge are in balance. When the rate of groundwater pumping greatly exceeds the rate of recharge, as is often the case in arid or even semi-arid regions, water is removed permanently from storage. The water that is lost from groundwater storage eventually reaches the ocean through the atmosphere or surface flow, resulting in sea level rise. Second, wetlands contain standing water, soil moisture and water in plants equivalent to water roughly 1 m deep. Hence, wetland destruction contributes to sea level rise. Over time scales shorter than a few years, diversion of surface waters for irrigation in the internally draining basins of arid regions results in increased evaporation. The water lost from the basin hydrologic system eventually reaches the ocean. Third, forests store water in living tissue both above and below ground. When a forest is removed, transpiration is eliminated so that runoff is favoured in the hydrologic budget.

On the other hand, impoundment of water behind dams removes water from the ocean and lowers sea level. Dams have led to a sea level drop over the past few decades of -0.5 to -0.7 mm yr^{-1} (Chao, 1994; Sahagian et al., 1994). Infiltration from dams and irrigation may raise the water table, storing more water. Gornitz (2001) estimated -0.33 to -0.27 mm yr^{-1} sea level change equivalent held by dams (not counting additional potential storage due to subsurface infiltration).

It is very difficult to provide accurate estimates of the net anthropogenic contribution, given the lack of worldwide information on each factor, although the effect caused by dams is possibly better known than other effects. According to Sahagian (2000), the sum of the above effects could be of the order of 0.05 mm yr^{-1} sea level rise over the past 50 years, with an uncertainty several times as large.

In summary, our assessment of the land hydrology contribution to sea level change has not led to a reduction in the uncertainty compared to the TAR, which estimated the rather wide ranges of -1.1 to $+0.4 \text{ mm yr}^{-1}$ for 1910 to 1990

and -1.9 to $+1.0$ mm yr^{-1} for 1990. However, indirect evidence from considering other contributions to the sea level budget (see Section 5.5.6) suggests that the land contribution either is small (<0.5 mm yr^{-1}) or is compensated for by unaccounted or underestimated contributions.

5.5.6 Total Budget of the Global Mean Sea Level Change

The various contributions to the budget of sea level change are summarised in Table 5.3 and Figure 5.21 for 1961 to 2003 and 1993 to 2003. Some terms known to be small have been omitted, including changes in atmospheric water vapour and climate-driven change in land water storage (Section 5.5.5), permafrost and sedimentation (see, e.g., Church et al., 2001), which very likely total less than 0.2 mm yr^{-1} . The poorly known anthropogenic contribution from terrestrial water storage (see Section 5.5.5.4) is also omitted.

For 1961 to 2003, thermal expansion accounts for only $23 \pm 9\%$ of the observed rate of sea level rise. Miller and Douglas (2004) reached a similar conclusion by computing steric sea level change over the past 50 years in three oceanic regions (northeast Pacific, northeast Atlantic and western Atlantic); they found it to be too small by about a factor of three to account for the observed sea level rise based on nine tide gauges in these regions. They concluded that sea level rise in the second half of the 20th century was mostly due to water mass added to the oceans. However, Table 5.3 shows that the sum of thermal expansion and contributions from land ice is smaller by 0.7 ± 0.7 mm yr^{-1} than the observed global average sea level rise. This is likely to be a significant difference. The assessment of Church et al. (2001) could allow this difference to be explained by positive anthropogenic terms (especially groundwater mining) but these are expected to have been

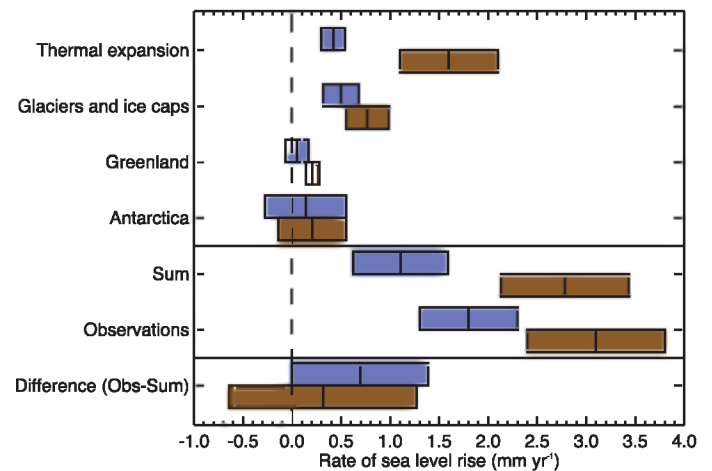


Figure 5.21. Estimates of the various contributions to the budget of the global mean sea level change (upper four entries), the sum of these contributions and the observed rate of rise (middle two), and the observed rate minus the sum of contributions (lower), all for 1961 to 2003 (blue) and 1993 to 2003 (brown). The bars represent the 90% error range. For the sum, the error has been calculated as the square root of the sum of squared errors of the contributions. Likewise the errors of the sum and the observed rate have been combined to obtain the error for the difference.

outweighed by negative terms (especially impoundment). We conclude that the budget has not yet been closed satisfactorily.

Given the large temporal variability in the rate of sea level rise evaluated from tide gauges (Section 5.5.2.4 and Figure 5.17), the budget is rather problematic on decadal time scales. The thermosteric contribution has smaller variability (though still substantial; Section 5.5.3) and there is only moderate temporal correlation between the thermosteric rate and the tide gauge rate. The difference between them has to be explained by ocean mass change. Because the thermosteric and climate-driven land water contributions are negatively correlated (Section 5.5.5.3.),

Table 5.3. Estimates of the various contributions to the budget of global mean sea level change for 1961 to 2003 and 1993 to 2003 compared with the observed rate of rise. Ice sheet mass loss of 100 Gt yr^{-1} is equivalent to 0.28 mm yr^{-1} of sea level rise. A GIA correction has been applied to observations from tide gauges and altimetry. For the sum, the error has been calculated as the square root of the sum of squared errors of the contributions. The thermosteric sea level changes are for the 0 to 3,000 m layer of the ocean.

Source	Sea Level Rise (mm yr^{-1})		Reference
	1961–2003	1993–2003	
Thermal Expansion	0.42 ± 0.12	1.6 ± 0.5	Section 5.5.3
Glaciers and Ice Caps	0.50 ± 0.18	0.77 ± 0.22	Section 4.5
Greenland Ice Sheet	0.05 ± 0.12	0.21 ± 0.07	Section 4.6.2
Antarctic Ice Sheet	0.14 ± 0.41	0.21 ± 0.35	Section 4.6.2
Sum	1.1 ± 0.5	2.8 ± 0.7	
Observed	1.8 ± 0.5		Section 5.5.2.1
		3.1 ± 0.7	Section 5.5.2.2
Difference (Observed – Sum)	0.7 ± 0.7	0.3 ± 1.0	

the apparent difference implies contributions during some 10-year periods from land ice, the only remaining term, exceeding 2 mm yr^{-1} (Figure 5.17). Since it is unlikely that the land ice contributions of 1993 to 2003 were exceeded in earlier decades (Figure 4.14 and Section 4.6.2.2), we conclude that the maximum 10-year rates of global sea level rise are likely overestimated from tide gauges, indicating that the estimated variability is excessive.

For 1993 to 2003, thermal expansion is much larger and land ice contributes $1.2 \pm 0.4 \text{ mm yr}^{-1}$. These increases may partly reflect decadal variability rather than an acceleration (Section 5.5.3; attribution of changes in rates and comparison with model results are discussed in Section 9.5.2). The sum is still less than the observed trend but the discrepancy of $0.3 \pm 1.0 \text{ mm yr}^{-1}$ is consistent with zero. It is interesting to note that the difference between the observed total and thermal expansion (assumed to be due to ocean mass change) is about the same in the two periods. The more satisfactory assessment for recent years, during which individual terms are better known and satellite altimetry is available, indicates progress since the TAR.

5.6 Synthesis

The patterns of observed changes in global heat content and salinity, sea level, steric sea level, water mass evolution and biogeochemical cycles described in the previous four sections are broadly consistent with known characteristics of the large-scale ocean circulation (e.g., ENSO, NAO and SAM).

There is compelling evidence that the heat content of the World Ocean has increased since 1955 (Section 5.2). In the North Atlantic, the warming is penetrating deeper than in the Pacific, Indian and Southern Oceans (Figure 5.3), consistent with the strong convection, subduction and deep overturning circulation cell that occurs in the North Atlantic Ocean. The overturning cell in the North Atlantic region (carrying heat and water downwards through the water column) also suggests that there should be a higher anthropogenic carbon content as observed (Figure 5.11). Subduction of SAMW (and to a lesser extent AAIW) also carries anthropogenic carbon into the ocean, which is observed to be higher in the formation areas of these subantarctic water masses (Figure 5.10). The transfer of heat into the ocean also leads to sea level rise through thermal expansion, and the geographical pattern of sea level change since 1955 is largely consistent with thermal expansion and with the change in heat content (Figure 5.2).

Although salinity measurements are relatively sparse compared with temperature measurements, the salinity data also show significant changes. In global analyses, the waters at high latitudes (poleward of 50°N and 70°S) are fresher in the upper 500 m (Figure 5.5 World). In the upper 500 m, the subtropical latitudes in both hemispheres are characterised by an increase in salinity. The regional analyses of salinity also

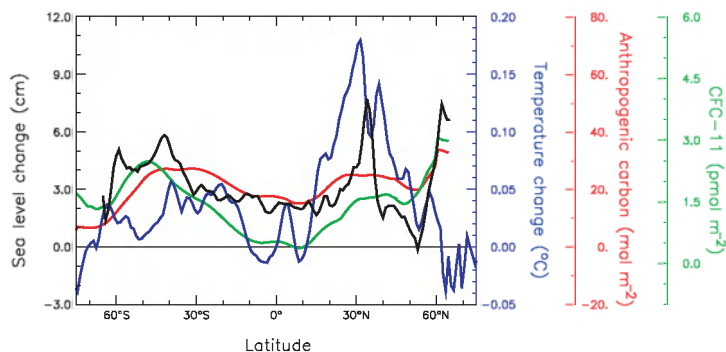


Figure 5.22. Averages of temperature change (blue, from Levitus et al., 2005a), anthropogenic carbon (red, from Sabine et al., 2004b) and CFC-11 (green, from Willey et al., 2004) along lines of constant latitude over the top 700-m layer of the upper ocean. Also shown is sea level change averaged along lines of constant latitude (black, from Cazenave and Nerem, 2004). The temperature changes are for the 1955 to 2003 period, the anthropogenic carbon is since pre-industrial times (i.e., 1750), CFC-11 concentrations are for the period 1930 to 1994 and sea level for the period 1993 to 2003.

show a similar distributional change with a freshening of key high-latitude water masses such as LSW, AAIW and NPIW, and increased salinity in some of the subtropical gyres such as that at 24°N . The North Atlantic (and other key ocean water masses) also shows significant decadal variations, such as the recent increase in surface salinity in the North Atlantic subpolar gyre. At high latitudes (particularly in the NH), there is an observed increase in melting of perennial sea ice, precipitation, and glacial melt water (see Chapter 4), all of which act to freshen high-latitude surface waters. At mid-latitudes it is likely that evaporation minus precipitation has increased (i.e., the transport of freshwater from the ocean to the atmosphere has increased). The pattern of salinity change suggests an intensification in the Earth's hydrological cycle over the last 50 years. These trends are consistent with changes in precipitation and inferred greater water transport in the atmosphere from low latitudes to high latitudes and from the Atlantic to the Pacific.

Figure 5.22 shows zonal means of changes in temperature, anthropogenic carbon, sea level rise and a passive tracer (CFC). It is remarkable that these independent variables (albeit with widely varying reference periods) show a common pattern of change in the ocean. Specifically, the close similarity of higher levels of warming, sea level rise, anthropogenic carbon and CFC-11 at mid-latitudes and near the equator strongly suggests that these changes are the result of changes in ocean ventilation and circulation. Warming of the upper ocean should lead to a decrease in ocean ventilation and subduction rates, for which there is some evidence from observed decreases in O_2 concentrations.

In the equatorial Pacific, the pattern of steric sea level rise also shows that strong west to east gradients in the Pacific have weakened (i.e., it is now cooler in the western Pacific and warmer in the eastern Pacific). This decrease in the equatorial temperature gradient is consistent with a tendency towards more prolonged and stronger El Niños over this same period (see Section 3.6.2).

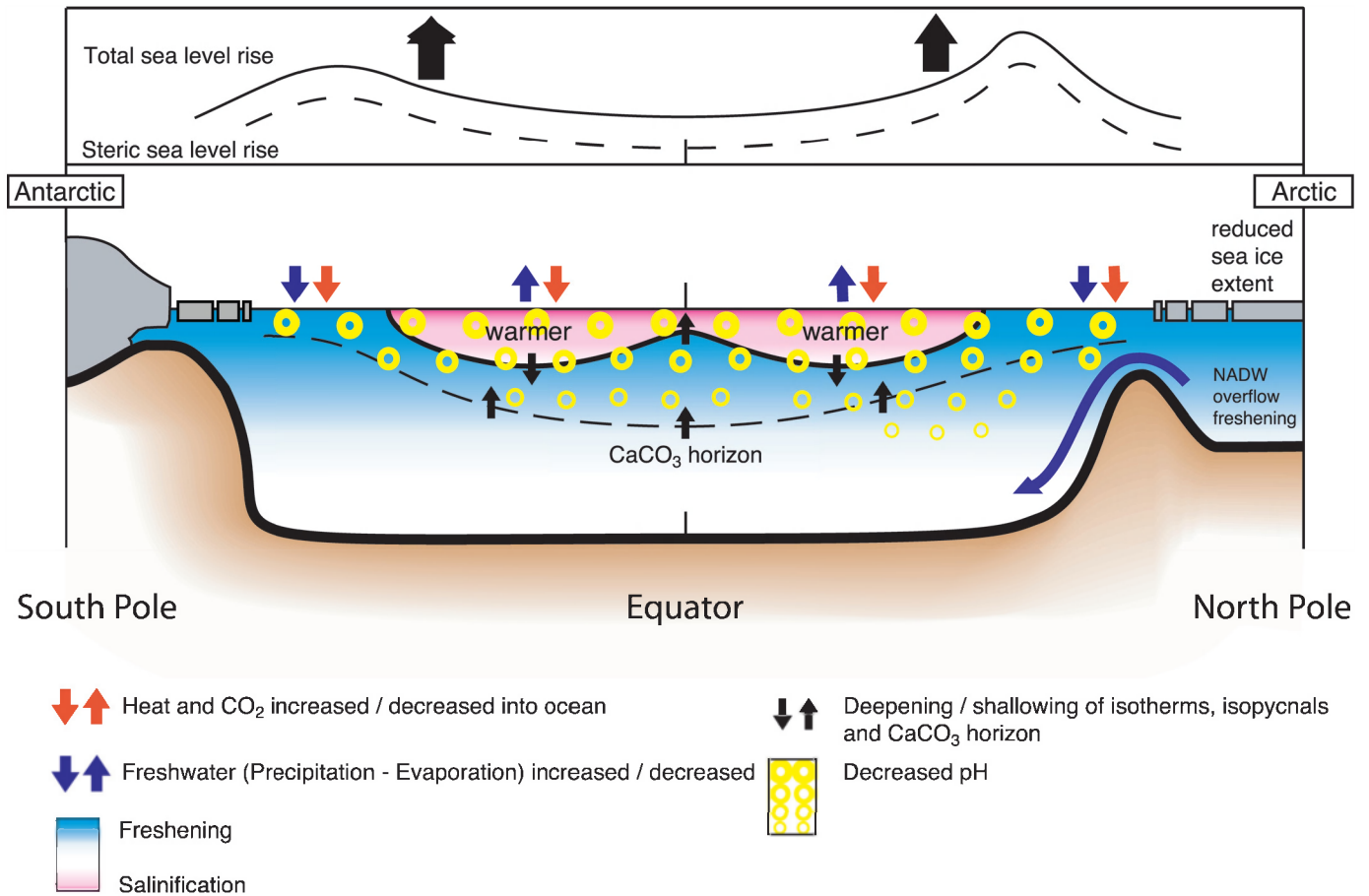


Figure 5.23. Schematic of the observed changes in the ocean state, including ocean temperature, ocean salinity, sea level, sea ice and biogeochemical cycles. The legend identifies the direction of the changes in these variables.

The subduction of carbon into the ocean has resulted in calcite and aragonite saturation horizons generally becoming shallower and pH decreasing primarily in the surface and near-surface ocean causing the ocean to become more acidic.

Since the TAR, the capability to measure most of the processes that contribute to sea level has been developed. In the 1990s, the observed sea level rise that was not explained through steric sea level rise could largely be explained by the transfer of mass from glaciers, ice sheets and river runoff (see Section 5.5). Figure 5.23 is a schematic that summarises the observed changes.

All of these observations taken together give high confidence that the ocean state has changed, that the spatial distribution of the changes is consistent with the large-scale ocean circulation and that these changes are in response to changed ocean surface conditions.

While there are many robust findings regarding the changed ocean state, key uncertainties still remain. Limitations in ocean sampling (particularly in the SH) mean that decadal variations in global heat content, regional salinity patterns, and rates of global sea level rise can only be evaluated with moderate confidence. Furthermore, there is low confidence in the evidence for trends

in the MOC and the global ocean freshwater budget. Finally, the global average sea level rise for the last 50 years is likely to be larger than can be explained by thermal expansion and loss of land ice due to increased melting, and thus for this period it is not possible to satisfactorily quantify the known processes causing sea level rise.

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Appendix 5.A: Techniques, Error Estimation and Measurement Systems

5.A.1 Ocean Temperature and Salinity

Sections 5.2 and 5.3 report on the changes in the oceans using two different approaches to the oceanic part of the climate system. Section 5.2 documents the changes found in the most comprehensive ocean data sets that exist for temperature and salinity. These data sets are collected from a wide range of organisations and are a composite of heterogeneous measurement systems, including mechanical and expendable bathythermographs, research ship measurements, voluntary observing ships, moored and drifting buoys and Argo floats for recent years. The advantage of these composite data sets is the greater spatial and temporal coverage that they offer for climate studies. The main disadvantage of these composite data sets, relative to the research data sets used in Section 5.3, is that they can have more problems related to the quality and heterogeneity of the measurements systems. This heterogeneity can lead to subtle biases and artificial noise and consequently difficulties in estimating trends at small regional scales (Harrison and Carson, 2006). On the other hand, Section 5.3 described the changes found in detailed analyses of very specific research voyages that consist mainly of very tightly calibrated and monitored temperature and salinity measurements (and other variables). The internal consistency of these research data sets is much higher than the composite data sets, and as a consequence they have significant advantages in their ease of interpretation and analysis. However, research quality oceanographic data sets are only collected occasionally and are focussed more frequently on regional rather than global issues. This means that in the poorly sampled oceans, such as the Indian, South Pacific and Southern Oceans, observational records only cover a relatively short period of time (e.g., the 1960s to present) with some decades poorly covered and highly heterogeneous in space (see Figure 5.A.1).

An example of the distribution of ocean temperature observations in both space and time is shown in Figure 5.A.1. This figure shows the *in situ* temperature data distribution for two five-year periods used to create estimates of global heat content change (e.g., Figure 5.1), one with a low (a) and one with a high (b) density of observations. It is clear that parts of the ocean, in particular in the SH, are not well sampled even in periods of high observation density. Hence, sampling errors resulting from the lack of data are potentially important but cannot easily be quantified.

Several different objective analysis techniques have been used to produce the gridded fields of temperature anomalies used to compute ocean heat content and steric sea level rise presented in this chapter. The technique used by Levitus et al. (2005b), Garcia et al. (2005) and Antonov et al. (2005) in their estimates of temperature (heat content), oxygen and the

thermoelectric component of sea level change is based on the construction of gridded (1° latitude by 1° longitude grid) fields at standard depth measurement levels. The objective analysis procedure used for interpolation (filling in data-void areas and smoothing the entire field) is described by Boyer et al. (2002). At each standard depth level, all data are averaged within each 1° square, and the deviation from climatology yields the observed anomaly. From all observations within the surrounding region of diameter 888 km, the analysed value is computed. Features with a wavelength of less than 500 to 600 km are substantially reduced in amplitude; in regions without sufficient data, it is essentially the climatological information that is used. Ishii et al. (2006) employed similar techniques, with a smaller decorrelation length scale of 300 km and a least-squares technique for estimating corrections to the climatological field. Willis et al. (2004) used a two-scale covariance function, but also used altimetric data in areas where ocean observations were lacking.

There are some differences in the data used in these studies. In addition to ocean temperature profile data, Ishii et al. (2006) also used the product of climatological mixed layer depth and individual SST measurements in their estimates of ocean heat content. Southern Hemisphere World Ocean Circulation Experiment profiling float temperature profiles for the 1990s were used by Willis et al. (2004) that were not used by Levitus et al. (2005a) and Ishii et al. (2006). The similarity of the three independently estimated heat content time series shown in Figure 5.1 to within confidence intervals indicates that the differences between analysis techniques and data sources do not substantially influence the estimates of the three global ocean heat content time series.

All analyses are subject to statistical errors and sampling errors. Statistical errors are estimated in a straightforward way. For example, for the Levitus et al. (2005a) fields, the uncertainty at any grid point is estimated from the variability of observations that contributed to the analysed value. In this way, 90% errors for all analysed variables are computed as a function of depth and horizontal position, and correspondingly for integrated variables such as heat content. Both Ishii et al. (2006) and Willis et al. (2004) used the interannual variability of heat content as the basis for error analyses.

5.A.2 Heat Transports

Estimates of meridional heat transport (MHT) derived from the surface heat balance involve the integration of the zonally averaged balances in the longitudinal direction. This integration also implies the integration of uncertainties in the zonally averaged estimates. For instance, an uncertainty in zonally averaged estimates of $\pm 10 \text{ W m}^{-2}$ results in an uncertainty of $0.5 \times 10^{15} \text{ W}$ in MHT in the Atlantic and nearly twice that value in the Pacific. Thus, all climatological estimates of MHT based on the surface heat balance have considerable uncertainties, and estimates of MHT variability are unlikely to be significant when derived from the surface heat balance.

In addition to the uncertainties in diagnostic computations of transports from vertical sections, estimates of MHT based on oceanic cross sections are largely influenced by sparse sampling of these sections during continuous time periods. As a result, there is no way to discriminate between the long-term signals and interannual variability using the estimates of MHT for individual years.

5.A.3 Estimates of Oxygen Changes

Estimates of changes in O_2 in the surface 100 m of the ocean between 1955 and 1998 were made for each pentad using a total of 530,000 O_2 profiles (Garcia et al., 2005). The measurement

method was not reported for all the cruises. Only the Winkler titration was reported, with only manual titrations prior to 1990. The Carpenter method to improve accuracy was reported for some cruises after 1970. An automated titration gives a significant improvement for measurement reproducibility but is not the essential solution for accuracy. Problems of O_2 leakage were reported from the older samples using Nansen bottles (generally before 1970). The Niskin bottles more widely used after 1970 are thought to be more reliable. There are no agreed standards for O_2 measurements because of reagent impurity and the difficulty in preparing a stable solution, which limits accuracy of these measurements to typically less than $10 \mu\text{mol kg}^{-1}$ for modern methods.

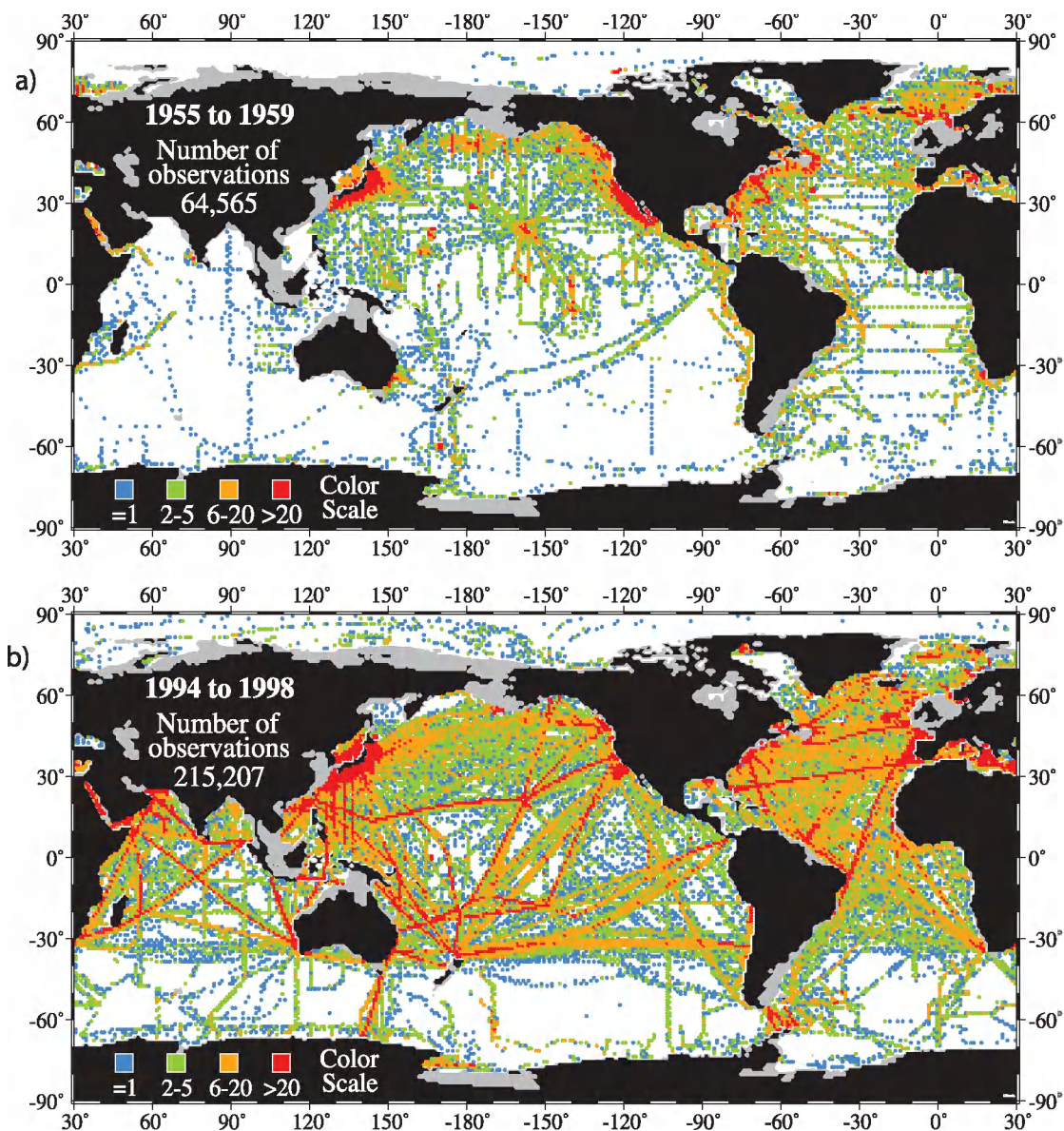


Figure 5.A.1 The number of ocean temperature observations in each 1° grid box at 250 m depth for two periods: (a) 1955 to 1959, with a low density of observations, and (b) 1994 to 1998, with a high density of observations. A blue dot indicates a 1° grid box containing 1 observation, a green dot 2 to 5 observations, an orange dot 6 to 20 observations, and a red dot more than 20 observations.

5.A.4 Estimation of Sea Level Change

5.A.4.1 *Satellite Altimetry: Measurement Principle and Associated Errors*

The concept of satellite altimetry measurement is rather straightforward. The onboard radar altimeter transmits a short pulse of microwave radiation with known power towards the nadir. Part of the incident radiation reflects back to the altimeter. Measurement of the round-trip travel time provides the height of the satellite above the instantaneous sea surface. The quantity of interest in oceanography is the height of the instantaneous sea surface above a fixed reference surface, which is computed as the difference between the altitude of the satellite above the reference ellipsoid and the altimeter range. The satellite position is computed through precise orbit determination, combining accurate modelling of the satellite motion and tracking measurements between the satellite and observing stations on Earth or other observing satellites. A number of corrections must be applied to obtain the correct sea surface height. These include instrumental corrections, ionospheric correction, dry and wet tropospheric corrections, electromagnetic bias correction, ocean and solid Earth tidal corrections, ocean loading correction, pole tide correction and an inverted barometer correction that has to be applied since the altimeter does not cover the global ocean completely. The total measurement accuracy for the TOPEX/Poseidon altimetry-based sea surface height is about 80 mm (95% error) for a single measurement based on one-second along-track averages (Chelton et al., 2001).

The above error estimates concern instantaneous sea surface height measurements. For estimating the mean sea level variations, the procedure consists of simply averaging over the ocean the point-to-point measurements collected by the satellite during a complete orbital cycle (10 days for TOPEX/Poseidon and Jason-1), accounting for the spatial distribution of the data using an equal area weighting. In effect, during this time interval, the satellite realises an almost complete coverage of the oceanic domain. The 95% error associated with a 10-day mean sea level estimate is approximately 8 mm.

When computing global mean sea level variations through time, proper account of instrumental bias and drifts (including the terrestrial reference frame) is of considerable importance. These effects (e.g., the radiometer drift onboard TOPEX/Poseidon used to correct for the wet tropospheric delay) are of the same order of magnitude as the sea level signal. Studies by Chambers et al. (1998) and Mitchum (1994; 2000) have demonstrated that comparing the altimeter sea level measurements to tide gauge sea level measurements produces the most robust way of correcting for instrumental bias and drifts. This approach uses a network of high-quality tide gauges, well distributed over the ocean domain. Current results indicate that the residual error in the mean sea level variation using the tide gauge calibration is about 0.8 mm yr^{-1} (a value resulting mainly from the uncertainties in vertical land

motion at the tide gauges). The current altimeter-inferred sea level measurements do not include modelling of the geocenter or mitigating the effect resulting from the potential drift of the terrestrial reference frame.

Detailed information about satellite altimetry, uncertainty and applications can be found in Fu and Cazenave (2001).

5.A.4.2 *Sea Level from Tide Gauge Observations*

Tide gauges are based on a number of different technologies (float, pressure, acoustic, radar), each of which has its advantages in particular applications. The Global Sea Level Observing System (GLOSS) specifies that a gauge must be capable of measuring sea level to centimetre accuracy (or better) in all weather conditions (i.e., in all wave conditions). The most important consideration is the need to maintain the gauge datum relative to the level of the Tide Gauge Bench Mark (TGBM), which provides the land reference level for the sea level measurements. The specifications for GLOSS require that local levelling must be repeated at least annually between the reference mark of the gauge, TGBM and a set of approximately five ancillary marks in the area, in order to maintain the geodetic integrity of the measurements. In practice, this objective is easier to meet if the area around the gauge is hard rock, rather than reclaimed land, for example. The question of whether the TGBM is moving vertically within a global reference frame (for whatever reason) is being addressed by advanced geodetic methods (GPS, Determination d'Orbite et Radiopositionnement Intégrés par Satellite (DORIS), Absolute Gravity). With typical rates of sea and land level change of order of 1 mm yr^{-1} , it is necessary to maintain the accuracy of the overall gauge system at the centimetre level over many decades. This demanding requirement has been met in many countries for many years (see IOC, 2002 for more information). The tide gauge observation system for three periods is shown in Figure 5.A.2, together with the evolution over time of the number of stations in both hemispheres. The distribution of tide gauge stations was particularly sparse in space at the beginning of the 20th century, but rapidly improved in the 1950s through to the current network of GLOSS standard instruments. This distribution of instruments through time means that confidence in the estimates of sea level rise has been improving and this certainty is reflected in the shrinking confidence intervals (Figure 5.13).

Figure 5.A.2. (a) Number of tide gauge stations in the Northern Hemisphere (NH) and Southern Hemisphere (SH) used to derive the global sea level curve (red and blue curves in Figure 5.13) as a function of time. Lower panels show the spatial distribution of tide gauge stations (denoted by red dots) for the periods (b) 1900 to 1909, (c) 1950 to 1959 and (d) 1980 to 1989.

