Chapter 5: Observations: Oceanic Climate Change and Sea Level

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Executive Summary

1. Greatly enhanced global databases of subsurface ocean temperature and salinity data for the last 50 years have recently become available, allowing continuous long term observation-based estimates of ocean warming and of the thermosteric contribution to sea level change.

2. Ocean temperatures are observed to have been rising since 1955. Global ocean heat content is estimated to have increased by 14.5 x 10^{22} J during the period 1955–1998. This amount is equivalent to an average warming of the entire ocean by 0.037°C, and an average heating rate of 0.28 W m^{-2} (per unit are of ocean surface). Current estimates of the warming rate from 1993 to 2003 in the 0–700 m ocean layer range from 0.70 ± 0.11 W m^{-2} to 0.86 ± 0.12 W m^{-2} (per unit area of ocean surface).

3. Global ocean heat content has strong interdecadal variations, superimposed on a near-linear trend. In the upper 700 m, the variability of the global heat content is associated with gyre and basin-scale variability, and is fairly robust to sampling variations.

4. The total carbon content of the oceans has increased by 118 ± 19 PgC since 1750 and continues to rise. The fraction of the CO_2 emitted that was taken up by the oceans decreased from 42 ± 7% during 1750–1994 to 36 ± 9% during 1993–2003. The surface CO_2 concentration increased nearly everywhere roughly following the atmospheric increase, with large regional and temporal variability. Surface ocean pH has decreased by 0.1 units since 1750, and continues to decrease. The depth at which $CaCO_3$ dissolves in the ocean has risen.

5. Surface oxygen concentration has strong interdecadal variations and no apparent trend between 1970 and 1995. Oxygen concentration of the ventilated thermocline (~100–1000 m) decreased in most ocean basins during the same period. The observed thermocline decrease is consistent with reduced rate of ventilation, although changes in biological activity may also play a role. Changes in surface chlorophyll and in deep ocean nutrients are indicative of changes in biological activity, but the available information is insufficient to identify any trends. The strength of the North Atlantic Deep Water overflows is unchanged but has freshened significantly.

6. Southern Ocean mode waters and Upper Circumpolar Deep Waters are warming, and a similar but weaker pattern of warming in the Gulf Stream and Kuroshio mode waters in the North Atlantic and North Pacific is observed. Overall, the warming is accompanied by a reduction in subtropical ventilation in the northern hemisphere. At least two marginal seas at subtropical latitudes (Mediterranean and Japan/East Sea) are warming.

7. Cooling is observed in the North Atlantic subpolar gyre (with increased convection), and in the central North Pacific, both creating colder water masses. Indirect evidence suggests that Atlantic meridional overturning circulation has considerable decadal variability but no confirmed trend.

8. Global sea level rise in the second half of the 20th century is estimated as 1.8 ± 0.3 mm yr^{-1}, which is consistent with the Third Assessment Report (TAR) estimate of 1.5 ± 0.5 mm yr^{-1} for the 20th century. Sea level rise measured by Topex/Poseidon satellite altimetry since 1993 has increased to 3.1 ± 0.4 mm yr^{-1}. It is however unclear whether the recent increase indicates an accelerating trend or whether it is associated with variability on decadal timescales.

9. The average steric contribution to sea-level rise for the last 50 years is 0.4 ± 0.1mm yr^{-1}, with significant decadal variations. For 1993 to 2003, estimate ranges from 1.3 to 1.8 mm yr^{-1}.

10. Sea level change is highly non-uniform spatially. In some regions, rates are up to 5 times the global mean rise, while in other regions sea level is falling. TP-derived absolute sea level trend patterns are well correlated with steric sea level trend over the same period.

11. The contribution of anthropogenic land water storage to sea level change (principally by impoundment in reservoirs and extraction of groundwater) is likely to be substantial but is the most

uncertain term. While the best estimate is now +0.05 mm yr^{-1} versus –0.35 in the TAR, the assessment of this uncertainty has not changed since the TAR. Decadal fluctuations in sea level (amplitude of 3–4 mm) due to land water storage appear to be negatively correlated with the change in ocean heat content.

12. Revised estimates of loss of mass from glaciers and ice caps show a higher contribution to sea level rise (0.76 ± 0.14 mm yr^{-1}) during 1993–2003, which is about twice as large as the mean over the last four decades, owing to recent warming. The contribution from the Greenland and Antarctica ice sheets during 1993–2003 is assessed as 0.0 ± 0.2 mm yr^{-1}.

13. Analysis of hourly tide gauge data since 1975 showed that there is evidence for an increase in occurrence of extreme high water worldwide and variations in extremes during this period are closely related to changes in regional climate.

14. As in the TAR, the sum of climate contributions to sea level rise during the last 50 years is significantly smaller than the observed value. For the last decade, the sum of thermal expansion and land ice melt is estimated as 2.6 ± 0.2 mm yr^{-1}. While the sea level budget is still not balanced, climate-related contributions are now relatively closer to the observations.

5.1 Introduction

The ocean has an important role in climate change. The ocean's heat capacity is about 1000 times larger than that of the atmosphere. The ocean's net heat uptake since 1955 is 21 times greater than that of the atmosphere (Levitus *et al*., 2005a). This large amount of heat, which has been mainly stored in the upper layers of the ocean, plays a dominant role for climate variations on seasonal to interannual time scales, such as El Niño. The transport of heat and freshwater by ocean currents can have an important effect for regional climates, and the large-scale meridional overturning circulation in the Atlantic and Southern Oceans (also referred to as thermohaline circulation) almost certainly influences the climate on a global scale (e.g., Vellinga and Wood, 2002). Life in the sea is dependent on the biogeochemical status of the ocean and is influenced by changes in the physical state and circulation. Changes in ocean biogeochemistry can directly feed back into the climate system, e.g., through changes in uptake or release of radiatively active gases such as carbon dioxide. Changes in sea level are of obvious importance for the human society, and are directly linked to changes in ocean circulation. Finally, oceanic parameters can be useful for climate change detection, in particular temperature and salinity changes in the deeper layers where the variability is smaller and signal-to-noise ratio is higher (e.g., Banks and Wood, 2002).

In the Third Assessment Report (TAR), several aspects of the ocean's role have been discussed. Folland *et al*. (2001) concluded that the global ocean has significantly warmed since the late 1950s. The warming is superimposed on strong global decadal variability. More than half of the warming was found in the upper 300m, equivalent to an average temperature increase of 0.037ºC per decade (Levitus *et al*., 2005a). Warming in many regions of the North Atlantic was found to be accelerating and likely to have contributed to parallel increases of near-surface air temperature over much of Europe. No information on ocean circulation changes (other than those related to ENSO) was presented in the TAR.

In the TAR, Prentice *et al*. (2001) presented a preliminary estimate of the total carbon increase in the ocean by 107 ± 27 PgC for the 1750–1990 period. This estimate was based entirely on indirect evidence. No other changes in ocean biogeochemistry (including pH) were discussed in the TAR.

In the TAR, Church *et al*. (2001) adopted a best estimate of 1.5 ± 0.5 mm yr^{-1} for the observed sea level rise in the 20th century. They also provided estimates of various climate-related contributions to the 20th century sea level rise, mainly based on models, and concluded that the largest positive contribution arises from thermal expansion due to warming of the oceans. Melting of mountain glaciers was found to contribute substantially, whereas the contributions from Greenland and Antarctica mass imbalance were more uncertain though likely also to be positive. The most uncertain contribution was identified as the change in terrestrial water storage resulting from human activities, which corresponds to sea level decrease. The sum of all climate contributions to sea level rise amounts to 0.7 ± 1.5 mm yr^{-1}. Church *et al*. (2001) note that this value is less than half of the observed value, although there is overlap between their respective uncertainties. It thus appeared that either the climate-related processes causing sea level rise have been underestimated or the rate of sea level rise observed with tide gauges is biased toward too high values.

This chapter will assess observations of changes in oceanic parameters. Among others, the following questions will be addressed:

1. How much, and where, are the oceans warming? Are changes in heat and freshwater content consistent with surface flux changes?

2. Is the oceanic circulation changing? Can the causes for observed changes be inferred?

3. Is the biogeochemical system in the ocean changing? Are these changes consistent with observed circulation changes?

4. At which rate, and where, is the sea level rising? Can the causes of observed sea level change be quantified? Can the discrepancy identified in the TAR be resolved?

5.2 Global Trends in Heat Content and Salinity

5.2.1 Background

Three of the major challenges for the climate-system community are quantifying the earth's heat balance, freshwater balance (hydrological cycle), and the carbon cycle. The contribution of the world ocean to each of these balances is substantial or dominant. Here we present observational evidence that directly or indirectly helps to quantify these balances.

The TAR included estimates of ocean heat content for the upper 3000 m of the world ocean. Here we report on updates of this estimate and present three new estimates for the upper ocean based on additional modern and historical data (Levitus *et al.*, 2005; Ishii *et al.*, 2005; and Willis *et al.*, 2005). We also present new estimates of the temporal variability of salinity. The data used for temperature and heat content estimates are based on the World Ocean Database 2001 and are described by Conkright *et al*. (2002), Boyer *et al*. (2002), Locarnini *et al*. (2002), and Stephens *et al*. (2002) and other additional sources. Temperature data include measurements from reversing thermometers, expendable bathythermographs (XBT), mechanical bathythermographs (MBT), conductivity-temperature-depth (CTD) instruments, profiling floats, moored buoys, and drifting buoys. The salinity data are described by Locarnini *et al*. (2002) and Stephens *et al*. (2002). The comprehensiveness of the World Ocean Database was made possible as a result of international data exchange under the aegis of the ICSU World Data Centre system and the IOC/IODE system. In particular, large amounts of historical data have been acquired as a result of the "Global Oceanographic Data Archaeology and Rescue (GODAR) Project" sponsored by the Intergovernmental Oceanographic Commission (Levitus *et al.,* 2005b). All of these data are available online at http://www.nodc.noaa.gov/OC5/indprod.html. This web site also includes data distribution plots for individual years and pentads (5-year periods) for temperature and by pentads for salinity.

5.2.2 Ocean Heat Content

Figure 5.2.1 shows time series of ocean heat content by years for the upper 700 m layer of the world ocean and by pentads for the upper 3000 m layer of the world ocean (Levitus *et al.,* 2005a). A total of 7.3 million profiles of temperature were used in constructing these time series. The lack of data with increasing depth forces us to composite data by five-year running pentads in order to have enough data for a meaningful analysis in the deep ocean. Even then, there is not enough deep ocean data to extend the time series for the upper 3000 m past the 1994–1998 pentad. However, visual inspection of the two curves in Figure 5.2.1 show close agreement and computations based on these time series (comparison of the two linear trends) indicate that about 69% of the increase in ocean heat content during 1955–1998 (the period when we have estimates from both time series) occurred in the upper 700 m of the world ocean. The time series shows an overall trend of increasing heat content in the world ocean with significant interdecadal variations superimposed on this trend. Based on the linear trend, for the 0–3000 m layer for the 1955–1998 period, there has been an increase of ocean heat content of approximately 14.5×10^{22} J corresponding to a mean temperature increase of 0.037ºC at a rate of 0.2 W m^{-2} (per unit area of the earth's surface). One feature of this figure is the decrease in heat content during 1980–1983. The 0–700 m layer cooled at a rate of 1.2 W m^{-2} during this period. Most of this cooling occurred in the Pacific Ocean. Some of this cooling could be due in part to the volcanic eruption of El Chichon in 1982 (Church *et al*., 2005a). Ocean cooling also occurred after 1963 and may in part be due to the eruption of Mount Agung. Notably, there was little if any cooling after the eruption of Mount Pinatubo in June 1991. The variation in the time series of global ocean heat content may result from internal variability of earth's climate system, external variability such as volcanic aerosols, and from anthropogenic forcings such as the increase of aerosols and greenhouse gases in earth's atmosphere. Deconvolving the global ocean heat content integral is critical for understanding the past, present, and future state of earth's climate system.

[INSERT FIGURE 5.2.1 HERE]

Figure 5.2.2 shows the linear trends of zonally integrated heat content for the world ocean and individual basins by 100 m thick layers based on yearly temperature anomaly fields (Levitus *et al*., 2005a). The strongest trends of this quantity are concentrated in the upper ocean. Warming occurs at most latitudes in all three of the ocean basins. Some areas show cooling, mainly in the shallow equatorial areas and at some high latitude regions. In the Indian Ocean cooling occurs at subsurface depths centred on 12ºS at 150 m depth and in the Pacific centred on the Equator and 150 m depth level. The cooling of the subsurface equatorial Pacific and the Indian Ocean at 12ºS may be due to changes in the thermocline structure associated with changes in the tropical wind field that occurred with the reversal in polarity of the Pacific Decadal Oscillation (PDO) in the late-1970s. This remains to be determined. Cooling also occurs in the 32–48ºN region of the Pacific Ocean and the 49–60ºN region of the Atlantic Ocean. The cooling of the upper ocean of the subarctic gyre of the North Atlantic during 1947–1985 has been documented by Levitus *et al*. (1994) based on time series data from Ocean Weather Station 'C'. The variability is characterized by a linear cooling trend of 0.19ºC per decade on which is superimposed strong quasidecadal oscillations with a range of approximately 2ºC. These oscillations appear to have been associated with the atmospheric variability over the East Atlantic. The cooling and freshening of the deep water of the subarctic since 1970 has been documented by Dickson *et al*. (2002) among others. The percentage variances explained by the linear trends shown in Figure 5.2.2 are given by Levitus *et al*. (2005a).

[INSERT FIGURE 5.2.2 HERE]

Results of EOF analysis of the global, yearly, 0–700 m heat content fields (Levitus *et al.,* 2005c) are shown in Figure 5.2.3 (spatial patterns and time series of EOFs 1 through 4). This analysis is a way of characterizing the time and space scales responsible for the global integrals of heat content described above. Due to the importance of ocean heat content as an indicator of the state of earth's climate system, and in particular for detection and attribution studies focusing on climate-system change, we describe these results in detail. It is important to recognize that the EOFs described here may not correspond to physical or dynamical modes of the ocean or ocean-atmosphere system although the results shown here bear resemblance to physical phenomena that have been described in the scientific literature. EOF analysis as used here only identifies the stationary modes of the heat content fields being analyzed. To identify physical and/or dynamical modes requires analyses by individual regions and/or basins since it is well-known that the results of multivariate analysis strongly depend on the region and time period of the data fields being analyzed.

We note that it is the product of the spatial pattern of the EOF and its related time series that yield the ocean heat content values associated with each EOF. Thus, the sign of the spatial pattern of any EOF can be reversed as long as its associated time series is reversed in sign. The first four EOFs account for 11.1, 8.7, 5.0, and 4.2 percent of the total variance. All four EOF fields are characterized by spatial variability on gyre and basin scales. EOF 1 is dominated by an east-west dipole between the eastern and western tropical Pacific Ocean. The negative polarity of the eastern tropical Pacific extends poleward on both sides of the equator in the eastern Pacific Ocean. The region of positive polarity in the western tropical Pacific extends to the southeast to approximately 40°S, 120°W. There is a region of positive polarity extending eastward from Japan centred along 40°N. The Indian Ocean is characterized by a region of positive polarity west of Australia with most of the rest of the Indian Ocean being of negative polarity. The Atlantic is characterized by a region of positive polarity in the subarctic with most of the rest of the Atlantic being of opposite polarity. The time series of EOF 1 exhibits relatively strong interannual variability, particularly after 1982, but is dominated by a change in sign during the late-1970s. The spatial and time series suggests that EOF 1 is associated with the PDO. Some of the interannual variability of this EOF and EOFs 2–4 appear to be associated in part with ENSO events. EOF 2 is characterized by a region of positive polarity extending westward from the equatorial eastern Pacific Ocean and a region of positive polarity centered on 40°N. A region of negative polarity extends northeastward from the western equatorial Pacific. The Indian Ocean is characterized by a region of positive polarity centred on 10°S with most of the remaining Indian Ocean characterized by negative polarity. Similar to EOF 1, the subarctic of the Atlantic is characterized by a region of the same polarity with most of the rest of the Atlantic being characterized by polarity of the opposite sign. The time series of EOF 2 clearly exhibits a negative linear trend. EOF 3 exhibits a trans-Pacific region of positive polarity south of approximately 4°N, a trans-Pacific region of negative polarity in the 4°N–15°N latitude belt and then a nearly trans-Pacific region of positive polarity centred on 40°N. The Indian Ocean is characterized by a region of positive polarity west of Australia and in the Arabian Sea. The subarctic Atlantic exhibits negative polarity that extends southward along the eastern Atlantic. Most of the rest of the North Atlantic is characterized by positive polarity between the equator and 40°N and the South Atlantic exhibits regions of mixed polarity. The time series of EOF 3 exhibits interannual variability on the order of 4–6 years. For the Pacific Ocean, EOF 4 is characterized by a series of zonal bands of alternating polarity extending southward from a band of negative polarity centred on 45°N to a band of negative polarity centred on 15°S. Most of the eastern Pacific Ocean is of negative polarity. The Indian Ocean exhibits positive polarity west of Australia. The Atlantic exhibits negative polarity in the central part of the subarctic gyre, in the western extension of this gyre north of the Gulf Stream and in the subtropics of the North Atlantic. South of the Gulf Stream there is a region of positive polarity. The time series of EOF 4 exhibits a positive linear trend after 1976.

Figure 5.2.4 shows four reconstructions of the 0-700 m global heat content integral based on EOF 1, EOFs 1–2, EOFs 1–3 and EOFs 1–4. The reconstruction using just EOF 1 shows an increase in ocean heat content in the late 1970s that appears to be associated with the PDO. However, there was only one reversal of the polarity of the PDO during the period we are examining so it is not possible to state with certainty that a change in ocean heat content occurs with each reversal of polarity of the PDO. If this were the case, then this might be an example of earth's heat balance varying due to internal variability of the ocean-atmosphere system. We note that the global integral of ocean heat content began increasing substantially around 1970. The reconstruction based on EOFs 1–2 captures the large increase in heat content that began around 1970 and which peaked around 1977–1980. The reconstruction based on EOFs 1–3 accounts for 24.8% of the variance-covariance matrix used in generating the EOF patterns shown here. These three EOFs do a reasonably good job of explaining the global heat content time series shown in Fig. 5.2.1. The reconstruction based on the first four EOFs account for 29.0% of the variance covariance matrix and only marginally improves on the reconstruction based on the first three EOFs.

It is clear from examining the linear trends of zonally integrated heat content in Fig. 5.2.2, the results of the EOF analysis shown in Figure 5.2.3, and the reconstructions (Fig. 5.2.4) of the 0–700 m global heat content fields using the EOFs, that global ocean heat content variability is associated with variability on large spatial scales and is not due to sampling mesoscale features such as eddies. Some of the temporal variability on interannual time scales seen in the results of the EOF analyses appears to be related to ENSO events. The reversal in polarity of the PDO that occurred in the late 1970s was associated with a step-like increase in ocean heat content.

[INSERT FIGURE 5.2.3 HERE]

[INSERT FIGURE 5.2.4 HERE]

Two other upper ocean temperature anomaly fields are available for estimating the available heat content. These alternative analyses allow the quantitative testing of the reliability of ocean heat content estimates. They are the Ishii *et al*. (2003) and Willis *et al*. (2005) analyses (Figure 5.2.5). The three heat content data sets cover different periods but where they overlap in time there is good agreement. The rms difference between the three data sets is 1.6 x 10^{22} J or 1.4 W m^{-2}. The two longest time series (using different methods on similar data sets) show very good agreement on trends and on decadal time scales. There are year-to-year differences but these differences are within the estimated errors. For the 1993–2003 period, the Levitus *et al*. (2005c) analysis has a linear global ocean trend of 0.70 ± 0.11 W m^{-2} compared with 0.86 ± 0.12 W m^{-2} and 0.58 ± 0.09 W m^{-2} respectively for Willis *et al.* (2005) and Ishii *et al.* (2005). (Note however that Levitus is from 0 to 700m and Willis is from 0 to 750m). All of these estimates are per unit area of ocean surface. While there are differences between these three ocean heat content estimates due to instrumental biases, temporal and spatial averaging and analysis methods (Appendix 5.A.1), we conclude that the available heat content estimates are consistent with each other giving a high confidence in their use in climate change studies.

[INSERT FIGURE 5.2.5 HERE]

5.2.3 Ocean Salinity

Ocean salinity is an indirect but potentially sensitive variable for detecting changes in precipitation (and evaporation) and hence changes in the earth's hydrological cycle. Figure 5.2.6 shows the linear trends (based on pentadal anomaly fields) of zonally averaged salinity for the world ocean and individual ocean basins (Boyer *et al.,* 2005). A total of 2.3 million salinity profiles were used in this analysis. These results are based on a data set about one third of the size of the ocean heat content estimates. The percent variance accounted for by the linear trend for each basin and the world ocean is given in the original paper. Between 15ºS to 42ºN of the Atlantic Ocean there is an increase in the upper 500 m layer. This region includes the North Atlantic sub-tropical gyre. In the 42–72ºN region, including the Labrador, Irminger and Icelandic Seas, there is a strong freshening (discussed further in Section 5.3) and South of 50ºS in the Polar region of the Southern Ocean a weaker freshening signal. Freshening occurs throughout most of the Pacific with the exception of the South Pacific Gyre between 32–8ºS where there is an increase in salinity. The near surface Indian Ocean is characterized mainly by increasing salinity. However, in the latitude band 42–5ºS, that is the South Indian Gyre, in the depth range of 200–1000m there is a freshening of the water column. The global zonal average trend shows that the mid-latitudes ocean in both hemispheres is tending to increased salinity (discussed further in Section 5.3) and that polar regions are tending to freshen at all depths (Figure 5.2.6d).

[INSERT FIGURE 5.2.6 HERE]

5.2.4 Implications for Earth's Heat Balance

To place the variability of ocean heat content in perspective, Figure 5.2.7 provides estimates of the change in heat content and heat of fusion associated with possible melting of different components of the cryosphere of various components of the earth's climate system for the 1955–1998 period (Levitus *et al.,* 2005c). The increase in ocean heat content dominates earth's heat balance on decadal time-scales. It accounts for approximately 84% of the possible increase in heat content of the earth system during this period. This possibility was first suggested by Rossby (1959) based on considerations of the specific heat of water and the mass of the ocean in comparison to similar characteristics of other parts of the climate system. These results demonstrate that ocean heat content variability is a critical metric for detecting the effects of the observed increase in greenhouse gases in the earth's atmosphere, as important as the related problem of understanding how carbon dioxide is stored in the global ocean.

[INSERT FIGURE 5.2.7 HERE]

5.2.5 Global Heat Balance and Meridional Fluxes

5.2.5.1. Surface ocean flux changes

Both reanalyses and VOS-based estimates of the surface net heat fluxes imply biases of order 30 W/m^2 on regional and global scales (e.g., Josey*,* 2001; Qiu *et al*.*,* 2004; Smith *et al.,* 2001). The imbalances can however be corrected using ocean heat transports from the oceanic cross-sections as constraints (Grist and Josey*,* 2003). Reanalyses of surface fluxes are able to reveal air-sea flux variability patterns associated with the major modes of atmospheric variability, such as NAO-associated SST tripole in the North Atlantic (Marshall *et al*.*,* 2001; Visbeck *et al.,* 2002). During the last 40–50 years both NCEP/NCAR and VOS net fluxes demonstrate decreasing fluxes (up to 1 W/m^2 per year) over the southern flank of the Gulf Stream and positive trends (up to 0.5 W/m^2 per year) in the Atlantic subtropics and central subpolar regions (Gulev *et al.,* 2005). However, in the Labrador Sea and the northwest Atlantic linear trends in the net flux in NCEP and VOS fluxes show disagreement in sign which can be attributed to the time-dependent sampling bias in the VOS fluxes in this area. Freshening of the North Atlantic in the extratropics and salinification in the tropics over the last 40 years (Curry *et al*., 2003) implies a strong link between the salinity changes and the surface fresh water flux. Josey and Marsh (2005) have shown that surface freshening of the eastern subpolar gyre since the late 1970s is strongly correlated with a major increase in precipitation to this region. In the North Pacific, Deser *et al*. (2004) have quantified changes in basic meteorological fields (air and sea surface temperature, precipitation and cloudiness) associated with a regime shift towards a deeper Aleutian low in the late 1970s. It is likely that ocean heat loss in the Eastern North Pacific has decreased since 1977 as a weaker northerly flow of cold air is accompanied by an increase in SST.

5.2.5.2 Ocean heat transport

Ocean Meridional Heat Transport (MHT) is a key mechanism of the re-distribution of heat energy in the climate system, dominating over atmospheric MHT between Equator and 17°N and accounting for 22% and 8% at 35°N and 35°S, where the peak total MHT in each hemisphere occurs (Trenberth and Caron, 2001). Although the uncertainties are still considerable (see Appendix 5.A.2), estimates of the climatologically implied mean ocean MHT (Trenberth and Caron, 2001; Grist and Josey, 2003) from different sources indicate a convergence in some regions. Global overviews of climatological MHT estimates exclusively based on oceanographic cross-sections (Ganachaud and Wunsch, 2003; Talley, 2003) qualitatively agree in the Atlantic, implying northward MHT with a maximum of greater than 1 PW at approximately 20°N, and showing less agreement in the Pacific and Indian Oceans, where even the sign is still unresolved for many latitudes.

At the best observed oceanographic section at 24° N in the Atlantic data from 1957, 1981 and 1992 indicate a steady value for the MHT of about 1.3 PW (Lavin *et al.*, 2003). At 43–48N in the Atlantic, Koltermann *et al.* (1999) reported interannual variability of the same order of magnitude as interdecadal variability, and implied that the MHT in the Atlantic at 48N has likely increased from the late 1950s to 1990s by approximately 0.2 ± 0.12 PW. Accurate box-model-based analysis of MHT using these data (Lumpkin *et al.*, 2005) shows, however no significant variations on interannual scale. Model studies show significant interannual to interdecadal variability in MHT which is mainly driven by surface heat flux variations associated with the North Atlantic Oscillation (Eden and Willebrand, 2001; Gulev *et al.*, 2003; Stammer *et al.*, 2003).

5.2.5.3 Meridional fluxes of freshwater (means and variations)

Quantitative estimates of the meridional freshwater transport (MFWT) are even more uncertain than those of MHT. Ganachaud and Wunsch (2003), based on inverse box model and the WOCE hydrographic data, imply a net evaporation of about $0.5 \times 10^9 \pm 0.3$ kg/s in the Atlantic Ocean. However, in the other basins, a little agreement between different estimates can be found.

Changes in the freshwater flux in the Nordic Seas and subpolar gyre imply an imbalance in the fluxes (oceanic freshwater fluxes; precipitation less evaporation; river runoff, ice sheet melt). E.g., the salinities in the northern North Atlantic have been declining in recent decades (Section 5.3.2.2), and it has been suggested that an increased hydrological cycle due to a warming climate is partly responsible for the declining salinities (Curry *et al.*, 2003). A recent quantitative estimate of the freshwater build-up during the past 50 years relative to the 1950s finds that the Nordic Seas increase is 4000 km^3 and the Subpolar Gyre increase is 15000 km^3 since 1965 (Curry and Mauritzen, 2005). Remarkably, half the excess freshwater entered in the early 1970s, corresponding to an extra flux of 0.07Sv in that time frame (i.e., a 50% increase over the average freshwater flux into the North Atlantic). In the subsequent decades the freshwater storage has been much more gradual, and it peaked in the mid-1990s.

5.3 Changes in Ocean Circulation and Water Mass Formation

5.3.1 Introduction

The oceanic processes of subduction, ventilation, advection and mixing control the ocean’s response to changes and variations in the main modes of climate variability. The main modes of climate variability are El Niño Southern Oscillation (ENSO), North Atlantic Oscillation (NAO), Pacific Decadal Oscillation (PDO) and the Southern Annular Mode (SAM). These modes drive the ocean, causing changes in ocean circulation through changed patterns of the winds, changes in the density and volume of key global water masses, and in combination lead to changes within deep ocean interior with time scales ranging from decades to millennia. This section assesses the observational evidence for ocean variation and change within the global oceans on regional and basin scales, and examines the climate modes driving these changes.

5.3.2 Atlantic Ocean

Atlantic Ocean climate variability in its northern part is likely to be dominated by superposition of anthropogenic change with the decadal signals from the North Atlantic Oscillation (NAO) and on a longer time-scale by the Atlantic Multi-decadal Oscillation (AMO) (Kerr, 2000). The linear trends in temperature and salinity shown in Section 5.2 (Figures 5.2.2 and 5.2.6) are broadly consistent with persistent positive NAO during the last several decades. During positive NAO, the North Atlantic subtropical gyre is warm while the subpolar gyre is colder, fresher and convection is respectively intensified in the Labrador and Irminger Seas (Dickson *et al.*, 1996) and weakened in the Nordic Seas. Observations also show a long-lasting freshening of Nordic Seas, which, with freshening of the sub-polar gyre, leads to freshening of the southward flowing North Atlantic Deep Water (NADW).

5.3.2.1 Upper ocean circulation and water property changes

Temperature changes in surface Atlantic waters (upper 200–300 meters) (Figure 5.2.2) are consistent with primarily warming tendencies identified from the global analyses of SST (cf. section 3.2.2.3). In the tropical Atlantic, part of the surface water changes are associated with the variability in the Atlantic Marine Inter-tropical Convergence Zone which has strong seasonal variability (Biasutti *et al*., 2003; Mitchell and Wallace, 1992; Stramma *et al*., 2003). Quasi-decadal variations of tropical SST, induced by the atmospheric variation, represent a superposition of the meridional gradient mode (Enfield *et al*., 1999) driven by rainfall variability during boreal spring (Nobre and Shukla, 1996; Ruiz-Barradas *et al*., 2000; Wang, 2002) and the Equatorial Atlantic El Niño, a mode which is likely to be forced by the boreal summer rainfall variability. The last manifestation of the El Niño mode occurred in 1999 (Xie and Carton, 2004). The gradient and El Niño modes can be also influenced by remote forcing from the Pacific ENSO (Chiang *et al*., 2002; Hastenrath, 2000) and southern hemisphere (Barreiro *et al*., 2004; Czaja *et al*., 2002). Because of the paucity of ocean observational datasets, decadal signals are still poorly detectable and some are at the level of model hypothesis.

In the Atlantic subtropical gyre, (1) SST, (2) temporal variations of the thick near-surface water mass, Subtropical Mode Water (STMW) (Hanawa and Talley, 2001), at Bermuda, and (3) thermocline ventilation are highly correlated with the NAO, with low thickness/production and warmer water during high NAO index (Gulev *et al*., 2003; Joyce *et al*., 2000; Marsh, 2000; Talley, 1996). The volume of STMW lags changes in the NAO by a couple of years, and low (high) volumes of STMW are associated with high (low) temperatures (Kwon and Riser, 2004). While quasi-cyclic variability in STMW renewal is apparent over the 1960–1980 period, the total volume of STMW appears to have halved since the early 1980's, associated with increasing NAO index (Gulev *et al*., 2003). The NAO index increased to the end of the 1990s and might have begun to decline since then.

The upper limb of the Meridional Overturning Circulation (MOC) brings Gulf Stream waters to the subpolar region and North Atlantic Current. Upper layer temperature and salinity changes in this system are coordinated with SST variability (which is discussed in Chapter 3), again mostly governed by the NAO (Deser and Blackmon, 1993; Masina *et al*., 2004). In addition to oscillations of order 0.2°C associated with local air-sea flux, SST is influenced by propagating ocean signals of order 0.5° to 1°C (Dickson *et al*., 1988; Hansen and Bezdek, 1996; Sutton and Allen, 1997). NAO-induced changes also result in south-north changes of the Gulf Stream position (Joyce *et al*., 2000; Molinari, 2004; Seager *et al*., 2001). At present it is uncertain whether these changes in the Gulf Stream position occur exactly in phase with the NAO (Joyce *et al*., 2000) or are lagged by 1–2 years (Taylor and Stephens, 1998).

At subpolar latitudes the surface layer changes are also closely correlated with NAO (Bersch, 2002; Flatau *et al*., 2003). Positive NAO implies strengthening of the subpolar gyre, intensified convection in the Labrador and Irminger Seas and weakened convection in the Nordic Seas. Advection of surface salinity anomalies can be a major factor in controlling the depth of convection, particularly in the Labrador Sea, since a fresh anomaly can severely limit convection, as happened in the 1960s–1970s (Dickson *et al*., 1988; Lazier, 1980; Talley and McCartney, 1982). The subpolar surface salinity variations are quasi-decadal; during high NAO, the subpolar gyre is strong and expanded towards the east (Bersch 2002; Flatau *et al*., 2003), resulting in lower salinity waters in the central subpolar region (Bersch, 2002; Levitus, 1989; Reverdin *et al*., 1997).

At multidecadal time scales, AMO subtropical SST anomaly variability has time scales from 50 to 80 years with a magnitude of ±0.2–0.4°C between the equator and 40°N (Delworth and Mann, 2000; Enfield *et al*., 2001). Starting from the mid 1990s the Atlantic has been in a positive (warm) AMO phase.

5.3.2.2 Intermediate and deep circulation and water property changes

Marked changes in NADW are apparent in the subtropical Atlantic over the past 50 years (Figure 5.2.2), reflecting changes in source waters in the Nordic Seas, Labrador Sea and Mediterranean Sea. At depths of 1000–2000 m, temperature has clearly increased since the late 1950s at Bermuda, at 24°N, and at 52°W and 66°W in the Gulf Stream (Bryden *et al*., 1996; Joyce and Robbins, 1996; Joyce *et al*., 1999). After the mid 1990s at greater depths (1500–2500 m), temperature and salinity decreased, reversing the previous warming trend. The initial trends reflected reduced production of Labrador Sea Water (LSW) (Lazier, 1995) and increased salinity and temperature of the Mediterranean Water (Potter and Lozier, 2004). At 500–1000 m at the same locations, a decade-long increase of temperature and salinity at depth stopped around 1999 (Phillips and Joyce, 2005). Curry *et al*. (2003) attributed the upper ocean salinity increase at 24ºN (Figure 5.3.1) to increased evaporation through air-sea exchange and the transport of the freshwater from the subtropics to the subpolar North Atlantic, where salinities were decreasing with time since the late 1950s (e.g., Dickson *et al.*, 2002) (Figure 5.3.2).

[INSERT FIGURE 5.3.1 HERE]

Net evaporation in the Mediterranean Sea transforms Atlantic Water into saltier, cooler, denser, Mediterranean Water (MW). MW affects the Atlantic and global thermohaline circulation (Gerdes *et al.*, 1999; Potter and Lozier, 2004; Talley, 1996) through injection of high salinity at mid-depths into the NADW (e.g., Artale *et al.*, 2002). In the last decade (1994–2003), a large warming (0.3°C) and salinity increase (0.6) were observed at Gibraltar (Millot *et al.*, 2005) and of similar magnitude in the Bay of Biscay (Vargas-Yanez *et al.*, 2002). From 1955 to 1993, the trends are ~0.1°C and 0.02/decade in a zone west of Gibraltar (Potter and Lozier, 2004) and of almost the same magnitude even west of the mid-Atlantic Ridge (Curry *et al.*, 2003). These changes are much larger than the trends of ~0.01°C/decade within most waters at a global scale (Levitus *et al.*, 2000). The large changes in outflow properties of the Mediterranean are due to a shift in the types and vigour of dense water formation, rather to property changes of each type of dense water formed in the Mediterranean. For instance, Western Mediterranean Deep Water (WMDW) has warmed since the 1960s (Rixen *et al.*, 2005), but its trends of ~0.035°C and ~0.01/decade are much less than those observed in the Mediterranean outflow. But since the mid 1990s the outflow has been composed of other Mediterranean water masses that are ~0.3°C warmer and ~0.6 saltier than WMDW. Thus changes within the Mediterranean that shift the volumetric importance of the different dense waters are critical for MOC properties (Millot *et al.*, 2005).

[INSERT FIGURE 5.3.2 HERE]

The sub-polar North Atlantic freshened at most depths during the past several decades (Figures 5.3.1, 5.3.2) (Curry *et al.*, 2003; Dickson *et al.*, 2002; Dickson *et al.*, 2003). Labrador Sea Water, as a major component of NADW, is of special interest. After several decades of warming and decreasing volume, interrupted briefly by a period of deep convection in 1976–1977, a longer period of winter convection over 6 years (1988/89 to 1994/95) produced an exceptionally large volume of cold, fresh, dense LSW (Sy *et al.*, 1997). From 1994 to 2004, the intensity of deep convection decreased, implying decreasing LSW volume with a short interruption in 1999-2000. This was associated with the declining of the North Atlantic subpolar gyre, seen also in the TOPEX/POSEIDON altimetry data (Häkkinen and Rhines, 2004). LSW volume and properties are governed mainly by the NAO, shown in observations (Dickson, 1997; Dickson *et al.*, 1996; Lazier *et al.*, 2002; Yashayaev *et al.*, 2003) and model studies (Eden and Willebrand, 2001; Gulev *et al.*, 2003).

The densest waters contributing to NADW and the deep limb of the MOC arise as overflows from the upper 1500 meters of the Nordic Seas. The overflow water masses exiting the Arctic have freshened markedly, associated with positive NAO and associated growing sea ice export from the Arctic and precipitation in the Nordic Seas (Dickson *et al*., 2002, 2003). In contrast, the transports of the overflow waters have been relatively stable (Dickson *et al*., 2002), with no clear variability in Denmark Strait and a decrease of 20% since the 1950s in the easternmost channel (Faroe Bank) where the total transport is smallest (Hansen *et al.*, 2001). The freshening overflows and small decrease in their transport suggests a potential weakening of the MOC, although increased salinity of the surface waters feeding the MOC may offset the weakening (Hansen *et al.*, 2004)**.**

[START OF BOX 5.1]

Box 5.1: Is the Thermohaline Circulation Changing?

The thermohaline circulation is a global 3-dimensional system of currents that redistributes heat and freshwater. In the North Atlantic it is manifested as a meridional overturning circulation (MOC) consisting of inflow of surface waters from the south that continuously densify through cooling as they move northward through the subtropical and subpolar gyres. The inflows reach the Nordic Seas (Greenland, Iceland and Norwegian Seas) and the Labrador Sea, where they are subject to deep convection. The convected Nordic

Seas waters return at depth to the North Atlantic as sill overflows where they are subject to vigorous mixing. The Mediterranean Sea is also a factor in transforming Atlantic surface waters, to mid-depth saline waters. Through these processes the main components of North Atlantic Deep Water are formed, constituting the southward-flowing lower limb of the MOC. The upper limb of the MOC transports sufficient heat and salt into the Nordic seas to keep them free of sea ice in winter, which has a moderating effect on European winter climate.

Analysis of sediment cores and corals shows major changes in the MOC during the past 120,000 years (Rahmstorf, 2002). There is also solid evidence for a link between these ocean circulation changes and abrupt changes in surface climate although the exact mechanism is not clear (Clark *et al.,* 2002). The most dramatic abrupt climate changes are the Dansgaard-Oeschger warm events, with a warming that can exceed 10ºC within a decade or so in central Greenland (Severinghaus *et al*., 2003). Proxy data show that the South Atlantic cooled when the north warmed (Voelker and workshop participants, 2002), supporting the hypothesis of an ocean heat transport change, and that salinity in the Irminger Sea increased strongly (Kreveld *et al.,* 2000), indicating northward advance of saline subtropical Atlantic waters. At the end of the last glacial, as the climate warmed and ice sheets melted, there were a number of abrupt oscillations, i.e., the Younger Dryas and the 8.2k cold event, which may have been caused by the ocean circulation. The variability of the thermohaline ocean circulation during the Holocene after the 8.2 k event is discussed in (Keigwin *et al*., 1994). Holocene variations are clearly much smaller than during glacial times.

Projections with coupled ocean-atmosphere models suggest that the MOC may gradually decrease in the 21st century as a consequence of anthropogenic warming and freshening in the North Atlantic (Cubasch *et al.*, 2001, see also Chapter 10). Direct observations of the MOC index do not exist, and one has to consider indirect evidence to infer changes of the MOC.

Changes in subpolar, Nordic Seas and Mediterranean Sea water mass formation are connected to changes in the MOC. Freshening of intermediate and deep waters throughout the subpolar gyre and Nordic Seas over the past several decades has been reported (Figure 5.3.2) (Dickson *et al*., 2002, 2003, Curry *et al.*, 2003). Meanwhile Mediterranean Water has become warmer and more saline. Decadal fluctuations in water masses in all of these regions are closely associated with the NAO. High NAO periods correspond with cold SST in the subpolar North Atlantic (Deser *et al*., 2002; Flatau *et al.,* 2003) and warm SST in the subtropics and Nordic Seas (Dickson *et al*., 1996). During high NAO, Labrador Sea convection is high, Nordic Seas convection is low, and Mediterranean Sea convection shifts in properties (Dickson *et al.*, 1996; Millot *et al.*, 2005). The NAO was exceptionally low in the 1960s, and has been in a protracted high state since the late 1980s, with occasional 1–2 year drops. The freshening and associated reduction in density of the overflow waters would lead to a weakening of the MOC; model results (Latif *et al*., 2005) suggest that the observed freshening would correspond to a reduction of 5–10% of the MOC strength. Model results (Eden and Willebrand, 2001) also suggest that the MOC index may have varied by 10–20% due to changes in Labrador Sea convection, with strong convection corresponding to high MOC. The observed basin-wide salinity changes are those expected from anthropogenic processes, but the available observational records are insufficient to distinguish between natural and anthropogenic processes.

A climate mode of longer timescale, the Atlantic Multidecadal Oscillation (AMO), spans the Atlantic hemispheres, with an index given roughly by North Atlantic minus South Atlantic SST. Coupled atmosphere-ocean models (Latif *et al.,* 2004) indicate that the MOC is closely related to the AMO. Knight *et al*. (2005) infer from the AMO that the MOC has increased since 1970, and conclude that even without anthropogenic influences a reduction of the MOC over the next 2–3 decades is very likely.

In summary, we conclude that up to the end of the 20th century the MOC properties are changing, but that it is so far not possible to distinguish between natural and anthropogenic causes.

[END OF BOX 5.1]

5.3.2.3 Adjacent seas: Arctic Ocean, Nordic Seas and Mediterranean Sea

The densest waters of the Atlantic MOC are formed in the Nordic Seas; sea ice cover in the Arctic is likely an important aspect of global climate. The Mediterranean Sea provides very high salinity to the mid-depth NADW and possibly also affects the salinity of inflow to the Nordic Seas. Climate change in the

Arctic/Nordic Seas is closely linked to the North Atlantic subpolar gyre (Østerhus *et al*., 2005), while climate change in the Mediterranean is also closely linked to the adjacent North Atlantic. Both are affected by the NAO.

Within the Arctic and Nordic Seas, surface temperature has increased since the mid-1980s and continues to increase (Comiso, 2003). In the Atlantic waters entering the Nordic Seas, a temperature increase in the late 1980s and early 1990s (Carmack *et al*., 1995; Quadfasel *et al*., 1991) has been associated with the shift in the 1980s from low to high NAO/AO. Salinity in the Nordic Seas has also decreased markedly since the 1970s (Dickson *et al*., 2003), directly affecting the salinity of the Nordic Seas overflow waters that contribute to NADW. Salinity increased in the upper layers of the Amundsen and Makarov Basins, while salinity of the upper layers in the Canada Basin decreased (Morison *et al*., 1998). Compared to the 1980s, the area of Pacific-derived upper waters has decreased (McLaughlin *et al*., 1996; Steele and Boyd, 1998).

The drastic change in Arctic ice cover through the 1990s has accelerated in the present decade (e.g., Comiso, 2003). In addition to its effect on albedo, melting changes ocean salinity structure and hence vertical stratification changing the conditions for further ice formation and convection. These surface freshening and decrease in surface stratification impact the MOC, contributing to the freshening of North Atlantic polar waters (Figure 5.3.2). During the 1990s redirection of river runoff from the Laptev Sea (Lena River etc.) reduced the low salinity layer in the Arctic Ocean covering the winter mixed layer (Steele and Boyd, 1998), thus allowing greater convection and heat transport into the surface Arctic layer from the Atlantic layer. This process could have been a factor in reduced ice formation (Martinson and Steele, 2001). Recently, however, the stratification in the central Arctic (Amundsen Basin) has increased and a low salinity mixed layer was again observed at the North Pole in 2001 (Bjork *et al*., 2002).

Within the Mediterranean Sea, water properties and circulation are affected by the long-term variability of surface fluxes (Krahmann and Schott, 1998), associated mainly with the NAO (Hurrell, 1995; Vignudelli *et al*., 1999), resulting in coordinated changes in surface heat fluxes in the Atlantic and Mediterranean Sea (Rixen *et al*., 2005). Biochemical and hydrographic databases show marked changes in thermohaline properties throughout the Mediterranean (Manca *et al*., 2002). In the western basin, the Western Mediterranean Deep Water (WMDW), formed in the Gulf of Lions, warmed during the last 50 years, interrupted by a short period of cooling in the early 1980s (Figure 5.3.3). The last decade was the warmest, in agreement with recent atmospheric (Luterbacher *et al*., 2004) and global ocean temperature (Levitus *et al*., 2001) results. WMDW salt content has been steadily increasing during the last 50 years, mainly attributed to decreasing precipitation since the 1940's (Krahmann and Schott, 1998; Mariotti *et al*., 2002) and man-induced reduction of the freshwater inflow (Rohling and Bryden, 1992).

[INSERT FIGURE 5.3.3 HERE]

Eastern Mediterranean climate variations have been dominated by sharp transients in the deep waters in the early-1990s (Roether *et al*., 1996) and intermediate waters; in particular the Levantine Intermediate Water (LIW) cooled from the late-1970s to the mid-1980s (Brankart and Pinardi, 2001). A negative correlation of SST and NAO (Tsimplis and Rixen, 2002) established the "Eastern Mediterranean Transient" (EMT) in the early 1990s (Klein *et al*., 2000) or in the late 1980s (Gertman *et al*., 2005), when the Aegean experienced strong heat losses (Demirov and Pinardi, 2002; Rupolo *et al*., 2003) and became the main source of Deep Water production in the Eastern Mediterranean instead of the Adriatic. It is less clear whether the EMT is to be considered "unique" or whether it is connected to some internal variability of the Eastern Mediterranean (Pisacane *et al*., 2005). Some observations suggest a major episode of deep water formation in the Aegean in the mid-1970s (Josey, 2003).

5.3.3 Pacific Ocean

The Pacific Ocean is the location of ENSO, which is one of the most intense naturally-occurring climate modes. ENSO interactions with ocean and atmosphere conditions at other latitudes and even other ocean basins are a major factor in climate modes with longer timescales, such as the PNA (associated with the strength of the Aleutian Low and relative strength of the subpolar and subtropical gyres) or PDO. The impact of anthropogenic forcing on these major determinants of interannual to decadal climate change must be closely followed. Climate change in the Pacific should be tracked not only through regional temperature and salinity, but also through climate modes that link the regions. The observational record, while it covers many decades in the upper ocean, is difficult to interpret in terms of causes for changes or shifts in ENSO intensity or PDO phase. Combining data with models, as done recently for the global ocean by Barnett *et al*. (2005), is more likely to provide diagnostics of the impact of anthropogenic forcing.

5.3.3.1 ENSO-related variability and climate change

Anthropogenic change might be manifested in changes in the strength of ENSO and the PDO, which are linked (Mantua *et al.*, 1997). Some coupled ocean-atmosphere model studies suggest that warming trends in the tropical Pacific Ocean in the second half of the 20th century were due to greenhouse gas forcing (Knutson and Manabe, 1998; Meehl and Washington, 1996). However, McPhaden and Zhang (2004) indicate that precise magnitude of anthropogenic influences in the tropical Pacific will be difficult to extract from observations given the rapidity with which observed warming trends can be reversed by natural variations.

A well-studied abrupt shift in the PDO occurred in 1976–1977 (cf. the discussion in Section 5.2). Tropical Pacific manifestations included a slowdown of the shallow meridional overturning circulation and a nearly 1°C warming of the sea surface in the cold tongue of the eastern and central equatorial Pacific (McPhaden and Zhang, 2002). These conditions resemble an enhanced El Niño period. Another "regime" shift occurred in the late 1990s in the North Pacific associated with the PDO (Chavez *et al*., 2003; Peterson and Schwing, 2003). It had strong manifestation in the tropics as a shift to a strong La Niña (mid-1998), persisting until at least 2003: lower SST in the central-eastern tropical Pacific, stronger trade winds, steeper east-west sea level slope, and stronger tropical meridional overturning (McPhaden and Zhang, 2004).

The observed changes in tropical Pacific oceanic and atmospheric conditions had measurable impacts on global climate. Cooler equatorial Pacific SSTs between 1998–2003 were accompanied by shifts in the pattern of tropical rainfall and deep convection, which affected the global atmospheric circulation through teleconnections to higher latitudes. These far-field effects contributed to the development of a globe girdling drought that gripped much of the subtropical Northern Hemisphere during 1998–2002 (Hoerling and Kumar, 2003).

5.3.3.2 North Pacific upper ocean changes

Most literature on climate in the upper North Pacific Ocean deals with interannual and decadal variability rather than long-term trends, which are difficult to discern given the length of the observational record and strength of decadal variability. Important exceptions are the basin-wide integrated views summarized in Section 5.2 above (e.g., Levitus *et al.*, 2005a). The zonally-integrated heat content trend from 1955 to 2003 (Figures 5.2.2 and 5.2.4) is dominated by the PDO regime "shift" in the mid 1970s. The figure shows the importance of the tropical Pacific in climate change as described above; the strong cooling between 50 and 200 m is due to relaxation and subsequent shallowing of the tropical thermocline, resulting from a decrease in the tropical meridional overturning circulation described in 5.3.3.1 (McPhaden and Zhang, 2004).

Warming in the Pacific subtropics, cooling in the subpolar region around 40°N, and slight warming farther north are precisely the pattern associated with a positive PDO (strengthened Aleutian Low) (Miller and Douglas, 2004). Thus the long-term trend in the tropical and North Pacific shown in Section 5.2 is most likely due to the prevalence or enhancement of positive PDO states in recent decades. Salinity variations in the North Pacific have similarly complicated spatial distributions, although Fig. 5.2.6 suggests that on the whole the region has freshened.

Can the multi-decade heat content and freshening trend in the Pacific be attributed to anthropogenic forcing, or is it simply a result of the naturally-varying PDO climate mode (or, as some recent authors would have it, natural variations in ENSO and the PNA)? Barnett *et al*. (2001) and Barnett *et al*. (2005) use ensembles of coupled climate models with and without anthropogenic forcing and a projection of the observed changes in heat content (as in Section 5.2) to conclude that the changed heat pattern is indeed symptomatic of anthropogenic climate change. Thompson and Solomon (2002) suggest that extended positive states of climate modes such as the PDO would result from anthropogenic forcing.

What does a positive PDO state (strong Aleutian Low and strong El Niño) look like within the North Pacific Ocean itself? For this we look to the many descriptions of the Pacific following the shift to positive PDO in

1976 ("regime shift"). The Kuroshio Extension strengthened with increased PDO index and its advection of temperature anomalies has been shown to be of similar importance in maintenance of the positive PDO as variations in ENSO and the Aleutian Low strength (Schneider and Cornuelle, 2004). The Oyashio penetrated farther southward along the coast of Japan during the 1980s than during the preceding two decades, consistent with a stronger Aleutian Low (Hanawa, 1995; Sekine, 1988, 1999). A shoaling of the halocline in the centre of the western subarctic gyre and a concurrent southward shift of the Oyashio extension front during 1976–1998 versus 1945–1975 has been detected (Joyce and Dunworth-Baker, 2003).

Temperature changes in upper ocean water masses in response to the stronger PDO after 1976 are well documented. (The PDO has also decreased in strength at times in recent decades, allowing observation of temperature and salinity changes associated with both strengthening and weakening. Since the long-term trend is towards positive PDO, we emphasize only these results here). The thick water mass just south of the Kuroshio Extension in the subtropical gyre (Subtropical Mode Water) warmed by 0.8°C from the mid-1970s to the late-1980s (Hanawa and Kamada, 2001), associated with stronger Kuroshio advection (Hanawa and Kamada, 2001; Yasuda *et al.*, 2000; Yasuda and Hanawa, 1997). The thick water mass along the subtropical-subpolar boundary near 40°N (North Pacific Central Mode Water) cooled by 1°C following the 1976 regime shift (Yasuda and Hanawa, 1997).

Salinity changes in response to a stronger PDO include freshened surface waters in the subpolar region (strong Aleutian Low), and higher evaporation in the central subtropical gyre (not apparent in Fig. 5.2.6). The latter is tracked by the salinity of a shallow salinity maximum layer in the subtropics. An abrupt salinity increase of 0.1 occurred at the 1976 regime shift, attributed to increased evaporation (Suga *et al.*, 2000). Wong *et al.* (2001) reported an interdecadal increase in salinity of the southern part of the salinity maximum layer and related it to redistribution of surface freshwater fluxes over the Pacific.

5.3.3.4 Intermediate and deep circulation and water property changes

Since the 1970s, the major mid-depth water mass in the North Pacific, North Pacific Intermediate Water (NPIW), has been freshening and it has become less ventilated, as measured by oxygen content. NPIW is formed in the subpolar North Pacific, with most influence from the Okhotsk Sea, so NPIW changes reflect northern North Pacific surface conditions. NPIW salinity decreased by 0.1 (0.02) psu in the subpolar (subtropical) gyres (Joyce and Dunworth-Baker, 2003; Wong *et al.*, 2001). An oxygen decrease and nutrient increase in the NPIWsouth of Hokkaido from 1970 to 1999 was reported (Ono *et al.*, 2001), along with a subpolar basin-wide oxygen decrease from the mid-1980s to the late 1990s (Watanabe *et al.*, 2001). Warming and freshening occurred in the Okhotsk Sea in the latter half of the 20th century (Hill *et al.*, 2003). The Okhotsk Sea intermediate water thickness was reduced and its density decreased in the 1990s (Yasuda *et al.*, 2001).

In the southwest Pacific, deeper waters originating from the North Atlantic and Antarctic, cooling and freshening of 0.07°C and 0.01 psu from 1968 to 1991 was observed (Johnson and Orsi, 1997). The authors suggested that the change was due to a warming at the source of these deep waters, most probably at the NADW source, using the Bindoff and McDougall (1994) model for deducing source water changes.

[INSERT FIGURE 5.3.4 HERE]

Bottom waters in the North Pacific are farther from the surface sources of the world's deep waters than any others, with an age of 500 to 1000 years. They are also the most uniform, in terms of spatial temperature and salinity variations. A large-scale, significant warming across the entire North Pacific of the bottom 1000 meters of 0.002°C occurred between 1985 and 1999, measurable because of the high accuracy of modern instruments (Figure 5.3.4, Fukasawa *et al.*, 2004). The cause of this surprising warming is uncertain, but could have resulted from warming of the deep waters in the South Pacific and Southern Ocean, where mid-depth changes since the 1950s are as high as 0.17°C (Gille, 2002).

5.3.3.5 Japan (East) Sea

A long-term trend of warming and salinity change is apparent in the Japan (or East) Sea. Since the 1930s, deep waters have warmed (by 0.1°C at 1000 m and 0.05°C below 2500 m since the 1960s). Since the 1950s salinity has also changed markedly: an increase at 300–1000 m depth and a decrease below 1500m with a trend about 0.06 and –0.02 psu/century, respectively (Kwon *et al.*, 2004). These changes are attributed to reduced surface heat loss and increased surface salinity, which have changed the mode of intermediate and deep ventilation (Kim *et al.,* 2004).

Deep water production slowed for many decades, as reflected in a dramatic decrease in dissolved oxygen in the deep waters (Gamo *et al.,* 1986; Kim and Kim, 1996; Kim *et al.,* 2004; Minami *et al.,* 1998; Talley *et al.,* 2003). Below 2000 m, oxygen continuously decreased at a rate of ~0.8 μmol kg^{-1} yr^{-1}, which would cause anoxia after 200 years. Mid-depth water mass formation was enhanced (Kang *et al.,* 2003), apparent in an increase in dissolved oxygen in the depth range 500–1500 m (Kim *et al.,* 2004). A simple box model predicts that the Japan Sea should remain well-oxygenated as its mid-waters expand over the next few decades (Kang *et al.,* 2004).

Because of weakening vertical stratification associated with the decades-long warming, conditions for convective winter mixing reappeared, resulting in formation of oxygen-rich bottom water in 2001 and subsequent years (Kim *et al.,* 2002; Senjyu *et al.,* 2002; Talley *et al.,* 2003; Tsunogai *et al.,* 2003).

5.3.4 Indian Ocean

For interdecadal time scales sensitivity studies on the role of Indian Ocean SSTs for precipitation elsewhere have been carried out by Giannini *et al*. (2003) and Bader and Latif (2003). In an atmospheric model forced by observed SST anomalies, the Indian Ocean was the main driver of Sahel rainfall anomalies, having an even larger importance than SSTs from the adjacent Atlantic. There are also important interdecadal variations of Indian monsoon rainfall and Indian Ocean SST that are atmospherically connected at the larger scale, e.g., Krishnamurthy and Goswami (2000) and Deser *et al*. (2004). From the sea-surface temperature data (HADISST, Rayner *et al*., 2003) tropical and eastern subtropical Indian Ocean (north of 10ºS) has experienced significant warming from 1900–1999. The warming trend in the period 1900–1970 is relatively weak but positive (Figure 5.3.5a), and increases significantly in the 1970–1999 period, with some regions exceeding 0.2°C/decade (Figure 5.3.5b).

[INSERT FIGURE 5.3.5 HERE]

One coupled mechanism where the Indian Ocean plays a role in modifying the monsoon rains is through the shallow cross-equatorial cell (Miyama *et al.,* 2003; Schott *et al.,* 2002). This cell connects the subduction regions of the southeastern subtropics and Indonesian Throughflow (ITF) via the South Equatorial and Somali Currents with the northern upwelling regions and is closed by southward Ekman transports that cross the equator. Large interannual variability of upper-layer stratification and SST in the Indian Ocean has been associated with the recently discovered Indian Ocean Dipole (IOD) or the Indian Ocean Zonal Mode (IOZM) (Saji *et al.,* 1999; Webster *et al.,* 1999). This mode manifests itself through a zonal gradient of tropical SST, which in one extreme phase, occurring in boreal fall, shows cooling off Sumatra and warming off Sudan (Figure 5.3.6a) in the west, combined with anomalous easterlies along the equator. It has been found that the thermocline variability associated with the IOZM is actually more pronounced than SST anomalies that are masked by unrelated air-sea interaction variability. Thus the development of the IOZM includes propagation of upper-layer thickness anomalies by Rossby waves (Feng and Meyers, 2003; Xie *et al.,* 2002; Yamagata *et al.,* 2004) in the 3°S–15°S latitude band, affecting the elongated upwelling regions northeast of Madagascar.

[INSERT FIGURE 5.3.6 HERE]

East African rainfall shows two maxima, the main one in April-June, and the secondary one in October to December ("short rains"). The magnitude of this rainfall maximum is strongly correlated with IOZM events (Figure 5.3.6a), in particular through increased mixed-layer depth and SST above the dome northeast of Madagascar (Xie *et al*., 2002).

Several recent IOZM events have occurred simultaneously with ENSO events (Figure 5.3.6b) and there is a significant debate going on whether the IOZM is an Indian Ocean Mode or whether it is triggered by ENSO in the Pacific Ocean. One convincing argument for an independent IOZM was the large episode of 1961 when no ENSO occurred (Saji *et al.,* 1999). Saji and Yamagata (2003), analyzing observations from 1958–1997, concluded that 11 out of the 19 episodes identified as moderate to strong IOZM events occurred independently of ENSO. Fischer *et al*. (2004) confirmed that the IOZM is either associated with ENSO or occurs independently. In their coupled model study that generated both ENSOs and IOZMs they suppressed the Pacific ENSO and showed that the IOZM still occurred.

The strongest ever observed IOZM episode occurred in 1997-98 and was associated with catastrophic flooding in East Africa. Latif *et al*. (1999) using an atmospheric model forced by observed SST found that applying seasonal-mean SST variations in the Pacific but observed SST variability in the Indian Ocean yielded approximately the same rainfall anomalies as applying the observed Pacific SST anomalies. Annamalai *et al*. (2005a) generalized these findings by applying ensemble-mean SST anomalies in a similar set of comparative runs as Latif *et al*. (1999). At interdecadal timescales, the SST patterns associated with the Indian monsoon rainfall are very similar to the SST patterns associated with the interdecadal variability of ENSO indices (Krishnamurthy and Goswami, 2000) and with the North Pacific interdecadal variability (Deser *et al.,* 2004). Krishnamurthy and Goswami (2000) find that the interannual variances of the ENSOs and Indian monsoon rainfall increase and decrease simultaneously, with the interannual variances of both, the monsoon rainfall and ENSO (Niño-3 SST regions) being high for the warm phase of the interdecadal SST mode in the Eastern Pacific.

[INSERT FIGURE 5.3.7 HERE]

An interesting decadal variability in the correlations between the IOZM and ENSO has recently been documented by Clark *et al*. (2003) who found high correlations between the IOZM and ENSO (given by Niño-3.4 SSTs) from 1960 to 1983 and again after 1993, but not in the decade in between (Figure 5.3.6c). Annamalai *et al*. (2005b) investigated these decadal changes of interannual correlations by a suite of ocean model experiments concentrating on the decadal variability of thermocline depths. They concluded that the reason for IOZMs to occur independently of ENSOs is the preconditioning of the eastern tropical thermocline to be anomalously shallow. From the single value decomposition (SVD) analysis they find that the EEIO thermocline was particularly shallow during 1952–1971 and 1990–1996, matching periods of IOZM development independent of ENSOs (Figure 5.3.7). The effect of the PDO is diagnosed by Annamalai *et al*. (2004b) to be partially caused by an atmospheric bridge of the PDO, and partially by advection of thinner or thicker mixed layers with the ITF.

South of the equatorial zone in the sub-tropical gyre, Bindoff and McDougall (2000) suggest a 20% slowdown of the subtropical gyre between 1962 and 1987, with a warming of the upper thermocline waters (observed as a cooling and freshening on density surfaces) and a decrease in dissolved oxygen. Between 1987 and 2002, McDonagh *et al*. (2005) find salinification on density surfaces of the upper thermocline, an increase in dissolved oxygen and inferred a 20% speed-up of the gyre over that period. Palmer *et al*. (2004) suggest that the speed up occurs mainly between 1995 and 2002. Thus there is an oscillatory pattern of variation in the circulation strength in this gyre over periods of decades.

5.3.5 Southern Ocean Water Masses

5.3.5.1 Upper ocean property changes

The upper ocean in the southern hemisphere is dominated by the thick mixed layers called Subantarctic Mode Water (SAMW). SAMW forms north of the Antarctic Circumpolar Current (ACC) and contains 40% of the total ocean inventory of anthropogenic carbon dioxide (Sabine *et al.,* 2004). SAMW is advected northwards into the Southern Hemisphere's subtropical gyres at the base of their ventilated layers. SAMW is the precursor to Antarctic Intermediate Water (AAIW), the low salinity layer underlying the thermocline throughout the southern hemisphere and parts of the Northern Hemisphere. Surface waters within and south of the ACC are strongly stratified due to a freshened surface layer from the summer melt of sea-ice and relatively high rainfall in this zone. Variations in salinity of this layer change the upwelling of deep water and formation of bottom waters in the Antarctic region.

SAMW in the Indian and Pacific Oceans and Tasman Sea has cooled and freshened on density surfaces since the 1960s (Fig. 5.3.8), consistent with the subduction of warmer surface waters in the SAMW source regions in the Indian and Pacific Oceans (Aoki *et al.,* 2003; Bindoff and McDougall, 1992; Johnson and Orsi, 1997; Wong *et al.,* 2001). The explanation for this paradoxical result is found in Bindoff and McDougall (1994, 2000). The SAMW isopycnals are now deeper implying that its heat content has increased. This is consistent with the warming of waters north of the Subantarctic Front and south of the centre of the subtropical gyres in the Southern Hemisphere since the 1950s (Levitus *et al*. 2005a; Willis *et al*. 2005).

Mid-depth waters of the Southern Ocean have warmed in recent decades. Temperatures increased near 900 m depth between the 1950s and the 1980s throughout most of the Southern Ocean (Gille, 2002; Aoki *et al*., 2003). The largest changes are found near the Antarctic Circumpolar Current, where the warming at 900 m depth is similar in magnitude to the increase in surface air temperatures (Gille, 2002). Analysis of altimeter data and Argo float profiles suggests that over the last ten years the zonally-averaged warming in the upper 400 m of the ocean near 40S is larger than at any other latitude (Willis *et al.,* 2005).

The major mid-depth water mass in the southern hemisphere, Antarctic Intermediate Water (AAIW), has been freshening since the 1960's (Aoki *et al.,* 2005; Bindoff and McDougall, 1992; Bindoff and McDougall, 2000; Wong *et al.,* 1999). The Atlantic freshening of AAIW is also supported by direct observations of a freshening of southern surface waters (Curry *et al.,* 2003). These changes in AAIW and NPIW suggest a global increase in the hydrological cycle with increased precipitation at high latitudes in the source regions of these key water-masses (Wong *et al.,* 1999) and as simulated in scenarios of climate change.

[INSERT FIGURE 5.3.8 HERE]

In the Upper Circumpolar Deep Water (UCDW) in the Indian and Pacific sectors of the Southern Oceans, temperature and salinity have been increasing and oxygen has been decreasing between the Subtropical Front at ~35°S and the Antarctic Divergence at ~60°S (Aoki *et al.,* 2005) (Fig. 5.3.8) consistent with the subduction or mixing of warmer and fresher surface waters with UCDW.

5.3.5.3 Variability in Antarctic regions

Although the Southern Ocean remains poorly observed, evidence for variability of the Southern Ocean is growing. The longest available records are from the Ross Sea, where high salinity shelf water has freshened by 0.003 per year over the last four decades, as a result of increased precipitation, reduced sea ice formation, and increased melting of ice shelves (Jacobs *et al*., 2002). The observed freshening of the Ross Sea shelf water may have contributed to an apparent shift in bottom water properties in the Australian-Antarctic Basin (Whitworth, 2002). The deep and bottom water properties of the Weddell Sea have also varied in the 1990s (Fahrbach *et al*., 2004; Robertson *et al*., 2002). Water at intermediate depth (the Warm Deep Water) became warmer and saltier in the early 1990s and has since cooled and freshened; dense Weddell Sea Bottom Water warmed by 0.01C per year from 1990 to 1996, after which temperatures have leveled off (Fahrbach *et al*., 2004). A combination of factors, including changes in atmospheric forcing, inflow to the Weddell gyre, and local effects such as iceberg calving, are believed to have contributed to the observed changes. Changes in bottom water properties have also been observed downstream of the source regions (Andrie *et al*., 2003; Hogg, 2001).

Antarctic Circumpolar Current

The transport of the ACC is about 130 x 10^6 m^3/s, with significant interannual variability. Measurements over 25 years at the South American choke point (Drake Passage) were reviewed by Cunningham (2003) and showed no evidence for a systematic trend in total volume transport between the 1970s and the present. However, Cunningham *et al*. (2003) also demonstrate that a large number of independent observations are required to determine a change in the mean transport of 10 or 5 percent.

5.3.6 Summary

The Southern Ocean contribution to the global overturning circulation has received increasing attention in recent years. The high latitude regions of the northern and southern hemispheres make roughly equal contributions to the ventilation of the deep ocean (Orsi *et al*., 2002; Orsi *et al*., 2001; Schmitz, 1995). Water mass transformations in the Southern Ocean contribute to the conversion of deep to intermediate waters to close the global overturning circulation associated with the formation and export of North Atlantic Deep Water (Rintoul *et al*., 2001; Sloyan and Rintoul, 2001a, 2001b; Speer *et al*., 2001). Export of relatively fresh mode and intermediate water from the Southern Ocean is the primary ocean pathway for the equatorward transport of freshwater in mid-latitudes of the Southern Hemisphere, as required to balance the excess of precipitation over evaporation at high southern latitudes (Ganachaud and Wunsch, 2003; Sloyan and Rintoul, 2001a; Wijffels *et al*., 2001). The export of mode water from the Southern Ocean also supplies nutrients to lower latitudes which support 75% of global primary production (Sarmiento *et al.,* 2004). The Southern Ocean and MOC therefore play a significant role in determining the mean structure of the global ocean circulation and the response of the climate system to changes in forcing.

5.4 Ocean Biogeochemical Changes

5.4.1 Introduction

The observed increase in atmospheric CO_2 (Chapter 2) and the observed changes in the physical properties of the ocean reported in this chapter can impact marine biogeochemical cycles (here mainly carbon, nutrients, and oxygen). In response to the atmospheric increase, CO_2 dissolves in the ocean. Changes in temperature and salinity impact the solubility and chemical equilibration of gases. Changes in circulation impact the supply of carbon and nutrients from below, the ventilation of oxygen-depleted waters, and the downward penetration anthropogenic carbon. The combined physical changes also impact biological activity, with further consequences for the biogeochemical cycles. In this section we report observed changes in biogeochemical cycles and assess their consistency with observed changes in physical properties.

5.4.2 Carbon

5.4.2.1 Change in air-sea CO_2 flux

The global mean air-sea CO_2 flux over the past decade is well constrained by observations. The most accurate method gives an oceanic CO_2 sink of 2.05 ± 0.5 PgC yr^{-1} for 1993–2003 based on atmospheric CO_2 and O_2 measurements combined to partition the land and ocean CO_2 sinks (Chapter 7). This method is consistent with other indirect methods based on observations of changes in carbon isotopes (Ciais *et al.,* 1995), atmospheric CO_2 inversions (Bousquet *et al.,* 2000)**,** and CFC changes (McNeil *et al.,* 2003), and with direct observations of the partial pressure of CO_2 (pCO_2) between the atmosphere and the ocean (Takahashi *et al.,* 2002) (see also Chapter 7 and TAR).

The trend in air-sea CO_2 flux is more difficult to constrain because of the large uncertainty in the data and methods (at best ±0.5 PgC yr^{-1}). Indirect methods based on O_2 measurements, atmospheric inversions, or CFC measurements suggest that the global CO_2 sink increased by 0.1 to 0.6 PgC yr^{-1} between the 1980s and 1990s (Le Quéré *et al.,* 2003; McNeil *et al.,* 2003). Direct measurements show that oceanic pCO_2 increased at a rate that roughly followed the atmospheric CO_2 during the past few decades; see Figure 5.4.3, summary table in Lefevre *et al*. (2005) and Takahashi *et al*. (2005). However direct pCO_2 measurements cannot be used to identify large-scale changes in the global CO_2 sink because the spatial coverage is incomplete, and because there are large decadal variability caused by changes in the underlying physics (Bates *et al.,* 2002; Takahashi *et al.,* 2003), precipitations (Dore *et al.,* 2003), and biological activity (Lefèvre *et al.,* 2005).

5.4.2.2 Total carbon change in the water column

The direct measure of dissolved inorganic carbon (DIC; includes CO_2 plus carbonate and bicarbonate) changes in the ocean reflects the anthropogenic CO_2 input plus the changes in carbon concentration due to changes in water masses and biological activity. To estimate the contribution of anthropogenic DIC alone, several corrections have to be applied. Changes in DIC were observed between the GEOSECS (1970s) and WOCE/JGOFS (1990s) surveys, from which an increase in anthropogenic DIC has been inferred down to a depth of 1100 m in the North Pacific (Peng *et al.,* 2003) and to a depth of 200–1200 m in the Indian Ocean (Peng *et al.,* 1998; Sabine *et al.,* 2004).

Indirect methods have been developed to estimate anthropogenic DIC based on measured DIC concentration corrected for organic matter decomposition and dissolution of carbonate minerals, minus an estimate of the pre-formed DIC. The pre-formed DIC is the DIC concentration of the water when it was last in contact with the atmosphere (Brewer, 1978). This method was applied by Chen (1993) using 930 profiles from the GEOSECS survey in the 1970s. The large preformed DIC component was estimated from empirical relationships with nutrients and temperature and an assumption that the deep ocean contains no anthropogenic CO_2. A global DIC increase of 90 ± 18 PgC was estimated between 1750 and 1978, including a contribution from marginal seas (Chen, 1993). The reported error in this estimate was criticized as being too low because of the errors in the estimate of pre-formed DIC and because of the limited quality and number of observations (Gruber *et al.,* 1996).

Gruber *et al*. (1996) improved the indirect method by defining a quasi-conservative tracer, C*, that separates the preformed DIC into an equilibrium component that can be calculated from thermodynamics, and a substantially smaller disequilibrium component. The C* approach is not as strongly influenced by mixing as previous approaches. This improved method was applied by Sabine *et al*. (2004) using 9618 profiles from the WOCE/JGOFS survey in the 1990s. A global DIC increase of 118 ± 19 PgC was estimated between 1750 and 1994 (Figure 5.4.1). Results are consistent with the direct measure of DIC changes (Peng *et al*., 1998; Peng *et al*., 2003). The uncertainty in the global inventory estimated by Sabine *et al*. (2004) is based on uncertainties in the anthropogenic DIC estimates and mapping errors. The potential biases in the technique that have been identified suggest an overestimate of the global uptake by ~10% mostly caused by assumptions about constant air-sea pCO_2disequilibrium, although not all potential sources of bias have been quantified (see Appendix on methods and errors).

[INSERT FIGURE 5.4.1 HERE]

Because of the limited mixing rate of the ocean, more than half of the anthropogenic carbon can still be found in the upper 400 meters, with deeper penetration in the Atlantic compared to other basins (Figure 5.4.2). The uptake of anthropogenic DIC by the ocean is only about 10% of the potential uptake, which can be computed by assuming that the entire ocean is at equilibrium with atmospheric CO_2 (Figure 5.4.2).

[INSERT FIGURE 5.4.2 HERE]

The increase in global anthropogenic carbon of 28 ± 26 PgC between 1978 (Chen, 1993) and 1994 (Sabine *et al*., 2004) is consistent with the indirect estimates of the ocean CO_2 sink (Section 5.4.2.1), although the uncertainty in the methods and data is large and may be underestimated in the early estimate.

The fraction of the CO_2 emissions that the ocean has taken up (the uptake fraction) was lower during 1993–2003 (0.36 ± 0.09) compared to 1750–1994 (0.42 ± 0.07), but still within the uncertainty (Table 5.4.1). This is consistent with our understanding that the ocean CO_2 sink is limited by the rate at which anthropogenic DIC is transported from the surface to the deep ocean, and with the non-linearity in carbon chemistry that reduce the CO_2 uptake capacity of water at high concentrations (Sarmiento *et al.,* 1995) .

Table 5.4.1. Fraction of CO_2 emissions taken up by the ocean for different time periods.

Time Period	Oceanic Increase	Net CO_2 Emissions[a]	Uptake Fraction	Reference
1750–1994	118 ± 19	283 ± 19	0.42 ± 0.07	Sabine *et al*., 2004
1993–2003	20.5 ± 5	57.6 ± 5.1	0.36 ± 0.09	Chapter 7

Notes:
(a) The net CO_2 emissions include emissions from fossil fuel burning, cement production, land use change, and the terrestrial biosphere response. It is equivalent to the sum of the atmospheric and oceanic increase.

5.4.2.3 *Ocean acidification by carbon dioxide*

CO_2 is a weak acid[1]. As CO_2 increases, pH decreases (i.e., acidity increases). pH can be computed from measurements of DIC and alkalinity. The input of anthropogenic DIC at the surface is estimated to have caused a decrease in pH by 0.1 pH units over the global ocean between 1750 and 1994. This calculation assumes that alkalinity and temperature remained constant. It is consistent with results from time-series stations (Figure 5.4.3). Changes in surface temperature may have induced an additional decrease in pH by <0.01 units.

[1]Acidity is a measure of the concentration of H^+ ions and is reported in pH units, where pH = –log(H^+). A pH decrease of 1 unit means a 10-fold increase in the concentration of H^+, or acidity.

[INSERT FIGURE 5.4.3 HERE]

5.4.2.4 Change in carbonate species

The injection of CO_2 in the ocean causes a shift in the distribution of carbon species. The availability of carbonate is particularly important because it controls the maximum amount of CO_2 that the ocean is able to absorb. Marine organisms use carbonate to produce shells of calcite and aragonite ($CaCO_3$). $CaCO_3$ dissolves either when it sinks below the saturation horizon (the shallowest depth where $CaCO_3$ is undersaturated) or under the action of biological activity.

Shoaling of the aragonite saturation horizon was measured in all ocean basins (Feely and Chen, 1982; Feely *et al*., 2002; Sabine *et al*., 2002; Sarma *et al*., 2002). Feely *et al*. (2004) calculated a shoaling of the aragonite saturation horizon between 1750 and 1994 by 30 to 200 m in the eastern Atlantic (50°S–15°N), the North Pacific, and in the North Indian Ocean, and a shoaling of the calcite saturation horizon by 40–100 m in the Pacific (north of 20°N). This calculation is based on the anthropogenic DIC increase estimated by Sabine *et al,* (2004), on a global compilation of biogeochemical data, and on carbonate chemistry equations. Shoaling of the aragonite and calcite saturation horizon is due to the combined effects of CO_2 increase and to respiration processes in the intermediate waters. Sarma *et al*. (2002) further reported measured increase in total alkalinity (primarily controlled by carbonate and bicarbonate) at the depth of the aragonite saturation horizon between 1970 and 1990, which is consistent with the calculated increase in $CaCO_3$ dissolution as a result of the shoaling of the aragonite saturation horizon. This study suggests that more than half of the observed changes in alkalinity are due to changes in biological activity.

5.4.3 Oxygen - Biogeochemical Aspects

O_2 is affected by the same physical processes that impact CO_2, but in contrast to CO_2, it is not affected by changes in its atmospheric concentration (which are only 10^{-6} its mean concentration). Thus O_2 has the potential to identify changes in biogeochemistry that would be caused by changes in physical or biological processes within the ocean. Furthermore, atmospheric O_2 is used to separate the land and ocean CO_2 sinks, a method which requires an estimate of the changes in oceanic O_2 concentration.

Decadal variations of ±0.5 umol/kg in the O_2 concentration of the upper 100 m of the world ocean were observed for the 1956–1998 period, with no clear trends (Garcia *et al*., 2005). 580,000 oxygen profiles were used in this analysis, with the largest data density during 1965–1990. The accuracy of the O_2 measurements in the early decades is difficult to determine. There were biases reported with the use of metal flasks prior to about 1960 (see Appendix on methods and errors), but the surface measurements from the GEOSECS and WOCE surveys are though to be accurate within the given uncertainty.

The surface changes in O_2 concentration are paralleled by opposite changes in Apparent Oxygen Utilisation (AOU). AOU is the difference between the observed O_2 and the O_2 that is at equilibrium with local the temperature, in effect removing the direct impact of temperature on O_2. AOU changes can only be caused by changes in ventilation or biological activity. However AOU changes in the surface ocean are difficult to interpret because O_2 equilibrates fast with the atmosphere. Thus any changes in ventilation or biological activity would be damped by the equilibration with the atmosphere. A similar analysis of O_2 observations in the deeper ocean does not yet exist.

In the ventilated thermocline (~100 to 1000 m), a large decrease in the O_2 concentration has been observed between about the early 1970s and late 1990s in the North and South Pacific, North Atlantic, and Southern Indian Oceans (see summary table in Emerson *et al*., 2004). The reported O_2 decrease ranges from 0.1 to 6 umol/kg yr^{-1}, superposed on decadal variations of ±2 umol/kg yr^{-1} (Andreev and Watanabe, 2002; Andreev and Kusakabe, 2001; Ono *et al*., 2001). The reported O_2 decrease ranges from 0.1 to 6 umol/kg yr^{-1}, superposed on decadal variations of ±2 umol/kg yr^{-1} (Johnson and Gruber, 2005; Ono *et al*., 2001; Watanabe *et al*., 2001). The recent O_2 decrease is paralleled by an increase in AOU and is consistent with changes in ventilation, although changes in biological activity cannot be ruled out (Figure 5.4.4 from Deutsch *et al*., 2005).

[INSERT FIGURE 5.4.4 HERE]

5.5 Sea Level

5.5.1 Introductory Remarks

Present-day sea level change in response to global warming is a topic of considerable interest because of its potential impact on human populations living in coastal regions and on islands. Besides, because sea level change integrates non-linear coupled responses of several components of the earth's system (i.e., oceans, atmosphere, ice sheets and glaciers, land water reservoirs, mantle and crust), measuring sea level variations and studying processes that cause them is highly interdisciplinary. This section will focus on global and regional sea level variations, over time spans ranging from the last decade to the past century; see Section 6.5.7 for sea level change on longer time-scales.

Depending on the measurement technique, observations of sea level change may be either relative (to the earth's crust) or absolute (i.e., referred to the earth's center of mass or to the geoid). While satellite altimetry provides absolute sea level information, tide gauges provide sea level relative to the crust. In some instances, vertical land motion is sufficiently well-known, hence tide gauge-based measurements corrected for land motion also provide absolute sea level. When comparing with climate-related contributions, only absolute sea level change is considered.

Two classes of processes are responsible for absolute global mean sea level variations. Firstly, temperature and salinity variations of ocean waters cause water density (specific volume) change. Sea level changes due to density variations are called steric. Changes induced by temperature only are called thermosteric while changes induced by salinity are called halosteric. Secondly, exchange of water between oceans and other reservoirs (e.g., ice sheets, mountain glaciers, land water reservoirs and atmosphere) causes ocean mass change.

Sea level change is not geographically uniform because, at the regional scale, several processes can affect sea level: ocean circulation changes, atmospheric loading, geoid change, etc.; see Section 5.5.4.1 below. These processes have however no contribution in terms of global mean.

The chapter on sea level change of the Third Assessment Report (TAR) (Church *et al.,* 2001) provided estimates of contributions to 20th century sea level rise, based mostly on climate models. The most uncertain contribution reported in the TAR was the change in terrestrial water storage that results from human activities, in the range of –1.1 to +0.4 mm yr^{-1} with a median value of –0.35 mm yr^{-1} (i.e., corresponding to sea level drop). The sum of these contributions ranges from –0.8 to 2.2 mm yr^{1}, with a median value of 0.7 mm yr^{-1}. For 20th century sea level rise, Church *et al*. (2001) adopted as a best estimate a value of 1.5 ± 0.5 mm yr^{-1} and noted that the sum of climate-related components (0.7 mm yr^{-1}) is low compared to observations. In effect, the observed value was more than twice as large as the TAR's estimate of the total climate contributions (even though uncertainty of the latter was quite large). It thus appeared that either the climate-related processes causing sea level rise had been underestimated or the rate of sea level rise observed with tide gauges was biased toward values too high. Munk (2002) referred to this as "Enigma".

Since the publication of the TAR, a number of new results have been reported in the recent literature. Sea level rise measured during the 1990s by Topex/Poseidon satellite altimetry is about 3 mm yr^{-1}, a value significantly larger than current estimates of the 20th century sea level rise that re-opens the question of sea level rise acceleration. It is however unclear whether this indicates an acceleration or whether this reflects decadal variability. Near-global ocean temperature and salinity data have been recently made available for the last 50 years, allowing the first observations-based estimate of the steric contribution to past decades sea level rise. Sea level change is highly nonuniform spatially, and observed patterns of sea level change are highly correlated to those of thermal expansion. Direct estimates of thermal expansion and land ice melting for the 1990s indicate that about 85% of the observed rate of sea level rise can be explained.

In this section, we summarize the current knowledge of present-day sea level rise. The observational results will be reviewed, followed by our current interpretation of these observations in terms of climate processes. Whereas estimates of sea level rise in the past few decades have not much changed since the TAR, new results on climatic contributions allow us to move closer to resolving the ‘enigma’ of the level change, at least for the recent years.

5.5.2 Observations of Sea Level Changes

5.5.2.1 Global sea level variations

Table 11.9 of the TAR listed 20th century global and regional sea level trends estimated by different analysts of 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, most long sea level records stemming from the Northern Hemisphere, and most inevitably from continental coastlines rather than ocean interiors. Recent studies include those of Holgate and Woodwort (2004), who estimated a rate of 1.7 ± 0.2 mm yr^{-1} for global-coastal sea level change during the period 1948–2002 based on data from 177 stations divided into 13 regions, and Church *et al*. (2004) (discussed further below), who determined a global rise of 1.8 ± 0.3 mm yr^{-1} during 1950–2000 based on a combined analysis of tide gauge and altimeter data. Church and White (2005) concluded from an analysis of tide-gauge observations that over the period 1870–2000 there is a clear acceleration in sea level rise. A new study of New Zealand data suggests a rise of 2.1 mm yr^{1} for the past 100 years (Hannah, 2004), while a regional study for the Russian Arctic based on 71 stations with data 1954–1989 obtained an overall rate of 1.85 mm yr^{-1} (Proshutinsky *et al.*, 2004). All of these reported rates have been adjusted for Glacial Isostatic Adjustment (GIA) (see below).

While recently published estimates of the 20th century rate of change remain within the range of the TAR (i.e., 1–2 mm yr^{-1}), there is an increasing consensus that the best estimate lies nearer to 2 than 1 mm yr^{-1}. A critical issue concerns how the records are adjusted for vertical movements of the land upon which the tide gauges (sea level stations) are located. Peltier (2001) demonstrated that in analyses which employ extrapolations of geological data obtained near the gauges, adjusted rates could be underestimated by several tenths of mm yr^{-1} if GIA is the only geological process involved. This argument applies particularly to some reported European rates, e.g., Shennan and Woodworth (1992), Europe being one of the regions having sufficient geological data to make the ‘direct correction’ method possible.

The TAR mentioned the developing geodetic technologies (especially the Global Positioning System, GPS) which hold the promise for measurement of the rates of vertical land movement at tide gauges, no matter if those movements are due to GIA or other geological processes. However, systematic problems with those techniques have not been resolved sufficiently such that measured rates of land movement are as yet available.

5.5.2.1.1 Satellite-based sea level change during the last decade: altimetry results

The era of precision satellite altimetry began with the launch of Topex/Poseidon (T/P) in 1992, although the road to success was paved by earlier less accurate missions such as GEOS-3, Seasat, Geosat, and ERS-1. T/P ushered in a new paradigm in satellite altimetry, largely due to advances in the instruments, the orbit determination, and the instrument calibration. With the launch of the Jason mission in 2001, the decade-long time series of precision satellite altimetry measurements is now being seamlessly extended. Estimating global mean sea level variations from the T/P measurements is a reasonably straightforward exercise if careful attention is paid to the measurement corrections and the instrument calibration. T/P and Jason make these measurements along a ground track that repeats once every 10 days. Global mean sea level can be computed at 10 day intervals by averaging the measurements over the ocean while accounting for the spatial distribution of the data using area-dependent weighting (Nerem and Mitchum, 2001). Each 10 day estimate of global mean sea level has a precision of approximately 5 mm. Numerous papers have been published on the scientific results (see Cazenave and Nerem, 2004, for a review), which currently show a rate of sea level rise of +3.1 ± 0.4 mm yr^{-1} over 1993–2004 (Figure 5.5.1). The observations show a 15 mm rise and fall of mean sea level accompanying the 1997–1998 El Niño-Southern Oscillation (ENSO) event. A significant fraction of this change has also been shown to arise from changes in the Southern Ocean (Cabanes *et al.*, 2001).

[INSERT FIGURE 5.5.1 HERE]

The accuracy needed to compute mean sea level change pushes the altimeter measurement system to its performance limits, and thus care must taken to ensure that the instrument is precisely calibrated (see

Appendix 5.A.4.1). The tide gauge calibration method developed by 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). Future monitoring of the tide gauges using geodetic techniques such as GPS and DORIS will be critical if the error in the instrument calibration is to be reduced. In summary, the altimetric results are considered to be extremely robust, and the estimate of sea level rise of 3.1 ± 0.4 mm yr^{-1} over the last decade is reliable within these error bars.

[INSERT FIGURE 5.5.2 HERE]

An important result of T/P altimetry is the mapping of the geographical distribution of sea level change (Figure 5.5.2). While in tide gauge-derived sea level studies, most investigators assumed uniform sea level change, now satellite altimetry has provided for the first time unambiguous evidence of regional variability of sea level change, with some regions exhibiting sea level trends about 10 times the global mean. It is in the western Pacific and eastern Indian Oceans that sea level rise shows the highest magnitude. It is also worth noting that the whole Atlantic Ocean shows sea level rise during the past decade. Besides, Figure 5.5.3 shows that sea level has been dropping in some regions (eastern Pacific and western Indian Oceans), even though in terms of global mean, sea level has been rising.

5.5.2.1.2 Past reconstructions based on Topex/Poseidon altimetry and tide gauges (last 50 years)

In an attempt to understand and reconcile the wide range of tide gauges-based sea level rise estimates, there is now a series of attempts 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. These sea-level reconstructions use the short altimeter record to determine the modes (empirical orthogonal functions, EOFs) of sea-level variability and the tide-gauge data to estimate the evolution of the amplitude of the EOFs over time. Chambers *et al*. (2002) used the EOF projection technique (similar to that used by Smith *et al*. (1996) for sea surface temperatures) but with the long-term trends removed from the tide gauge data; i.e. they focused on the variability of global mean sea level. Error assessments indicated the accuracy of the annual mean global mean sea level from the tide gauge reconstruction was 2–4 mm. Several El Niño/La Niña events are evident in the global mean sea level time series and there was significant low-frequency variability. Church *et al*. (2004) used the optimal interpolation technique developed by Kaplan *et al*. (2000), and used by Rayner *et al*. (2003) for sea-surface temperature reconstructions, to combine EOFs from nine years of T/P data with the tide gauge data. The method assumes that the geographical pattern of decadal sea level trends can be represented by a superposition of the patterns of variability which are manifest in interannual variability. As the tide gauges are all on different and unknown datums, they used first differences between monthly values of mean sea level obtained from the PSMSL web site (Woodworth and Player, 2003) and then integrated backwards in time to get sea-level changes. These additions to the reconstruction technique allowed them to focus on global and regional mean sea-level trends and they found good agreement with the satellite altimeter results over the period of overlap between the two data sets (January 1993 to December 2000). Over the 51-year period from January 1950 to December 2000, they found a global mean sea-level rise of 1.8 ± 0.3 mm yr^{-1} (Figure 5.5.3). Both the average rate of rise and the decadal variability are similar to the results of Holgate and Woodworth (2004) based on an average of decadal tide-gauge trends (see Figure 5.5.3).

[INSERT FIGURE 5.5.3 HERE]

The trends in the EOF amplitudes (and the implied global correlations) are responsible for a spatially variable rate of sea-level rise over this 51-year period. The results help reconcile and substantiate earlier, apparently inconsistent, estimates of regional variations in tide gauge-based sea-level rise. For example, the minimum in rise along the north-west Australian coast is consistent with the results of Lambeck (2002) and smaller rates of sea-level rise and indeed sea-level fall off north-western Australia over the last few decades. Also, for the North Atlantic Ocean, the rate of rise reaches a maximum (over 2 mm yr^{-1}) in a band running east-north-east from the US east coast. The trends are lower in the east Atlantic than the west, as suggested by Woodworth *et al*. (1999), Lambeck *et al*. (1998) and Mitrovica *et al*. (2001).

5.5.2.1.3 Interannual/decadal variability and recent accelerations in sea level

Sea level records contain a considerable amount of interannual and decadal variability, the existence of which is a major reason why no definite long term acceleration of sea level has been identified using 20th century data alone (Douglas, 1992; Woodworth, 1990). Another possibility is that the sparse tide-gauge network may have been inadequate to detect it if present (Gregory *et al.*, 2001). Interannual variability is coherent throughout extended parts of the ocean. This is the case for that related to ENSO as shown clearly in altimetric maps of sea level anomalies (Cazenave and Nerem, 2004). Over the past few decades, the time series of the first EOF of Church *et al*. (2004) represents El Niño variability and there is a good (negative) correlation with the Southern Oscillation Index.

As far as trends are concerned, Holgate and Woodworth (2004) concluded that the 1990s experienced one of the fastest recorded rates of global-coastal sea level rise (~4 mm yr^{-1}), slightly higher than the altimetry-based open ocean sea level rise (3 mm yr^{-1}). However, these analyses also show that some previous decades experienced comparably large rates of sea level rise. This recent higher value can go some way to explaining why the rate of truly-global sea level change observed in the decade by satellite altimetry is closer to 3 mm yr^{-1} than the 1–2 mm yr^{-1} reported for the 20th century as a whole obtained from gauge data. White *et al*. (2005) confirmed the larger sea level rise during the 1990s around continental coastlines compared to the open ocean but concluded that over a much longer period (the last 50 years) the coastal and open ocean rates of change were essentially the same.

5.5.2.1.4 Long term accelerations

The longest records available from Europe and North American were shown to contain accelerations of order 0.4 mm yr^{-1} per century between the 19th and 20th century (Ekman, 1988; Woodworth *et al.*, 1999). These data were consistent with available information from North American geological sources, with an inference that the onset of acceleration occurred during the 19th century. Recently, Church and White (2005) applied their reconstruction method (see 5.5.2.1.2) to provide a sea level curve back to 1870. They find a significant acceleration, of 0.012 ± 0.006 mm yr^{-2}, over the period 1870–2000, slightly lower than accelerations in climate change model estimates for the 20th century.

There have been several recent analyses of either archaeological or geological data in combination with nearby tide gauge information. The use of proxy sea level data from archaeological sources is well established in the Mediterranean, where many Roman and Greek constructions are relatable to the level of the sea. Lambeck *et al*. (2004) used sea level data derived from Roman fish ponds, considered to be a particularly reliable source of such information, together with nearby tide gauge records and concluded that the onset of the modern sea level rise occurred ~100 ± 53 years before present (year 2000). Donnelly *et al*. (2004) and Gehrels *et al*. (2004) employed geological data from Connecticut, Maine and Nova Scotia salt-marshes together with nearby tide gauge records to demonstrate that the sea level rise observed during the 20th century was significantly in excess of that averaged over the previous several centuries.

The importance of further 'data archaeology' cannot be stressed too highly, as it is vital to place the 20th century sea level rise in a proper historical context. Small amounts of additional 18th and 19th tide gauge information occasionally become available. For example, Hunter *et al*. (2003) estimated sea level change averaged over the one and a half centuries since 1841 at Port Arthur, Tasmania, to be 1.0 ± 0.3 mm yr^{-1} using tide gauge data collected over a two year period at the penal settlement. These data were lost to science, until they were rediscovered recently after hard work in archives. Without tide gauge, archaeological or geological data, one can resort to the use of various types of proxy-information. In another example, Camuffo and Sturaro (2003) estimated changes in Venice during the last three centuries by comparison of the heights of algae evident in paintings by Canaletto and his pupils to heights observed today. This result is likely to be primarily of local, rather than regional or global, relevance. Nevertheless, it demonstrates the sort of ingenuity which must be applied to the acquisition of longer-term sea level information.

5.5.2.2 Regional sea level change

In this section, we present a few examples of regional sea level change, focusing on a semi-enclosed sea (the Mediterranean Sea), on Arctic regions (for which results are available for the first time) and on small Pacific Islands which are the subject of much concern in view of their potential vulnerability to sea level rise.

5.5.2.2.1 Sea level changes in the Mediterranean Sea

The Mediterranean Sea is a semi-enclosed sea in which the loss of water through the evaporation minus precipitation and river runoff is balanced by influx of water through the Gibraltar Strait. As a result, in addition to basin wide steric variations, addition of water mass and oceanic circulation (inclusive of changes in intermediate and deep water formation), sea level depends on the hydraulic control of the water exchange at the Strait of Gibraltar.

Only a few long, good quality sea level records spanning to the beginning of the 1900s exist in the Mediterranean Sea and these are located at the Northern coasts of the Western Mediterranean (Marseille and Genoa) and at the northern coasts of the Adriatic Sea (Trieste) (Tsimplis and Baker, 2000). The sea level trends for these three stations are presently in the range 1.1–1.3 mm yr^{-1}, thus lower than the estimated global value for sea level rise. Besides this long term trend, sea level in the Mediterranean is affected by interdecadal/decadal variability. For example, between 1960 and the beginning of the 1990s sea level in the Mediterranean Sea was either not changing or decreasing (Tsimplis and Baker, 2000) due to atmospheric pressure changes during the winter period (Tsimplis and Josey, 2001; Woolf *et al.*, 2003) as well as temperature reduction and salinity changes linked to the North Atlantic Oscillation (NAO) (Tsimplis and Rixen, 2002). During the 1990s, fast sea level rise was observed by T/P at the Eastern Mediterranean Sea while sea level fall was observed in the Ionian Sea (Cazenave *et al.*, 2001; Fenoglio-Marc, 2002). During the same period of time a reduction in the sea level gradient across Gibraltar Strait has been observed and the change was suggested as caused by varied hydraulic conditions in the Strait (Ross *et al.*, 2000) or by changes in the density difference between the Mediterranean and the Atlantic (Brandt *et al.*, 2004). The extent to which the Mediterranean Sea can have long-term sea level variability different from the global ocean remains an open question. The whole question turns around the gradients which can be sustained across Gibraltar Strait coupled with the density changes within the basin.

5.5.2.2.2 Arctic Ocean

Proshutinsky *et al*. (2004) have analyzed monthly relative sea level data (1954–1989) from the 71 tide gauges in the Barents, Kara, Laptev, East Siberian and Chukchi Seas in order to estimate the rate of sea level change and major factors responsible for this process in the Arctic Ocean. It was found that the Arctic Ocean sea level time series have well pronounced decadal variability which corresponds to the variability of the NAO index. A similar conclusion was later published by Hughes and Stepanov (2004). These studies concluded that during the period 1954–1989 the average rate of relative sea level rise over the seas of the Russian Arctic has been 1.85 mm yr^{-1}. The contribution to the observed rate of sea level rise from the steric (temperature and salinity) effect was estimated as 0.64 mm yr^{-1}. In the Arctic Ocean, changes in salinity are more important for sea level variability than changes in temperature, and the combination of freshening of the Arctic Seas with warming and salinization of the Atlantic layer therefore leads to the rise of sea level along coastlines and the fall of sea level in the central parts of the Arctic Basin. The contribution of atmospheric loading to the Russian Arctic Ocean sea level rise was estimated as 0.56 mm yr^{-1}. The estimated rate of sea level rise due to the effect of wind was 0.18 mm yr^{-1} but it varied significantly from region to region. In the Arctic, this effect is due to the gradual decrease of the sea level atmospheric pressure over the Arctic Ocean and therefore to the more strongly cyclonic atmospheric circulation.

In summary, by subtracting the influence of these factors from the observed regional sea level trends, Proshutinsky *et al*. (2004) speculated that the residual term of the sea level rise balance assessment (0.48 mm yr^{-1}), was associated with the increasing of the Arctic Ocean and global ocean mass due to melting of ice caps and small glaciers and with adjustments of the Greenland and Antarctica ice sheets to the observed climate change.

5.5.2.2.3 Pacific and Indian Oceans

The Pacific Ocean region is the centre of the strongest interannual variability of the climate system, the coupled ocean-atmosphere ENSO phenomenon. On time scales of months to years and space scales of several hundred kilometers and longer, the largest sea-level variations occur in the equatorial Pacific Ocean (mostly related to ENSO events). During El Niños, sea level is anomalously high (by tens of centimeters) in the eastern tropical Pacific and low in the western tropical Pacific.

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 relative sea-level rise was 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 was 0.7 mm yr^{-1}. Using a slightly updated (and therefore longer) data set for a similar set of gauges, Church *et al*. (Church *et al.,* 2005b) found a similar average, 0.9 mm yr^{-1}. However, both of these data sets contain a large range of rates of relative sea-level change, presumably as a result of poorly quantified vertical tectonic motions.

[INSERT FIGURE 5.5.4 HERE]

An example of the large interannual variability in sea level is Kwajalein (8º44'N, 167º44'E) (Marshall Archipelago). Here, the relative sea-level data, the reconstructed sea level of Church *et al*. (2005b) and the short satellite altimeter record (Figure 5.5.4) indicate interannual variations associated with ENSO events are greater than 20 cm. The Kwajalein data also suggest increased variability in sea level after the mid-1970s, consistent with the trend to more frequent, persistence and intense ENSO events since the mid 1970's (Folland *et al.,* 2001). For the Kwajalein record, the rate of relative sea level rise is 1.3 ± 0.4 mm yr^{-1} (all error estimates are one standard deviation) and after correction for GIA land motions and isostatic response to atmospheric pressure changes is 1.9 ± 0.4 mm yr^{-1}. However, the uncertainty of 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 at Tuvalu Islands (Western Pacific) has been the subject of intense interest as a result of Tuvalu Islands' low lying nature and reports that flooding is becoming increasingly common. There are two records available at Funafuti, Tuvalu; the first record commences in 1977 and the second (with rigorous datum control) in 1993. The most thorough analysis of the Tuvalu sea-level data is described by Hunter (2004) (the original report is available at *http://www.antcrc.utas.edu.au/~johunter/tuvalu.pdf*). Leveling data since 1993 suggest that the first record is contaminated by subsidence. After allowing for this subsidence, Hunter (2004) provides an estimate of relative sea-level rise at Tuvalu of 1.2 ± 0.8 mm yr^{-1}, in agreement with the reconstructed rate of sea level rise by Church *et al*. (2005b).

5.5.2.3 Changes in extreme sea level

Impacts of sea level change on the coast occur primarily via the extreme levels rather than as a consequence of mean sea level changes. Studies of variations in extreme sea levels during the 20th century based on tide gauge data are fewer than those of changes in mean sea level for several reasons. Firstly, records of century-time scale are few; secondly, the hourly sampling interval, which is often employed, has clearly a poorer chance of recording the true extreme than higher frequency sampling. Finally, there is the problem of different authors using different types of high water extreme such as annual maximum high water, annual maximum surge, annual maximum surge-at-high-water, surge at annual maximum high water etc. Annual maximum surge is clearly the best indicator of climatic trends; however, for study of long records extending back to 19th century or before, in which high waters were recorded rather than the full tidal curve, one is forced to use a parameter such as annual maximum surge at high water.

Studies of the longest records of extremes are inevitably restricted to a small number of locations. Woodworth and Blackman (2002) conducted a study of extremes at Liverpool since 1768 finding annual maximum high water and surge at annual maximum high water to vary considerably from year to year, but to exhibit no long term change. On the other hand, values of annual maximum high water and annual maximum surge at high water were found to be 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. Bouligand and Pirazzoli (1999), while analysing the tide gauge record at Brest spanning from 1860 to 1994 found an increasing trend for annual maxima and 99th percentile of surges; however, when analysed for the period 1953–1994, a decreasing trend was noticed, which is attributed to a decrease in the frequency and duration of storms during this period. Bromirski *et al*. (2003) studied non-tidal residuals' (essentially same as 'surges') at San Francisco since 1858, concluding that extreme winter residuals had exhibited a significant increasing trend since about 1950. This increasing trend is attributed to an increase in storminess during this period. Zhang *et al*. (2000) studied trends in annual maximum high water since 1900 at ten stations along the east coast of USA and found that the rise in the level of extremes closely followed that 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 extreme dailies at Honolulu, (the highest ever value being due an anti-cyclonic eddy system in 2003) if measured relative to a fixed datum, 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 the recent decades (Woodworth and Blackman, 2004) showed that there is evidence for an increase in extreme high water 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 such as ENSO in the Pacific, NAO in the Atlantic and IOD in the Indian Ocean.

5.5.3 Ocean Density Changes - by Thermal and Haline Expansion

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 3D distribution of ocean temperature change. If the equation of state of sea water were linear, the effect on global average sea level of redistributing the heat and salt within it, while not changing their integrals, would be zero. Owing to the non-linearity, mixing of sea water with differing temperature, salinities and pressure may result in small global average sea level changes (Gille, 2004). Gregory and Lowe (2000) found this effect to be about 1% of the global average thermosteric sea level rise in a simulation of the 21st century.

Antonov *et al*. (2002) attribute about 10% of the global average steric sea level rise during recent decades to halosteric expansion due to the dilution by added freshwater. While it is of interest to quantify this effect, note that this term is compensated by a decrease in volume of the added water when its salinity is raised to the mean ocean value; the compensation is exact for a linear state equation. Hence this term cannot be counted separately in global sums from the volume of added freshwater (which Antonov *et al*. also calculate, see Section 5.5.5.1).

In regional sea level change, thermosteric and halosteric contributions can however be comparably important. Regional changes of absolute sea level are closely coupled to ocean circulation changes via geostrophy. To the extent that the barotropic circulation (or equivalently the level of no motion) is unchanged, regional steric sea level change patterns must hence coincide with the patterns of absolute sea level change.

Over the previous decades, temperature and salinity have been collected by buoys, commercial ships and oceanographic cruises. However, these data suffer considerable inhomogeneity both in time and space. Thus for easier handling of these observations, data interpolation at standard ocean depths and geographical positions is indicated. Recently, Levitus *et al*. (2000), Levitus *et al*. (2005a), Ishii *et al*. (2003) and Ishii *et al*. (2005) provided global gridded temperature data sets for 1950–2003 based on objective analysis methods applied to the raw data. Analysis of the last half-century of temperature profiles indicates that the ocean has warmed in all basins (see Section 5.2). The rate of averaged (84S–90N) thermosteric sea-level rise caused by ocean heating is estimated to be 0.40 mm yr^{-1} over 1955–1995 (Antonov *et al*., 2005), based on pentadal temperature data down to 3000 m. For the 0–700 m layer, the averaged (84S–90N) thermosteric trends (based on yearly mean temperature data from Levitus *et al*., 2005a) are 0.34 mm yr^{-1} and 1.23 mm yr^{-1} for the 1955–2003 and 1993–2003 periods, respectively (Antonov *et al*., 2005). For the same two periods and same depth range (0-700 m), the mean thermosteric rate based on monthly ocean temperature data from Ishii *et al*. (2005) are 0.38 ± 0.04 mm yr^{-1} and 1.8 ± 0.2 mm yr^{-1}. Figure 5.5.5 shows the thermosteric sea level curve over 1950–2003 for both the Levitus and Ishii data sets.

[INSERT FIGURE 5.5.5 HERE]

[INSERT FIGURE 5.5.6 HERE]

For the last decade, there is another estimate of the thermosteric sea level rise (Willis *et al*., 2005). These authors used a combination of approximately 1,000,000 in situ temperature profiles and satellite-based measurements of altimetric height. Between 1993 and 2003, they found that thermosteric sea level rose at a rate of 1.6 ± 0.3 mm yr^{-1} (Figure 5.5.6). By including the satellite data, the authors were able to reduce 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.

Table 5.5.1 summarizes the various estimates of the steric sea level rates available for the past 50 years and last decade. For 1993–2003, the mean steric rate is 1.55 ± 0.3 mm yr^{-1}.

Table 5.5.1. Recent estimates for steric sea level trends from different studies.

Reference	Steric sea level change with rms errors (mm yr^{-1})	Period	Depth range (m)	Data
Antonov *et al*. (2005)	0.40 ± 0.05	1955–1998	0–3000	Levitus *et al*. (2005b)
Antonov *et al*. (2005)	0.34 ± 0.04	1955–2003	0–700	Levitus *et al*. (2005b)
Ishii *et al*. (2005)	0.38 ± 0.04	1955–2003	0–700	Ishii *et al*. (2005)
Antonov *et al*. (2005)	1.23 ± 0.2	1993–2003	0–700	Levitus *et al*. (2005b)
Ishii *et al*. (2005)	1.8 ± 0.2	1993–2003	0–700	Ishii *et al*. (2005)
Willis *et al*. (2005)	1.6 ± 0.3	1993–2003	0–750	Willis *et al*. (2005)

The Antonov *et al*. (2005), Lombard *et al*. (2005a) and Willis *et al*. (2005) results contradict those of an earlier study by Cabanes *et al*. (2001), which was based on the former Levitus *et al*. (2000) upper ocean temperature data set. In the Cabanes *et al*. (2001) study, authors found a rapid rise in thermosteric sea level during 1997 and 1998 that was large enough to account for all of the sea level rise observed by T/P. The new Levitus *et al*. (2005b) temperature data now lead to reduced ocean thermosteric sea level rise during the 1997–1998 ENSO event, which agrees well with the Willis *et al*. (2005) and Ishii *et al*. (2005) results.

[INSERT FIGURE 5.5.7 HERE]

The rate of thermosteric sea-level rise is clearly not constant in time (Figure 5.5.5). There are large decadal fluctuations; a rise of more than 20 mm occurred from the late 1960s to the late 1970s with a smaller drop afterwards. Another large rise occurred in the 1990s and appears to be continuing. Moreover, as for the T/P-based observed sea level trends, the global distribution of thermosteric sea-level trends are not spatially uniform. This is illustrated by Figure 5.5.7 which shows the spatial distribution of thermosteric sea level trends over the 50-year (1950–2000) period. Some regions experience sea-level rise while others experience a decrease in sea level, often with rates that are several times the global mean (Lombard *et al.,* 2005a). However, the patterns of thermosteric sea level rise over the 40-year period are different than those seen in the 1990s (Figure 5.5.8). This occurs because the spatial patterns, like the global average, are also subject to decadal variability (Lombard *et al*., 2005a). In other words, variability on different timescales may have different characteristic patterns. EOF analysis of gridded thermosteric sea level time series since 1950 (Lombard *et al*., 2005a) displays a spatial pattern that is very similar to the spatial distribution of thermosteric sea level trends over the same time span (Figure 5.5.9). In addition, the temporal curve is highly correlated with the Southern Oscillation Index (SOI). It appears thus that ENSO-related ocean variability accounts for the 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. The thermosteric sea level data also indicate a change in regime around the late 1970s-early 1980s, a behaviour that originates in the Atlantic Ocean. For the recent years (last decade), the spatial patterns of thermosteric sea level change show remarkable correlation with the geographic distribution of observed sea level trends shown in Figure 5.5.2. This indicates that much 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 the ocean circulation. Moreover, this suggests that the spatial pattern of sea level trends observed by T/P is likely not a long-lived feature, a fact which potentially is a cause for concern with methods which use the 10-year sea level patterns from satellite altimetry to reconstruct the geographical distribution of past sea level change.

[INSERT FIGURE 5.5.8 HERE?]

[INSERT FIGURE 5.5.9 HERE?]

Cabanes *et al*. (2001) argued that sampling error due to the sparse global distribution of tide gauge data had caused 20th century total sea-level rise to be over-estimated. As a result, they concluded that sea-level rise over the past half century could be attributed entirely to the 0.5 mm yr^{-1} thermosteric signal observed by Antonov *et al*. (2002). This point was refuted by Miller and Douglas (2004) who claimed that the bias suggested by Cabanes *et al*. (2001) was actually due to problems in the Gulf Stream region of the analysed temperature fields (Levitus *et al*., 2000) that were used to calculate the thermosteric sea-level signal. Miller and Douglas (2004) averaged raw ocean temperature and salinity data over the past 50 years in three oceanic regions (northeast Pacific, northeast Atlantic and western Atlantic) and found that the inferred steric sea level was much too low (by a factor of about 3) to account for the observed sea level rise at a few tide gauges sites located in these regions. They concluded that the second-half of 20th century sea level rise was mostly due to water mass added to the oceans. This conclusion was recently confirmed by Lombard *et al*. (2005b) who used the Ishii *et al*. (2003) and new Levitus *et al*. (2005a) ocean temperature data to estimate the thermosteric sea level rise over the past 40 years at the location of historical tide gauges. They found that about 1.4 mm yr^{-1} sea level rise of the last 50 years is unexplained by thermal expansion.

Much of the remaining error in these estimates of globally averaged, thermosteric sea level rise is due to inadequate sampling by in situ profiles, particularly in the Southern Hemisphere. As the ARGO array of profiling floats approaches its operational density of 1 float every 3 x 3 degrees of ocean, sampling error is expected to be dramatically reduced.

5.5.4 How to Interpret Regional Variations in the Rate of Sea Level Change

The deviation of local sealevel from the geoid is dynamically closely linked to the ocean circulation. In addition, melting of large ice sheets and mountain glaciers can cause spatially non-uniform change of relative sea level due to isostatic adjustment.

5.5.4.1 Ocean circulation changes

Changes in ocean circulation are a consequence of changes in atmospheric forcing which is primarily by surface wind stress and buoyancy flux (i.e., heat and freshwater flux). The geographical distribution of steric sea level trends is highly non-uniform. At least for the 1990s, spatial patterns of thermosteric sea level change are remarkably correlated with observed sea level trends (see 5.5.3), suggesting that much of the non-uniform pattern of sea level change observed by T/P altimetry over the past decade can be attributed to changes in the ocean thermal structure. Likewise, Antonov *et al*. (2002) have shown that the halosteric effect can be quite significant at regional scale, and e.g. in sub-polar areas of the North Atlantic, especially in the Labrador Sea, it nearly counteracts the thermosteric contribution. This observational result is supported by recent model results (e.g., Stammer *et al*., 2003). Note however that density changes can result both from wind and buoyancy forcing which cannot be separated in simple ways.

Surface atmospheric pressure forcing is dynamically less relevant but causes sea level to deviate from a globally uniform value. On time scales above a few days, the ocean adjusts isostatically to changes in atmospheric pressure, i.e., per 1 hPa sea level pressure increase the ocean is depressed by approximately 1 cm, shifting the underlying mass sideways to other regions. On a time mean, regional changes in sea level caused by atmospheric pressure loading reach about 20 cm (e.g., between the subtropical Atlantic and the subpolar Atlantic).

5.5.4.2 Melt water flux and GIA

The solid earth and oceans continue to respond to the ice and complementary water loads associated with the Late Pleistocene - Early Holocene glacial cycles. While this Glacial Isostatic Adjustment (GIA) process drives large crustal uplift and relative sea level changes near the location of former ice complexes, a world-wide GIA signature results from gravitational, deformational and rotational effects: as mantle material flows to restore isostasy during and after the last deglaciation, uplift occurs under the former centers of the ice sheets while the surrounding peripheral bulges experience a subsidence. The return of the meltwater to the oceans produces an ongoing subsidence of the ocean basins and an upwarping 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 geoid (and hence sea level).

Estimates of the GIA signal on both absolute and relative - to the crust - sea level are generally obtained through computer modeling. Two sources of uncertainty affect the modeling : the viscoelastic structure of the earth mantle and history of ice loads.

GIA studies assume that the surface of the ocean is always constrained to an equipotential surface. However, the value of that equipotential must vary with time in order to account for the exchange of water with the continents and changes in the shape of the ocean. Thus, these two mechanisms, along with changes to the geoid, contribute to a change of absolute sea level. (It should be noted that some of the GIA literature use the terms “geoid” and “absolute sea level” interchangeably. In these cases, “geoid” encompasses the three contributions previously listed.) While geoid variations contribute to the regional changes of absolute sea level, an average of the absolute sea level change over the global oceans due to GIA is controlled by the continuing variation of ocean basin volume. Averaging the GIA contribution over oceanic regions sampled by the T/P--Jason mission yields a value close to –0.3 mm yr^{-1} (Peltier, 2001), with results from plausible suite of viscosity models suggesting an uncertainty of 0.15 mm yr^{-1} (Tamisiea *et al.,* 2004).

The GIA also leads to systematic bias in the average value of sea level change inferred from tide gauges. Unfortunately, uncertainties in GIA predictions can lead to a range in the inferred (GIA-corrected) average of up to 0.5 mm yr^{-1} for tide gauges in the far-field of the former ice complexes (Mitrovica and Davis, 1995). Historically, GIA corrections have been checked by their ability to remove geographic variations present in a set of tide gauge rates. However, variations in GIA-corrected rates are to be expected. For example, rapid mass loss in an ice sheet causes a relative sea level fall in the near field due to the decrease in the geoid height and the uplift of the crust, and a larger-than-expected rise in the far field. Thus, residual geographic trends provide a tool for estimating the source(s) of the sea-level change (Mitrovica *et al*., 2001; Plag and Jüttner, 2001) (see also Section 5.5.2).

5.5.5 Ocean Mass Change

5.5.5.1 Salinity change and fresh water added to the oceans

Salinity data suggest that the oceans have freshened over the last 50 years (Antonov *et al*., 2002). Because salt in the ocean is conserved on shorter than geologic time scales, the only way for the ocean’s salinity to decrease is through the addition of fresh water, either from melting sea ice (which does not affect sea level), or changes in land ice (5.5.5.3) and terrestrial water storage (which do). Antonov *et al*. (2002), Munk (2003) and Wadhams and Munk (2004) used the salinity changes computed by Antonov *et al*. (2002) to estimate the global average sea level change due to fresh water input. Precise assumption on the exact amount of sea ice melting is critical to such an estimate. Several studies have reported a net decline of Northern Hemisphere sea ice volume over the recent decades (see Chapter 4). Assuming 430 ± 130 km^3 yr^{-1} sea ice volume decrease, Wadhams and Munk (2004) estimate to 0.6 ± 0.18 mm yr^{-1}, the rate of sea level rise inferred from fresh water input. These authors concede, however, that large uncertainties remain in their estimate, from both the estimate of ocean freshening as well as the estimates of sea ice melt. It is worth mentioning that Wadhams and Munk's (2004) estimate of fresh water input is a factor of about 2 smaller than indirect estimates deduced from the excess of observed sea level rise over thermal expansion (Section 5.5.6).

5.5.5.2 Land ice

During the 20th century, glaciers and ice caps generally experienced considerable mass losses (Section 4.5), with strong retreats since 1990 being a response to global warming after 1970. Including the glaciers and ice caps fringing the Greenland and Antarctic ice sheets, the estimated contribution to sea level rise is 0.43 ± 0.06 mm yr^{-1} during 1961–1998 and 0.88 ± 0.13 mm yr^1 during 1993-2003 (Table 4.5.3, with ±15% uncertainties reflecting the spread in Table 4.5.2, as discussed in Section 4.5.2).

Altimetric surveys for recent years indicate that the Greenland ice sheet is also losing mass and contributed 0.1 to 0.2 mm yr^1 during 1993–2003, while the Antarctic ice sheet may have had a net mass gain, giving a sea level contribution of –0.2 to –0.0 mm yr^{-1} (Section 4.7). In the absence of direct measurements for earlier decades, we assume the recent tendencies result from a combination of mass-balance changes in response to recent climate change, and long-term imbalance from earlier climate change. The former can be assumed to lie between zero and the 1990s rate, while modelling studies give the latter as –0.1 to 0.0 mm yr^{-1} from Greenland and 0.1 to 0.4 mm yr^{-1} from Antarctica (Huybrechts *et al.*, 1998), which will be the same in the 1990s and earlier decades. Hence the contributions during earlier decades are –0.1 to 0.2 mm yr^{-1} for Greenland and –0.2 to 0.4 mm yr^{-1} for Antarctica.

Constraints on the ice contributions to sea level rise implied by observations of changes in the geoid, earth rotation, polar wander, and earth flattening, have been proposed in a number of studies (Mitrovica *et al.*, 2001; Munk, 2002; Nakada and Okuno, 2003; Sabadini and Vermeersen, 2002). Munk (2002) constrains the sum of the Greenland and Antarctica contributions to ~1.0 mm yr^{-1}. Mitrovica *et al.* (2001) suggest a contribution from Greenland of 0.54 ± 0.13 mm yr^{-1}, which is too large to be consistent with altimetric observations; the method assumes a globally uniform steric sea level change, which could perhaps affect the results. In general, these approaches remains inconclusive because they are strongly dependent on models used to account for GIA (Section 5.5.4.2).

5.5.5.3 Land hydrology: natural variability in land water storage

Interannual/decadal change in land water storage is another contributor to global mean sea level change. Continental water storage includes water (both liquid and solid) stored in subsurface saturated (groundwater) and unsaturated (soil water) zones, in the snow pack, and in surface water bodies (lakes, artificial reservoirs, rivers, floodplains and wetlands). Variations in land water storage result from variations in the climatic conditions that control storage and from direct human intervention in the water cycle or human modification of the land surface. Changes associated with climate variations can be estimated by detailed physical models, whereas the great uncertainty in the direct anthropogenic factors justifies only relatively simple calculations at this time. Changes in concentrated stores, most notably very large lakes, are relatively well known from direct observation. In contrast, global estimates of changes in soil water, groundwater, and small surface stores rely on computations with hydrological models either coupled with global ocean-atmosphere circulation models and/or forced by observations.

5.5.5.3.1 Climate-driven changes in land-water storage

Global land surface models estimate the variation in land water storage (soil moisture, ground water, snow depth and surface waters) by solving the water balance equation. The Land Dynamics (LaD) model developed by Milly and Shmakin (2002) provides global 1° x 1° monthly gridded time series of root-zone soil water, ground water and snow pack for the last two decades. These data were used to quantify the contributions of time-varying storage of terrestrial waters to sea level rise in response to climate change (Milly *et al.*, 2003). A small positive sea level trend, of ~0.12 mm yr^{-1}, was estimated for the last two decades, with larger interannual/decadal fluctuations. Ngo-Duc *et al.* (2005) used a land model forced by a global climatic data set based on the NCEP/NCAR reanalysis and on observations, to estimate land water changes for the past 5 decades. They found a low-frequency variability of about 2 mm in amplitude but no significant trend. The variations are related to ground waters and caused by precipitation variations, which are strongly anti-correlated to the detrended thermosteric sea level. This suggests that the land water contribution to sea level and thermal expansion partly compensate at decadal time scales, perhaps because warmer climatic conditions are associated with greater precipitation (Chapter 10).

5.5.5.3.2 Anthropogenic change in land water storage

The amount of anthropogenic exchange of water with land water storage systems cannot be estimated with much confidence, as already discussed by Church *et al.* (2001).

Ground water. Natural ground water systems typically are in a condition of dynamic equilibrium where, over long times, recharge and discharge are in balance. When the rate of ground water 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 ground water storage eventually reaches the ocean through the atmosphere or surface flow, resulting in sea level rise.

Surface water. 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.

Deforestation. Forests store water in living tissue both above and below ground. When a forest is removed, transpiration is eliminated so that runoff is more favored in the hydrologic budget.

Wetlands 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.

Dams. Impoundment behind dams for agricultural irrigation, flood control, hydroelectric power, and municipal use removes water from the ocean. Enough water has been impounded behind dams to measurably affect global sea level (Chao, 1994; Sahagian *et al*., 1994). Infiltration may raise the water table in the area around the reservoir, storing more water.

5.5.6 Total Budget of the Global Mean Sea Level

The various contributions to the budget of sea level change are summarised in Table 5.5.2 and Figure 5.5.10, both for the 1990's and for the last 4–5 decades. Some terms known to be small have been omitted (atmospheric water vapour, permafrost, sedimentation, cf. Church *et al*., 2001), probably totalling less than 0.2 mm yr^{-1}.

[INSERT FIGURE 5.5.10 HERE]

For the 1990s, the sum of the climatic contributions observed independently explain a large fraction (≥ 75%) of the observed rate of sea level rise (3.1 ± 0.4 mm yr^{-1}), but 0.5 ± 0.4 mm yr^{-1} remains unexplained. Thermal expansion is in the range 1.35–1.75 mm yr^{-1}, while the total land ice contribution is 0.7–1.4 mm yr^{-1}.

For 1950–2000, observed sea level rise amounts to 1.8 ± 0.3 mm yr^{-1}, and is not significantly different for 1961–1998. The trend in thermal expansion is ~0.4 mm yr^{-1}, smaller than the value for the last decade. Part of this difference may reflect decadal variability in thermal expansion rather than a trend. The contribution of mountain glaciers melting is smaller as well, consistent with the recent increase in warming. The sum of the climate contributions is 0.4–1.5 mm yr^{-1}, leaving 0.9 ± 0.4 mm yr^{-1} unexplained.

Thus for both periods, the sea level budget remains unclosed, even though progress has been made since the TAR for the thermal expansion and land ice contributions. The discrepancy is similar for the two periods and could perhaps be the net contribution of anthropogenic terrestrial water storage (Section 5.5.5.3.2), but is at the upper limit of the range estimated by Church *et al*. (2001).

Table 5.5.2. Estimates for the various contributions to the budget of global-mean sea level change for 1993–2003 and for the last 4–5 decades compared with the observed rate of rise.

		Period	Rate of sea level rise with rms errors (mm yr^{-1})	Sources
1	Observed (Tide gauges)	1950–2000	1.8 ± 0.3	Church *et al*. (2004)
2	Observed (Altimetry)	1993–2003	3.1 ± 0.4	Leuliette *et al*. (2004), corrected and adjusted for GIA following Tamisiea *et al*. (2004)
3	Thermosteric	1961–1998	0.35 ± 0.22	Section 5.3
4	Thermosteric	1993–2003	1.55 ± 0.10	Section 5.3
5	Glaciers and ice caps	1961–1998	0.43 ± 0.06	Section 4.5
6	Glaciers and ice caps	1993–2003	0.88 ± 0.06	Section 4.5
7	Greenland ice sheet	20th century	0.05 ± 0.08	Section 5.5.2.2
8	Greenland ice sheet	1993–2003	0.15 ± 0.03	Section 4.7
9	Antarctic ice sheet	20th century	0.10 ± 0.15	Section 5.5.2.2
10	Antarctic ice sheet	1993–2003	0.00 ± 0.10	Section 4.7
	Sum of climate-related terms (3+5+7+9)	1961–1998	1.0 ± 0.3	Uncertainties from the terms have been combined in quadrature
	Sum of climate-related terms (4+6+8+10)	1993–2003	2.6 ± 0.2	Uncertainties from the terms have been combined in quadrature

5.5.7 Comparison with Model Estimates of Contributions to Global Mean Sea Level

In order to have confidence in model-based projections of future sea level change resulting from climate change, we have to demonstrate that model-based estimates are consistent with the observational estimates of historical global mean sea level rise. We have considered results from 15 AOGCMs, of which four have an average rate of global average surface air temperature change over the 20th century lying outside the range of 0.6 ± 0.2 K century^{-1} derived from observations (Section 3.2.2). This means their sea level rise due to thermal expansion and to changes in glaciers and ice caps (G&IC) will be unrealistic in some respect, so we omit them from further consideration. Agreement with observed global average temperature change does not guarantee a realistic simulation of thermal expansion, as there may be compensating errors among climate sensitivity, ocean heat uptake and radiative forcing (cf. Raper *et al.,* 2002). We use results for global average thermosteric sea level rise from the remaining 11 models (cf. Section 10.6.1) and calculate the sea level rise contribution from G&IC by the methods used for projections (Section 10.6.3).

The model range of 0.32–0.99 mm yr^{-1} for simulated thermal expansion for 1961-1998 is somewhat above the observational range of 0.35 ± 0.22 mm yr^{-1} (Section 5.5.3). This is likely to be because most of the models do not include volcanic forcing; several large volcanoes cooled the climate during the 1960s, and the models with this effect generally have smaller ocean heat uptake in recent decades (Gleckler *et al.,* 2005). For 1993–2003 the eleven models lie in the range 0.5-1.8 mm yr^{-1}and are generally below the observational range of 1.2–1.6 mm yr^{-1}. The discrepancy could be partly covered by internally generated variability, which is 0.1-0.6 mm yr^{-1} for 10-year trends in model control runs, but may be underestimated by models (Section 5.3, Gregory *et al.,* 2005). Using the PCM1 AOGCM, Church *et al*. (2005a) suggest that 0.5 mm yr^{-1} of the trend in the last decade may result from warming as a recovery from the Pinatubo eruption of 1991, and this is supported by Gregory *et al*. (2005) using UKMO-HadCM3. Volcanoes on shorter and longer periods may have opposite influences because the short-term response is determined by the cooling and recovery of the upper ocean, whereas the multi-decadal response is related to much longer persistence of cool anomalies in the deep ocean (Gleckler *et al*., 2005).

The range of G&IC contributions is 0.0-0.2 mm yr^{-1} for 1961–1975 mm yr^{-1} and 0.0–0.4 mm yr^{-1} for 1988–1998, the upper values being consistent with those given by Dyurgerov (2003) of about 0.15 and 0.41 mm yr^{-1} for these periods. More recent evaluations of G&IC (Section 4.5) give larger numbers, in particular for recent years, the consequence being that the AOGCM-derived estimates lie about 0.3 and 0.7 mm yr^{-1} below the values given in Section 5.5.6 for 1961–1998 and 1993–2003; the reasons for the discrepancies are not clear.

Adding the thermal expansion and G&IC terms we find ranges of 0.3–1.2 and 0.7–2.0 mm yr^{-1} for the AOGCMs, which lie below the observed rates. This is not surprising because the observational and model-based estimates of the terms are roughly similar, so the deficit is the same as discussed in Section 5.5.6 for the observational budget.

The increase in rate of rise over recent decades is consistent with observations and with rising anthropogenic forcing (Woodworth *et al.,* 2004). Almost all models show a statistically significant acceleration during the 20th century, ranging from 0.006 to 0.016 mm yr^{-2}. The observational estimate of 0.012 ± 0.006 mm yr^{-2} made by Church and White (2005) is in the range of model accelerations. Since the models have global coverage, unlike tide gauges, an acceleration in sea-level rise is more easily detectable in models (Gregory *et al.,* 2001).

[START OF QUESTION 5.1]

Question 5.1: Is the Sea Level Rising?

Geological and archeological observations indicate that, during the past ~3 thousand years, sea level did not rise by more than 0.1–0.2 mm yr^{-1} globally. There is some evidence for onset of acceleration during the 19th century. Modern sea level measurements by tide gauges show indisputable evidence of significant sea level rise during the 20th century. Recent estimates for the last half of the 20th century (1950–2000) give ~2 mm yr^{-1} global mean sea level rise.

New satellite observations available since the early 1990s provide very precise 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 3.1 ± 0.4 mm yr^{-1}, a rate significantly higher than during the previous decades. However, it is presently unclear whether the higher rate of sea level rise in the 1990s is part of an acceleration due to anthropogenic global warming, or a result of climate variability, or a combination of both effects.

Satellite data also show that sea level is not rising uniformly over the world. While in some regions (e.g., western Pacific) sea level rise since 1993 is up to 5 times the global mean, in other regions (e.g., eastern Pacific) sea level is falling. Substantial spatial variation in rates of sea level change is also inferred from hydrographic observations, and expected from climate models. Spatial variability of sea level rates is mostly due to non uniform thermal expansion.

A recent re-evaluation of climate factors causing sea level rise (published since the TAR) leads to reasonable quantitative agreement with sea level observations for the past decade. The change in the total mass of ocean water (caused by mass changes in glaciers and ice sheets) is estimated to ~1 mm yr^{-1} sea level rise for the recent years. The thermal expansion due to increase in ocean heat content amounts to 1.55 ± 0.10 mm yr^{-1} over 1993–2003 (i.e., ~50% of the observed rate). For the past 50 years, the contribution of thermal expansion to sea level rise was ~0.4 mm yr^{-1}, i.e. only ~25% of the observed rate, and the land ice contribution more uncertain, estimated as 0.2–0.9 mm yr^{-1}. The substantial difference of ~0.9 mm yr^{-1} between observed sea level rise and thermal expansion requires an explanation.

[END OF QUESTION 5.1]

5.6 Synthesis

The patterns of observed changes in global heat content and salinity, sea-level, steric sea-level, water mass evolution and bio-geochemical cycles described in the previous four sections are consistent with known characteristics of the large scale ocean circulation (Figure 5.6.1).

[INSERT FIGURE 5.6.1 HERE?]

There is compelling evidence that the heat content of the world ocean has increased since 1955. The North Atlantic has warmed (south of the 45°N) and the warming is penetrating deeper in this ocean basin than in the Pacific, Indian and Southern Oceans (Figure 5.2.2), 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 high Anthropogenic Carbon Content as observed (Figure 5.4.1). The Southern Ocean has both a deep and shallow overturning circulation. The shallow overturning circulation is characterised by subduction of Sub-Antarctic Mode Waters and a northward circulation of heat anomaly (<1000 m). The 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 Sub-Antarctic Water masses (Figure 5.4.1). The Southern Hemisphere deep overturning circulation shows little evidence of change based on presently available data.

The subduction of carbon into the ocean has meant that Calcium Carbonate and Aragonite saturation horizons have generally shallowed, and that pH has decreased primarily in the surface and near surface ocean causing the ocean to become more acidic..

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 poleward of 70°S) are fresher in the upper 500m (Figure 5.2.6c). In the upper 500m, the sub-tropical latitudes in both hemispheres are characterised by an increase in salinity. The regional analyses of salinity also show a similar distributional change with a freshening of key high latitudes water masses such as Labrador Sea Waters, and Antarctic and North Pacific Intermediate Waters, and increased salinity in some of the subtropical gyres such as 24°N. We are confident that these changes in salinity imply a change in the atmospheric hydrological cycle over the oceans. At high latitudes (particularly in the North Hemisphere) there is an observed increase in the melt of perennial sea-ice, increased precipitation, and glacial meltwaters

(Chapter 4), all of which act to freshen high latitude surface waters. At mid-latitudes it would seem likely that Precipitation-Evaporation has decreased (i.e., increase in the transport of freshwater from the ocean to the atmosphere). Together the pattern of salinity change suggests an increase in the Earth's hydrological cycle over the last 50 years.

The transfer of heat into the ocean via processes such as convection and subduction leads to sea-level rise, mainly through thermal expansion (thermosteric component of sea level change). Over the last 50 years the greatest coherent steric sea-level rises occur in the water masses of Southern Ocean and North Atlantic (see Figure 5.5.7). These are regions that showed strongest change in the zonal average of temperature change and largest changes in Anthropogenic Carbon Content. We now have the capability to measure most components of sea-level. In the 1990's the observed sea-level rise that is not explained through steric sea level rise is largely explained by the transfer of mass from glaciers, ice sheets, and river runoff (Section 5.5).

In the Equatorial Pacific the pattern of steric sea-level rise also shows that strong west to east gradients in the Pacific has weakened (i.e., now cooler in the Western Pacific and warmer in the Eastern Pacific). This decrease in the equatorial temperature gradient is consistent with the increased frequency and duration of El Niño over this same period (Figure 5.5.9b). The spatial gradients in sea-level mean that surface ocean currents have changed, (e.g., the Antarctic Circumpolar Current is slightly stronger, the North Atlantic sub-tropical gyre has strengthened), but typically these changes in horizontal currents are small compared with the mean circulation.

There is some evidence that the rate of change of the state of the ocean is increasing. The increase in global heat content, steric sea-level, and absolute sea-level are all higher in the 1990–2000 period, than in the period from 1955. There is some evidence that the fraction of the CO_2 emission into the atmosphere that the ocean can absorb is decreasing, although the uncertainty in the estimates is also too large to prevent a stronger conclusion. Therefore it is possible that the changes of the ocean state could be accelerating; the presence of decadal variations such as the Pacific Decadal Oscillation and North Atlantic Oscillation, and the lack of sufficient data prevent a stronger conclusion.

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

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

5.A.1 Ocean Heat Content and Salinity

Several different objective analysis techniques have been used to produce the gridded fields of temperature anomalies used to compute ocean heat content presented in this section. Similarly, the methods used to compute uncertainty of the estimates of heat content are different. However, as described in the text and Figure 5.2.5, the results of three different estimates of ocean heat content are quite similar. 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 thermosteric component of sea level change) is based on the construction of gridded (1° latitude-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 an iterative difference-correction method based on the work of Barnes (1973) and its implementation is described by Boyer *et al*. (2002). Generally the objective analysis scheme works as follows. At each standard depth level all data are averaged within each 1° square (ODSQ) and a first-guess value (which is taken from climatology) is subtracted to produce a mean anomaly value. An "influence" region is defined based on an influence radius R (decorrelation length scale = 555 km) around each ODSQ, and a "correction" is computed using all ODSQ values in the influence region based on a Gaussian-shaped, distance-related weight function. At each ODSQ the correction is added to the first-guess field to produce an "analyzed" value. Both the climatologies and anomaly fields produced in this way are characterized by a response function such that features with wavelength less than 555 km are substantially (and deliberately) reduced in amplitude.

Ishii *et al*. (2005) employed similar techniques as above, with a smaller decorrelation length scale of 300 km and a least-squares technique for estimating corrections to the first-guess field. Willis *et al*. (2005) used a two-scale covariance function. The similarity of the three independently estimated heat content time series shown in Figure 5.2.5 indicates that the differences in analysis techniques 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 which are briefly described in the following.

5.A.1.1 Statistical errors

For the Levitus *et al*. (2005a) estimates (all profile data and analyses are available at *http://www.nodc.noaa.gov*), the uncertainty at any gridpoint is estimated from the variability of corrections to that gridpoint contributed by all ODSQs containing data within the influence region surrounding that gridpoint, using the rules of error propagation (e.g., Taylor, 1997). The method is based on the assumption that all standard deviations within the influence region are equal, which has been shown by Stephens *et al*. (2002) to be a good assumption. In this way, standard (rms) errors for all analysed variables (e.g., temperature, salinity ..) are available a function of depth and horizontal position, and for integrated variables as function of longitude and/or latitude (i.e., for basin or zonal mean values).

Both Ishii *et al*. (2005) and Willis *et al*. (2005) used the interannual standard deviation of their heat content as the basis for error analyses.

5.A.1.2 Sampling errors

In view of the highly inhomogeneous distribution of ocean observations in space, errors resulting from spatial and temporal variations in the distributions of data are important. Although the linear trend of ocean heat content has been reproduced by several AOGM simulations forced by all or some of the estimated changes in atmospheric concentrations of greenhouse gases, anthropogenic aerosols, volcanic aerosols, and the effect of solar variability (Barnett *et al*., 2001; Barnett *et al*., 2005; Gregory *et al*., 2004; Levitus *et al*., 2001; Sun and Hansen, 2003), the observed interdecadal variability of the global integral of ocean heat content has not been simulated. In particular the observations (e.g., Figure 5.2.1) indicate a decrease of ocean heat content of approximately 6 x 10^{22}J between 1980 and 1983. This has led to the question (Gregory *et al*., 2004) as to whether the models or the data (or both) are deficient, and whether AOGCMs should be judged based on their ability to simulate this interdecadal variability. Gregory *et al*. (2004) and AchutaRao *et al*. (2005) have conducted studies which examine this problem and present evidence suggesting that deficiencies in the comprehensiveness of the ocean temperature database (in space and time) may be responsible for the observed interdecadal variability.

It is not clear that a lack of data is specifically responsible for the observed interdecadal variability in question. Levitus *et al*. (2005a, Supplemental Figure S4) have shown that most of the decrease in ocean heat content between 1980 and 1983 occurs in the Pacific Ocean, specifically north of 20°S. Data distribution plots of the number of observations at 400 m depth for the 1976–1980 and 1984–1988 pentads show a reasonably good global data coverage for estimating the global ocean heat content integral. Equally important in determining whether or not such interdecadal variability may be real is an examination of data from an independent observing system. Satellite altimetric observations of sea level variability (Nerem *et al.*, 1999) indicate that during the 1997–1998 El Niño, sea level rose by 15 mm during 1997 and then decreased by a similar amount during 1998. In contrast, the drop in the thermosteric component of sea level during 1980–1983 associated with the heat content change in question, was approximately 9 mm (Antonov *et al*., 2005). Considering the change in sea level during 1997–1998, the data distributions, the large-scale nature of the heat content variability associated with the global heat content integral (Figs. 5.2 and 5.3), and the similarity of the Levitus *et al*. (2005a) and Ishii *et al*. (2005) analyses, it appears likely that the observed interdecadal variability in ocean heat content is indeed real.

5.A.2. Heat Fluxes and Transports

Surface meteorological and subsurface hydrographic observations are inhomogeneously distributed in space and in time. For instance, in the mid-latitudinal North Atlantic in the area of the major ship routes the number of observations may be 10–100 times higher than that in the Southern Ocean. Similarly, in the period from 1900 to 1950 the number of observations is 3 to 30 times smaller than during decades of 1960s–1990s. This results in the lack of representativness of the estimates based on the limited number of reports in comparison to the well sampled areas or periods. The magnitude of sampling uncertainty of surface heat fluxes in the poorly sampled Southern Ocean can amount to 20–50 W/m^2, which is higher than the magnitude of interannual variability of fluxes (Gulev *et al*. 2005). In the Northern Hemisphere locally high sampling uncertainties are observed in the Labrador sea and the North-West Pacific, where they also amount to 50 W/m^2.

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 implies also the integration of uncertainties of the zonally averaged estimates. For instance, the uncertainty of zonal averaged estimates of ±10 W m^{-2} can result in 0.5 PW uncertainty in MHT in the Atlantic and in nearly twice that value in the Pacific. Thus, all climatological estimates of MHT based on the surface heat balance should be taken with great care, which is especially true for the estimates of MHT variability.

5.A.3 Estimates of Anthropogenic Carbon and Oxygen Changes

Estimates of anthropogenic carbon in the ocean since 1750 are made from an indirect method that uses measurements of dissolved inorganic carbon (DIC) and removes from these measurements an estimate of the change in DIC that result from biological activity, and the change in DIC that are caused by the CO_2 disequilibrium at the ocean surface. Although the anthropogenic carbon is not directly measured, the method is based on well known processes that control the distribution of natural DIC in the ocean. The method combined with existing data gives a global estimate of 118 ± 19 PgC (Sabine *et al.*, 2004), where the reported uncertainty is based on measurement errors and potential biases. Most potential biased have been quantified by other studies and they suggest that the global anthropogenic carbon is overestimated by ~10%, mainly because of the assumption of constant air-sea disequilibrium with time by the method (Matsumoto and Gruber, 2005), and because of the impact of recent changes in temperature and ocean circulation, although this later correction is more uncertain. Biases may not be additive, and potential biases from assumptions of constant ratios for biological activity have not been assessed.

Estimates of changes in oxygen between 1955 and 1998 were made for each pentad using data compiled by the World Ocean Atlas 2001 (Garcia *et al*., 2005). Objectively analyzed monthly climatologies of oxygen and AOU were prepared using quality-controlled oceanographic data on seven vertical levels (0, 10, 20, 30, 50, 75, and 100 m) on a grid of 1 x 1 degree. The database includes 0.5 million profiles per pentads between 1965 and 1990, and 0.3 to 0.5 million profiles before 1965 and after 1990. The standard error of the data is constant for the different time periods at ±1–3 umol/kg. The measurement method was not reported for all the cruises. For the cruises where the measurement method was reported, only the Winkler titration was used. Cruises from 1955 to 1990 reported only manual titrations. The Carpenter method to improve the accuracy was reported on some cruises after 1970. Other improvements in the titration method and automated titrations were reported after 1990 only. There are no standards for oxygen measurements because of the difficulty in preparing a stable solution. Problems of oxygen 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 reliable.

5.A.4 Estimation of Sea Level Change

5.A.4.1 Satellite altimetry : measurement principle and associated errors

The concept of the 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 R of the satellite above the instantaneous sea surface. The quantity of interest in oceanography is the height h=H-R of the instaneous sea surface above a reference fixed surface (typically a conventional reference ellipsoid) which is computed as the difference between the altitude of the satellite above the reference ellipsoid H and the altimeter range R. The satellite altitude H is computed through precise orbit determination, a long standing approach in space geodesy which combines accurate modelling of the dynamics of the satellite motion and tracking measurements (GPS, DORIS or satellite laser ranging) between the satellite and observing stations on Earth or other observing satellites. The range R from the satellite to mean sea level must be corrected for various components of the atmospheric refraction and biases between the mean electromagnetic scattering surface and mean sea level at the air-sea interface. A number of corrections must be applied to obtain the correct height h. 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. They also include the inverted barometer correction which has to be applied since the altimeter does not cover the global ocean completely.

State of the art satellite altimetry has more than 2 decades of heritage. Over the years, technological improvements have considerably decreased the instrumental noise, to 1.7 cm for Topex/Poseidon (launched in 1992) and Jason-1 (launched in 2001) for point-to-point measurement. Thanks to a concerted effort in precise modelling of the geophysisical and environmental corrections, the rms of these various corrections errors has been reduced to 2.7 cm for a single Topex/Poseidon point measurement. Similarly, precision orbit determination has reduced the rms satellite height H error to 2.5 cm. For Jason-1, the orbit error is even smaller, in the range 1-2 cm. The total rms measurement accuracy for the Topex/Poseidon altimetry-based sea surface height is about 4 cm for a single measurement (see Table 5.A.1).

Table 5.A.1. The rms system measurement accuracy for the NASA Topex/Poseidon dual frequency altimeter (after Chelton *et al.,* 2001)

Single-pass sea surface height accuracy	
Topex/Poseidon radar altimeter noise	1.7 cm
Ionosphere	0.5 cm
Electromagnetic bias	2.0 cm
Dry troposphere	0.7 cm
Wet troposphere	1.1 cm
Orbit	2.5 cm
Total (rms) sea surface height	4.1 cm

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-day for Topex/Poseidon and Jason-1), accounting for the spatial distribution of the data using an equi-area weighting. In effect, during this time interval, the satellite realizes an almost complete coverage of the oceanic domain. The error associated with a 10-day mean sea level estimate is approximately 4 mm.

Of considerable importance when computing global mean sea level variations through time is proper account of instrumental bias and drifts. These effects (e.g., the radiometer drift onboard Topex/Poseidon used to correct for the wet tropospheric delay) are indeed of the same order of magnitude as the sea level signal. Studies by Chambers (1998) and Mitchum (1994; 2000) have demonstrated that comparing the altimeter sea level measurements to tide gauges 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 on the mean sea level variation using the tide gauge calibration is about 0.4 mm/yr (a value resulting mainly from the uncertainties in vertical land motion at the tide gauges).

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

5.A.4. *Estimating 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, and for which procedures must be followed to ensure good quality data for a long term record. For example, pressure-based gauges are robust devices but their main components (the pressure transducers) can be prone to slow drifts which must be regularly corrected for. The Global Sea Level Observing System (GLOSS) specifies that a gauge must be capable of measuring sea level to cm accuracy (or better) in all weather conditions (which in practice means in all wave conditions). However, the most important consideration, common to all types of gauge, 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. GLOSS specifications require that local levelling must be repeated at least annually between the reference mark of the gauge (sometimes called the Contact Point), 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, DORIS, Absolute Gravity) for which specifications also exist with regard to monumentation and benchmark controls. With typical rates of sea and land level change of 1 mm/year, it is necessary to maintain the accuracy of the overall gauge-system (which includes the benchmark network) at the cm level over many decades. This demanding requirement has been met in many countries for many years; the challenge now is to have similar standards throughout the global network. See IOC (2002) for more information.