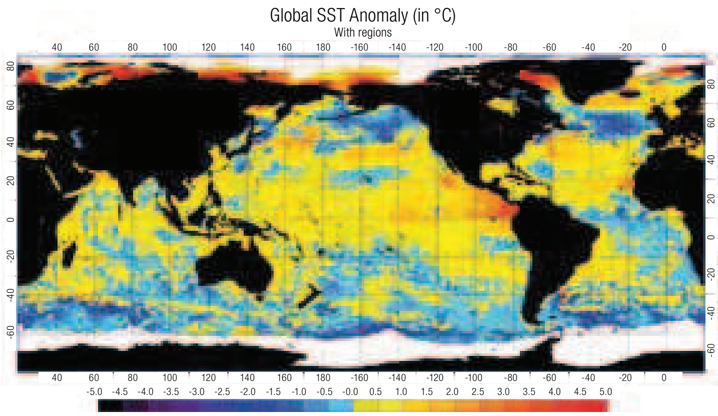
CLIMATE CHANGE 2009
SCIENCE COMPENDIUM

EARTH'S OCEANS

Earth's Oceans

Over the last five decades, the world's oceans have been subjected to fishery overharvesting, seafloor damage from bottom trawling, and habitat loss around margins from coastal development schemes. Climate change further threatens oceans with higher temperatures, increased acidification, and altered circulation and nutrient supplies.

Figure 3.1: Global sea surface temperature anomaly on 10 September, 2009



This sea surface temperature (SST) map is generated by subtracting the long-term mean SST, for that location and in that time of year, from the current value. A positive anomaly means that the current sea surface temperature is warmer than average and a negative anomaly means it is cooler than average. *Source: NOAA 2009b*

After 20 years of targeted research into how climate change is affecting Earth Systems, enormous challenges remain in understanding balances, feedbacks, and relations among sub-systems in the world's oceans. Research about sea-level rise, circulation shifts, and chemical responses to anthropogenic inputs often seems to raise more questions than answers.

INCREASED TEMPERATURES

The seasonal variations in heating penetrate into the ocean through a combination of radiation, convective overturning, and mechanical stirring by winds. These processes move heat through the mixed layer, which, on average, involves about 90 metres of the ocean's upper layer. The thermal inertia of a 90 metre layer can add a delay of about six years to the temperature response to an immediate change. With its huge volume and mean depth of about 3,800 metres, the total ocean would take 230 years to fully respond to a temperature change if it were rapidly mixed. However, mixing is not a rapid process for most of the ocean so in reality the response depends on the rate of ventilation of water between the well-mixed upper layers of the ocean and the deeper, more isolated layers that are separated by the thermocline—the ocean layer exhibiting a strong vertical temperature gradient. The rate of such mixing is not well established and varies greatly geographically. An overall estimate of the delay in surface temperature response caused by the oceans is from 10 to 100 years. The slowest response should be in high latitudes where deep mixing and convection occur, and the fastest response is expected in the tropics. Consequently, the oceans are a great moderating effect on climate changes (Trenberth 2001).

Changes in the climate system's energy budget are predominantly revealed in ocean temperatures and the associated thermal expansion contribution to sea-level rise. Climate models, however, do not reproduce the large decadal variability in globally averaged ocean heat content inferred from the sparse observational database, even when volcanic and other variable climate forcings are included. The sum of the observed contributions has also not adequately explained the overall multi-decadal rise. But now improved estimates of near-global ocean heat content and thermal expansion for the upper 300 metres and 700 metres of the ocean for 1950–2003 have been reported, using statistical techniques that allow for sparse data coverage and that apply corrections to reduce systematic biases in the most common ocean temperature observations. These adjusted ocean warming and thermal expansion trends for 1961–2003 are about 50 per cent larger than earlier estimates but about 40 per cent smaller for 1993–2003, which is consistent with the recognition that previously estimated rates for the 1990s had a positive bias as a result of instrumental errors. On average, the decadal variability of the climate models with volcanic forcing now agrees approximately with the observations, but the modelled multi-decadal trends are smaller than observed (Domingues *et al.* 2008).

SEA-LEVEL RISE

Global average sea level is rising predominantly as a consequence of three factors—thermal expansion of warming ocean water, addition of new water from the ice sheets of Greenland and Antarctica and from glaciers and ice caps, and the addition of water from land surface runoff. All three potential sources are undergoing changes of anthropogenic origin. Regionally, sea level is affected by isostatic responses to the unloading of burden from bedrock, by coastal subsidence in response to removal of materials or to new loads, and by gravitational and ocean current effects causing the ocean surface to deviate from a consistent elevation (Pfeffer *et al.* 2008, Milne *et al.* 2009, Lettenmaier and Milly 2009, Bamber *et al.* 2009).

Since at least the 19th century, sea-level changes have been measured directly by tide gauge records and, since the 1990s, by satellite altimetry. Sea-level changes over longer periods of time, thousands to millions of years, are inferred from geologic evidence (Rohling *et al.* 2009). The average rate of global mean sea-level rise over the 20th century was about 1.7 millimetres (mm) per year. In the period 1993-2003 global mean sea level rose about 3.1 mm per year, and since 2003 the rate of rise has been about 2.5 mm per year. The relative importance of the three factors contributing to global average sea-level rise has varied during this time (Jevrejeva *et al.* 2008, Church 2008, Lettenmaier and Milly 2009, WCRP 2009).

Contributions to sea-level rise are measured by a variety of methods. Synthesis analyses, referred to as sea-level budgets, are conducted periodically

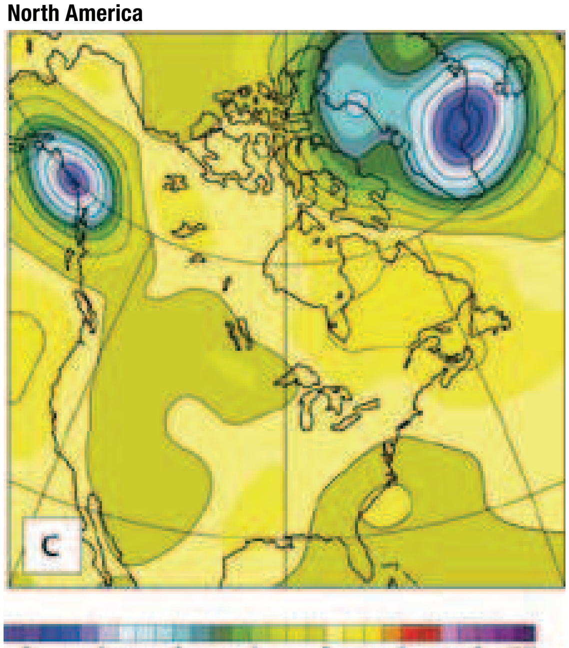
In Hawaii, as in other islands or low-lying regions, sea-level rise combined with high rainfall events pose a flooding risk due to storm sewers backing up with saltwater. Another associated hazard is accelerated beach erosion. *Source: L. Carey* to compare direct observations of sea-level rise with models and evaluations of the component contributions. Thermal expansion is determined by ocean temperature measurements from large numbers of automated buoys, while glacier and ice sheet contributions are determined by geodetic measurements of land ice volume change as well as by mass budget fluxes observed in individual ice bodies. Since 2003, changes in land ice and land hydrology (surface and underground water flows) have been detected gravitationally by the Gravity Recovery and Climate Experiment (GRACE) satellite system. GRACE observations are also used in combination with satellite altimetry to measure thermal expansion (NASA JPL 2009). Changes in all three components are also estimated by modelling and limited observational data are upscaled to global values by a variety of statistical methods.

Prior to about 1990, ocean thermal expansion accounted for slightly more than 50 per cent of global sea-level rise. Since then, the contribution from thermal expansion has declined to about 15 per cent but this decrease has been countered by increases in glacier, ice cap, and ice sheet contributions. By 2006, glaciers and ice caps contributed about 32 per cent of the total sea-level rise, while the ice sheets on Greenland and Antarctic together contributed about 20 per cent (Milne *et al.* 2009, Hock *et al.* 2009).

While glaciers and ice caps exclusive of the ice sheets dominate present-day contributions to sea-level rise, they collectively constitute a far smaller total sea-level rise owing to their much smaller global volume. If current trends in ice loss continue, the glacier and ice cap reservoir will be exhausted by 2200. On the time scale of decades to the next century, however, glaciers and ice caps will remain a source of sea-level rise equal to or greater than the ice sheets (Meier *et al.* 2007, Bahr *et al.* 2009).

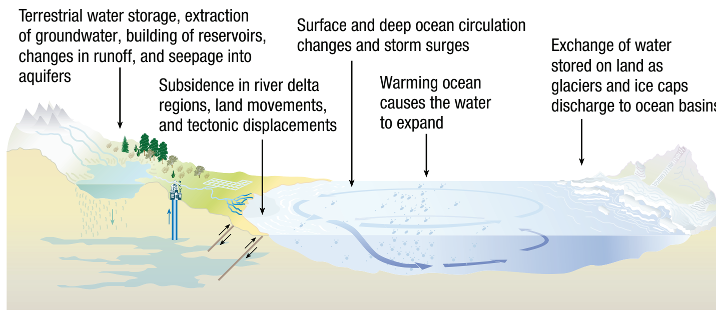
The impacts of sea-level rise will be felt through both an increase in mean sea level and through an increase in the frequency of extreme sea-level events such as storm surges. These impacts include increased frequency and severity of flooding in low-lying areas, erosion of beaches, and damage to infrastructure and the environment, including wetlands and inter-tidal zones, and mangroves, with significant impacts on biodiversity and ecosystem

Figure 3.2: Equivalent water thickness variations over North America



Gravity Recovery and Climate Experiment (GRACE) surface mass-rate field corrected for glacio-isostatic rebound and showing current ice loads of Alaska and Greenland. The twin satellites detect the gravitational fields characterizing different masses on Earth's near surface. *Source: Peltier 2009*

Figure 3.3: What causes sea level to change?



Source: Griggs 2001

Table 3.1: Components of Global Average Sea-Level Rise (millimetres per year)

Time period	IPCC 2007b 1993-2003	Meier *et al.* 2007, 2006	Cazenave and Nerem 2004, 1993-2003	Cazenave *et al.* 2009, 2003-2008
Thermal expansion	1.6±0.5	—	1.6±0.3	0.34±0.12[3]
Greenland	0.21±0.07	0.50±0.10	0.20±0.04	0.38±0.05
Antarctica	0.21±0.35	0.17±0.11	0.55±0.06	0.56±0.06
Other Glaciers and Ice Caps	0.5±0.18	1.1±0.24	0.8±0.1	1.1±0.25[4]
Land hydrology[5]	—	—	—	0.17±0.1
Sum of components	**2.8±0.72[2]**	**1.8±0.50[1,2]**	**3.0±0.5[2]**	**2.2±0.28**

Notes:
1: Sum does not include thermal expansion
2: Sum does not include land hydrology
3: Average of two estimates
4: Taken from Meier *et al.* 2007
5: Land hydrology contribution available from GRACE measurements only since 2003

Source: Pfeffer 2009

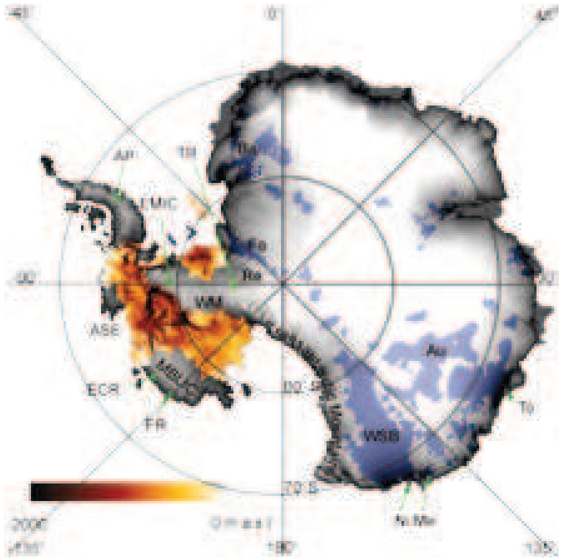
function. Millions of people in low-lying nations such as Bangladesh, along deltas and river systems like the Mekong, and on islands such as Tuvalu will have to respond to rising sea levels during the 21st century and beyond. Developing and developed countries alike have significant challenges ahead imposed by sea-level rise that will continue for hundreds of years (Church *et al.* 2008, Heberger *et al.* 2009, Karl *et al.* 2009, Solomon *et al.* 2009).

With growing population and infrastructure development human exposure to natural hazards is inevitably increasing. This is particularly true as the strongest population growth is located in coastal areas with greater exposure to floods, cyclones, and high tidal surges. To make matters worse, any land remaining available for urban growth is generally risk-prone, for instance along flood plains or on steep slopes subject to landslides (Nelleman *et al.* 2008). Currently, about 100 million people worldwide live within 1 metre of sea level and that number is growing every day (Anthoff *et al.* 2006).

Infrastructure has been built along many vulnerable coastlines in developing countries, as well as in developed countries, because slopes are gentle enough for buildings, ocean waters are used to cool power plant turbines and industrial processes, and sewage systems discharge to ocean outfalls. Rising sea levels—that will continue to rise for centuries at rates that are not yet well constrained—will be a major determinant in relocating and building new transportation routes, as well as power and waste treatment plants. Relocating business districts and residential areas will become another vast challenge to coastal communities and to governments at every level (Heberger *et al.* 2009).

Estimates of how much regional and global sea levels will rise over particular periods of time have been vigorously discussed since the IPCC AR4 estimated a rise of only 18–59 centimetres (cm) over the 21st century. The discussions focus on the dynamic ice changes that were excluded from AR4 estimates because no consensus could be reached based on published literature available at that time (Solomon *et al.* 2009). Since the

Figure 3.4: West Antarctic Ice Sheet vulnerability to collapse



West Antarctic above-sea level surface topography in grey shading and below-sea level topography in browns defining the areas subject to rapid ice collapse. The browns range from 0 to 2000 metres below sea level. For clarity, the ice shelves in West Antarctica are not shown. In East Antarctica, areas more than 200 metres below sea level are indicated by blue shading. *Source: Bamber et al. 2009*

publication of the IPCC AR4, climatological modelling, without dynamic effects explicitly included, suggests that 21st century sea level could rise to 0.5 to 1.4 metres above the 1990 level (Horton *et al.* 2008, Rahmstorf *et al.* 2009). It is clear that while the estimates produced by different modelling studies agree on the general projected trend in global average sea level, they vary in the estimated magnitude of future sea-level rise.

As discussed in the section on Earth's Ice, there are indications of a larger contribution than had been estimated to sea-level rise from dynamic changes of glaciers, ice caps, and the Greenland and West Antarctic Ice Sheets over the last decade. In the shorter term—decades to centuries—glaciers and ice caps may contribute significantly faster to sea level than changes in melt rate alone would indicate. In the longer term—centuries to millennia—the Greenland and West Antarctic Ice Sheets could potentially raise sea level by 6 metres and 3.3 metres, respectively. There is abundant geologic evidence that melting ice sheets have raised sea level by very large amounts in decades to centuries, but it is unclear whether this is possible today given the present configurations of bedrock topography and of ice on Greenland and Antarctica (Pfeffer *et al.* 2008, Bamber *et al.* 2009, Dutton *et al.* 2009).

At this time there is still no robust method for modelling future dynamic glacier and ice cap or ice sheet contributions to sea level, but limiting values have been estimated for the next century. By considering rates of discharge from melt and from iceberg fluxes required to drain ice through existing marine outlets, it can be shown that a combined sea-level rise in excess of 1.15 metres from Greenland and Antarctica by 2100 is physically very unlikely. Similarly, glaciers and ice caps are realistically limited to no more than about 0.55 metres by 2100. Introduction of realistic future melt and discharge values into the same analysis suggests that plausible values of total global average sea-level rise, including all land-ice sources plus thermal expansion, may reach 0.8 to 2.0 metres by 2100, although no preferred value was established within this range (Pfeffer *et al.* 2008).

As discussed in Earth Systems (Chapter One), published estimates for sea-level rise beyond 2100 agree that global mean sea levels will continue to rise regardless of changes in the driving forces of ocean thermal expansion and melting of ice (Solomon *et al.* 2009, Siddall *et al.* 2009).

Immediate implications of sea-level rise are already daunting: According to the Institute for Public Works in Australia, for every 20 cm of sea-level rise the frequency of any extreme sea level of a given height increases by a factor of about 10. According to this approach, by 2100, a rise of sea level of 50 cm would produce events every day that now occur once a year and extreme events expected once during the whole of the 20th century will occur several times every year by the end of the 21st century (Hunter 2009). It is obvious that stringent measures will be needed to adapt to sea-level rise.

CIRCULATION

Fresh water from the melting Arctic sea-ice and from the Greenland Ice Sheet enter the North Atlantic and encounter warmer and saltier currents arriving from more temperate latitudes. Changes in quantities and other characteristics of the fresh water could affect dynamics of the thermohaline convection that sink into the deep ocean as a distinct mass and are a driving force of circulation patterns in the Atlantic Ocean.

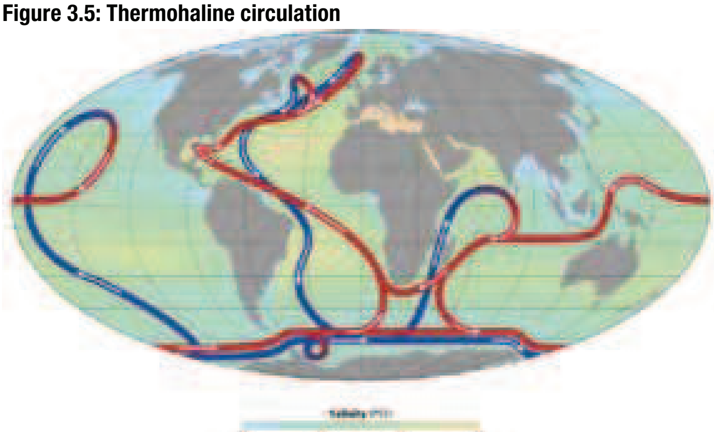
The deep mixing of ocean waters at high latitudes is important for the heat and carbon uptake of the oceans. This overturning is usually triggered by strong heat loss during the winter season. But with expectations of warming surface waters and the increased influx of fresh water in the high latitudes from sea-ice and glacier melt, it has been suggested that deep mixing may diminish or perhaps even cease in the near future: Both effects, warming and freshening, make the top layer of water less dense and therefore increasingly resistant to deep mixing (Lozier 2009).

As water is removed from the surface, it carries not only heat and salinity anomalies to great depths, but also anthropogenic carbon dioxide, absorbed when the water was still at the surface. The carbon dioxide concentration, like other water-mass properties, is transported to the deep ocean where it remains for hundreds of years. Therefore, the amount of carbon dioxide that has been—and will be—stored in the deep ocean is critically linked to the production of water masses through deep overturning events (Sabine *et al.* 2004).

The vertical exchange that feeds the North Atlantic Deep Water current moving south along the ocean floor seemed to slow for a few years in the early 21st century (Bryden *et al.* 2005, Alley 2007, Lozier 2009). More recently, strong ocean convection in gyres of the sub-polar North Atlantic seems to have returned (Våge *et al.* 2009, Yashayaev and Loder 2009).

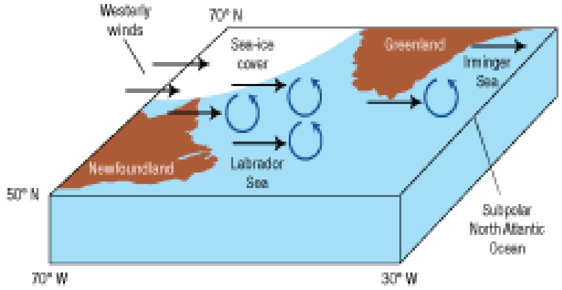
The strong mixing documented in the Irminger Sea to the east of Greenland's southern tip and in the Labrador Sea to the southwest is attributed to

Figure 3.5: Thermohaline circulation



Map shows general location and direction of the warm surface (red) and cold deep-water (blue) currents of the thermohaline circulation. Salinity is represented by colour in units of the Practical Salinity Scale (the conductivity ratio of a sea water sample to a standard KCl solution). Low values (blue) are less saline, while high values (orange) are more saline. *Source: NASA 2005*

Figure 3.6: Deep convection in the subpolar ocean



An extensive sea-ice cover maintained the strong, cold, westerly winds until they reached warmer open waters in the central basin of the Labrador Sea and in the Irminger Sea during the winter 2007-2008. There, the unusually cold winds rapidly cooled the surface water, leading to mixing of the water column to depths that had not been reached in the last 15 years. *Source: Lozier 2009*

cold air arriving from Canada that initiates a heat transfer from the ocean to the air, with a consequential sinking mass of cold water. In recent winters, higher temperatures of water flowing south through the Davis Strait have warmed the cold air from the west. However, in the winter of 2007 to 2008 after record Arctic sea-ice loss, the surface water flowing south was melt from that loss, colder and fresher than usual, so with the winter it froze quickly over the Davis Strait. The cold air from the west stayed chilly until it reached the relatively warm water off Greenland, where the subsequent energy exchange triggered renewal of a vertical exchange (Våge *et al.* 2009). This unexpected feedback from colder fresher currents delivered to the west of Greenland demonstrates the complexity of Earth Systems involved in the distribution of heat in a changing climate.

In the Southern Ocean, circulation is closely coupled with the dominant westerly winds that ring Antarctica. Observations show a significant intensification of the Southern Hemisphere westerlies, the prevailing winds between the latitudes of 30° and 60° S, over the past decades. The response of the Antarctic Circumpolar Current and the carbon sink in the Southern Ocean to changes in wind stress and surface buoyancy fluxes is under debate: Do enhanced winds support more upwelling or are they dissipated at levels near the surface? Analysis of data from the Argo network of profiling floats and historical oceanographic records detected coherent hemispheric-scale warming and freshening trends that extend to depths of more than 1,000 metres. The warming and freshening is partly related to changes in the properties of the water masses that make up the Antarctic Circumpolar Current, which are consistent with the anthropogenic changes in heat and freshwater fluxes suggested by climate models (Böning *et al.* 2008).

Beyond meridional overturning and creation of deep water, ocean water from different depths mixes through upwelling processes. Southern Ocean upwelling not only mixes water of differing salinity and temperature but it brings carbon-rich deeper water to the surface and delivers CO_2 to the atmosphere. Upwelling may have been a major contributor to the increase in atmospheric carbon during Pleistocene deglaciation (Anderson *et al.* 2009).

However, absorption of CO_2 by the oceans accounts for 30 to 40 per cent of the excess that has been emitted from anthropogenic sources since the beginning of the industrial revolution (Canadell *et al.* 2007). Recent research has reported a possible slowing in the uptake of CO_2 by the Southern Ocean (Le Quéré *et al.* 2007, Lenton *et al.* 2009).

ACIDIFICATION

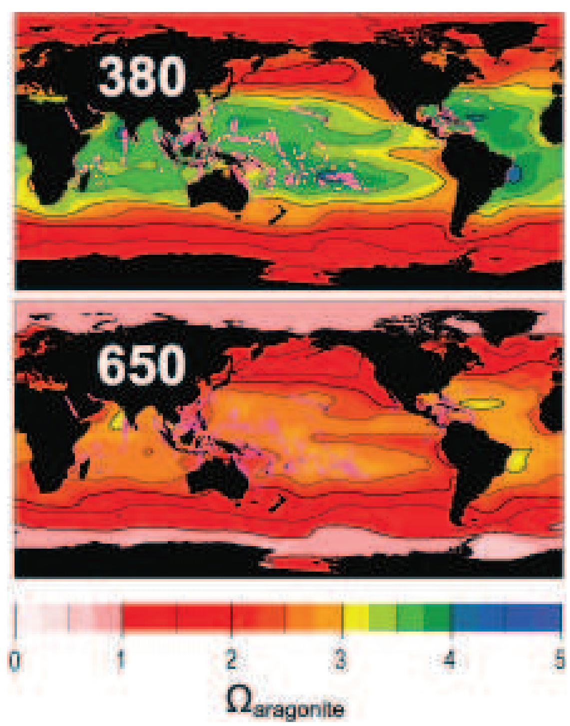
Further physical changes in the world's oceans can be attributed to mounting concentrations of CO_2 in the atmosphere. While increases in water temperature and fluctuations between fresh and saline water affect circulation at the surface and with vertical exchange, the repercussions from increasing concentrations of CO_2 in the oceans introduce a separate but related threat. The ocean's role in absorbing anthropogenic CO_2 released into the atmosphere has been underway for over two centuries. This has altered the chemistry of the global ocean fundamentally, by acidifying the top 2,000 metre layer of the oceans' waters and thus shrinking the total amount of ocean habitat where organisms that incorporate calcium carbonate ($CaCO_3$) into their shells and skeletons can thrive (Caldeira and Wickett 2003, Sabine *et al.* 2004, Orr *et al.* 2005, Denman *et al.* 2007, Feely *et al.* 2008, Ilyina *et al.* 2009, Silverman *et al.* 2009).

Box 3.1: Dense shelf water cascades

Understanding of the environmental, ecological, and societal importance of dense shelf water cascades (DSWC) and the role that submarine canyons play is progressively increasing. Matter and energy transfer from shallow to deep waters are the main fuel supplies for deep-water ecosystems and are essential for their maintenance. Contrary to the classical view of vertical settling of particles, or pelagic rain, new research results in the Mediterranean Sea have demonstrated that slope-determined horizontal injection of particles plays a key role in sustaining deep ecosystems and in moving chemical compounds, such as organic and mineralized carbon, into deepwater basins where they may remain sequestered over the long-term. Such injections often take the form of energetic cascades that last for weeks and can occur every few years. In addition to the northwestern Mediterranean area, cascades of dense shelf water have been identified in the Adriatic Sea and deduced from bed-form assemblages in the Aegean Sea. Because DSWC occurs in many ocean margins of the world it is very likely that these findings have global implications. Clearly, further studies in other regions of the world ocean are needed to fully resolve the likely cause and effect relations between DSWC and climate change.

Source: Puig et al. 2008, Allen and Durrieu de Madron 2009

Figure 3.7: Aragonite saturation and ocean pH change



Changes in surface ocean pH relative to pre-industrial values for different atmospheric CO_2 stabilization levels, 380 ppm and 650 ppm plotted over existing shallow-water coral reef locations (shown as magenta dots). Results are obtained by adding model-predicted perturbations in geochemical fields to modern observations, except for the Arctic Ocean where results are model simulations only due to a lack of observations. *Source: Cao and Caldeira 2008*

Box 3.2 The chemistry of acidification

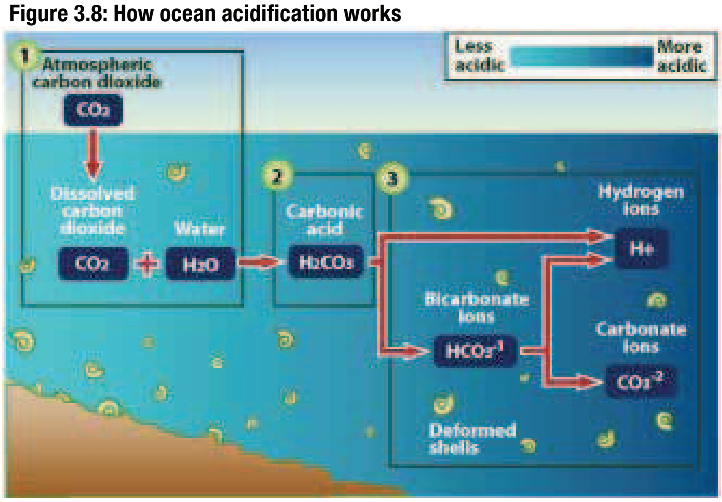
The oceanic uptake of anthropogenic CO_2 occurs through a series of well-known chemical reactions that increase aqueous CO_2, lower seawater pH, and lower carbonate ion levels. To the beginning of the 21st century, anthropogenic CO_2 has reduced average surface ocean acidity to 8.1 pH units from a pre-industrial value of 8.2 pH units on a logarithmic scale, a 30 per cent increase in acidity (Caldeira and Wickett 2003, Caldeira 2009). Acidification decreases the concentration of carbonate (CO_3), decreasing the saturation state of the $CaCO_3$ mineral calcite in the upper ocean that many marine organisms need to metabolize the shells and skeletons that support their functions.

Projected increase in anthropogenic CO_2 emissions will accelerate these chemical changes to rates unprecedented in the recent geological record. At current emission rates, atmospheric CO_2 concentrations will increase from 385 parts per million (ppm) in 2008 to 450–650 ppm by 2060, which would decrease average ocean surface acidity to an average of 7.9–7.8 pH units and reduce the saturation states of calcite and aragonite, two more $CaCO_3$ minerals, by 25 per cent—further shrinking optimal regions for biological carbonate formation (Doney and Schimel 2007, Doney *et al.* 2009, Steinacher *et al.* 2009, Cooley and Doney 2009).

Seasonal acidification events are already appearing—water that can corrode aragonite is welling up during the summer months along the California coastline, decades earlier than models predict (Feely *et al.* 2008). Researchers are anticipating the same degree of corrosive water in some high-latitude polar and subpolar locations by 2050 or earlier (Steinacher *et al.* 2009). But these model predictions and logical anticipations may be too conservative because they are based on scenarios that expected some decrease in CO_2 emissions by the early 21st century. Estimated fossil-fuel CO_2 emissions in 2005 exceeded those predicted by the most extreme scenario from the 1990s implying that future atmospheric CO_2 levels may exceed current model predictions, and the oceans may acidify faster than presently forecast.

Ongoing ocean acidification may harm a wide range of marine organisms and the food webs that depend on them, eventually degrading entire marine ecosystems (Fabry *et al.* 2008, Silverman *et al.* 2009, Doney *et al.* 2009). Laboratory studies suggest that molluscs, including species that support valuable marine fisheries such as mussels and oysters, and especially their juveniles, are particularly sensitive to these changes (Gazeau *et al.* 2007, Kurihara *et al.* 2007, Kurihara *et al.* 2009, Cooley and Doney 2009).

Figure 3.8: How ocean acidification works



1. Up to one half of the carbon dioxide (CO_2) released by burning fossil fuels over the past 200 years has been absorbed by the world's oceans
2. Absorbed CO_2 in seawater (H_2O) forms carbonic acid (H_2CO_3), lowering the water's pH level and making it more acidic
3. This raises the hydrogen ion concentration in the water, and limits organisms' access to carbonate ions, which are needed to form hard outer shells

Source: Adapted from University of Maryland 2009

Organisms' net responses to rising CO_2 will vary, depending on sensitivities to decreasing seawater pH, carbonate concentration, and carbonate saturation state and to increasing oceanic total inorganic carbon and gaseous CO_2. Shell-forming marine organisms create carbonate structures using one of two approaches:

Organisms that exert low biological control over calcification directly deposit $CaCO_3$ along their inner shell walls. Consequently, they depend on a sufficient ambient carbonate concentration to accumulate shells successfully. Commercially valuable molluscs such as scallops and oysters and some gastropods such as conchs use this method to build shells. Shells deposited in this manner are more likely to contain aragonite, a more soluble mineral form of $CaCO_3$. Corals form aragonite skeletons around their exterior, while coralline algae secrete aragonite or magnesium calcite, a moderately soluble form of $CaCO_3$ (Fabry *et al.* 2008, Doney *et al.* 2009).

Organisms that exert high biological control over calcification typically accumulate intracellular stocks of carbonate ions, gradually hardening their chitin and protein exoskeletons from within by depositing $CaCO_3$, the least soluble form of calcite. Sea urchins and crustaceans, including lobsters, shrimp, and crabs, follow this model and therefore do not require specific seawater chemistry to form shells. An organism's ultimate responses will also depend on factors such as individual history or genetic variability (Doney *et al.* 2009).

Many organisms, some of which are commercially valuable, also exhibit a range of damages to functions such as metabolism, reproduction, development, and immunity (Fabry *et al.* 2008, Holman *et al.* 2004, Burgents *et al.* 2005). Still unknown are the effects of acidification on the ability of fish to grow internal carbonate structures, which are important because they determine their advantages for feeding and migration.

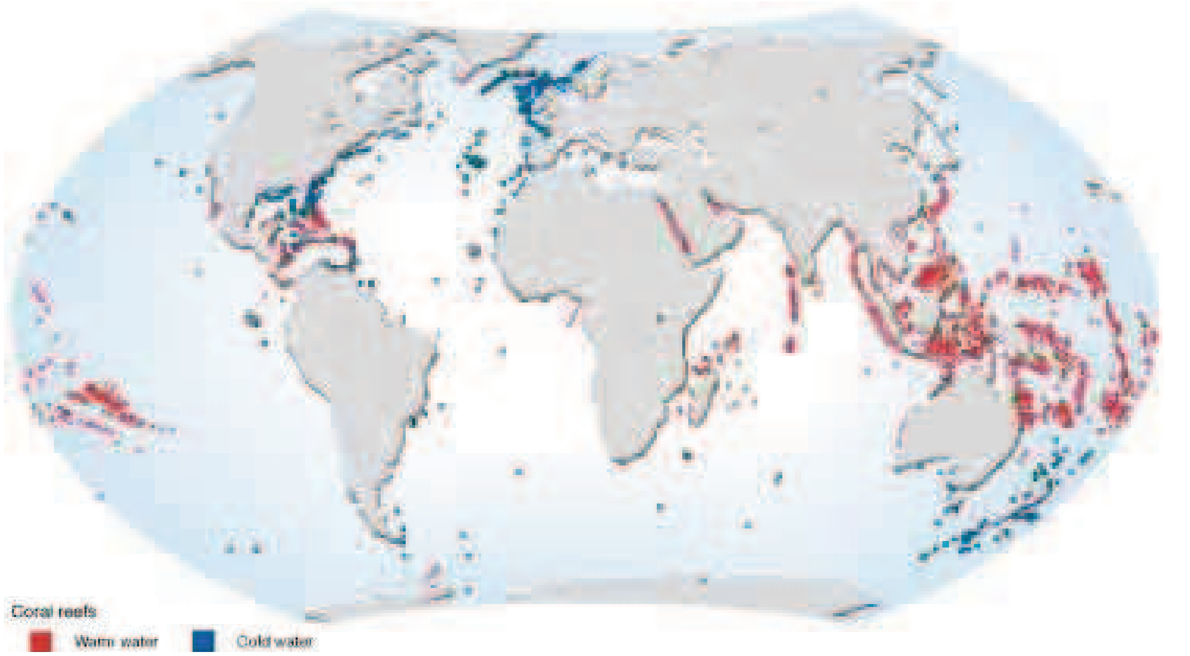
However, crabs, lobsters, shrimp, some planktons, and other organisms increase calcification or photosynthesis in seawater that is high in CO_2 (Ries *et al.* 2008a, Ries *et al.* 2008b, Doney *et al.* 2009). Whether the observed cases of increased calcification or photosynthesis result in any kind of advantage is not known. However, decreases in calcification and biological function due to ocean acidification are capable of decreasing the fitness of commercially valuable groups by directly damaging shells or by compromising early development and survival (Kurihara *et al.* 2007, Kurihara *et al.* 2009, Gazeau *et al.* 2007).

Ocean acidification's total effects on the marine environment will depend also on ecosystem responses. Even if carbonate-forming organisms do form shells and skeletons in elevated CO_2 conditions, they may encounter high energy costs that could reduce survival and reproduction (Wood *et al.* 2008, Kleypas *et al.* 2006). Losses of plankton, juvenile shellfish, and other organisms at the bottom of marine food chains have the potential to reduce harvests of economically important predator species. At the same time, acidic conditions will damage coral and prevent its re-growth, destroying crucial marine 'nursery' habitats and disrupting feeding and reproduction processes in a range of species (Kleypas *et al.* 2006, Lumsden *et al.* 2007).

Ecological shifts to algal overgrowth and decreased species diversity sometimes follow after coral disturbances, creating new ecosystem states that are stable but are then dominated by herbivores and less commercially valuable species. Ocean acidification has been implicated in similar ecological shifts from corals and other calcifying organisms to sea grasses and algae in communities with decreasing pH (Norström *et al.* 2009, Hall-Spencer *et al.* 2008, Wootton *et al.* 2009, Scheffer *et al.* 2001, Hoegh-Guldberg *et al.* 2007).

Ocean acidification will affect coral reefs and the ecosystems that depend on them in the more temperate latitudes because the processes involved are more robust in colder water. Coral reefs in warmer waters are subjected to the threat of coral bleaching.

Figure 3.9: Distribution of coldwater and tropical coral reefs



The coldwater reefs are highly susceptible to ocean acidification from climate change, which has its greatest impacts at high latitudes, while tropical reefs will become severely damaged by rising sea temperatures. In addition, both reef types must adapt to sea-level rise. *Source: UNEP 2008b*

CUMULATIVE EFFECTS

Since the early 1980s, episodes of coral reef bleaching and mortality, due primarily to climate-induced ocean warming, have occurred almost annually in one or more of the world's tropical or subtropical seas. Bleaching is episodic, with the most severe events typically accompanying coupled ocean–atmosphere phenomena, such as the El Niño Southern Oscillation, which result in sustained regional elevations of ocean temperature. Bleaching episodes have resulted in catastrophic loss of coral cover in some locations, and have changed coral community structure in many others, with a potentially critical influence on the maintenance of biodiversity in the marine tropics (Donner *et al.* 2007).

Bleaching has also set the stage for other declines in reef health, such as increases in coral diseases, the breakdown of reef framework by bioeroders, and the loss of critical habitat for associated reef fishes and other biota. Secondary ecological effects, such as the concentration of predators on remnant surviving coral populations, have also accelerated the pace of decline in some areas. Although bleaching severity and recovery have been variable across all spatial scales, some reefs have experienced relatively rapid recovery from severe bleaching impacts. Bleaching disturbances are likely to become a chronic stress in many reef areas in the coming decades, but if coral communities were exposed to the stress less intensely and over longer periods of time, they could adapt to changing conditions (Donner *et al.* 2007). Some reefs degraded by multiple stressors may already be approaching their end, although to date there has not been any global extinction of individual coral species as a result of bleaching events (Baker *et al.* 2008). Should such bleaching events increase in intensity and frequency, as has been projected by the best models, they will seriously degrade these important ecosystems.

If coral reefs were only threatened by rising sea levels they could possibly grow at the accelerated rates that are likely for the next century at least. But acidification and warming, as well as pollution and physical destruction, are weakening reefs further and they are unlikely to continue to provide 'fish nursery services' at the optimal rates required for healthy marine ecosystems (Hoegh-Guldberg *et al.* 2009).

Box 3.3 The Coral Triangle

Coral bleaching results when colonies of colorful photosynthesizing zooxanthellae algae abandon the calcareous structures built by coral polyps. Once bleached, the structures may or may not be re-colonized to re-establish the symbiotic relationships that characterize healthy reefs. Over the last three decades, coral reefs have experienced severe mass bleaching events in many tropical regions, including the Coral Triangle that stretches across six countries of Southeast Asia and Melanesia.

A recent review of coral reef health in the Coral Triangle indicates that acidification, warmer temperatures, and rising sea levels have already damaged coastal ecosystems. These three climate change-related processes act as stresses in addition to other anthropogenic threats such as destructive and unsustainable fishing techniques, increased chemical pollution, and higher levels of sedimentation due to onshore land use practices.

Source: Baskett et al. 2009, Hoegh-Guldberg et al. 2009, Oliver and Palumbi 2009