Chapter 5: Observations: Oceanic Climate Change and Sea Level

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Executive Summary

1. Consistent with the Third Assessment Report (TAR), global ocean average temperature has risen 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 represents an average warming of the upper 3000 m of the world ocean by 0.037°C.

2. Global ocean heat content has considerable interannual and interdecadal variability superimposed on a longer-term trend.

3. Large-scale, coherent trends of salinity are observed for 1955–1998, and are characterized by a global freshening in subpolar latitudes and a salinification of shallower parts of the tropics and subtropics. Freshening is pronounced in the Pacific, increasing salinities prevail over most of Atlantic and Indian Oceans. These trends suggest that changes in earth's hydrological cycle are occurring. While we are confident of basin-scale regional changes in distribution of salinity, the observations do not allow for a reliable estimate of the global average change in salinity.

4. Southern Ocean mode waters and Upper Circumpolar Deep Waters waters are warming. 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 accompagnied 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.

5. Cooling is observed in the North Atlantic subpolar gyre (with increased convection), and in the central North Pacific. North Atlantic Deep Water has freshened significantly.

6. Indirect evidence suggests that Atlantic meridional overturning circulation has considerable decadal variability, but there is low confidence in its long-term trend.

7. Oxygen concentrations in the ventilated thermocline (~100–1000 m) have decreased in most ocean basins from the early 1970s to the late 1990s, consistent with reduced rates of ventilation. 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.

8. The total carbon content of the oceans has increased by 118 ± 19 GtC between 1750 and 1994 and continues to increase. The fraction of the CO_2 emitted that was taken up by the oceans appears to have decreased from 42 ± 7% during 1750–1994 to 37 ± 7% during 1980–2005. This would be consistent with the limited rate at which the oceans can absorb carbon, but the uncertainty in this estimate does not allow firm conclusions. 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 decreased.

9. Global sea level rise in the second half of the 20th century is estimated as 1.8 ± 0.5 mm yr^{-1}, consistent with the TAR estimate of 1.5 ± 0.5 mm yr^{-1} for the 20th century. Sea level rise measured by satellite altimetry since 1993 is estimated as 3.1 ± 0.8 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.

10. The average thermal expansion contribution to sea-level rise for the last 50 years is is 0.4 ± 0.1 mm yr^{-1}, with significant decadal variations. For 1993 to 2003, the estimate is 1.6 ± 0.6 mm yr^{-1}, but it is unclear whether the increase indicates an accelerated rise or it is associated with decadal or longer variability of the ocean-atmosphere system.

11. Revised estimates of loss of mass from glaciers, ice caps and the Greenland and Antarctica ice sheets show a contribution to sea level rise of 1.2 ± 0.6 mm yr^{-1} during 1993–2003.

12. The climate contribution (sum of thermal expansion and land ice melting) to sea level rise during the last 50 years is significantly smaller than the observed value, as in the TAR. For the last decade, the

sum is estimated as 2.8 ± 0.8 mm yr^{-1} and constitutes the main contribution to the sea level budget which is closed within known errors.

13. Sea level change is highly non-uniform spatially. In some regions, rates are up several times the global mean rise, while in other regions sea level is falling. Sea level trend patterns derived from altimetry are well correlated with the thermal expansion contribution over the same period.

14. 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 related to rise in mean sea level and variations in regional climate.

15. 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 this chapter are broadly consistent with known characteristics of the large scale ocean circulation.

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 oceans net heat uptake since 1955 is around 20 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 crucial role for climate, in particular for variations on seasonal to decadal time scales such as e.g. related to 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 feedback on 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 also linked to changes in ocean circulation. Finally, oceanic parameters can be useful for detecting climate change, in particular temperature and salinity changes in the deeper layers and in different regions where the short-term variability is smaller and signal-to-noise ratio is higher (e.g., Banks and Wood, 2002).

In the Third Assessment Report (TAR), some 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 300 m, equivalent to an average temperature increase of 0.037°C per decade. No information on ocean circulation changes (other than those related to El Niño) 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. In this assessment we report on new direct evidence for changes in total carbon increase and on changes ocean biogeochemistry (including pH).

In the TAR, Church et al. (2001) adopted a best estimate of 1.5 ± 0.5 mm yr^{-1} as the average rate for the observed sea level rise in the 20th century, and also discussed possible causes for sea level rise, such as thermal expansion through ocean warming, melting of mountain glaciers and ice sheets, and changes in terrestrial water storages. They concluded that the estimated sum of climate-related contributions was less than half of the observed value,, and it was unclear whether the climate-related processes had been underestimated, or the rate of sea level rise had been overestimated.

This chapter will assess observations of changes in oceanic parameters and update the current state of knowledge. 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? (see Sections 5.2 and 5.3)

2. Is the ocean circulation changing? Can the causes for observed changes be inferred? (see Section 5.3)

3. Is the ocean biogeochemistry changing? Are these changes consistent with observed physical changes? (see Section 5.4)

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? (see Section 5.5)

5.2 Changes in Global-Scale Temperature 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 (Willis et al., 2004; Levitus et al., 2005b; Ishii et al., 2006). 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).

5.2.2 Ocean Heat Content

5.2.2.1 Long-term temperature changes

Figure 5.2.1 shows two time series of ocean heat content for the 0–700 m layer of the world ocean (Levitus et al., 2005a; Ishii et al., 2006) for 1955–2003 and a time series for 0–750 m for 1993–2003 (Willis et al., 2004). Approximately 7.3 million profiles of temperature were used in constructing the two longer time series. The three heat content analyses cover different periods but where they overlap in time there is good agreement. The existence of three analyses increases our confidence in the reliability of ocean heat content estimates. The time series shows an overall trend of increasing heat content in the world ocean with interannual and interdecadal variations superimposed on this trend. The root mean square difference between the three data sets is 1.6×10^{22} J or 0.99 W m^{-2}. The two longest time series (using independent criteria for, selection, quality control, interpolation and analysis on similar data sets) show very good agreement on trends and on decadal time scales. There are year-to-year differences which are due to differences in quality control and data used. For the 1993–2003 period, the Levitus et al. (2005c) analysis has a linear global ocean trend of 0.50 ± 0.16 W m^{-2} compared with 0.61 ± 0.17 W m^{-2} and 0.41 ± 0.13 W m^{-2} respectively for Willis et al. (2004) and Ishii et al. (2006). For the 0–700 m layer and the period 1955–2003 the heat content change is 11.2×10^{22} J or 0.15 W m^{-2}. All of these estimates are per unit area of earth surface and the error bars are reported as 95% confidence intervals. Despite the fact that there are differences between these three ocean heat content estimates due to the data used, quality control applied, instrumental biases, temporal and spatial averaging and analysis methods (Appendix 5.A.1), they are consistent with each other giving a high degree of confidence for their use in climate change studies. Note that the global increase in ocean heat content during 1993–2003 period is considerably larger than the observational estimates in two ocean models constrained by assimilating altimetric sea level and other observations (Stammer et al., 2003; Carton et al., 2005). We conclude that the available heat content estimates since the 1955 show very significant increasing trends in ocean heat content.

[INSERT FIGURE 5.2.1 HERE]

The data used in estimating the ocean temperature fields (for the above heat content estimates) do not include SST observations discussed in Chapter 3. However, comparison of the global, annual mean time series of near-surface temperature and the corresponding SST based on the ICOADS database (Smith and Reynolds, 2003) show a correlation of 0.92 (Levitus et al., 2005a). The ICOADS database contains approximately 159 million SST observations. The consistency between these two sea-surface temperature data sets gives confidence in the ocean temperature data set used for estimating depth integrated heat content, and supports the trends in SST reported in Chapter 3.

There is a contribution to the global heat content integral from depths greater than 700 m as documented by Levitus et al. (2000; 2005a). However, due to the lack of data with increasing depth the data have to be composited 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. There is close agreement between the 0–700 and 0–3000 m time series of Levitus et al. (2005a). A comparison of the two linear trends from these two series 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. 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 global ocean volume mean temperature increase of 0.037°C during this period. This corresponds to an average heating rate of 0.2 W m^{-2} for the earth's surface.

The geographical distribution of the linear trend of 0–700 m heat content for 1955–2003 for the world ocean is shown in Figure 5.2.2. Most of the Atlantic Ocean exhibits warming with a major exception being the sub-Arctic gyre. The Atlantic Ocean accounts for approximately half of the global linear trend of ocean heat content (Levitus et al., 2005a). Much of the Indian Ocean has warmed with a major exception being the 5-20ºS latitude belt. The Southern Ocean (south of 35ºS) in the Atlantic, Indian and Pacific sectors has warmed. The Pacific Ocean is characterized by warming with major exceptions along 40ºN and the western Tropical Pacific. The Pacific Ocean variability in heat content is dominated by PDO and El Niño (Stephens et al., 2001) and Levitus et al. (2005c).

[INSERT FIGURE 5.2.2 HERE]

Figure 5.2.3 shows the linear trends (1955–2003) of zonally averaged temperature anomalies (0–1500 m) for the world ocean and individual basins based on yearly 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. The regions that exhibit cooling occur mainly in the shallow equatorial areas and in some high latitude regions. In the Indian Ocean cooling occurs at subsurface depths centred on 12°S at 150 m depth and in the Pacific centred on the Equator and 150 m depth level. Cooling also occurs in the 32–48°N region of the Pacific Ocean and the 49–60°N region of the Atlantic Ocean. The upper ocean cooling 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 0.19°C per decade on which is superimposed strong quasi-decadal oscillations with a range of approximately 2°C. These oscillations appear to have been associated with the East Atlantic Oscillation in sea level pressure. The cooling (and freshening) of the deep water of the Atlantic sub-Arctic gyre since 1970 has been documented by Dickson *et al*. (2002) among others. Although we focus on the upper ocean we note that warming of the deep water in the Weddell Sea has occurred since the 1970s (Robertson et al., 2002; Smedsrud, 2005).

[INSERT FIGURE 5.2.3 HERE]

5.2.2.2 Variability of heat content

Examination of short-term variability of global mean sea level from satellite altimeter measurements (Figure 5.5.1) documents that during the 1997–1998 El Niño sea level rose by approximately 12 mm during 1997 and fell by approximately the same amount after reaching its peak value. The direct implication of such variations in ocean heat content during El Niño is that the net radiation at the top-of-the-atmosphere (TOA) is consistent with ocean heat storage (Wong et al., 2006). The reversal in polarity of the PDO that occurred in the mid-1970s exhibits changes in SST that are very similar to SST changes that occur during El Niño events (Zhang et al., 1997).

A major feature of Figure 5.2.1 is the relatively large increase in heat content during 1969–1980 and a sharp decrease 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 and appears to be associated with the reversal in polarity of the Pacific Decadal Oscillation (PDO) (Stephens et al., 2001; Levitus et al., 2005c, see also Chapter 3, Section 3.6.3).

Figure 5.2.4 shows the difference in 0–700 m heat content between the 1977–1981 and 1965–1969 pentads (a) and the 1986–1990 and 1977–1981 pentad (b). During these two periods much of the world ocean first warmed and then cooled. The pattern of warming and cooling has spatial scales of entire ocean basins and is also found in similar analyses by Ishii et al. (2006). It is the Pacific Ocean that dominates the decadal variations of global heat content during these two periods. Note also that the thermosteric component of sea level during 1980–1983 decreased by approximately 9 mm (Antonov et al., 2005).

[INSERT FIGURE 5.2.4a HERE]

[INSERT FIGURE 5.2.4b HERE]

Considering the large-scale nature of the heat content variability, and the similarity of the Levitus et al. (2005a) and Ishii et al. (2006) analyses, we are confident that there is substantial interdecadal variability in ocean heat content. However, even in periods with overall good coverage in the observing system (Figure 5.A.1), large regions in Southern Hemisphere are not well sampled, and their contribution to global heat content variability is less certain.

5.2.2.3 Implications for Earth's heat balance

To place the changes of ocean heat content in perspective, Table 5.2.1 and Figure 5.2.5 provide updated estimates of the change in heat content of various components of the earth's climate system for the 1961–2003 period (Levitus et al., 2005c). This includes changes in heat content of the lithosphere (Beltrami et al., 2002), the atmosphere (e.g., Trenberth et al., 2001), and the total heat of fusion associated with estimates of the maximum amount of melting of glaciers and small ice-caps, Antarctic and Greenland ice-sheets (see Chapter 4) and the total heat of fusion due to melting of Arctic sea ice (Hilmer and Lemke, 2000). The increase in ocean heat content is much larger than any other component of the earth's heat balance over the two periods 1961 to 2003 and 1993–2003, and accounts for more than 90% of the possible increase in heat content of the earth system during these periods. Ocean heat content variability is thus a critical variable for detecting the effects of the observed increase in greenhouse gases in the earth's atmosphere and for resolving the Earth's overall energy balance. It is noteworthy that while ice melt from glaciers, ice caps and ice sheets is very important in the sea level budget, the energy associated with ice melt contributes only about 1% to the Earth's energy budget.

Table 5.2.1. Energy balance for different components of the Earth System for two periods (1961–2003) and (1993–2003). Ocean heat content change are from this Section for 1993–2003 and from Levitus et al. (2005) for 1961–2003, glaciers and small ice caps and ice sheets from Chapter 4, continental heat content from Beltrami et al. (2002), atmospheric energy content from Trenberth et al. (2001) and Arctic sea-ice release from Hilmer and Lemke (2000). Positive change means an increase in stored energy for that component . All error estimates are 95% confidence intervals. No estimate of confidence was available for the continental heat gain. Some of the results have been scaled from published results for the two respective periods. Ocean heat content change for the 1961–2003 period is for the 0–3000m ocean layer. Ocean heat content change for the 1993–2003 period is for the 0–700m (or 0–750m) layer and is computed as average of the trends from Ishii et al. (2006), Levitus et al. (2005), and Willis et al. (2004).

Energy Content	Past half-century *Energy content change*		Past decade *Energy content change*	
	1961–2003 (10^{22}J)*	*1961–2003 (10^{-3} W m^{-2})*	*1993–2003 (*10^{22}J)*	*1993–2003 (10^{-3} W m^{2})*
Ocean heat content	14.2 ± 2.90	210 ± 42.9	8.32 ± 1.02	517.3 ± 0.634
Glaciers & small ice caps	0.22 ± 0.15	3.3 ± 2.2	0.08 ± 0.05	5.0 ± 0.031
Greenland Ice sheet	0.02 ± 0.05	0.3 ± 0.7	0.02 ± 0.01	1.2 ± 0.6
Antarctic Ice Sheet	0.06 ± 0.18	0.9 ± 2.7	0.02 ± 0.04	1.2 ± 2.5
Continental heat content	0.76	11	0.18	11.2
Atmospheric energy content	0.5 ± 0.1	7.4 ± 1.5	0.2 ± 0.1	12.4 ± 6.2
Arctic Sea-Ice	0.15 ± 0.05	0.7 ± 0.7	0.07 ± 0.28	4.4 ± 17.4
Sum of Energy terms	15.8 ± 2.9	234 ± 43.1	8.9 ± 1.1	552.7 ± 66.2

[INSERT FIGURE 5.2.5 HERE]

5.2.3 Ocean Salinity

Ocean salinity changes are an indirect but potentially sensitive indicator for detecting changes in precipitation, evaporation, river runoff and ice melt. The patterns of salinity change can be used to infer changes in the earth's hydrological cycle over the oceans (Wong et al., 1999; Curry et al., 2003) and are an important complement to atmospheric measurements. 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, about one third of the amount of data used in the ocean heat content estimates (in Section 5.2.2). Between 15°S and 42°N 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 42–72°N region, including the Labrador, Irminger and Icelandic Seas, there is a strong freshening (discussed further in Section 5.3). The increase in salinity north of 72°N (Arctic Ocean) is highly uncertain because of the paucity of data in this region.

South of 50°S in the Polar region of the Southern Ocean there is a weaker freshening signal. Freshening occurs throughout most of the Pacific with the exception of the South Pacific sub-tropical gyre between 32–8°S and above 300m where there is an increase in salinity. The near surface Indian Ocean is characterized mainly by increasing salinity. However, in the latitude band 42–5°S, (South Indian Gyre) in the depth range of 200–1000 m there is a freshening of the water column.

[INSERT FIGURE 5.2.6 HERE]

The results shown here document that ocean salinity and hence freshwater are changing on gyre and basin scales, with the near surface waters in the more evaporative regions increasing in almost all ocean basins. In the high latitude regions in both hemispheres the surface waters are freshening consistent with these regions having greater precipitation, although higher run-off and ice melting may also contribute. In additon to these meridional changes the Atlantic is becoming saltier over much of the water column (Figure 5.2.6). Although the South Pacific subtropic region is becoming saltier, the salinity over the whole water column in the Pacific Basin is becoming fresher. The increasing difference in volume averaged salinity between the Atlantic and Pacific Oceans suggests changes in freswater transport between these two ocean basins. While the available data and their analyses are insufficient to identify in detail the origin of these changes, the patterns are consistent with a change in the earth's hydrological cycle, in particular with with a larger water transport in the atmosphere from low latitudes to high latitudes and from the Atlantic to the Pacific (see Chapter 3, Section 3.3.2).

5.2.4 Air-Sea Fluxes and Meridional Transports

The changes in heat content discussed above are driven by changes in the air-sea net energy flux. In the global average, we have high confidence that the ocean has been warming at a rate between 0.5 and 1 W m^{-2} from 1993-2003, a value which is computed from the ocean's heat budget (see Section 5.2.2.1) and cannot be directly observed. Observations of changes in air-sea heat fluxes are discussed in Chapter 3, Section 3.5.6.

Estimates of the climatological mean oceanic meridional heat transport derived from atmospheric observations (e.g., Trenberth and Caron, 2001) and from oceanographic cross-sections (e.g., Ganachaud and Wunsch, 2003) are in fair agreement, despite considerable uncertainties (see Appendix 5.A.2). Significant interannual to interdecadal variability in the Atlantic meridional heat transport has been reported in models (e.g., Stammer et al., 2003). Changes of heat transport have been estimated for two latitudes in the North Atlantic. A 30% decrease of heat transport at 25°N during the 1957–2004 has been reported (Bryden et al., 2005), however it is unclear whether this estimate which is based on the analysis of 5 sections is robust. Koltermann et al. (1999) reported an increase of the MHT at 43–48°N from the late 1950s to 1990s by approximately 0.2 PW, with interannual variability being of the same order of magnitude as interdecadal changes; it is however likely that interpretation of these data in the framework of an inverse model would result in less significant variations. Overall, we have low confidence in the observed changes in heat transport.

For the climatological mean meridional freshwater transport, estimates based on an inverse model (Ganachaud and Wunsch, 2003) imply a net evaporation of about $0.5 \times 10^9 \pm 0.3$ kg/s in the Atlantic Ocean. In the other basins, less agreement between different estimates can be found. Overall, we have very low confidence in changes in freshwater transport.

5.3 Changes in Ocean Circulation and Water Masses

5.3.1 Introduction

The large-scale three-dimensional ocean circulation and the formation of water masses that ventilate the thermocline are important oceanic processes that influence the climate system and its response to changes in atmospheric forcing. The processes of water mass formation and the large-scale circulation together create pathways for the transport of heat, freshwater, and dissolved gases such as carbon dioxide from the surface ocean into the interior of the ocean where they are isolated from further interaction with the atmosphere.

Changes in these ocean processes were not covered in the TAR, except in the context of sea-level rise. Since then there has been a significant increase in the number of analyses that report on regional and global change in ocean heat content and water masses, and for the first time changes in other quantities such as the distribution of salinity, oxygen concentration, carbon dioxide and ocean alkalinity. The changing patterns of salinity and emerging patterns of oxygen decrease in thermocline waters in both the Northern and Southern Hemisphere are all new in the AR4.

However, significant uncertainties remain. The main modes of interannual to decadal climate variability are El Niño Southern Oscillation (ENSO), Northern Annual Mode and the related North Atlantic Oscillation (NAO), Pacific Decadal Oscillation (PDO) and the Southern Annular Mode (SAM). These modes which are described in Chapter 3 drive the ocean, causing changes in ocean circulation through changed patterns of the winds and changes in seawater density, and may in turn be affected by ocean dynamics.

Evidence for change in circulation, temperature and salinity is described for each of the major oceans. Three marginal seas are included to provide examples of regional climate change to complement the regional emphasis of Chapter 11. In the cases of the Arctic and Mediterranean, the regional changes also impact the open ocean; while the Japan Sea does not have such impact, it provides an excellent record of climate change. Relation of the changes to climate change and patterns of variability is primarily reserved for the assessment of changes (see Section 5.3.6).

5.3.2 Atlantic and Arctic Oceans

Atlantic Ocean climate variability is to a considerable extent connected to atmospheric forcing associated with the NAO which is part of the Northern Annular Mode (see Chapter 3, Section 3.6.4). On longer time-scales, there is also a relation to the Atlantic Multi-decadal Oscillation (AMO, see Chapter 3, Section 3.6.6.1). The NAO index has increased from minimum values in the late 1960s to strongly positive NAO index values in the early and mid-1990s, and has somewhat declined since.

The linear trends in heat content (Figure 5.2.2) are consistent with the warming tendencies identified from the global analyses of SST (cf. Chapter 3, Section 3.2.2.3), and show a strong warming of the subtropical gyre and a cooling of the subpolar gyre, broadly consistent with predominantly positive NAO during the last several decades. The warming reaches down to over 1,000m, deeper than anywhere in the world oceans (Figure 5.2.3), and is particular pronounced under the Gulf Stream/North Atlantic Current system near 40°N.

Trends in salinity towards freshening in the subpolar regions and increased salinity in the subtropics (Figure 5.3.1) are consistent with the global tendencies for fresher and saltier regions, respectively. These salinity variations reach to the bottom of the North Atlantic and Nordic Seas (Norwegian, Greenland and Iceland Seas), deeper and more pronounced than in any other region of the world.

5.3.2.1 Upper ocean circulation and water property changes

Temperature changes in surface Atlantic waters (upper 200–300 m) (Figure 5.2.2) are consistent with the tendencies identified from the global SST (see Chapter 3, Section 3.2.2.3). In the tropical Atlantic, the surface water changes are partly associated with the variability in the Atlantic Marine Inter-tropical Convergence Zone which has strong seasonal variability (Mitchell and Wallace, 1992; Biasutti et al., 2003; Stramma et al., 2003). Because of the paucity of ocean observational datasets, decadal signals are still poorly detectable and some are at the level of model hypothesis.

Tropical Atlantic variability on interannual to decadal time scales can be influenced by a South Atlantic dipole in SST (Venegas et al., 1998), associated with latent heat fluxes related to changes in the subtropical high (Sterl and Hazeleger, 2003). The South Equatorial Current provides subduction of the water masses

(Hazeleger et al., 2003) and may also maintain propagation of the water mass anomalies toward the north (Lazar et al., 2002).

In the North Atlantic subtropical gyre, SST, the thickness of near-surface Subtropical Mode Water (STMW) (Hanawa and Talley, 2001) and thermocline ventilation are highly correlated with the NAO. Low thickness/production and warmer water occur during high NAO index (Talley, 1996; Hazeleger and Drijfhout, 1998; Joyce et al., 2000; Marsh, 2000). 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 North Atlantic Current carries subtropical waters to the subpolar region. Upper layer temperature and salinity changes in this system are coordinated with SST variability, again mostly governed by the NAO (Deser and Blackmon, 1993; Masina et al., 2004). In addition to oscillations of up to 0.5°C associated with local air-sea flux (Seager et al., 2000), SST is influenced by propagating oceanic signals on the order of 0.5° to 1°C (Dickson et al., 1988; Hansen and Bezdek, 1996; Sutton and Allen, 1997). The decadal changes in the Florida Current transport derived from direct measurements (Baringer and Larsen, 2001) are significantly correlated to the NAO. NAO-induced changes also result in south-north changes of the Gulf Stream position (Joyce et al., 2000; Seager et al., 2001; Molinari, 2004). At 25°N, Bryden et al. (2005) found warming of the upper ocean layer using 5 snapshots from 1957 to 2004 of 1° to 2°C, consistent with results in Section 5.2, a thickening of this warm layer, and a strengthening of the horizontal upper ocean gyre circulation.

Advection of surface salinity anomalies can be a major factor in controlling the depth of convection, particularly in the Labrador Sea, since a freshwater anomaly can severely limit convection. Observed changes are associated with both the NAO and sporadic pulses of fresh water, the latter sometimes called "Great Salinity Anomalies" (GSA) (Dickson et al., 1988; Belkin et al., 1998; Belkin, 2004). During high NAO the subpolar gyre strengthens and expands towards the east, resulting in lower surface salinity in the central subpolar region (Levitus, 1989; Reverdin et al., 1997; Bersch, 2002; Flatau et al., 2003). Three GSAs have been thoroughly documented: 1968-1978, in the 1980s, and in the 1990s. Salinity time series data provided by Dickson et al. (1988) suggest that a GSA occurred as early as 1910. Observational and modelling studies show that the relative influence of local events and advection differ between different GSA events and regions (Houghton and Visbeck, 2002; Josey and Marsh, 2005).

The eastern half of the subpolar North Atlantic has freshened over the past several decades. About two-thirds of the freshening is due to an increase in precipitation associated with a climate pattern known as the Eastern Atlantic Oscillation and the remainder due to GSAs (Josey and Marsh, 2005). The NAO played only a minor role in the freshening. From 1965 to 1995 the subpolar storage anomaly amounted to an equivalent freshwater layer ~3 m thick spread evenly over its total area (Lazier, 1980; Talley and McCartney, 1982; Dickson et al., 1988; Belkin, 2004; Curry and Mauritzen, 2005).

5.3.2.2 Intermediate and deep circulation and water property changes

Marked changes in North Atlantic Deep Water (NADW) in the subtropical North Atlantic over the past 50 years reflect 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).

The Mediterranean Water (MW) found at intermediate depth (800–1200 m) in the North Atlantic reflects the net evaporation and heat exchange within the Mediterranean basin (Section 5.3.2.3). MW affects the Atlantic and global thermohaline circulation (Talley, 1996; Gerdes et al., 1999; Potter and Lozier, 2004) through injection of high salinity at mid-depths into the NADW (e.g., Calmanti et al., 2006). Intermediate waters in the mid-latitude eastern North Atlantic have warmed during last decade (1994–2003) at rates of more than 0.2°C/decade and 0.4°C/decade in some levels (Vargas-Yáñez et al., 2004). In the southern Bay of Biscay (43º N), similar warming rates were observed through the thermocline and into the core of the MW

(González-Pola et al., 2005). In the last decade (1994–2003), a large warming (0.3°C) and salinity increase (0.06) were observed at Gibraltar (Millot et al., 2006). From 1955 to 1993, the trends are around 0.1°C/decade and 0.02 psu/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), much larger than the trends of ~0.01°C/decade within most waters at a global scale (Levitus et al., 2000).

The subpolar North Atlantic freshened at most depths during the past several decades (Dickson et al., 2002; Curry et al., 2003; Dickson et al., 2003, see also Figure 5.3.1). LSW, as a major component of NADW, was decreasing in volume due to several decades of warming with a short interruption in 1976–1977. From 1988/1989 to 1994/1995 6-year long period of winter convection produced an exceptionally large volume of cold, fresh, dense LSW (Sy et al., 1997). From 1994 to 2004, the intensity of deep convection and the production of LSW decreased with a short interruption in 1999–2000. This was associated with the weakening 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 et al., 1996; Dickson, 1997; Lazier et al., 2002; Yashayaev et al., 2003) and diagnosed in models, driven with observed fluxes (Eden and Willebrand, 2001; Gulev et al., 2003; Marsh et al., 2005).

[INSERT FIGURE 5.3.1 HERE]

The densest waters contributing to NADW and the deep limb of the MOC arise as overflows from the upper 1500 m of the Nordic Seas. The marked freshening of the overflow water masses exiting the Arctic was associated with growing sea ice export from the Arctic and precipitation in the Nordic Seas (Dickson et al., 2002; 2003). The transports of the overflow waters have been relatively stable, with no clear variability in Denmark Strait where transport is largest. In the easternmost channel (Faroe Bank), a 20% decrease of the overflows since the 1950s (Hansen et al., 2001) has only a small impact on the total overflow transport since these eastern transports are only a small part of the total. Although the freshening overflows and small decrease in their transport may imply a potential weakening of the MOC, the increased salinity of the surface waters feeding the MOC (Hátún et al., 2005) may offset or even prevent the weakening (Curry and Mauritzen, 2005).

NADW flows southward through the subtropical North Atlantic, where its properties and transport have been measured over the past several decades. Southward transport in layers containing the deep portion of the NADW at 25°N has decreased from 15 Sv to 7 Sv ± 6 Sv (1 Sv = 10^6 m^3/sec) from 1957 to 2004, as derived from hydrographic data collected five times during the period (Bryden et al., 2005). No slowdown in the 10 Sv transport in the layers containing the Labrador Sea component of the NADW was reported.

Not only the coldest water masses of northern sub-polar regions contribute to the variability of the MOC of the whole Atlantic. Surface waters in high latitudes of the South Atlantic set the initial conditions for bottom water in the southern hemisphere. This extremely dense Antarctic Bottom Water (AABW) which is formed by deep-reaching convection in the Weddell Sea (see Section 5.3.5.3) spreads equatorward and enters the Brazil Basin through the narrow Vema Channel of the Rio Grande Rise at 31°S. Ongoing observations of the lowest bottom temperatures there have revealed a slow but consistent increase of the order .002°C/year in the abyssal layer over the last 30 years (Hogg and Zenk, 1997).

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 an important aspect of global climate (see Chapter 4). 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 Sea 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 (Quadfasel et al., 1991; Carmack et al., 1995) has been associated with the shift in the 1980s from low to high NAO. Warm waters have been observed to enter the Arctic from the Atlantic waters as pulses that propagate from the Norwegian Sea through Fram Strait and then along the slope to the Laptev

Sea (Polyakov et al., 2005); the increased heat content and increased transport in the pulses both contribute to net warming (Schauer et al., 2004). 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. Multidecadal variability in the temperature of the Atlantic Water core in the Arctic Ocean has been documented (Polyakov et al., 2004, Figure 5.3.2).

Subsurface salinity in the Nordic Seas has decreased markedly since the 1970s (Dickson et al., 2003), directly affecting the salinity of the Nordic Seas overflow waters that contribute to NADW. In contrast to overall freshening in the subpolar North Atlantic, the salinity of Atlantic inflow to the Nordic Seas has increased to the highest values since observations started in 1948, largely due to a change in the shape of the subpolar gyre which allowed more warm water into the Nordic Seas (Hátún et al., 2005). Within the Arctic, 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 upper waters of Pacific origin has decreased (McLaughlin et al., 1996; Steele and Boyd, 1998).

The large change in Arctic ice cover through the 1990s has accelerated in the present decade (e.g., Comiso, 2003, see Chapter 4, Section 4.4). Melting changes ocean salinity structure and hence vertical stratification, thus changing the conditions for further ice formation and convection. Surface freshening and increase in surface stratification contribute to the freshening of North Atlantic subpolar waters (Figure 5.3.1). 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. 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).

[INSERT FIGURE 5.3.2 HERE]

Within the Mediterranean Sea, water properties and circulation are affected by the long-term variability of surface fluxes (Krahmann and Schott, 1998)), partially associated with the NAO (Hurrell, 1995; Vignudelli et al., 1999), resulting in coordinated changes in surface heat fluxes in the Atlantic and Mediterranean Sea temperatures (Rixen et al., 2005). Marked changes in thermohaline properties have been observed 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 5.3.2), the latter reflected in cooling of the Levantine Intermediate Water between the late 1970s and mid 1980s (Brankart and Pinardi, 2001). 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).

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. 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. This shift in dense water properties has been traced to the Eastern Mediterranean.

Box 5.1: Has the Meridional Overturning Circulation in the Atlantic Changed?

The global Meridional Overturning Circulation (MOC) consists primarily of dense waters that sink to the seafloor at high-latitudes in the North Atlantic Ocean and near Antarctica. These dense waters then spread across the equator with comparable flows of approximately 17 and 14 Sv respectively (e.g., Orsi et al., 2002; L.D. Talley et al., 2003). The North Atlantic MOC is characterized by an inflow of upper ocean waters from the south that gradually densify from 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, sill overflows and vigorous mixing. Through these processes North Atlantic Deep Water is formed, constituting the southward-flowing lower limb of the MOC.

Climate models show that the Earth's climate system responds to changes in the MOC (e.g., Vellinga and Wood, 2002), and also suggest that the MOC might gradually decrease in the 21st century as a consequence of anthropogenic warming and freshening in the North Atlantic (Bi et al., 2001; Gregory et al., 2005), see also Chapter 10). However, only indirect estimates of the MOC strength and variability exist, and the best evidence for observational changes in the overturning circulation comes from the North Atlantic.

There is evidence for a link between MOC and abrupt changes in surface climate during the past 120,000 years, although the exact mechanism is not clear (Clark et al., 2002). At the end of the last glacial, as the climate warmed and ice sheets melted, there were a number of abrupt oscillations, e.g., the Younger Dryas and the 8.2 ka cold event (see Chapter 6, Section 6.3), which may have been caused by the ocean circulation. The variability of the MOC during the Holocene after the 8.2 ka cooling event is clearly much smaller than during glacial times (Keigwin et al., 1994, see Section 6.4).

Based on the analysis of five hydrographic sections and Gulf Stream transport observations, Bryden et al. (2005) concluded that the MOC transport in the North Atlantic at 25°N has decreased by 30% between 1957 and 2004, indicating a stronger mid-ocean return flow in the upper km though not a decrease in Gulf Stream strength. Note however that this result is based on 5 snapshots in time, and it is not clear whether the trend estimate can be viewed as robust in the presence of considerable variability.

Indirect evidence from observed changes water mass formation in the subpolar North Atlantic and Nordic Seas also reveals changes of the MOC. While direct measurements of the sill overflows (see Section 5.3.2.2) have been interpreted as a sign of decreasing MOC (Hansen et al., 2004), due to the short-term observation periods (a few years) it is unclear whether these observations reflect long-term trends. The observed freshening and associated reduction in density of intermediate and deep waters in the subpolar gyre and Nordic Seas (see Section 5.3.2) would in principle lead to a weakening of the MOC; based on results from various models, Latif et al. (2006) concluded that the observed freshening would correspond to a reduction of not more than 5–10% of the MOC strength. Wu et al. (2004) found fresehening but not MOC decrease in a coupled model simulation.

MOC changes can also be caused by changes in Labrador Sea convection, with strong convection corresponding to higher MOC. Convection has been strong since the 1970s, but after 1995 the Labrador Sea has warmed and re-stratified (Lazier et al., 2002), and convection has been weaker. From direct current meter observations at the exit of the subpolar North Atlantic, Schott et al. (2004) found that the Deep Water outflow, while varying at shorter time scales, had no significant trend during the 1993–2005 period.

Based on a model-based relation of MOC transport with interdecadal SST patterns, it has been concluded that the MOC has increased since the 1970s (Knight et al., 2005; Latif et al., 2006), in contrast to the Bryden et al. (2005) results. Note that the interpretation of these SST patterns could be ambiguous due to the presence of an anthropogenic climate change signal, however Latif et al. (2004) maintain that the relation of their SST index to MOC at 30°N is maintained in model scenarios of rising CO_2.

While it is very likely that up to the end of the 20th century the MOC has been changing at interannual to decadal time scales, so far there is only a low level of confidence that the strength of deep limb of the MOC in the North Atlantic MOC has actually decreased. This low level of confidence results from structural uncertainties in the observational records and in the models/techniques needed to establish changes in the MOC.

5.3.3 Pacific Ocean

Increased heat content and decreased salinity in the upper Pacific are revealed in global heat and freshwater analyses (see Section 5.2). The subtropical North and South Pacific have been warming; in the Southern Hemisphere the major warming footprint is associated with the thick mode waters north of the Antarctic Circumpolar Current. The North Pacific has cooled around 40°N, which along with subtropical warming, stronger Kuroshio and more penetrating Oyashio, suggests a positive Pacific Decadal Oscillation (PDO) state (see Chapter 3, Section 3.6.3). Warming of the Pacific deep waters has also occurred, and while the ocean process causing these changes is uncertain, it could be associated with warming mid-depth Antarctic waters and/or warming North Atlantic Deep Water.

5.3.3.1 Pacific upper ocean changes

Most literature on climate in the upper 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). In the North Pacific, the zonally-averaged temperature trend from 1955 to 2003 (Figure 5.2.3b) is dominated by the PDO shift in the mid 1970s. The figure shows the importance of the tropical Pacific in climate change; 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 Section 5.3.3.1 and a relaxation of the equatorial thermocline (McPhaden and Zhang, 2002).

Warming in the North Pacific subtropics, cooling 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 heat content trend in the tropical and North Pacific shown in Section 5.2 is most likely related to the prevalence or enhancement of positive PDO states in recent decades. Salinity variations in the North Pacific have similarly complicated spatial distributions, although Figure 5.2.6b suggests that on the whole the region has freshened.

Trends towards increased heat content, indicative of global change, include a major signal in the subtropical South Pacific, within the thick mixed layers just north of the Antarctic Circumpolar Current (Willis et al., 2004, Section 5.3.5). The strength of the South Pacific subtropical gyre circulation has increased more than 20% over the 1990s, peaking in 2003, and subsequently declined (e.g., Roemmich et al., 2006). This spinup is linked to an increase of Ekman pumping over the gyre due to an increase in the Southern Annular Mode (SAM).

Within the North Pacific Ocean, a positive PDO state such as followed the shift in 1976, is characterized by a strengthened Kuroshio Extension. After 1976, the Kuroshio Extension and North Pacific Current transport increased by 8% and expanded southward (Parrish et al., 2000). The Kuroshio's advection of temperature anomalies has been shown to be of similar importance in maintaining the positive PDO as are variations in ENSO and the Aleutian Low strength (Schneider and Cornuelle, 2005). 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 (Sekine, 1988; Hanawa, 1995; Sekine, 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). Similarly, mixed layer depth decreased throughout the eastern subarctic gyre, with a trend verified over 50 years (Freeland et al., 1997; Li et al., 2005).

Temperature changes in upper ocean water masses in response to the stronger PDO after 1976 are well documented. 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 (Yasuda and Hanawa, 1997; Yasuda et al., 2000; Hanawa and Kamada, 2001). 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).

5.3.3.2 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 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 (Wong et al., 2001; Joyce and Dunworth-Baker, 2003). An oxygen decrease and nutrient increase in the NPIW south 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.

Bottom waters in the North Pacific are farther from the surface sources than any other of the world's deep waters. They are also the most uniform, in terms of spatial temperature and salinity variations. A large-scale, significant warming of the bottom 1000 meters across the entire North Pacific on the order of 0.002°C occurred between 1985 and 1999, measurable because of the high accuracy of modern instruments (Figure 5.3.3, 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), or from declining transport of bottom waters into the deep North Pacific (Johnson et al., 1994).

[INSERT FIGURE 5.3.3 HERE]

5.3.3.3 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; Minami et al., 1998; L. D. Talley et al., 2003; Kim et al., 2004). 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).

Because of weakened vertical stratification at mid-depths 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; L. D. Talley et al., 2003; Tsunogai et al., 2003).

5.3.4 Indian Ocean

The Indian Ocean is sparsely observed with respect to documenting change in the general circulation and subsurface water properties. Changes in upper ocean temperature and salinity averaged zonally across the Indian Ocean were presented in Section 5.2, showing a warming trend throughout, except in a band centered at about 12°S (the South Equatorial Current). Model results suggest that upper ocean warming in the southern Indian Ocean can be attributed to anthropogenic forcing (Barnett et al., 2005), whereas in the northern Indian Ocean warming was attributed to a change in advection (Lee, 2004).

In the tropical and eastern subtropical Indian Ocean (north of 10°S), warming in the upper 100m (Figure 5.2.3c) is consistent the significant warming of the sea surface from 1900–1999 (see Chapter 3, Section 3.2.2.3). The surface warming trend in the period 1900–1970 was relatively weak, and increased significantly in the 1970–1999 period, with some regions exceeding 0.2°C/decade.

Transport of warm, relatively fresh waters from the Pacific to the Indian Ocean through the Indonesian passages is part of the global scale overturning circulation. Flow into the Indonesian Seas is primarily through Makassar and Lifamatola Straits. Transport measured at Makassar from 1996 to 1998 was 9–10 Sv (Vranes et al., 2002), matching transports exiting the Indonesian Seas (e.g., Sprintall et al., 2004). Large variability is associated with varying tropical Pacific and Indian winds (Wijffels and Meyers, 2004) and may be associated with changes in SST in the tropical Indian Ocean.

Changes in Indian subtropical gyre circulation since the 1960s include a 20% slowdown from 1962 to 1987 (Bindoff and McDougall, 2000) and a 20% speedup from 1987 to 2002 (Bryden et al., 2003; McDonagh et al., 2005), with the speedup mainly between 1995 and 2002 (Palmer et al., 2004). Associated with these circulation changes are changes in oxygen on density surfaces: a decrease during the slowdown and increase during the speedup, suggesting a direct ventilation response. During the slowdown, the upper thermocline warmed and became fresher on density surfaces (Bindoff and McDougall, 2000; Bryden et al., 2003). The opposite occurred during speedup, suggesting that the earlier slowdown was part of an oscillatory pattern of variation in upper part of this gyre over periods of decades. On the other hand, the lower thermocline (< 10°C) has continued to freshen and warm consistently from 1936 to 2002 (Bryden et al., 2003) consistent with heat content increases discussed in Bindoff and McDougall (2000) and Section 5.2.

5.3.5 Southern Ocean

The Southern Ocean connects the Atlantic, Indian and Pacific Oceans together allowing inter-ocean exchange. This ocean is also active in the formation and subduction of waters that contributed strongly to the storage of Anthropogenic Carbon Dioxide and heat (see Section 5.2). Note that some observed changes found in the Atlantic, Indian and Pacific Oceans are related to changes in the Southern Ocean waters but have largely been described in those sections.

5.3.5.1 Upper ocean property changes

The upper ocean in the Southern Hemisphere is dominated by the thick near-surface layers called Subantarctic Mode Water (SAMW). Analyses of SAMW in the Indian and Pacific Oceans and Tasman Sea have shown that it has warmed since the 1960s. The observed warming of SAMW is consistent with the subduction of warmer surface waters from the south in the Southern Ocean (Bindoff and Church, 1992; Johnson and Orsi, 1997; Wong et al., 2001; Aoki et al., 2003). The upper-thermocline layer (including the lighter layers of SAMW) has cooled since 1987 (Bryden et al., 2003; McDonagh et al., 2005). The general increase of temperature of deeper (and denser layers) of SAMW from these regional analyses is also consistent with the results from global trends of warming at these depths (<700 m) between the Subantarctic Front and the centre of the subtropical gyres in the Southern Hemisphere since the 1950s (Levitus et al., 2005a) and in the 1990's (Willis et al., 2004).

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. 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 40°S is larger than at any other latitude (Willis et al., 2004). This warming is also found in the long-term temperature trends (Section 5.2, Figure 5.2.3d).

The major mid-depth water mass in the Southern Hemisphere, Antarctic Intermediate Water (AAIW), has been freshening since the 1960's (Bindoff and Church, 1992; Wong et al., 1999; Bindoff and McDougall, 2000; Aoki et al., 2005a). The Atlantic freshening of AAIW is also supported by direct observations of a freshening of southern surface waters (Curry et al., 2003).

In the Upper Circumpolar Deep Water (UCDW) in the Indian and Pacific sectors of the Southern Oceans, temperature and salinity have been increasing (on density surfaces) and oxygen has been decreasing between the Subtropical Front at ~35°S and the Antarctic Divergence at ~60°S (Aoki et al., 2005a). These changes just below the mix layer (~100 to 300 m) are consistent with the mixing of warmer and fresher surface waters with UCDW, suggesting an increase in stratification in the surface layer in this region.

5.3.5.2. Antarctic regions

The deep and bottom water properties of the Weddell Sea varied in the 1990s (Robertson et al., 2002; Fahrbach et al., 2004). Water at intermediate depth (Warm Deep Water) became warmer and saltier in the early 1990s and has since cooled and freshened; dense Weddell Sea Bottom Water warmed by 0.01°C per year from 1990 to 1996, after which temperatures leveled off. Meredith and King (2005) presented evidence that the upper ocean adjacent to the West Antarctic Peninsula has also warmed and become more saline. 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 these source regions (Hogg, 2001; Andrie et al., 2003).

There is also growing evidence for the changes in the Antarctic Bottom Waters in the Ross and Australian-Antarctic basins. Significant decreases in the salinity of 0.003 per year (and density decreases) over the last four decades of waters adjacent to the Ross Ice Shelf (Jacobs et al., 2002) have been observed. Downstream of the Ross Ice Shelf in the Australian-Antarctic basin Antarctic Bottom Waters have also cooled and freshened (Aoki et al., 2005b). These observed decreases are significantly greater than earlier reports of Antarctic Bottom Water variability (Whitworth, 2002) and suggests that changes in the Antarctic shelf waters can be quite quickly communicated to deep waters. Jacobs et al. (2002) concluded that the freshening appears to have resulted from a combination of factors including increased precipitation, reduced sea ice production, and increased melting of the West Antarctic Ice Sheet.

5.3.5.3 Antarctic circumpolar current
The transport of the ACC through Drake Passage is about 130 Sv, with significant interannual variability. Measurements over 25 years at the South American choke point (Drake Passage) showed no evidence for a systematic trend in total volume transport between the 1970s and the present (Cunningham et al., 2003).

5.3.6 Assesment of Changes

5.3.6.1 Changes in global water mass properties
The regional analyses described in the previous section also have a global scale. The water mass changes in the Pacific and Atlantic oceans from the World Ocean Circulation Experiment have a strikingly similar pattern of change (Figure 5.3.4). These two North–South sections that span the whole of the Atlantic (Figure 5.3.4a) and Pacific Oceans (Figure 5.3.4 b) cover roughly the same period of time from the 1960s to 1990s. They show that the subtropical waters have increased in salinity in both hemispheres and also in both the Atlantic and Pacific Ocean basins. The waters that underly the near surface subtropical waters have freshened, and in particular the fresh intermediate water layer (at ~1000 m) in the Southern Hemisphere has freshened in both the Atlantic and Pacific Oceans. In the Northern hemisphere the Pacific intermediate waters have freshened, and the underlying deep waters are unchanged consistent with no sources of bottom waters in the North Pacific. In the North Atlantic, as discussed in Section 5.3.2, the Labrador Seas show a distinct freshening signal. These similiarity of the salinity differences, over a wide range of depths and water masses between different ocean basins illustrate the global scale of the water mass changes.

[INSERT FIGURE 5.3.4a HERE]

[INSERT FIGURE 5.3.4b HERE]

5.3.6.2 Consistency of regional and global analyses
The available global comprehensive data sets (see Section 5.2) and the many regional research oceanographic analyses in the Atlantic, Pacific, Indian and Southern Ocean (where available) tend to show similar changes over the last 30–50 years. The similarity of changes analysed from different data sets, using widely varying methods gives us greater confidence that the changes found in both the global analyses and regional analyses are indeed real.

The global and regional analyses of ocean warming generally show a pattern of increased ocean temperature in the regions of mode water formation (with clearest signals of change in North Atlantic and North Pacific and Indian Sector of the Southern Ocean (Figure 5.2.2). There are also regions of decreased ocean temperature in both the global and regional analyses in parts of the subpolar and equatorial regions.

Both the global analyses and the regional analyses show a strong freshening in the subpolar waters in the North Atlantic and Labrador Sea since the 1960's; these extend throughout the entire water column (Figure 5.2.6a and Figure 5.3.4a). A similar freshening is observed in the North Pacific regional analyses of the subpolar gyre (north of 45°N, Section 5.3.3.2) and also from the global analyses (Figure 5.2.6b). The evidence for freshening of intermediate depth waters (>300 m) from Southern Ocean sources (see Section 5.3.5) is supported in both the global and regional analyses (e.g., Figure 5.2.6 and Figure 5.3.4a and b). The increase in the near surface salinity fields (<100 m deep) in the mid-latitudes in the regional analyses supports the results from the global analyses, except in the case of the North Pacific sub-tropical gyres (as discussed in Section 5.3.3.4). Freshening in the high latitudes of both hemispheres and a salinity increase in the upper ocean at low latitudes suggest an increase in the atmospheric hydrological cycle.

Many of the observed changes in the temperature and salinity fields have been linked to atmospheric forcing through correlations with atmospheric indices associated with NAO, PDO and SAM. Indeed, of the few time series ocean measurements or repeat measurements of long sections (see Sections 5.3.2 and 5.3.4), most show evidence of significant decadal variability. Because of the long timescales of these natural modes, it is difficult to discern if observed oceanic variability is natural or a climate change signal; indeed changes in these natural modes themselves could well be related to climate change. In the North Atlantic, freshening at high latitudes and increased evaporation at subtropical latitudes may be associated with an extended positive NAO. Likewise in the Pacific, freshening at high latitudes and increased evaporation in the subtropics, cooling in the central North Pacific, warming in the eastern and tropical Pacific, and reduced ventilation in the Kuroshio region, Japan and Okhotsk Seas could be associated with an extended positive PDO.

On a global scale, the observed long term patterns of zonal temperature and salinity field tend to be approximately symmetric around the equator (Figure 5.3.4 a and b) and occurring simultaneously in different ocean basins (e.g., Section 5.2, Figure 5.2.4, Figure 5.2.6, this Section, Figures 5.3.4 a and b). The global nature of these patterns and their scale which extends beyond the regions of influence normally associated with NAO, PDO and SAM suggests that these coherent changes between hemispheres are associated with a global phenomenon (see Chapter 9).

5.3.6.3 Consistency with the large-scale ocean circulation

The observed changes are broadly consistent with our understanding of the circulation of the global oceans. The regions where the oceans ventilate the deep waters on short time-scales (<50 years) show strong evidence of change over the instrumental record. For example the deep overturning circulation in the North Atlantic shows evidence of a deep penetrative warming and freshening. There is evidence of change in the Southern Ocean bottom waters consistent with the sinking of fresher waters Antarctic shelf waters. The deep waters, remote from interaction with the atmosphere, and with replenishment rates that are long compared with the instrumental record typically show no significant changes. Mode waters, a key global water mass, found in every ocean basin equator-ward of major oceanic frontal systems or separated boundary currents have a relatively rapid formation and ventilation rates (<20 years) and provide a pathway for heat (and salinity) to be transported into the main subtropical gyres of the global oceans as observed.

5.4 Ocean Biogeochemical Changes

5.4.1 Introduction

The observed increase in atmospheric CO_2 (see 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.

The increase in surface ocean CO_2 has consequences for the chemical equilibration of the ocean. As CO_2 increases, the acidity of the water increases and the concentration of carbonate ions decreases. The consequences of this change in chemical equilibration for the carbon cycle are poorly known primarily because of the unknown response of marine ecosystems, see Chapter 7, Section 7.3.2.2.

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 of the order of 10^{-6} its mean concentration).

Changes in O_2 thus provide information on the changes in the physical or biological processes that occur within the ocean, such as ventilation, mode water formation, upwelling, or biological export and respiration. Furthermore, changes in the oceanic O_2 content are needed to estimate the CO_2 budget from atmospheric measurements, a method which currently estimates the O_2 changes based on heat fluxes alone (see Section 7.3.3).

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 Total carbon change in the water column and air-sea CO_2 flux

Links between the main modes of climate variability and the marine carbon cycle have been observed on interannual time-scales in several regions of the world (see Chapter 7, Section 7.3.2.4 for quantitative estimates). In the equatorial Pacific, the reduced upwelling associated with El Niño events decreases the ventilation of Dissolved Inorganic Carbon (DIC) and the regional outgas of CO_2 to the atmosphere (Feely et al., 1999). In the sub-tropical North Atlantic, the reduced water formation during the positive phase of the North Atlantic Oscillation increases the storage of carbon in the intermediate ocean (Bates et al., 2002). These observations show that variability in the content of carbon in the ocean has occurred in association with climate variability although their variability occurs on relatively short time scales.

Longer observations exist for the partial pressure of CO_2 (pCO_2) at the surface only. Over more than two decades the oceanic pCO_2 increase has generally followed the atmospheric CO_2 within the given uncertainty, although regional differences have been observed (Feely et al., 1999; Inoue and Ishii, 2005; Takahashi et al., 2006). The three longest time-series stations all in the northern sub-tropics show pCO_2 increases at a rate varying between 1.6 and 1.9 μatm yr^{-1} (Figure 5.4.1), undistinguishable from the atmospheric increase of 1.5 to 1.9 μatm yr^{-1}. Variability on the order of 20 μatm over periods of five years was observed in the three times series, as well as in other data sets, and has been associated with regional changes in ocean circulation and climate variability (Gruber et al., 2002; Dore et al., 2003) or with variations in biological activity (Lefèvre et al., 2004). The air-sea CO_2 flux remains constant as long as oceanic and atmospheric CO_2 increase at the same rate. The difference between the increase in oceanic and atmospheric CO_2 indicates if there has been a change in the air-sea CO_2 fluxes, and thus in the oceanic sink of CO_2. It is not yet possible to detect if there has been large-scale changes in the global oceanic CO_2 sink from direct observations because of the large influence of climate variability, although inverse methods that estimate the oceanic CO_2 sink from the spatio-temporal distribution of atmospheric CO_2 suggest that the global CO_2 sink increased by 0.1 to 0.6 GtC yr^{-1} between the 1980s and 1990s, consistent with results from ocean models (Le Quéré et al., 2003).

[INSERT FIGURE 5.4.1 HERE]

5.4.2.2 Anthropogenic carbon change in the water column

The recent uptake of *anthropogenic* carbon in the ocean is well constrained by observations to a decadal mean of 2.2 ± 0.4 GtC yr^{-1} for the 1990s (see Chapter 7). The uptake of anthropogenic carbon over longer time scales can be estimated from measurements in the water column. The direct measure of changes in DIC between two time periods reflects the anthropogenic carbon uptake plus the changes in carbon concentration due to changes in water masses and biological activity. To estimate the contribution of anthropogenic carbon 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; Sabine et al., 2004b), 200–1200 m in the Indian Ocean (Peng et al., 1998; Sabine et al., 1999), and 1900 m in the Southern Ocean (McNeil et al., 2003).

Indirect methods have been developed to estimate anthropogenic carbon based on observations from one single time period. The method corrects the observed DIC concentration for organic matter decomposition and dissolution of carbonate minerals, and removes an estimate of the pre-formed DIC (Gruber et al., 1996). The pre-formed DIC is the DIC concentration of the water when it was last in contact with the atmosphere. With this method, Sabine et al. (2004a) estimated a global DIC increase of 118 ± 19 GtC between 1750 and

1994, using 9618 profiles from the WOCE/JGOFS survey in the 1990s (Figure 5.4.2). The uncertainty in the global inventory estimated by Sabine et al. (2004a) is based on uncertainties in the anthropogenic DIC estimates and mapping errors. Potential biases in the technique have been identified mostly caused by assumptions about constant air-sea pCO_2 disequilibrium. While magnitude and direction of all potential biases are not yet clear, the given uncertainty of ±16% appears realistic compared to the biases already identified (see Appendix 5.A.3).

[INSERT FIGURE 5.4.2 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, and it is undetectable in the deep ocean (Figure 5.4.3). The vertical penetration of anthropogenic carbon from Sabine et al. is consistent with the DIC changes observed between two cruises (Peng et al., 1998; 2003). Anthropogenic carbon has penetrated deeper in the North Atlantic and sub-Antarctic Southern ocean compared to other basins, due to a combination of i) high surface alkalinity (in the case of the Atlantic) which favors the uptake of CO_2, and ii) more active vertical exchanges caused by intense winter mixing and by the formation of deep waters (Sabine et al., 2004a). The deeper penetration of anthropogenic carbon in these regions is consistent with similar features observed in the oceanic distribution of CFCs of atmospheric origin (Willey et al., 2004), confirming that it takes decades to many centuries to transport carbon from the surface into the thermocline and the deep ocean. Deeper penetration in the North Atlantic and sub-Antarctic Southern Ocean is also observed in the changes in heat content shown in Figure 5.2.2. The largest storage of anthropogenic carbon is observed in the subtropical gyres because of the lateral transport of carbon from the region of mode water formation towards the lower latitudes (Figure 5.4.2).

[INSERT FIGURE 5.4.3 HERE]

The fraction of the CO_2 emissions taken up by the ocean (the uptake fraction) was possibly lower during 1980–2005 (0.37 ± 0.07) compared to 1750–1994 (0.42 ± 0.07); note however that the uncertainty in the estimates is larger than the difference between the estimates (Table 5.4.1). This would be consistent with our understanding that the ocean CO_2 sink is limited by the transport rate of anthropogenic carbon from the surface to the deep ocean, and also with the non-linearity in carbon chemistry that reduces the CO_2 uptake capacity of water at high concentrations (Sarmiento et al., 1995). Thus over the 245 years spanning the 1750–1994 period, the ocean had more time to equilibrate with the atmosphere than during the 25 years spanning the later 1980-2005 period. The uptake fraction diminishes as the rate of atmospheric CO_2 increases.

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

Time Period	Oceanic Increase (GtC)	Net CO_2 Emissions[a] (GtC)	Uptake Fraction	Reference
1750–1994	118 ± 19	283 ± 19	0.42 ± 0.07	Sabine et al., 2004
1980–2005 [b]	51 ± 9	136.5 ± 9	0.37 ± 0.07	Chapter 7 [c]

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.
(b) The longest possible time-period was used for the recent decades to minimise the effect of the large variability in atmospheric CO_2.
(c) Sum of the estimates for the 1980s, 1990s and 2000–2005 from Table 7.3.3.

5.4.2.3 Ocean acidification by carbon dioxide

The uptake of anthropogenic carbon by the ocean changes the chemical equilibration of the ocean. Dissolved CO_2 forms a weak acid[1]. As CO_2 increases, pH decreases (i.e., acidity increases). pH can be computed from measurements of DIC and alkalinity. A decrease in pH by 0.1 over the global ocean was calculated from the uptake of anthropogenic carbon between 1750 and 1994 (Raven et al., 2005), with the lowest decrease of 0.06 in the tropics and subtropics, and the highest decrease of 0.12 at high latitudes. For comparison, the

[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.

mean pH of surface waters ranges between 7.7 and 8.3 in the open ocean, pH was higher by 0.1 unit during glaciations, and there are no evidence of pH lower than 0.6 units in the past 300 million years (Caldeira and Wickett, 2003). A decrease in ocean pH of 0.1 units corresponds to a 30% increase in the concentration of H^+ in seawater, assuming that alkalinity and temperature remain constant. Changes in surface temperature may have induced an additional decrease in pH by <0.01. The calculated anthropogenic impact on pH is consistent with results from time-series stations which observed a decrease in pH of 0.02 per decade (Figure 5.4.1). Results from time-series stations include not only the increase in anthropogenic carbon, but also other changes due to local physical and biological variability. The consequences of changes in pH are poorly known; see Chapter 7, Section 7.3.2.2.

5.4.2.4 Change in carbonate species

The uptake of anthropogenic carbon occurs through the injection of CO_2 and causes a shift in the distribution of carbon species (i.e. the balance between CO_2, carbonate, and bicarbonate). 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 calcite or aragonite saturation horizons (the shallowest depth where $CaCO_3$ is under-saturated) or under the action of biological activity.

Shoaling of the aragonite saturation horizon was observed 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 that the uptake of anthropogenic carbon has caused 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), based on the anthropogenic DIC increase estimated by Sabine et al. (2004b), 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. These results are consistent with the calculated increase in $CaCO_3$ dissolution as a result of the shoaling of the aragonite saturation horizon, but with large uncertainty. Carbonate decreases at high latitudes and particularly in the Southern Ocean may have consequences for marine ecosystems because the current saturation horizon is closer to the surface than in other basins (Orr et al., 2005), see Chapter 7, Section 7.3.2.2.

5.4.3 Oxygen - Biogeochemical Aspects

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 or later in the North and South Pacific, North Atlantic, and Southern Indian Oceans (see summary table in Emerson et al., 2004). The reported O_2 decreases range from 0.1 to 6 μmol kg^{-1} yr^{-1}, superposed on decadal variations of ±2 μmol kg^{-1} yr^{-1} (Andreev and Kusakabe, 2001; Ono et al., 2001; Andreev and Watanabe, 2002).

The recent O_2 decreases are everywhere paralleled by an increase in Apparent Oxygen Utilisation (AOU, Deutsch et al., 2005). AOU is the difference between the O_2 that is at equilibrium with the local temperature and the observed O_2, in effect removing the direct impact of temperature on O_2. An increase in AOU can only be caused by reduction in ventilation or by increased biological activity. All studies indicate that the O_2 decrease and AOU increase are consistent with reduction in ocean ventilation from physical processes. A few studies have quantified the change in ventilation using estimates of changes in apparent CFC ages (Doney, 1998; Watanabe et al., 2001; Mecking et al., 2006). In nearly all cases, the increase in AOU could entirely be accounted for by the reduced apparent CFC age which resulted from reduced ventilation of intermediate waters. Changes in biological processes were only significant at the coast of California and may result from assumptions in the method (Mecking et al., 2006).

It is unclear whether the recent changes in O_2 are indicative of trends or of variability. Recent data in the Indian ocean have shown a reversal of the O_2 decrease between 1987 and 2002 in the South Indian ocean (McDonagh et al., 2005), as well as large variability on decadal time-scales in the North Atlantic (Johnson and Gruber, 2006).

In the upper 100 m of the ocean surface, decadal variations of 0.5 mol kg^{-1} in O_2 concentration were observed for the 1956–1998 period, with no clear trends (Garcia et al., 2005). The surface changes in O_2 concentration are paralleled by opposite changes in AOU. Surface AOU changes are damped by the equilibration with the atmosphere and thus they are smaller than the changes in the intermediate waters.

Changes in surface AOU are difficult to interpret. They can be caused by changes in biological activity or by changes in the physical transport of AOU from intermediate waters. The changes in physical transport can originate either from a change in the amount of water that is mixed with the surface, and/or from a change in the AOU concentration of the underlying water being mixed. These processes can oppose each other and we have no independent information to assess if one process dominates over the others. The accuracy of the O_2 measurements in the early decades is difficult to determine because of biases reported with the use of metal flasks prior to about 1970 (see Appendix on methods and errors). Because we have little confidence in the early measurements and cannot explain the reported changes with known processes, we cannot say if the absence of a long-term trend in surface O_2 is realistic or not.

5.4.4 Nutrients

Only a few studies reported decadal changes in nutrient concentrations. In the North Pacific, the concentration of N and P decreased at the surface (Freeland et al., 1997; Watanabe et al., 2005) and increased at the sub-surface (Emerson et al., 2001; Ono et al., 2001; Keller et al., 2002)in the past two decades. Nutrient changes were observed in the deep ocean but no clear pattern emerges from available observations. Pahlow and Riebesell (2000) used a large dataset and found changes in the ratio of nutrients in the North Pacific and Atlantic oceans, and no significant changes in the South Pacific. In the North Pacific, Keller et al. (2002) observed a decrease in N associated with the increase in O_2 between 1970 and 1990 at 1050 m, opposite to Pahlow and Riebesell's longer study. Using the same data set extended to the world, Li (2002) observed large regional changes in nutrient ratios but no consistent basin-scale patterns. Uncertainties in deep ocean nutrient observations may be responsible for the lack of coherence in the nutrient changes. Source of inaccuracy include the limited number of observations, and the limited comparability of measurements from different laboratories at different times.

In same cases, the observed trends can be explained by either a change in thermocline ventilation or a change in biological activity (Pahlow and Riebesell, 2000: Emerson, 2001 #167),, but in other cases are mostly consistent with a reduction in thermocline ventilation (Freeland et al., 1997; Ono et al., 2001; Watanabe et al., 2005).. Thus all trends reported are consistent with a physical explanation of the observations, although changes in biological activity cannot be ruled out.

5.4.5 Biological changes

The changes in biological activity are difficult to quantify at the global scale. Marine export production (the fraction of primary production that is not respired in the ocean surface and thus sinks to depth) is the biological process that has the largest influence on the cycles of elements. There are no global observations on changes in export production, and neither of large-scale changes in respiration or of changes in the total Dissolved Organic Matter (DOM) pool in the ocean. However, estimates of changes in primary production provide partial information. A reduction in global primary production by ~6% between the early 1980s and the late 1990s was estimated based on the comparison of chlorophyll data from two satellites (Gregg et al., 2003). The largest changes occurred at high latitudes and were suggested to be associated with warming and decreased iron deposition in the North hemisphere, and with increased wind stress in the South Hemisphere. Although the potential errors in this estimate are large, a change in biological fluxes of this order of magnitude is fully plausible considering that biological production is controlled primarily by nutrient input from intermediate waters, and that a large decrease in ventilation has been observed during that period as indicated by the decrease in oxygen content. This decrease in biological production is comparable in amplitude to the 10% increase estimated between 1997 and 2000 also based on satellite observations (Behrenfeld et al., 2001). Shifts and trends in plankton biomass have been observed for instance in the North Atlantic (Beaugrand and Reid, 2003), the North Pacific (Karl, 1999; Chavez et al., 2003), and in the Southern Indian Ocean (Hirawake et al., 2005), but the spatial and temporal coverage is limited. The exact processes that link environmental variables to changes in plankton biomass and primary production are not well known. The potential impact of changes in marine ecosystems or DOM on climate are discussed in Chapter 7, Section 7.3.4, and the impact of climate on marine ecosystems in Chapter 4 of WGII.

5.4.6 Consistency with Physical Changes

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

Large scale changes in the O_2 content of the thermocline have been observed between the 1970s and late 1990s. These changes are everywhere consistent with the local changes in ocean ventilation as identified either by changes in density gradients or by changes in apparent CFC ages. Nevertheless an influence of changes in marine biology cannot be ruled out. The available data is insufficient to say if the changes in O_2 are caused by natural variability or are trends that are likely to persist in the future, but it does point out that large scale changes in ocean physics have an influence on natural biogeochemical cycles, and thus that the cycles of O_2 and CO_2 are likely to undergo changes if ocean circulation (and particularly ventilation) persist in the future.

The observed variability in surface O_2 concentration is not fully explained. It could be caused by large-scale changes in biological activity, which have not been quantified, or by changes in physical circulation that would have larger impact on O_2 than on temperature. The observed changes in surface temperature and salinity (see Section 5.3.2) are indicative of changes in the surface mixing, which has important consequences for ocean biogeochemistry (see Chapter 7, Section 7.3.2.2). In most of the Pacific Ocean, surface warming and freshening act in the same direction and contribute to reduced mixing (Figures 5.2.2 and 5.2.6). Regional observations indicate that mixing indeed decreased in the North Pacific since 1970 in the west (Watanabe et al., 2005) and since 1960 in the East (Freeland et al., 1997), although a large-scale analysis is missing. In the Atlantic and Indian Oceans, temperature and salinity trends generally act in opposite direction and changes in mixing have not been quantified regionally.

Oceanic *changes* in N_2O and CH_4. have not been assessed because of the lack of large scale observations. Observations of the mean fluxes of N_2O and CH_4 (including CH_4 clathrates) are covered in Chapter 7.

5.5 Changes in 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. This section focuses on global and regional sea level variations, over time spans ranging from the last decade to the past century; a brief discussion of sea level change in previous centuries is given in Section 5.5.2.5, and changes over previous millennia are discussed in Chapter 6, Section 6.3.3.

Sea level change integrates non-linear coupled responses of several components of the earth's system, so that measuring sea level variations and studying processes that cause them is a highly interdisciplinary endeavour. On decadal and longer time scales, global mean sea level change results from two major processes that alter the volume of the global ocean: i) thermal expansion which is the largest effect, and ii) the exchange of water between oceans and other reservoirs (e.g., glaciers and ice caps, ice sheets, other land water reservoirs and atmosphere). In addition to factors related to recent climate change, other processes affect the mean sea level such as anthropogenic change in land hydrology or vertical land movements such as resulting from Glacial Isostatic Adjustment (GIA). All these processes cause geographically non-uniform sea level change as well as changes in the global mean; some oceanographic factors (e.g., changes in ocean circulation or atmospheric pressure) also affect sea level at the regional scale, while not contributing to changes of the global mean.

Measurements of present-day sea level change rely on two different techniques: tide gauges and satellite altimetry. Tide gauges provide sea level variations with respect to the land on which they lie. However, the Earth's crust is subject to various vertical motions due to tectonics, local subsidences etc.; these motions need to be removed from the tide gauge measurement to extract the sea level signal. Sea level change based on satellite altimetry is measured with respect to the earth's center of mass, and thus is not distorted by land motions, except for a small component due to large scale deformation of ocean basins from Glacial Isostatic Adjustment (GIA).

The chapter on sea level change of the Third Assessment Report (TAR) (Church et al., 2001) provided estimates of climate and other anthropogenic contributions to 20th century sea level rise, based mostly on models. The sum of these contributions ranged from –0.8 to 2.2 mm yr^{-1}, with a median value of 0.7 mm yr^{-1}, and a large part of this uncertainty was due to the lack of information on anthropogenic land water change. For observed 20th century sea level rise, Church et al. (2001) adopted as a best estimate a value of 1.5 ± 0.5 mm yr^{-1} and noted that the observed value was more than twice as large as the TAR's estimate. It thus appeared that either the processes causing sea level rise had been underestimated or the rate of sea level rise observed with tide gauges was biased toward higer values. Munk (2002) referred to this as the "Enigma".

Since the publication of the TAR, a number of new results have been published. Sea level rise can be measured directly since the early 1990s by Topex/Poseidon and Jason satellite altimetry. The global coverage of satellite altimetry has also revealed the complex geographical patterns of sea level change in open oceans. Near-global ocean temperature data have been recently made available for the last 50 years, allowing the first observations-based estimate of the thermal expansion contribution to past decades sea level rise. For the recent years, direct estimates of the land ice melt contribution to sea level are available from mass balance observations of glaciers and ice sheets.

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 and other processes.

5.5.2 Observations of Sea Level Changes

5.5.2.1 20th century sea level rise from tide gauges

Table 11.9 of the TAR listed several estimates for global and regional 20th century sea level trends based on the Permanent Service for Mean Sea Level (PSMSL) data set (Woodworth and Player, 2003). The concerns about geographical bias in the PSMSL data set remain, most long sea level records stemming from the Northern Hemisphere, and most from continental coastlines rather than ocean interiors. Based on a small number (~25) of high-quality tide gauges records from stable land regions, Douglas (2001) and Peltier (2001) estimate the rate of sea level rise as ~ 1.8 mm yr^{-1} for the past 70 years, and Miller and Douglas (2004) find a range of 1.5 – 2.0 mm yr^{-1} from 9 stable tide gauge sites. Holgate and Woodworth (2004) estimate a rate of 1.7 ± 0.4 mm yr^{-1} for global-coastal sea level change during the period 1948–2002 based on data from 177 stations divided into 13 regions. Based on combined analysis of tide gauge and altimetry data, Church et al. (2004) (discussed further below) determine a global rise of 1.8 ± 0.3 mm yr^{-1} during 1950–2000. Changes in global sea level as derived from analyses of tide gauges are displayed in Figure 5.5.1. Considering the results of Holgate and Woodworth (2004), Church et al. (2004) and Church and White (2006), and allowing for the ongoing higher trend in recent years shown by altimetry (see Section 5.5.2.2), we estimate the rate for 1961–2003 as 1.8 ± 0.5 mm yr^{-1}.

[INSERT FIGURE 5.5.1 HERE]

While the recently published estimates of last decades sea level rise remain within the range of the TAR values (i.e., 1–2 mm yr^{-1}), there is an increasing consensus that the best estimate lies closer to 2 than 1 mm yr^{-1}. The lower bound reported in the TAR resulted from local/regional studies; local/regional rates may differ significantly from the global mean, as discussed below (see Section 5.5.2.6).

A critical issue concerns how the records are adjusted for vertical movements of the land upon which the tide gauges are located. Recent global estimates are corrected for GIA using models, but not for other land motions. Peltier (2001) demonstrated that adjusted rates could be underestimated by several tenths of mm yr^{-1} in analyses which employ extrapolations of geological data obtained near the gauges, if GIA is the only geological process involved. This argument applies particularly to some reported European rates, e.g., Shennan and Woodworth (1992). The TAR mentioned the developing geodetic technologies (especially the Global Positioning System, GPS) which hold the promise of measuring rates of vertical land movement at tide gauges, no matter if those movements are due to GIA or to other geological processes. However, systematic problems with such techniques, including too short data spans, have not been resolved.

5.5.2.2 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, precision orbit determination, and 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- and/or latitude-dependent weighting (Nerem and Mitchum, 2001). Each 10 day estimate of global mean sea level has an accuracy of approximately 5 mm. Numerous papers have been published on the altimetry results (see Cazenave and Nerem, 2004, for a review), which currently show a rate of sea level rise of $+3.1 \pm 0.8$ mm yr^{-1} over 1993–2005 (Leuliette et al., 2004, Figure 5.5.2). 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 the 3 mm yr^{-1} rate of change has also been shown to arise from changes in the Southern Ocean (Cabanes et al., 2001).

[INSERT FIGURE 5.5.2 HERE]

The accuracy needed to compute mean sea level change pushes the altimeter measurement system to its performance limits, and thus care must be taken to ensure that the instrument is precisely calibrated (see Appendix 5.A.4.1). The tide gauge calibration method 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 ± 0.8 mm yr^{-1} over the last decade is reliable within these error bars.

Note that altimetry-based sea level measurements record variations of the global ocean basins volume which include changes due to GIA. Averaged over the oceanic regions sampled by the altimeter satellites, this effect yields a value close to -0.3 mm yr^{-1} (Peltier, 2001), with plausible uncertainty of 0.15 mm yr^{-1} (Tamisiea et al., 2006). This number is removed from altimetry-derived global mean sea level, in order to allow comparison with climate and anthropogenic contributions.

An important result of T/P altimetry is the mapping of the geographical distribution of sea level change (Figure 5.5.3a). Although regional variability in coastal sea level change had been reported from tide gauge analyses (e.g., Douglas, 1992; Lambeck, 2002), the global coverage of satellite altimetry provides unambiguous evidence of non uniform sea level change in open oceans, some regions exhibiting rates of sea level change about 5 times the global mean. For the past decade, it is in the western Pacific and eastern Indian oceans that sea level rise shows the highest magnitude. It is also worth noting that except for the Gulf Stream region, most of the Atlantic Ocean shows sea level rise during the past decade. Despite the global-mean rise, Figure 5.5.3a shows that sea level has been dropping in some regions (eastern Pacific and western Indian Oceans). Empirical Orthogonal Functions (EOF) analyses of altimetry-based sea level maps over

1993–2003 show a strong signature of the 1997–1998 El Niño, with the geographical patterns of the dominant mode being very similar to those of the sea level trend map (e.g., Nerem et al., 1999).

[INSERT FIGURE 5.5.3A HERE]

[INSERT FIGURE 5.5.3B HERE]

[INSERT FIGURE 5.5.3C HERE]

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

Attempts have been made to reconstruct historical sea-level fields by combining the near global coverage from satellite altimeter data with the longer but spatially sparse tide-gauge records (Chambers et al., 2002; Church et al., 2004). These sea-level reconstructions are based on a similar approach to sea surface temperature reconstructions, using the short altimeter record to determine the principal EOF of sea-level variability and the tide-gauge data to estimate the evolution of the amplitude of the EOFs over time. The method assumes that the geographical patterns of decadal sea level trends can be represented by a superposition of the patterns of variability which are manifest in interannual variability. As a caveat, note however that variability on different timescales may have different characteristic patterns (see Section 5.5.4.1). The sea level computed by Church and White (2006) which is shown in Figure 5.5.1 for the 1870–2000 period is based on this approach.

The trends in the EOF amplitudes (and the implied global correlations) are responsible for a spatially variable rate of sea-level rise. Figure 5.5.4a (updated from Church et al., 2004) shows the geographical distribution of linear sea level trends for 1955–2003 based on this reconstruction technique. Comparison with the altimetry-based trend map for the shorter period (1993–2003) indicates quite different geographical patterns. These differences mainly arise from thermal expansion changes through time (see Section 5.5.3)

[INSERT FIGURE 5.5.4A HERE]

[INSERT FIGURE 5.5.4B HERE]

The results help reconcile 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) in having 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 noted by Woodworth et al. (1999), Lambeck et al. (1998) and Mitrovica et al. (2001).

5.5.2.4 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 coherent throughout extended parts of the ocean. 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 mentioned above, the signature of the 1997–1998 El Niño is also clear in the in altimetric maps of sea level anomalies (see Section 5.5.2.2). Model results suggest that large volcanic eruptions produce interannual to decadal fluctuations of the global mean sea level (Church et al., 2005).

Interannual variability is a major reason why no long term acceleration of sea level has been identified using 20th century data alone (Woodworth, 1990; Douglas, 1992). Another possibility is that the sparse tide-gauge network may have been inadequate to detect it if present (Gregory et al., 2001).

As far as coastal 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. White et al. (2005) confirmed the larger sea level rise during the 1990s around continental coastlines compared to the open ocean but found that in some previous periods the coastal rate was smaller, concluding that over the last 50 years the coastal and open ocean rates of change were the same on average. The global reconstruction of Church et al. (2004) also exhibits large decadal variability in the rate of global mean sea level rise, with the 1993–2003 rate having been exceeded in some previous decades.

5.5.2.5 Long term sea level change

The longest records available from Europe and North American 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 support an inference that the onset of acceleration occurred during the 19th century. Recently, Church and White (2006) applied their reconstruction method (see Section 5.5.2.3) to provide a sea level curve back to 1870 (Figure 5.5.1). They find a significant acceleration, of 1.3 ± 0.6 mm yr^{-1} per century over the period 1870–2000.

Geological observations indicate that during the last 2000 years (i.e., before the recent rise recorded by tide gauges), sea level rose at an average rate of 0.0–0.2 mm yr^{-1} only (see Chapter 6, Section 6.4.3). The use of proxy sea level data from archaeological sources is well established in the Mediterranean. Oscillations in sea level from 2000 to 100 yr BP did not exceed ±25 cm, based on the Roman-Byzantine-Crusader well-data (Sivan et al., 2004). 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 between 1850 and 1950. Donnelly et al. (2004) and Gehrels et al. (2004), employing geological data from Connecticut, Maine and Nova Scotia salt-marshes together with nearby tide gauge records, demonstrated that the sea level rise observed during the 20th century was significantly in excess of that averaged over the previous several centuries.

5.5.2.6 Regional sea level change

In this section, we present three examples of regional sea level change, focusing on northeast Atlantic (where several studies show the link between sea level and NAO), Arctic coastal regions (for which results are available for the first time and where future sea level change is expected to be the largest, see Chapter 10, Section 10.6) and on small Pacific Islands which are the subject of much concern in view of their potential vulnerability to sea level rise.

5.5.2.6.1 Northeast Atlantic

North-East Atlantic sea level records are notable for their lower 20th century secular trends than the global average (Woodworth et al., 1999; Douglas, 2001), for which possible explanations have been proposed (e.g. the GIA effect, Mitrovica et al., 2001). The interannual variability in regional sea level demonstrates a clear relationship to the air pressure and wind changes associated with the NAO, with the magnitude and sign of the response depending primarily upon latitude. The relationship has been shown in North Atlantic deep ocean sea levels obtained from satellite altimetry, in northern European coastal (including Baltic) mean sea level obtained from tide gauges, and with the use of barotropic (storm surge) numerical models (Andersson, 2002; Wakelin et al., 2003; Woolf et al., 2003). The signal of the NAO can also be observed to some extent in ocean temperature records, suggesting a possible, smaller NAO influence on regional mean sea level via steric (density) changes (Tsimplis et al., 2006).

5.5.2.6.2 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 sea level rise (GIA corrected) over the seas of the Russian Arctic has been 1.85 mm yr^{-1}. It is interesting to notice that this rate is roughly of the same order of magnitude as the global mean rate inferred from tide gauges over the past few decades. This may be fortuitous however since in this particular region, wind stress and atmospheric pressure loading contribute to nearly half of the observed Arctic sea level rise (Proshutinsky et al., 2004).

5.5.2.6.3 Pacific Ocean Islands

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

[INSERT FIGURE 5.5.5 HERE]

An example of the large interannual variability in sea level is Kwajalein (8º44'N, 167º44'E) (Marshall Archipelago). Here, the local tide gauge data, the sea level reconstruction based Church et al. (2004) and Church and White (2006), and the short satellite altimeter record (Figure 5.5.5) agree and indicate that 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, persistent and intense ENSO events since the mid 1970's (Folland et al., 2001). For the Kwajalein record, the rate of sea level rise, after correction for GIA land motions and isostatic response to atmospheric pressure changes, is 1.9 ± 0.8 mm yr^{-1}. However, the uncertainty in rates of sea-level change increase rapidly with decreasing record length and can be several mm yr^{-1} for decade-long records (depending on the magnitude of the interannual variability). Sea-level change in the Tuvalu Islands (Western Pacific) has been the subject of intense interest as a result of their low lying nature and increasing incidence of flooding. There are two records available at Funafuti, Tuvalu; the first record commences in 1977 and the second (with rigorous datum control) in 1993. After allowing for subsidence affecting the first record, Hunter (2004) estimates sea-level rise at Tuvalu to be 1.2 ± 0.8 mm yr^{-1}, in agreement with the reconstructed rate of sea level rise.

5.5.2.7 Changes in extreme sea level

Impacts of sea level change 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, multidecadal records are few. Secondly, the normally used hourly sampling interval is not always sufficient to capture the true extreme. Finally, different authors use different types of high water extreme such as annual maximum high water, annual maximum surge, annual maximum surge-at-high-water, and surge at annual maximum high water. Annual maximum surge is a good 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. From sea level extremes at Liverpool since 1768, Woodworth and Blackman (2002) concluded that the annual maximum surge at high water was larger in the late-18th, late-19th and late-20th centuries than for most of the 20th century, qualitatively consistent with the long term variability in storminess from meteorological data. From the tide gauge record at Brest from 1860 to 1994, Bouligand and Pirazzoli (1999) found an increasing trend for annual maxima and 99th percentile of surges; however, during the period 1953–1994 a decreasing trend was noticed which has been attributed to a decrease in the frequency and duration of storms during this period. From non-tidal residuals ('surges') at San Francisco since 1858, Bromirski et al. (2003) concluded that extreme winter residuals have exhibited a significant increasing trend since about 1950. This increasing trend is attributed to an increase in storminess during this period. Zhang et al. (2000) concluded from ten stations along the east coast of USA since 1900 that the rise in extreme sea level 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, (interestingly, the highest ever value being due an anti-cyclonic oceanic eddy system in 2003), but no evidence for an increase relative to the underlying upward mean sea level trend.

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

The above studies indicate that changes in the occurrence of extreme sea level are caused by mean sea level changes as well as changes in surges, which are caused by regional storminess. More studies including regional storm surge modeling are required to understand the trends in surges and means separately and their combined effect in producing the observed extreme sea level.

5.5.3 Ocean Density Changes

Sea level will rise if the ocean warms and fall if it cools, since the density of the water column will change. If the thermal expansivity were constant, global sea level change would parallel the global ocean heat content discussed in Section 5.2. However, since warm water expands more than cold water (with the same input of heat), and water at higher pressure expands more than at lower pressure, the global sea level change depends on the 3D distribution of ocean temperature change.

Analysis of the last half-century of temperature observations indicates that the ocean has warmed in all basins (see Section 5.2). The average rate of thermosteric sea-level rise caused by heating of the global ocean is estimated to be 0.40 ± 0.1 mm yr^{-1} over 1955–1995 (Antonov et al., 2005), based on 5-year mean temperature data down to 3000 m. This is ~25% of the observed rate of total sea level rise over that period (~1.8 mm yr^{-1}).

For the 0–700 m layer and the 1955–2003 period (the time span used for the total sea level budget; Section 5.5.6), the averaged thermosteric trend, based on yearly mean temperature data from Levitus et al. (2005a) is 0.33 ± 0.08 mm yr^{-1} (Antonov et al., 2005). For the same period and depth range, the mean thermosteric rate based on monthly ocean temperature data from Ishii et al. (2006) is 0.36 ± 0.14 mm yr^{-1}. Figure 5.5.6 shows the thermosteric sea level curve over 1955–2003 for both the Levitus and Ishii data sets. The rate of thermosteric sea-level rise is clearly not constant in time and shows considerable fluctuations. A rise of more than 20 mm occurred from the late 1960s to the late 1970s with a smaller drop afterwards. Another large rise began in the 1990s.

The Levitus and Ishii datasets both give 0.32 ± 0.10 mm yr^{-1} for the upper 700 m during 1961–2003, but the Levitus dataset of temperature down to 3000 m ends in 1998. From the results of Antonov et al. (2005) for thermal expansion, the difference between the trends in the upper 3000 m and the upper 700 m for 1961–1998 is ~0.1 mm yr^{-1}. Assuming that the ocean below 700 m continues to contribute beyond 1998 at a similar rate, with an uncertainty similar to that on the upper-ocean contribution, we assess the thermal expansion of the ocean down to 3000 m during 1961–2003 as 0.42 ± 0.14 mm yr^{-1}.

[INSERT FIGURE 5.5.6 HERE]

For the recent period 1993–2003, a value of 1.2 ± 0.6 mm yr^{-1} for thermal expansion in the upper 700 m is estimated both by Antonov et al. (2005) and Ishii et al. (2006). Willis et al. (2004) estimate thermal expansion to be 1.6 ± 0.6 mm yr^{-1}, based on combined sparse in situ temperature profiles down to 750 m and satellite measurements of altimetric height. Including the satellite data reduces the error caused by the inadequate sampling of the profile data. Error bars were estimated to be about 2 mm for individual years in the time series, with most of the remaining error due to inadequate profile availability. A close result (1.8 ± 0.4 mm yr^{-1} steric sea level rise for 1993–2003) has been recently obtained by Lombard et al. (2006), using global gridded temperature and salinity fields from Guinehut et al. (2004). The latter fields are also based on a combined analysis of in situ hydrographic data and satellite sea surface height and sea surface temperature data. It presently unclear why the latter two estimates are significantly larger than the thermosteric rates by Levitus et al. (2005a) and Ishii et al. (2006) over the same period. It is possible that the in situ only data underestimate thermal expansion because of poor coverage in southern oceans, and it is interesting to note that the value from a model based on assimilation of hydrographic data yields a somewhat higher estimate of 2.3 mm yr^{-1} (Carton et al., 2005). However it is as well possible that combined satellite and in situ analyses overweigh the satellite information, especially in the Austral Ocean where sea level trends from satellite altimetry appear quite high (see Figure 5.5.3a). We assess the thermal expansion of the upper 700 m during

1993–2003 as 1.5 ± 0.6 mm yr^{-1}, and of the upper 3000 m as 1.6 ± 0.6 mm yr^{-1}, allowing for the ocean below 700 m as for the earlier period.

[INSERT FIGURE 5.5.7 HERE]

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. A similar result is obtained by Ishii et al. (2006) who estimate the halosteric contribution to 1955–2003 sea level rise as 0.04 ± 0.02 mm yr^{-1}. 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 (by mixing) to the mean ocean value; the compensation is exact for a linear state equation (Lowe and Gregory, 2006). 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). For regional changes of sea level, thermosteric and halosteric contributions can however be comparably important (cf. Section 5.5.4.1). Estimates of the steric sea level rates available for 1955–2003 and 1993–2003 are summarized in Table 5.5.1.

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 ± 0.10	1955–1998	0–3000	Levitus et al. (2005b)
Antonov et al. (2005)	0.33 ± 0.08	1955–2003	0–700	Levitus et al. (2005b)
Ishii et al. (2006)	0.36 ± 0.07	1955–2003	0–700	Ishii et al. (2006)
Antonov et al. (2005)	1.2 ± 0.6	1993–2003	0–700	Levitus et al. (2005b)
Ishii et al. (2006)	1.2 ± 0.6	1993–2003	0–700	Ishii et al. (2006)
Willis et al. (2004)	1.6 ± 0.6	1993–2003	0–750	Willis et al. (2004)
Lombard et al. (2006)	1.8 ± 0.4	1993-2003	0-700	Guinehut et al.(2004)

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

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

5.5.4.1 Steric sea-level changes

Like the sea level trends observed by satellite altimetry (see Section 5.5.2.3), the global distribution of thermosteric sea-level trends is not spatially uniform. This is illustrated by Figure 5.5.3b and Figure 5.5.4b which show the geographical distribution of thermosteric sea level trends over two different periods, 1993–2003 and 1955–2003 respectively (updated from Lombard et al., 2005). Some regions experience sea-level rise while others experience a fall, often with rates that are several times the global mean. However, the patterns of thermosteric sea level rise over the ~50-year period are different than those seen in the 1990s. This occurs because the spatial patterns, like the global average, are also subject to decadal variability. In other words, variability on different timescales may have different characteristic patterns.

An EOF analysis of gridded thermosteric sea level time series since 1955 (updated from Lombard et al., 2005) displays a spatial pattern that is very similar to the spatial distribution of thermosteric sea level trends over the same time span (Figure 5.5.8). In addition, the principal component is negatively 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.

[INSERT FIGURE 5.5.8 HERE]

For the recent years (1993–2003), the spatial patterns of thermosteric sea level change show remarkable correlation with the geographic distribution of observed sea level trends (compare Figures 5.5.3a and 5.5.3b). This suggests that at least part of the non-uniform pattern of sea level rise observed in the altimeter data over the past decade can be attributed to changes in the ocean's thermal structure, which is itself driven by the ocean circulation. Note that the steric changes due to salinity changes have not been included in these figures due to insufficient salinity data in parts of the world ocean.

Ocean salinity changes, while unimportant for sea level at the global scale, can have a significant effect on regional sea level (e.g., Antonov et al., 2002; Ishii et al., 2006). E.g., in the sub-polar gyre of the North Atlantic, especially in the Labrador Sea, the halosteric contribution nearly counteracts the thermosteric contribution. This observational result is supported by results from data assimilation into models (e.g., Stammer et al., 2003). Since density changes can result not only from surface buoyancy fluxes but also from the wind, a simple attribution of density changes to buoyancy forcing is not possible.

While much of the non-uniform pattern of sea level change can be attributed to steric volume changes, the difference map (Figure 5.5.3c) show high 'residual signal' in a number of regions, especially in the southern oceans. Part of these residuals is likely due to the lack of ocean temperature coverage in remote oceans as well as in deep layers (below 700 m).

5.5.4.2 Ocean circulation changes

The highly non-uniform geographical distribution of steric sea level trends is closely connected, through geostrophic balance, with changes in ocean circulation. Density and circulation changes result from changes in atmospheric forcing which is primarily by surface wind stress and buoyancy flux (i.e., heat and freshwater flux). In a recent investigation based on ocean data assimilation in an OGCM, Carton et al. (2005) satisfactorily reproduce the spatial structure of sea level trends for the past decade. In particular, they show that the tropical Pacific pattern results from decadal fluctuations in the depth of the tropical thermocline and change in equatorial trade winds. The similarity of the patterns of steric and actual sea-level change indicates that density changes near the surface are the dominant influence. Discrepancies may indicate a significant contribution from changes in the wind-driven barotropic circulation, especially at high latitudes.

5.5.4.3 Surface Atmospheric Pressure Changes

Surface atmospheric pressure also causes regional sea level variations. On time scales above a few days, the ocean adjusts nearly isostatically to changes in atmospheric pressure (inverted barometer effect), 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). Such effects are generally corrected for in tide gauge and altimetry-based sea level analyses. The inverted barometer has a negligible effect on the global mean sea level; however, since the coverage of the altimetric data is not quite global, it contributes –0.06 mm/yr according to Ponte (2006).

5.5.4.4 Solid earth and geoid changes.

Geodynamical processes related to the deformable Earth's response to spatio-variable ice melt loading (due to last deglaciation and present-day land ice melt) cause also non uniform sea level change (e.g., Mitrovica et al., 2001; Peltier, 2001; Plag, 2006). The solid earth and oceans continue to respond to the ice and complementary water loads associated with the Late Pleistocene - Early Holocene glacial cycles through Glacial Isostatic Adjustment (GIA). This process not only drives large crustal uplift near the location of former ice complexes, but also produces a world-wide signature in sea level that results from gravitational, deformational and rotational effects: as the viscous mantle material flows to restore isostasy during and after the last deglaciation, uplift occurs under the former 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 the geoid (an equipotential surface of the Earth's gravity field that coincides with the mean surface of the oceans).

Because GIA induces vertical movements of the crust, it affects tide gauge-based sea level measurements. This correction (estimated from models) ranges from ~ 1 mm yr^{-1} (or more) in the near field to a few tenths mm yr^{-1} in the far field (Peltier, 2001; Tamisiea et al., 2006). GIA-induced change of the geoid and shape of the ocean basins causes regional changes in the sea level up to a few tenth mm yr^{-1} (Peltier, 2001; Plag, 2006; Tamisiea et al., 2006). Self-gravitation and deformation of the Earth's surface in response to the load of land ice melt from glaciers and ice sheets is another cause of regional sea level variations. Model predictions show quite different patterns of non uniform sea level change depending on the source of the ice melt (Mitrovica et al., 2001; Plag, 2006), and associated regional sea level variations reach up to a few 0.1 mm yr^{-1}.

5.5.5 Ocean Mass Change

As remarked in 5.5.3, only ~25% of sea level rise in recent decades was due to thermal expansion. Miller and Douglas (2004) reached a similar conclusion by averaging raw ocean temperature and salinity data over the past 50 years in three oceanic regions (northeast Pacific, northeast Atlantic and western Atlantic). They 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 9 tide gauges sites located in these regions. They concluded that sea level rise in the second half of 20th century was mostly due to water mass added to the oceans. It is worth noting that during recent years (1993–2003), a roughly similar value (between 1.3 and 1.9 mm yr^{-1}, out of 3.1 mm yr^{-1}) is estimated also due to 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; Ishii et al., 2006). 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 (see Section 5.5.5.2) and terrestrial water storage (which do). Antonov et al. (2002), Munk (2003) and Wadhams and Munk (2004) used the global salinity changes 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 (e.g., Hilmer and Lemke, 2000, see Chapter 4, Section 4.4). Assuming 430 ± 130 km^3 yr^{-1} sea ice volume decrease, Wadhams and Munk (2004) estimate 0.6 ± 0.18 mm yr^{-1} as the rate of sea level rise inferred from fresh water input. However, large uncertainties remain in their estimate, from both the estimate of ocean freshening as well as the estimates of sea ice melt. Their 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 (see Section 5.5.6).

5.5.5.2 Land ice

During the 20th century, glaciers and ice caps have experienced considerable mass losses, with strong retreats in response to global warming after 1970. For 1961–2003 their contribution to sea level is assessed as 0.51 ± 0.32 mm yr^{-1} and for 1993–2003 as 0.81 ± 0.43 mm yr^{-1} (see Chapter 4, Section 4.5.2).

The Greenland ice sheet has also been losing mass in recent years, contributing 0.05 ± 0.12 mm yr^{-1} during 1961–2003 and 0.21 ± 0.07 mm yr^{-1} in 1993–2003 (see Chapter 4, Section 4.6.2.2). Assessments for the Antarctic ice sheet are less certain, especially before the advent of satellite measurements, , and are 0.14 ± 0.41 mm yr^{-1} for 1961–2003 and 0.21 ± 0.35 mm $yr^{-1}$for 1993–2003 (see Chapter 4, Section 4.6.2.2). Geodetic data on Earth rotation and polar wander allow a late-20th century sea-level contribution of up to ~1 mm yr^{-1} from land ice (Mitrovica et al., 2006, Chapter 4, Section 4.6.2.2).

5.5.5.3 Land hydrology

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 coupled to global ocean-atmosphere circulation models 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° by 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 surface 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 (decadal) 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 tion over land and consequent transient increase in storage (see Chapter 10).

5.5.5.3.2 Anthropogenic change in land water storage

The amount of anthropogenic change in land water storage systems cannot be estimated with much confidence, as already discussed by Church et al. (2001). A number of factors can contribute to sea level rise:

1) 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.

2) 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

3) 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.

4) 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.

Note that the last two processes have a small effect on time scales longer than a few years.

On the other hand, impoundment of water behind dams removes water from the ocean and lowers sea level. Chao (1994) and Sahagian et al. (1994) estimate that dams have led to sea level drop over the past few decades, by –0.5 to –0.7 mm yr^{-1}. Infiltration from dams and irrigation may raise the water table, storing more water.

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

In summary, our assessment of the land hydrology contribution to sea level change has not led to a reduction of the uncertainty compared to the TAR, which estimated the rather wide ranges of –1.1 to +0.4 mm yr^{-1} for 1910–1990 and –1.9 to +1.0 mm yr^{-1} for 1990. However, indirect evidence from considering other contributions to the sea level budget (see Section 5.5.6) suggests that the land contribution is either small (<0.5 mm yr^{-1}) or is compensated with unaccounted contributions (e.g., deep thermal expansion).

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 5.5.9, for 1961–2003 and 1993–2003. Some terms known to be small have been omitted (changes in atmospheric water vapour, soil moisture, permafrost, sedimentation, cf. Church et al., 2001), probably totalling less than 0.2 mm yr^{-1}. The poorly known anthropogenic contribution from terrestrial water storage (see Section 5.5.5.3.2) is also omitted.

[INSERT FIGURE 5.5.9 HERE]

For 1961–2003 the sum of thermal expansion and land ice leaves 0.7 ± 0.8 mm yr^{-1} unexplained, barely consistent with zero. The assessment of Church et al. (2001) could allow it to be covered by positive anthropogenic terms (especially groundwater mining) but these are expected to have been outweighed by negative terms (especially impoundment). We conclude that the budget has not yet been closed satisfactorily.

For 1993–2003, thermal expansion is much larger and land ice contributes 1.2 ± 0.6 mm yr^{-1}. These increase may partly reflect decadal variability rather than an acceleration. (Attribution of changes in rates and comparison with model results are discussed in Chapter 9 , Section 9.5.2) The sum is still less than the observed trend but the discrepancy of 0.3 ± 1.1 mm yr^{-1} is consistent with zero. This more satisfactory assessment for recent years, during which individual terms are better known, indicates progress since the TAR.

Table 5.5.2. Estimates for the various contributions to the budget of global-mean sea level change for 1961–2003 and 1993–2003 compared with the observed rate of rise. Ice sheet mass loss of 100 Gt yr^{-1} is equivalent to 0.28 mm yr^{-1} of sea level rise. A GIA correction has been applied to observations from tide gauges and altimetry. For the sum of terms, uncertainties of the terms have been combined in quadrature.

	Sea level rise (mm yr^{-1})		
	1961–2003	1993–2003	
Thermal expansion	0.42 ± 0.14	1.6 ± 0.6	Section 5.5.3
Glaciers and ice caps	0.51 ± 0.32	0.81 ± 0.43	Chapter 4, Section 4.5
Greenland ice sheet	0.05 ± 0.12	0.21 ± 0.07	Chapter 4, Section 4.6.2
Antarctic ice sheet	0.14 ± 0.41	0.21 ± 0.35	Chapter 4, Section 4.6.2
Sum	1.1 ± 0.6	2.8 ± 0.8	
Observed	1.8 ± 0.5		Section 5.5.2.1
		3.1 ± 0.8	Section 5.5.2.2

Question 5.1: Is the Sea Level Rising?

While global sea level rose by ~120 m during the several millennia that followed the end of the last glacial maximum, it stabilized between 3000 and 2000 years ago (Lambeck et al., 2004). Since then, paleo sea level indicators suggest that global sea level did not change significantly: the average rate of change from 2000 and ~ 100 years ago is near zero (Peltier, 2002; Lambeck et al., 2004).

The instrumentally-based estimates of modern sea level change show evidence for onset of acceleration at the end of the 19th century (Church and White, 2006). Recent estimates for the last half of the 20th century (1950–2000) give ~2 mm yr^{-1} global mean sea level rise (Church et al., 2004; Holgate and Woodworth, 2004).

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 mm yr^{-1} (Cazenave and Nerem, 2004; Leuliette et al., 2004), 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 indicates an acceleration due to anthropogenic global warming, or a result of natural climate variability, or a combination of both effects.

Satellite data also confirm 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.

Near-global ocean temperature data sets made recently available for the past 50 years (Levitus et al., 2005b; Ishii et al., 2006) allow a direct calculation of thermal expansion. This contribution to sea level rise is ~0.4 mm yr^{-1}. For the recent years (1993-2003), thermal expansion accounts for ~1.5 mm yr^{-1} and is the largest part of sea-level rise.

On average over the past four decades, loss of mass by glaciers and ice caps has accounted for ~0.5 mm yr^{-1} sea level rise, and more in recent years. Observations beginning in the 1990s indicate that the Greenland and Antarctic ice sheets are also contributing substantially.

For 1993–2003 there is fair agreement between the observed rate of sea level rise and the sum of known contributions. This sets constraints on poorly estimated effects such as anthropogenic land water storage.

[INSERT QUESTION 5.1, FIGURE 1 HERE]

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 broadly consistent with known characteristics of the large scale ocean circulation.

There is compelling evidence that the heat content of the world ocean has increased since 1955 (Figure 5.6.1) and in summary of changes (Figure 5.6.2). 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 5.2.4), 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 5.4.2 and Figure 5.6.2). 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 (around 50 to 40°S) (Figure 5.4.2 and Figure 5.6.2). The Southern Hemisphere deep overturning circulation shows some evidence of change from new but sparse results around the Antarctic Continent. The transfer of heat into the ocean also leads to sea-level rise through thermal expansion, and the horizontal pattern of sea level change since 1955 (Figure 5.5.4a) is largely consistent with thermal expansion (Figure 5.5.4b) and with the change in heat content (Figure 5.2.2).

[INSERT FIGURE 5.6.1 HERE]

[INSERT FIGURE 5.6.2 HERE]

The similarity in mid-latitudes between the zonal patterns of changes in temperature, Anthropogenic Carbon, and sea-level rise with a passive tracer (CFC) (Figure 5.6.1) is strong evidence that these independent data sets (albeit with widely varying reference periods) show a common pattern of change in the ocean. These changes are the result of ocean ventilation and subduction.

The subduction of carbon into the ocean has meant that Calcite 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 5.2.6). 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. 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 (see 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.

Since the TAR, we have now developed 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 (see 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 (see Chapter 3, Section 3.6.2). The spatial gradients in sea-level mean suggest that surface ocean currents have changed, (e.g., both the Antarctic Circumpolar Current and the North Atlantic sub-tropical gyre have strengthened), but these changes in horizontal currents are small compared with the mean circulation and have not been verified by direct observations.

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 1993–2003 period, than in the average trends over the period from 1961–2003. 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. The presence of decadal variations in the data mean that we have very low confidence that such increases in the rate of change have been detected.

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 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 Temperature and Salinity

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

[INSERT FIGURE 5.A.1a HERE]

[INSERT FIGURE 5.A.1b HERE]

An example of the distribution of ocean temperature observations in both space and time is shown in Figure 5.A.1. This figure the insitu temperature data distribution for two 5-year periods used to create estimates of global heat content change (e.g., Figure 5.2.1), one with low (a) and one with high (b) density of observations. It is clear that parts of the ocean, in particular in the Southern Hemisphere, are not well sampled even in periods of high observation density. Hence sampling errors resulting from the lack of data are potentially important but cannot easily be quantified.

Several different objective analysis techniques have been used to produce the gridded fields of temperature anomalies used to compute ocean heat content and steric sea level presented in this chapter. The technique used by Levitus et al. (2005b), Garcia et al. (2005), and Antonov et al. (2005) in their estimates of temperature (heat content), oxygen, and the 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 described by Boyer et al. (2002). At each standard depth level all data are averaged within each 1° square, and a first-guess value from climatology is subtracted to produce an anomaly value. The analyzed value is computed from all observations within the surrounding region of 500–600 km,. Features with wavelength less than 500–600 km are substantially reduced in amplitude; in regions without sufficient data, essentially the climatological information is used. Ishii et al. (2006) employed similar techniques, with a smaller decorrelation length scale of 300 km and a least-squares technique for estimating corrections to the first-guess field. Willis et al. (2004) used a two-scale covariance function, but also use altrimetric data in areas where ocean observations are lacking.

There are some differences in the data used in these studies. In addition to ocean temperature profile data, Ishii et al. (2006) also used the product of climatological mixed layer depth and individual SST measurements in their estimates of ocean heat content. Southern hemisphere WOCE profiling float temperature profiles for the 1990s were used by Willis et al. (2004) that were not used by Levitus et al. (2005) and Ishii et al. (2006). The similarity of the three independently estimated heat content time series shown in Figure 5.2.1 to within confidence intervals indicates that the differences between analysis techniques and data sources do not substantially influence the estimates of the three global ocean heat content time series.

All analyses are subject to statistical errors and sampling errors. Statistical errors are estimated in a straightforward way. E.g., for the Levitus et al. (2005a) fields, the uncertainty at any gridpoint is estimated from the variability of observations that contributed to the analyzed value. In this way, 95% errors for all analysed variables are computed as function of depth and horizontal position, and correspondingly for integrated variables such as heat content. Both Ishii et al. (2006) and Willis et al. (2004) used the interannual variability of heat content as the basis for error analyses.

5.A.2. Heat Fluxes and Transports

Surface meteorological and subsurface hydrographic observations are inhomogeneously distributed in space and in time. E.g., the number of observations in the mid-latitude North Atlantic may be 10–100 times higher than in the poorly sampled 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 in poorly sampled areas or periods. The magnitude of sampling uncertainty of surface heat fluxes in the Southern Ocean can amount to 20–50 W m^{-2}, which is higher than the magnitude of interannual variability of fluxes. 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, an uncertainty of zonal averaged estimates of ±10 W m^{-2} results in an uncertainty of 0.5 PW 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, and estimates of MHT variability are hardly possible in this way.

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 reported uncertainty of the global estimate of ±19 PgC (Sabine et al., 2004a) is based on measurement errors and potential biases. Newer studies suggest that the global anthropogenic carbon could be overestimated by ~10% because of potential biases caused by assumptions on the time evolution of CO_2, the age or the identification of water masses (Matsumoto and Gruber, 2005) and by the recent changes in surface warming and stratification (Keeling, 2005). However, 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 a total of 530,000 oxygen profiles (Garcia et al., 2005). The measurement method was not reported for all the cruises. Only the Winkler titration was reported with only manual titrations prior to 1990. The Carpenter method to improve the accuracy was reported on some cruises after 1970. Other improvements in the titration method and automated titrations were reported after 1990 only. 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. There are no standards for oxygen measurements because of the difficulty in preparing a stable solution.

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 of the satellite above the instantaneous sea surface. The quantity of interest in oceanography is the height of the instaneous sea surface above a fixed reference surface which is computed as the difference between the altitude of the satellite above the reference ellipsoid and the altimeter range. The satellite position is computed through precise orbit determination, combining accurate modelling of the satellite motion and tracking measurements between the satellite and observing stations on Earth or other observing satellites. A number of corrections must be applied to obtain the correct the sea surface height. These include instrumental corrections, ionospheric correction, dry and wet tropospheric corrections, electromagnetic bias correction, ocean and solid earth tidal corrections, ocean loading correction, pole tide correction, and also an inverted barometer correction which has to be applied since the altimeter does not cover the global ocean completely.

The total measurement accuracy for the Topex/Poseidon altimetry-based sea surface height is about 8 cm (95% error) for a single measurement based on 1-sec along-track averages (Chelton et al., 2001).

The above error estimates concern instantaneous sea surface height measurements. For estimating the mean sea level variations, the procedure consists of simply averaging over the ocean the point-to-point measurements collected by the satellite during a complete orbital cycle (10-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 95%-error associated with a 10-day mean sea level estimate is approximately 8 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 et al. (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 0.8 mm yr^{-1} (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.4.2 Sea level from tide gauge observations

Tide gauges are based on a number of different technologies (float, pressure, acoustic, radar), each of which has its advantages in particular applications. The Global Sea Level Observing System (GLOSS) specifies that a gauge must be capable of measuring sea level to cm accuracy (or better) in all weather conditions (i.e., in all wave conditions). The most important consideration is the need to maintain the gauge datum relative to the level of the Tide Gauge Bench Mark (TGBM), which provides the land reference level for the sea level measurements. GLOSS specifications require that local levelling must be repeated at least annually between the reference mark of the gauge, TGBM, and a set of approximately five ancillary marks in the area, in order to maintain the geodetic integrity of the measurements. In practice, this objective is easier to meet if the area around the gauge is hard rock, rather than reclaimed land, for example. The question of whether the TGBM is moving vertically within a global reference frame (for whatever reason) is being addressed by advanced geodetic methods (GPS, DORIS, Absolute Gravity). With typical rates of sea and land level change of order of 1 mm yr^{-1}, it is necessary to maintain the accuracy of the overall gauge-system at the 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. The tide gauge observation system for three periods is shown in Figure 5.A.2, together with the evolution over time of the number of stations in both hemispheres. The distribution of tide gauge stations is particularly sparse in space at the beginning of the 20th century, but rapidly improves in the 1950's through to the current network of GLOSS standard instruments. This distribution of instruments through time means that our confidence in the estimates of sea-level rise has been improving and this certainty is reflected in the shrinking confidence intervals (Figure 5.5.1).

[INSERT FIGURE 5.A.2 HERE]