

Chapter 5. Sea-Air Interactions

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1. Introduction

From the physical point of view, the interaction between these two turbulent fluids, the ocean and the atmosphere, is a complex, highly nonlinear process, fundamental to the motions of both. The winds blowing over the surface of the ocean transfer momentum and mechanical energy to the water, generating waves and currents. The ocean in turn gives off energy as heat, by the emission of electromagnetic radiation, by conduction, and, in latent form, by evaporation.

The heat flux from the ocean provides one of the main energy sources for atmospheric motions. This source of energy for the atmosphere is affected by the turbulence at the air/sea interface, and by the spatial distribution of the centres of high and low energy transfer affected by the ocean currents. This coupling takes place through processes that fundamentally occur at small scales. The strength of this coupling depends on air-sea differences in several factors and therefore has geographic and temporal scales over a broad range. At these small scales on the sea-surface interface itself, waves, winds, water temperature and salinity, bubbles, spray and variations in the amount of solar radiation that reaches the ocean surface, and other factors, affect the transfer of properties and energy.

In the long term, the convergence and divergence of oceanic heat transport provide sources and sinks of heat for the atmosphere and partly shape the mean climate of the earth. Analyzing whether these processes are changing due to anthropogenic influences and the potential impact of these changes is the subject of this chapter. Following guidance from the Ad Hoc Working Group of the Whole, much of the information presented here is based on or derives from the very thorough analysis conducted by the Intergovernmental Panel on Climate Change (IPCC) for its recent Fifth Assessment Report (AR5).

The atmosphere and the ocean form a coupled system, exchanging at the air-sea interface gases, water (and water vapour), particles, momentum and energy. These exchanges affect the biology, the chemistry and the physics of the ocean and influence its biogeochemical processes, weather and climate (exchanges affecting the water cycle are addressed in Chapter 4).

From a biogeochemical point of view, gas and chemical exchanges between the oceans and the atmosphere are important to life processes. Half of the Global Net Primary Production of the world is by phytoplankton and other marine plants, uptaking CO₂ and releasing oxygen (Field et al., 1998; Falkowski and Raven, 1997). Phytoplankton is

therefore also responsible for half of the annual production of oxygen by plants and, through the generation of organic matter, is at the basis of most marine food webs in the ocean. Oxygen production by plants is a critical ecosystem service that keeps atmospheric oxygen from otherwise declining. However, in many regions of the ocean, phytoplankton growth is limited by a deficit of iron in seawater. Most of the iron alleviating this limitation reaches the ocean through wind-borne dust from the deserts of the world.

Gas and chemical exchanges between the atmosphere and ocean are also important to climate change processes. For example, marine phytoplankton produces dimethyl sulphide (DMS), the most abundant biological sulphur compound emitted to the atmosphere (Kiene et al., 1996). DMS is oxidized in the marine atmosphere to form various sulphur-containing compounds, including sulphuric acid, which influence the formation of clouds. Through this interaction with cloud formation, the massive production of atmospheric DMS over the ocean may have an impact on the earth's climate. The absorption of CO₂ from the atmosphere at the sea surface is responsible for the fundamental role of the ocean as a carbon sink (see section 3 below).

2. Heat flux and temperature

2.1 Sea-Surface Temperature

Sea-surface temperature (SST) has been measured in surface waters by a variety of methods that have changed significantly over time. Furthermore the spatial patterns of SST change are difficult to interpret. Nevertheless a robust trend emerges from these historical series after careful inspection and analysis of the datasets. Figure 1 shows the historical SST trend instrumentally observed using the best datasets of spatially interpolated products, contrasted against the 1961 – 1990 climatology. Changes in SST are reported in this section and in Chapter 2 of the IPCC (Hartmann et al., 2013).

The IPCC in AR5 concluded that 'recent' warming (since the 1950s) is strongly evident in SST at all latitudes of each ocean. Prominent spatio-temporal structures, including the El Niño Southern Oscillation (ENSO), decadal variability patterns in the Pacific Ocean, and a hemispheric asymmetry in the Atlantic Ocean, were highlighted as contributors to the regional differences in surface warming rates, which in turn affect atmospheric circulation (Hartmann et al., 2013).

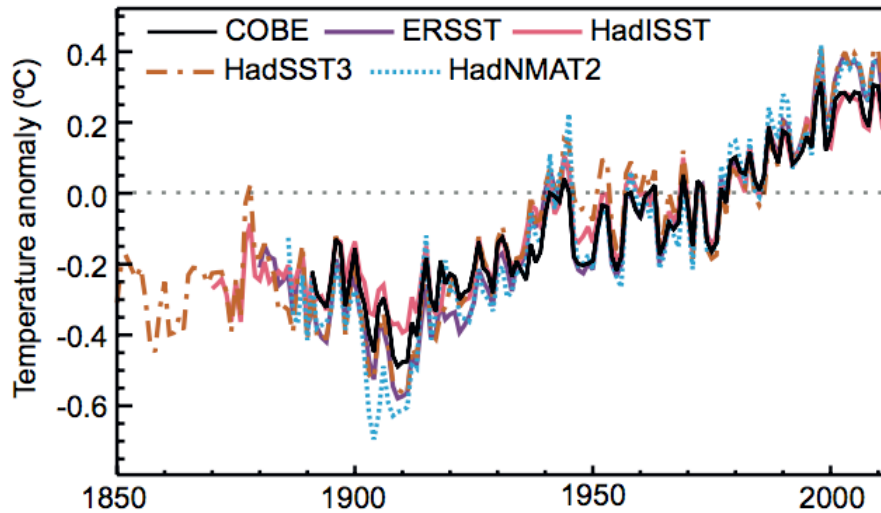


Figure 1. Global annual average sea surface temperature (SST) and Night Marine Air Temperature (NMAT) relative to a 1961–1990 climatology from state of the art data sets. Spatially interpolated products are shown by solid lines; non-interpolated products by dashed lines. From Hartmann et al. 2013, Fig. 2.18.

“It is certain that global average sea surface temperatures (SSTs) have increased since the beginning of the 20th century. (...) Intercomparisons of new SST data records obtained by different measurement methods, including satellite data, have resulted in better understanding of uncertainties and biases in the records. Although these innovations have helped highlight and quantify uncertainties and affect our understanding of the character of changes since the mid-20th century, they do not alter the conclusion that global SSTs have increased both since the 1950s and since the late 19th century.” (Hartmann et al., 2013).

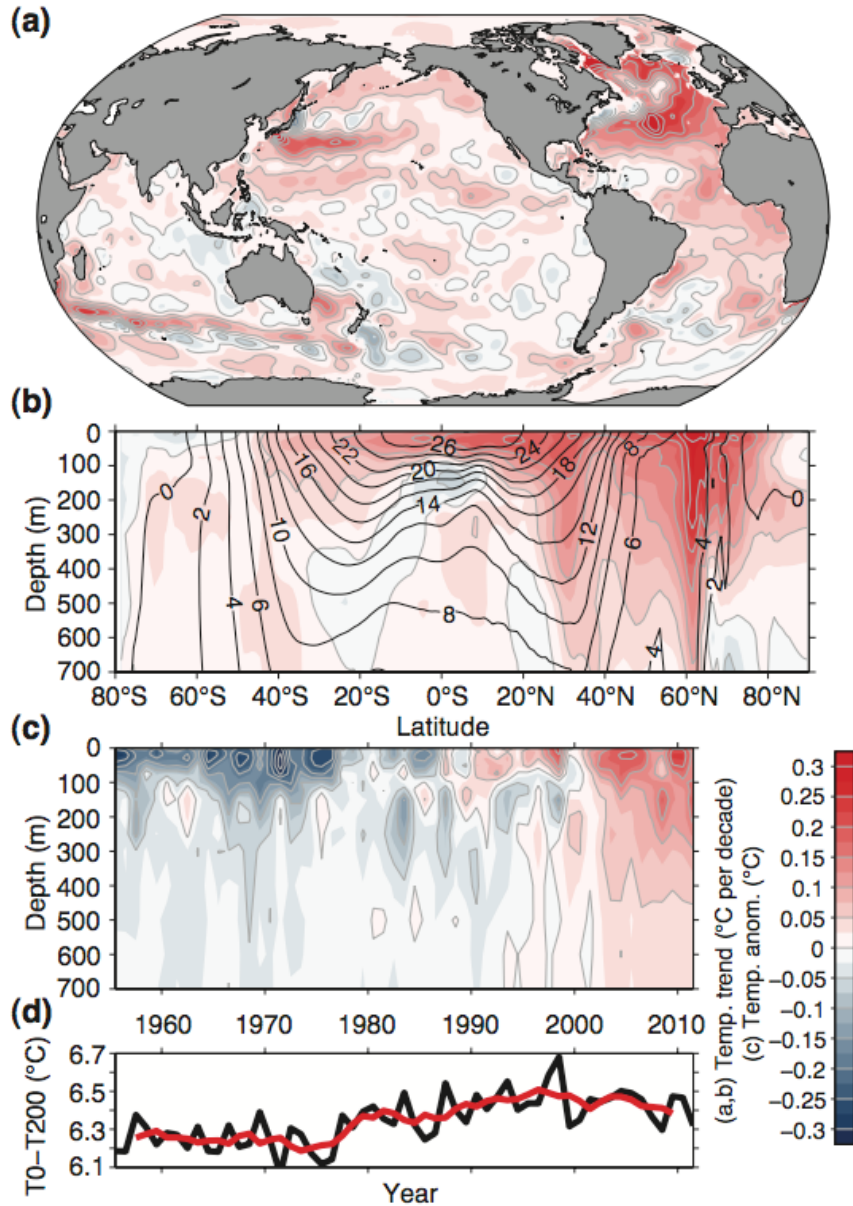
2.2 Changes in sea-surface temperature (SST) as inferred from subsurface measurements.

Upper ocean temperature (hence heat content) varies over multiple time scales, including seasonal, interannual (e.g., associated with El Niño), decadal and centennial (Rhein et al., 2013). Depth-averaged (0 to 700 m) ocean-temperature trends from 1971 to 2010 are positive over most of the globe. The warming is more prominent in the Northern Hemisphere, especially in the North Atlantic. This result holds true in different analyses, using different time periods, bias corrections and data sources (e.g., with or without XBT or MBT data¹) (Rhein et al. 2013). Zonally averaged upper-ocean temperature trends show warming at nearly all latitudes and depths (Figure 2a). However, the greater volume of the Southern Hemisphere ocean increases the contribution of its warming to the global heat content (Rhein et al., 2013). Strongest warming is found closest to the sea surface, and the near-surface trends are consistent

¹ XBT are expendable bathythermographs, probes that using electronic solid-state transducers register temperature and pressure while they free fall through the water column. MBT are their mechanical predecessors, that lowered on a wire suspended from a ship, used a metallic thermocouple as transducer.

with independently measured SST (Hartmann et al., 2013). The global average warming over this period is 0.11 [0.09 to 0.13] °C per decade in the upper 75 m, decreasing to 0.015°C per decade by 700 m (Figure 2c) (Rhein et al 2013).

The globally averaged temperature difference between the ocean surface and 200 m increased by about 0.25°C from 1971 to 2010. This change, which corresponds to a 4 per cent increase in density stratification, is widespread in all the oceans north of about 40°S. Increased stratification will potentially diminish the exchanges between the interior and the surface layers of the ocean; this will limit, for example, the input of nutrients from below into the illuminated surface layer and of oxygen from above into the deeper layers. These changes might in turn result in reduced productivity and increased anoxic waters in many regions of the world ocean (Capotondi et al., 2012).



The boundaries and names shown and the designations used on this map do not imply official endorsement or acceptance by the United Nations.

Figure 2. (a) Depth-averaged (0 to 700) m ocean-temperature trend for 1971-2010 (longitude vs. latitude, colours and grey contours in degrees Celsius per decade); (b) Zonally averaged temperature trends (latitude vs. depth, colours and grey contours in degree Celsius per decade) for 1971-2010 with zonally averaged mean temperature over-plotted (black contours in degrees Celsius). Both North (25-65°N) and South (south of 30°S), the zonally averaged warming signals extend to 700 m and are consistent with poleward displacement of the mean temperature field. Zonally averaged upper-ocean temperature trends show warming at nearly all latitudes and depths (Figure 2 (b)). A relative maximum in warming appears south of 30°S. (c) Globally averaged temperature anomaly (time vs. depth, colours and grey contours in degrees Celsius) relative to the 1971-2010 mean; (d) Globally averaged temperature difference between the ocean surface and 200 m depth (black: annual values, red: 5-year running mean). All panels are constructed from an update of the annual analysis of Levitus et al. (2009). From Rhein et al. (2013) Fig 3.1.

2.3 Upper Ocean Heat Content (UOHC)

The ocean's large mass and high heat capacity allow it to store huge amounts of energy: more than 1000 times that found in the atmosphere for an equivalent increase in temperature. The earth is absorbing more heat than it is emitting back into space, and nearly all this excess heat is entering the ocean and being stored there.

The upper ocean (0 to 700 m) heat content increased during the 40-year period from 1971 to 2010. Published rates range from 74 TW to 137 TW (1 TW = 10^{12} watts), while an estimate of global upper (0 to 700 m depth) ocean heat content change, using ocean statistics to extrapolate to sparsely sampled regions and estimate uncertainties (Domingues et al., 2008), gives a rate of increase of global upper ocean heat content of 137 TW (Rhein, et al. 2013). Warming of the ocean accounts for about 93 per cent of the increase in the Earth's energy inventory between 1971 and 2010 (high confidence), Melting ice (including Arctic sea ice, ice sheets and glaciers) and warming of the continents and atmosphere account for the remainder of the change in energy (Rhein et al. 2013). Global integrals of 0 to 700 m upper ocean heat content (UOHC) (Figure 3.) estimated from ocean temperature measurements all show a gain from 1971 to 2010 (Rhein et al. 2013).

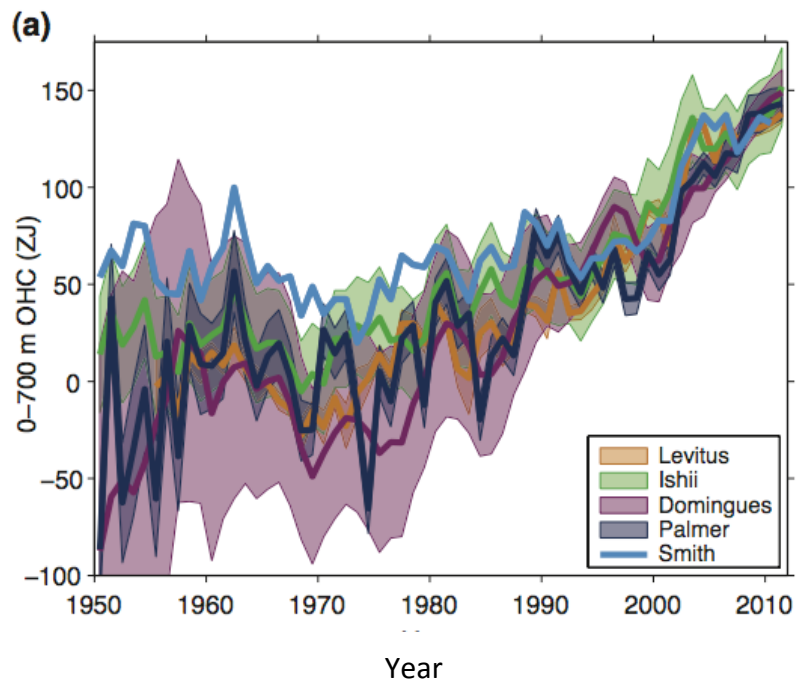


Figure 3. Observation-based estimates of annual global mean upper (0 to 700 m) ocean heat content in ZJ (1 ZJ = 10^{21} Joules) updated from (see legend): Levitus et al. (2012), Ishii and Kimoto (2009), Domingues et al. (2008), Palmer et al. (2009; ©American Meteorological Society. Used with permission.) and Smith and Murphy (2007). Uncertainties are shaded and plotted as published (at the one standard error level, other than one standard deviation for Levitus, with no uncertainties provided for Smith). Estimates are shifted to align for 2006-2010, 5 years that are well measured by the ARGO Program of autonomous profiling floats, and then plotted relative to the resulting mean of all curves for 1971, the starting year for trend calculations.

2.4 *The ocean's role in heat transport*

Solar energy is unevenly distributed over the earth's surface, leading to excess heat reaching the tropics and a heat deficit in latitudes poleward of about 40° in each hemisphere. The heat balance, and therefore a relatively stable climate, is maintained through the meridional redistribution, or flux, of heat by the atmosphere and the ocean. Quantification and understanding of this heat content and its redistribution have been achieved through diverse methods, including international programmes maintaining instrumented moorings, transoceanic lines of XBTs, satellite observations, numerical modelling and, more recently, the ARGO Program of autonomous profiling instruments (Abraham et al., 2013; von Schuckmann and Le Traon, 2011).

In the latitude band between 25°N and 25°S, the atmospheric and oceanic contributions to the meridional heat fluxes are similar, and the atmosphere dominates at higher latitudes. In the ocean, the heat flux is accomplished by contributions from the wind-driven circulation in the upper ocean, by turbulent eddies, and by the Meridional Overturning Circulation (MOC). The MOC is a component of ocean circulation that is driven by density contrasts, rather than by winds or tides, and one which exhibits a pronounced vertical component, with dense water sinking at high latitudes, offset by broadly distributed upwelling at lower ones. As distinct circulation patterns characterize each of the ocean basins, their individual contributions to the meridional heat flux differ significantly. Estimates indicate that, on a yearly average, the global oceans carry 1-2 PW (1PW=10¹⁵W) of heat from the tropics to higher latitudes, with somewhat higher transports to the northern hemisphere (Fasullo and Trenberth, 2008).

Most of the heat excess due to increases in atmospheric greenhouse gases goes into the ocean (IPCC, 2013). Although all ocean basins have warmed during the last decades, the increase in heat content is not uniform; the increase in heat content in the Atlantic during the last four decades exceeds that of the Pacific and Indian Oceans combined (Levitus et al., 2009; Palmer and Haines, 2009). Enhanced northward heat flux in the subtropical South Atlantic, which includes heat driven from the subtropical Indian Ocean through the Agulhas Retroflexion, may have contributed to the larger increase in heat content in the Atlantic Ocean compared with other basins (Abraham et al., 2013; Lee et al., 2011).

Numerical simulations also indicate that changes in ocean heat fluxes are the main mechanism responsible for the observed temperature fluctuations in the subtropical and subpolar North Atlantic (Grist et al., 2010).

Meridional heat flux estimates inferred from the residual of heat content variations suggest that the heat transferred northward throughout the Atlantic is transferred to the atmosphere in the subtropical North Atlantic (Kelly et al., 2014). Observations from the Rapid/Mocha instrument array at 26°N in the North Atlantic indicate that the mean Atlantic meridional heat flux at this latitude is 1.33 PW, with substantial variability due to changes in the strength of the MOC (Cunningham et al., 2007; Kanzow et al., 2007; Johns et al., 2011; McCarthy et al., 2012). Moreover, recent studies show that interannual changes in the MOC (and the associated heat flux measured at 26°N) lead to temperature anomalies in the subtropical North Atlantic which, in turn, can have a

strong impact on the northern hemisphere climate (Cunningham et al., 2013; Buchan et al., 2014).

2.5 *Air-sea Heat fluxes*

Heat uptake by the ocean can be substantially altered by natural oscillations in the earth's ocean and atmosphere. The effects of these large-scale climate oscillations are often felt around the world, leading to the rearrangement of wind and precipitation patterns, which in turn substantially affect regional weather, sometimes with devastating consequences.

The ENSO is the most prominent of these oscillations and is characterized by an anomalous warming and cooling of the central-eastern equatorial Pacific. The warm phase is called El Niño and the cold, La Niña. During El Niño events, a weakening of the Pacific trade winds decreases the upwelling of cold waters in the eastern equatorial Pacific and allows warm surface water that generally accumulates in the western Pacific to flow east.

As a consequence, El Niños release heat into the atmosphere, causing an increase in globally averaged air temperature. However, the “recharge oscillator theory” (Ren and Jin, 2013) indicates that a buildup of upper-ocean heat content is a necessary precondition for the development of El Niño events. La Niñas are associated with a strengthening of the trade winds, which leads to a strong upwelling of cold subsurface water in the eastern Pacific. In this case, the ocean uptake of heat from the atmosphere is enhanced, causing the global average surface temperature to decrease (Roemmich and Gilson, 2011).

The cycling of ENSO between El Niño and La Niña is irregular. In some decades El Niño has dominated and in other decades La Niña has been more frequent, also seen in phase shifts of the Interdecadal Pacific Oscillation (Meehl et al., 2013), which is related to build up and release of heat. A strengthening of the Pacific trade winds in the past two decades has led to a more frequent occurrence of La Niñas (England et al., 2014). Consequently, the heat uptake by the subsurface ocean was enhanced, leading to a slowdown of the surface warming (Kosaka and Xie, 2013). This is one of the factors affecting the global mean temperature, expected to increase by 0.21°C per decade from 1998 to 2012, but which instead warmed by just 0.04°C (the so-called recent warming hiatus, IPCC, 2013). Although there are several hypotheses on the cause of the global warming hiatus, the role of ocean circulation in this negative feedback is certain. Drijfhout et al. (2014) have shown that the North Atlantic, Southern Ocean and Tropical Pacific all play significant roles in the ocean heat uptake associated with the warming hiatus.

Chen and Tung (2014) analyzed the historical and recent record of sea surface temperature and Ocean Heat Content (OHC), and found distinct patterns at the surface and in deeper layers. On the surface, the patterns conform to the El Niño/La Niña patterns, with the Pacific Ocean playing a dominant role by releasing heat during an El Niño (or capturing heat during La Niña). At depth, the dominant pattern shows heating

taking place in the Atlantic Ocean and in the Circumpolar Current region. Coinciding in time, changes in OHC could help to explain the observed slowdown in global warming. It is anticipated that the mechanisms involved may at some point reverse, releasing large amounts of heat to the atmosphere and accelerating global warming (e.g., Levermann, et al., 2012).

Many other naturally occurring ocean-atmosphere oscillations in the Pacific, Atlantic, and Indian Oceans have also been recognized and named. The ENSO as a global phenomenon, has an expression in the Atlantic basin called the Atlantic Niño. In the last six decades, this mode has weakened, leading to a warming of the equatorial eastern Atlantic of up to 1.5°C (Tokinaga and Xie, 2011). Although the role of the Atlantic Niño on the global heat budget is not significant, this Atlantic warming trend has led to an increase in precipitation over the equatorial Amazon, Northeast South America, Equatorial West Africa and the Guinea coast, and a decrease in rainfall over the Sahel (Gianinni et al., 2003; Tokinaga and Xie, 2011; Marengo et al., 2011; Rodrigues et al., 2011). Moreover, recent studies have shown that the Atlantic Niño can have an effect on ENSO (Rodríguez-Fonseca et al., 2009; Keenlyside et al., 2013).

In the Indian Ocean, the dominant basin-wide oscillation is the Indian Dipole Mode (Saji et al., 1999). A positive phase is characterized by cool surface-temperature anomalies in the eastern Indian Ocean, warm-temperature anomalies in the western Indian Ocean, and easterly wind-stress anomalies along the equator. Similarly to ENSO, meridional heat transport and the associated buildup of upper-ocean heat content are a possible precondition for the development of the Indian Ocean Dipole event (McPhaden and Nagura, 2014). The warm surface temperatures in the western Indian Ocean are associated with an increase in subsurface heat content and vice-versa for the east (Feng et al. 2001; Rao et al., 2002). This zonal contrast of ocean heat content is induced by anomalies of zonal wind along the equator and the resulting variability in zonal mass and heat transport (Nagura and McPhaden 2010). The warm surface temperatures in the western Indian Ocean are associated with an increase in subsurface heat content and *vice-versa* for the east; the positive dipole causes above-average rainfall in eastern Africa and droughts in Indonesia and Australia (Behera et al., 2005; Yamagata et al., 2004; Ummenhofer et al., 2009; Cai et al., 2011; Section 5 below). Although the phenomena discussed here are global, many of the most significant impacts are on the coastal environment (see following Section).

2.6 *Environmental, economic and social impacts of changes in ocean temperature and of major ocean temperature events*

Coastal waters are valuable both ecologically and economically because they support a high level of biodiversity. They act as nursery areas for many commercially important fish species, and are the marine areas most accessible to the public. Because inshore habitats are shallow, water temperatures in coastal areas are closely linked to the regional climate and its seasonal and long-term fluctuations. Coastal waters also host some of the most vulnerable marine habitats, because they are intensively exploited by (including, but not limited to) the fishing industry and recreational craft, and because of

their proximity to outlets of pollution, such as rivers and sewage outfalls. Coastal development and the threat of rising sea level may also impinge upon these valuable habitats (Halpern et al., 2008). Ecological degradation can lower the socio-economic value of coastal regions, with negative impacts on commercial fisheries, aquaculture facilities, damage to coastal infrastructure, problems with power-station cooling, and exert a dampening effect on coastal tourism from degraded ecological services.

It has been recently shown that when compared with estimates for the global ocean, decadal rates of SST change are higher at the coast. During the last three decades, approximately 70 per cent of the world's coastline has experienced significant increases in SST (Lima and Wethey, 2012). This has been accompanied by an increase in the number of yearly extremely hot days along 38 per cent of the world's coastline, and warming has been occurring significantly earlier in the year along approximately 36 per cent of the world's temperate coastal areas (defined as those between latitudes 30° and 60° in both hemispheres) at an average rate of 6.1 ± 3.2 days per decade (Lima and Wethey, 2012).

The warming of coastal waters can have many serious consequences for the ecological system (Harley et al., 2006). This can include changes in the distribution of important commercial fish and shellfish species, particularly the movement of species to higher latitudes due to thermal stress (Perry et al., 2005). Warming of coastal waters also can lead to more favourable conditions for many organisms, among them marine invasive species that can devastate commercial fisheries and destroy marine ecosystem dynamics (Occhipinti-Ambrogi, 2007). Water quality might also be impacted by higher temperatures that can increase the severity of local outbreaks by pathogenic bacteria or the occurrence of Harmful Algal Blooms (HABs). These in turn would cause harm to seafood, consumers and marine organisms (Bresnan et al., 2013). Increased coral reef bleaching and mortality from warming seas (combined with ocean acidification, see next sections) will lead to the loss of important marine habitats and associated biodiversity.

Changes in ocean temperatures have global impacts. As ocean temperatures warm, species that prefer specific temperature ranges may relocate – as has been observed, for instance, in copepod assemblages in the North Atlantic (Hays et al., 2005). Some organisms, like corals, are sedentary and cannot relocate with changing temperatures. If the water becomes too warm, they may experience a bleaching event. Higher sea level and warmer ocean temperatures can alter ocean circulation and current flow and increase the frequency and intensity of storms, leading to changes in the habitat of many species worldwide.

Changes in ocean temperatures affect not only marine ecosystems, but also the climate over land, with devastating economic and social implications. Many natural oceanic oscillations are known to have an impact on (terrestrial) climate, but these oscillations and the response of the climate to them are also changing during recent decades. For instance, an El Niño phase of ENSO (see previous Section for more details on ENSO) displaces great amounts of warm water from the western to the eastern Pacific, leading to more evaporation over the latter. As a consequence, western and southern South America and parts of North America experience wetter conditions. At the same time, Australia, Brazil, India, Indonesia, the Philippines, parts of Africa and the United States

of America suffer droughts. La Niña events usually cause the opposite patterns. However, in the last several decades, ENSO events have changed their spatial and temporal characteristics (Yeh et al., 2009; McPhaden, 2012).

During recent decades, the warm waters of El Niño events have been displaced to the central Pacific instead of to the eastern Pacific. It is not clear yet whether these changes are linked to anthropogenic climate change or natural variability (Yeh et al., 2011). In any case, the effects on climate of an ENSO event centred in the central Pacific (a central Pacific ENSO) are in sharp contrast to that associated with one centred in the eastern Pacific.

For instance, northeastern and southeastern Australia experience a reduction in rainfall during the eastern Pacific El Niños and there is a decrease in rainfall over northwestern and northern Australia during central Pacific events (Taschetto and England, 2009; Taschetto et al., 2009). The Indian monsoon fails during eastern Pacific El Niños, but is enhanced during central Pacific El Niños (Kumar et al., 2006). Over the semi-arid region of northeast Brazil, eastern Pacific El Niños/La Niñas cause dry/wet conditions; central Pacific El Niños have the opposite effect, with the worst drought in the last 50 years associated with the strong 2011/12 La Niña and not with El Niños as in the past (Rodrigues et al., 2011; Rodrigues and McPhaden, 2014). This drought caused the displacement of 10 million people and economic losses on the order of 3 billion United States dollars in relation to agriculture and cattle raising alone. In contrast to drought in Brazil, the 2011/12 La Niña caused floods across southeastern Australia.

In other ocean basins, changes in oceanic oscillations and temperatures have also had an impact on climate. For instance, in the Indian Ocean, a positive phase of the Indian Dipole Mode (warm/cold temperatures in the western/eastern equatorial Indian Ocean) leads to flooding in east Africa and droughts in Indonesia, Australia, and India (Saji et al., 1999; Ashok et al., 2001; Gadgil et al., 2004; Yamagata et al., 2004; Behera et al., 2005; Ummenhofer et al., 2009; Cai et al., 2011). The counterpart of ENSO in the Atlantic (Atlantic Niño) has weakened during the last six decades, leading to an increase in SST in the eastern equatorial Atlantic. As a consequence, rainfall has been enhanced over the equatorial Amazon and West Africa (Tokinaga and Xie, 2011). On the other hand, an unusual warming of the tropical North Atlantic in 2005 was responsible for one of the worst droughts in the Amazon River basin and a record Atlantic hurricane season. Hurricanes Rita and Katrina caused the loss of almost 2000 lives and an estimated economic toll of 150 billion —135 billion US dollars from Katrina and 15 billion US dollars from Rita. (<http://www.datacenterresearch.org/data-resources/katrina/facts-for-impact/>). Anomalous warm conditions also occurred in the tropical North Atlantic in 2010 leading to two once-in-a-century droughts in less than five years in the Amazon River basin (Marengo et al., 2011).

Ocean warming will stress species both through thermic changes in their environmental envelope and through increased interspecies competition. These shifts become all the more important in shelf seas once they reach terrestrial boundaries, i.e., the shifting species runs out of shelf. For example, changes in the coastal currents in south-eastern Australia cause changes to primary production through to fisheries productivity. This then feeds through to local and regional socio-economic impacts (Suthers et al., 2011).

The IPCC AR5 concluded that “it is unlikely that annual numbers of tropical storms, hurricanes and major hurricanes counts have increased over the past 100 years in the North Atlantic basin. Evidence, however, is for a virtually certain increase in the frequency and intensity of the strongest tropical cyclones since the 1970s in that region” (Hartmann et al. 2013, Section 2.6.3). Moreover, the IPCC AR5 states that “it is difficult to draw firm conclusions with respect to the confidence levels associated with observed trends prior to the satellite era and in ocean basins outside of the North Atlantic” (Hartmann et al. 2013, Section 2.6.3). Although a strong scientific consensus on the matter does not exist, there is some evidence supporting the hypothesis that global warming might lead to fewer but more intense tropical cyclones globally (Knutson et al., 2010). Evidence exists that the observed expansion of the tropics since approximately 1979 is accompanied by a pronounced poleward migration of the latitude at which the maximum intensities of storms occur at a rate of 1° of latitude per decade (Kossin et al., 2014; Hartmann et al., 2013; Seidel et al., 2008). If this trend is confirmed, it would increase the frequency of events in coastal areas that are not exposed regularly to the dangers caused by cyclones. Hurricane Sandy in 2012 may be an example of this (Woodruff et al., 2013).

3. Water flux and salinity

3.1 Regional patterns of salinity, and changes in salinity² and freshwater content

The ocean plays a pivotal role in the global water cycle: about 85 per cent of the evaporation and 77 per cent of the precipitation occur over the ocean (Schmitt, 2008). The horizontal salinity distribution of the upper ocean largely reflects this exchange of freshwater: high surface salinity is generally found in regions where evaporation exceeds precipitation, and low salinity is found in regions of excess precipitation and runoff. Ocean circulation also affects the regional distribution of surface salinity.

The Earth’s water cycle involves evaporation and precipitation of moisture at the Earth’s surface. Changes in the atmosphere’s water vapour content provide strong evidence that the water cycle is already responding to a warming climate. Further evidence comes from changes in the distribution of ocean salinity (Rhein et al. 2013; FAQ. 3.2). Diagnosis and understanding of ocean salinity trends are also important, because salinity changes, like temperature changes, affect circulation and stratification, and therefore the ocean’s capacity to store heat and carbon as well as to change biological productivity.

Seawater contains both salt and fresh water, and its salinity is a function of the weight of dissolved salts it contains. Because the total amount of salt does not change over

² ‘Salinity’ refers to the weight of dissolved salts in a kilogram of seawater. Because the total amount of salt in the ocean does not change, the salinity of seawater can be changed only by addition or removal of fresh water.

human time scales, seawater's salinity can only be altered—over days or centuries—by the addition or removal of fresh water.

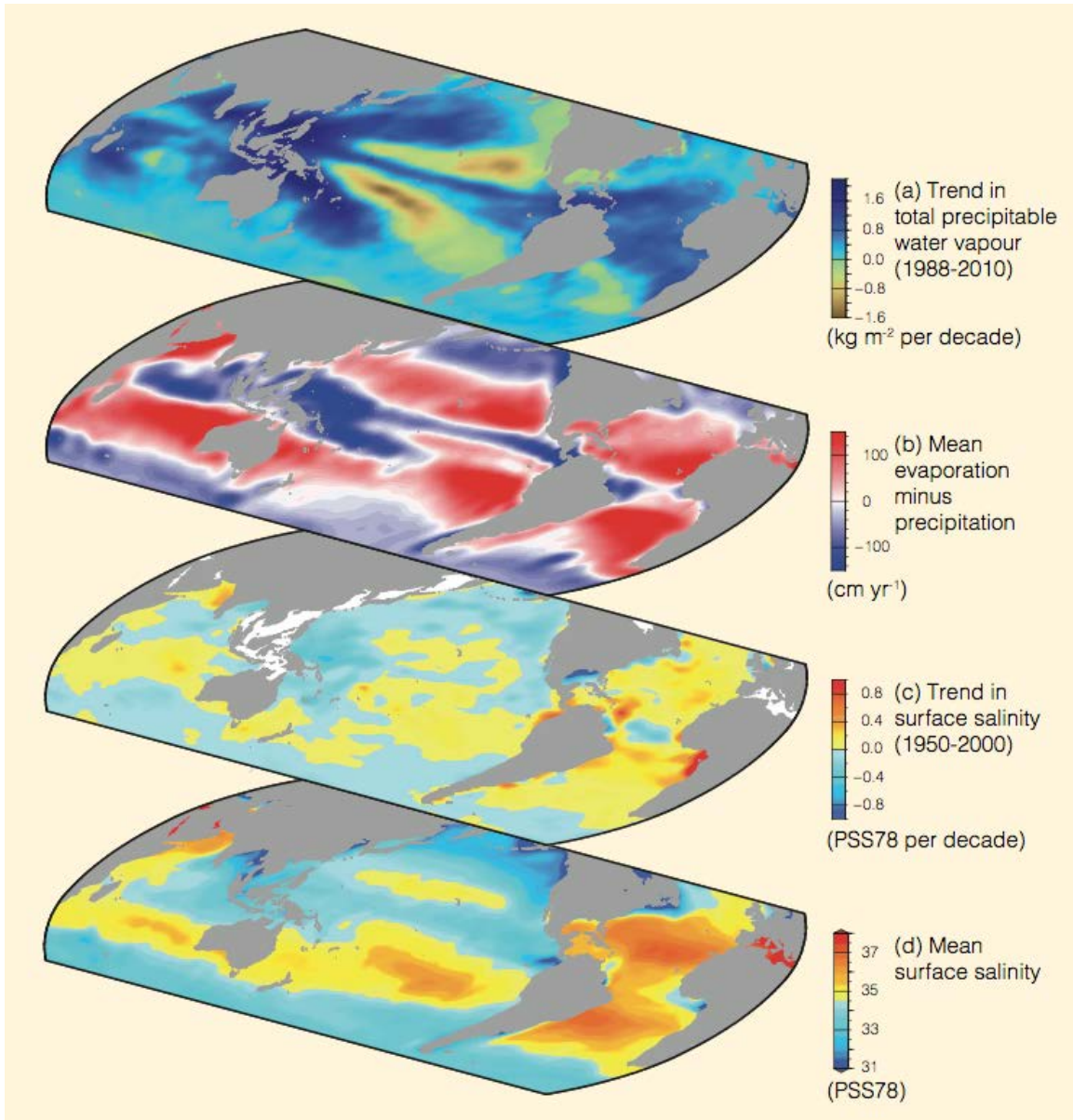
The water cycle is expected to intensify in a warmer climate. Observations since the 1970s show increases in surface and lower atmospheric water vapour (Figure 4a), at a rate consistent with observed warming. Moreover, evaporation and precipitation are projected to intensify in a warmer climate. Recorded changes in ocean salinity in the last 50 years support that projection (Rhein et al. 2013; FAQ. 3.2).

The atmosphere connects the ocean's regions of net fresh water loss to those of fresh water gain by moving evaporated water vapour from one place to another. The distribution of salinity at the ocean surface largely reflects the spatial pattern of evaporation minus precipitation (Figure 4b), runoff from land, and sea ice processes. There is some shifting of the patterns relative to each other, because of the ocean's currents. Ocean salinity acts as a sensitive and effective rain gauge over the ocean. It naturally reflects and smoothes out the difference between water gained by the ocean from precipitation, and water lost by the ocean through evaporation, both of which are very patchy and episodic (Rhein et al. 2013; FAQ. 3.2). Data from the past 50 years show widespread salinity changes in the upper ocean, which are indicative of systematic changes in precipitation and runoff minus evaporation.

(Figure 4b). Subtropical waters are highly saline, because evaporation exceeds rainfall, whereas seawater at high latitudes and in the tropics—where more rain falls than evaporates—is less so. The Atlantic, the saltiest ocean basin, loses more freshwater through evaporation than it gains from precipitation, while the Pacific is nearly neutral, i.e., precipitation gain nearly balances evaporation loss, and the Southern Ocean is dominated by precipitation. (Figure 4b; Rhein et al. 2013; FAQ. 3.2). Changes in surface salinity and in the upper ocean have reinforced the mean salinity pattern (4c). The evaporation-dominated subtropical regions have become saltier, while the precipitation-dominated subpolar and tropical regions have become fresher. When changes over the top 500 m are considered, the evaporation-dominated Atlantic has become saltier, while the nearly neutral Pacific and precipitation-dominated Southern Ocean have become fresher (Figure 4d; Rhein et al. 2013; FAQ. 3.2).

Observed surface salinity changes also suggest a change in the global water cycle has occurred (Chapter 4). The long-term trends show a strong positive correlation between the mean climate of the surface salinity and the temporal changes in surface salinity from 1950 to 2000. This correlation shows an enhancement of the climatological salinity pattern: fresh areas have become fresher and salty areas saltier.

Ocean salinity is also affected by water runoff from the continents, and by the melting and freezing of sea ice or floating glacial ice. Fresh water added by melting ice on land will change global-averaged salinity, but changes to date are too small to observe (Rhein et al. 2013; FAQ. 3.2).



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Figure 4. Changes in sea surface salinity are related to the atmospheric patterns of evaporation minus precipitation ($E - P$) and trends in total precipitable water: (a) Linear trend (1988–2010) in total precipitable water (water vapour integrated from the Earth’s surface up through the entire atmosphere) (kg m^{-2} per decade) from satellite observations (Special Sensor Microwave Imager) (after Wentz et al., 2007) (blues: wetter; yellows: drier). (b) The 1979–2005 climatological mean net $E - P$ (cm yr^{-1}) from meteorological reanalysis (National Centers for Environmental Prediction/National Center for Atmospheric Research; Kalnay et al., 1996) (reds: net evaporation; blues: net precipitation). (c) Trend (1950–2000) in surface salinity (PSS78 per 50 years) (after Durack and Wijffels, 2010) (blues freshening; yellows-reds saltier). (d) The climatological-mean surface salinity (PSS78) (blues: <35 ; yellows–reds: >35). From Rhein et al. 2013; FAQ. 3.2. Fig 1.

In conclusion, according to the last IPCC AR5, “It is very likely that regional trends have enhanced the mean geographical contrasts in sea surface salinity since the 1950s: saline

surface waters in the evaporation-dominated mid-latitudes have become more saline, while relatively fresh surface waters in rainfall-dominated tropical and polar regions have become fresher” (Stocker et al., 2013). “The mean contrast between high- and low-salinity regions increased by 0.13 [0.08 to 0.17] from 1950 to 2008. It is very likely that the inter-basin contrast in freshwater content has increased: the Atlantic has become saltier and the Pacific and Southern Oceans have freshened. Although similar conclusions were reached in AR4, recent studies based on expanded data sets and new analysis approaches provide high confidence in this assessment” (Stocker et al., 2013). “The spatial patterns of the salinity trends, mean salinity and the mean distribution of evaporation minus precipitation are all similar. These similarities provide indirect evidence that the pattern of evaporation minus precipitation over the oceans has been enhanced since the 1950s (medium confidence)” Stocker et al., (2013). “Uncertainties in currently available surface fluxes prevent the flux products from being reliably used to identify trends in the regional or global distribution of evaporation or precipitation over the oceans on the time scale of the observed salinity changes since the 1950s” (Stocker et al., 2013).

4. Carbon dioxide flux and ocean acidification

4.1 Carbon dioxide emissions from anthropogenic activities

Since the start of Industrial Revolution, human activities have been releasing large amounts of carbon dioxide into the atmosphere. As a result, atmospheric CO₂ has increased from a glacial to interglacial cycle of 180-280 ppm to about 395 ppm in 2013 (Dlugokencky and Tans, 2014). Until around 1920, the primary source of carbon dioxide to the atmosphere was from deforestation and other land-use change activities (Ciais et al., 2013). Since the end of World War II, anthropogenic emissions of CO₂ have been increasing steadily. Data from 2004 to 2013 show that human activities (fossil fuel combustion and cement production) are now responsible for about 91 per cent of the total CO₂ emissions (Le Quéré et al. 2014).

CO₂ emissions from fossil fuel consumption can be estimated from the energy data that are available from the United Nations Statistics Division and the BP Annual Energy Review. Data in 2013 suggests that about 43 per cent of the anthropogenic CO₂ emissions were produced from coal, 33 per cent from oil and 18 per cent from gas, and 6 per cent from cement production (Figure 5).

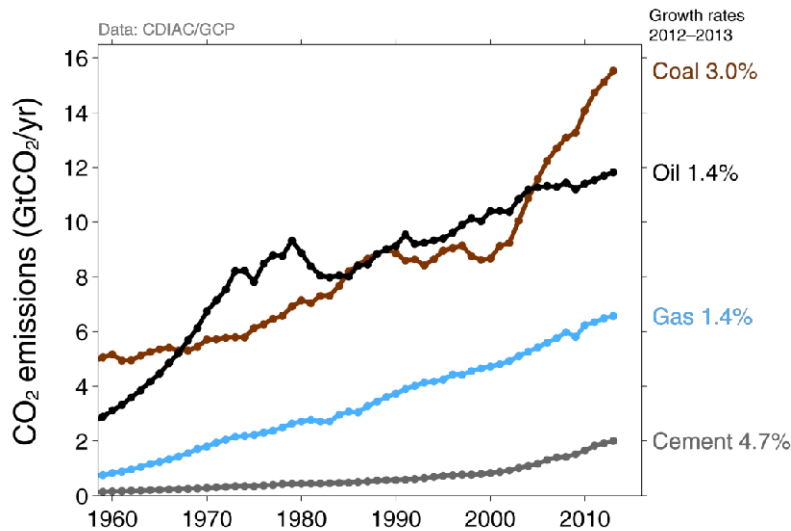


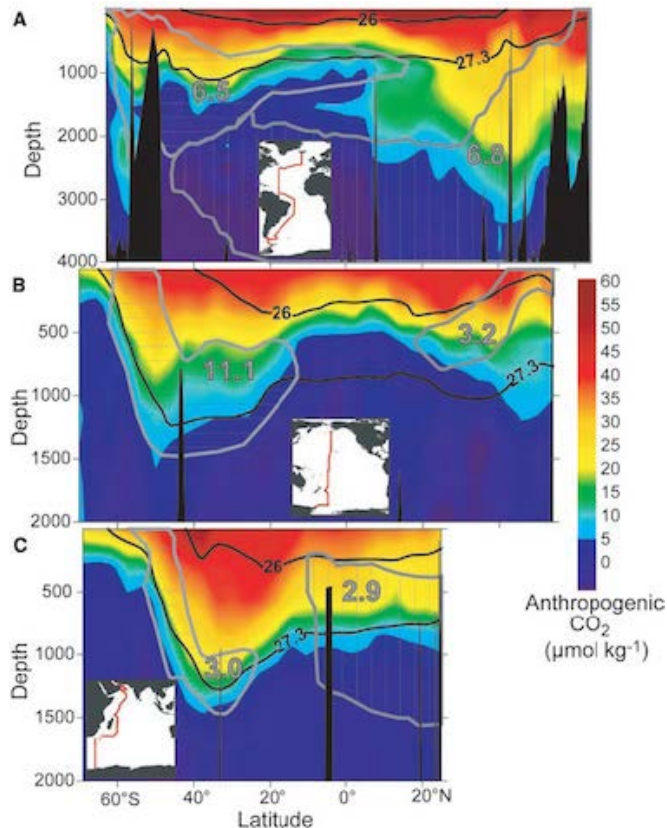
Figure 5. CO₂ emissions from different sources from 1958 to 2013 (Le Quéré et al. 2014).

Coal is an important and, recently, growing proportion of CO₂ emissions from fossil fuel combustion. From 2012 to 2013, CO₂ emissions from coal increased 3.0 per cent, compared to the increase rate of 1.4 per cent for oil and gas (Le Quéré et al. 2014). Coal accounted for about 60 per cent of the CO₂ emission growth in the same period. This is largely because many large economies of the world have recently resorted to using coal as an energy source for a wide variety of industrial processes, instead of oil, gas and other energy sources.

4.2 The ocean as a sink for atmospheric CO₂

The global oceans serve as a major sink of atmospheric CO₂. The oceans take up carbon dioxide through mainly two processes: physical air-sea flux of atmospheric CO₂ at the ocean surface, the so called "solubility pump" and through the active biological uptake of CO₂ into the biomass and skeletons of plankters the so-called "biological pump". Colder water can take up CO₂ more than warm water, and if this cold, denser water sinks to form intermediate, deep, or bottom water, there is transport of carbon away from the surface ocean and thus from the atmosphere into the ocean interior. This "solubility pump" helps to keep the surface waters of the ocean on average lower in CO₂ than the deep water, a condition that promotes the flux of the gas from the atmosphere into the ocean.

Phytoplankton take up CO₂ from the water in the process of photosynthesis, some of which sinks to the bottom in the form of particles or is mixed into the deeper waters as dissolved organic or inorganic carbon. Part of this carbon is permanently buried in the sediments and other part enters into the slower circulation of the deep ocean. This "biological pump" serves to maintain the gradient in CO₂ concentration between the surface and deep waters.



The boundaries and names shown and the designations used on this map do not imply official endorsement or acceptance by the United Nations.

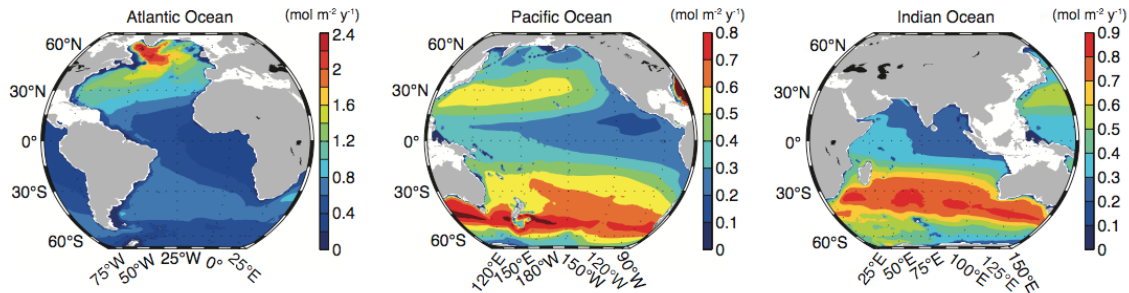
Figure 6. Anthropogenic CO₂ distributions along representative meridional sections in the Atlantic, Pacific, and Indian oceans for the mid-1990s (Sabine et al. 2004).

Because the ocean mixes slowly, about half of the anthropogenic CO₂ (C_{ant}) stored in the ocean is found in the upper 10 per cent of the ocean (Figure 6.). On average, the penetration depth is about 1000 meters and about 50 per cent of the anthropogenic CO₂ in the ocean is shallower than 400 meters.

Globally, the ocean shows large spatial variations in terms of its role as a sink of atmospheric CO₂ (Takahashi et al. 2009). Over the past 200 years the oceans have absorbed 525 billion tons of CO₂ from the atmosphere, or nearly half of the fossil fuel emissions over the period (Feely et al. 2009). The oceanic sink of atmospheric CO₂ has increased from 4.0 ± 1.8 GtCO₂ ($1 \text{ GtCO}_2 = 10^9$ tons of carbon dioxide) per year in the 1960s to 9.5 ± 1.8 GtCO₂ per year during 2004-2013. During the same period, the estimated annual atmospheric CO₂ captured by the ocean was 2.6 ± 0.5 Gt of CO₂ compared with around 1.9 Gt of CO₂ during the sixties (Le Queré et al., 2014). However, due to the decreased buffering capacity, caused by this CO₂ uptake, the proportion of anthropogenic carbon dioxide that goes into the ocean has been decreasing.

Estimates of the global inventory of anthropogenic carbon, C_{ant} (including marginal seas) have a mean value of 118 PgC and a range of 93 to 137 PgC in 1994 and a mean of 160 PgC and range of 134 to 186 PgC in 2010 (Rhein et al 2013). When combined with model results Khatiwala et al. (2013) arrive at a “best” estimate of the global ocean inventory (including marginal seas) of anthropogenic carbon from 1750 to 2010 of 155 PgC with an uncertainty of ± 20 per cent (Rhein et al 2013).

The storage rate of anthropogenic CO₂ is assessed by calculating the change in C_{ant} concentrations between two time periods. Regional observations of the storage rate are in general agreement with that expected from the increase in atmospheric CO₂ concentrations and with the tracer-based estimates. However, there are significant spatial and temporal variations in the degree to which the inventory of C_{ant} tracks changes in the atmosphere (Figure 7, Rhein et al 2013).



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Figure 7. Maps of storage rate distribution of C_{ant} in (mol m⁻² yr⁻¹) averaged over 1980-2005 for the three ocean basins (left to right: Atlantic, Pacific and Indian Ocean). From Khatiwala et al 2009, a slightly different colour scale is used for each basin.

Comprehensive evaluation of available data shows that in the context of the global carbon cycle, it is only the ocean that has acted as a net sink of carbon from the atmosphere. The land was a source early in the industrial age, and since about 1950 has trended toward a sink, but it is not yet clearly a net sink. (Ciais et al. 2013 and Khatiwala et al. 2009, Khatiwala et al. 2013). Latest data from 2004 to 2013 show that the global oceans take up about one-fourth (26 per cent, Le Quéré, 2014) of the total annual anthropogenic emissions of CO₂. This is a very important physical and ecological service that the ocean has performed in the past and performs today, that underpins all *strategies to mitigate the negative impacts of global warming*.

4.3 Ocean acidification

As already seen in the previous section, the global oceans serve as an important sink of atmospheric CO₂, effectively slowing down global climate change. However, this benefit comes with a steep bio-ecological cost. When CO₂ reacts with water, it forms carbonic acid, which then dissociates and produces hydrogen ions. The extra hydrogen ions consume carbonate ions (CO₃²⁻) to form bicarbonate (HCO₃⁻). In this process, the pH and concentrations of carbonate ions (CO₃²⁻) are decreasing. As a result, the carbonate mineral saturation states are also decreasing. Due to the increasing acidity, this process is commonly referred to as “ocean acidification (OA)”. According to the IPCC AR 4 and 5, “Ocean acidification refers to a reduction in pH of the ocean over an extended period, typically decades or longer, caused primarily by the uptake of carbon dioxide (CO₂) from the atmosphere.” (...) “Anthropogenic ocean acidification refers to the component of pH reduction that is caused by human activity” (Rhein et al. 2013).

Although the average oceanic pH can vary on interglacial time scales, the changes are usually on the order of ~ 0.002 units per 100 years; however, the current observed rate of change is ~ 0.1 units per 100 years, or roughly 50 times faster. Regional factors, such as coastal upwelling, changes in riverine and glacial discharge rates, and sea-ice loss have created “OA hotspots” where changes are occurring at even faster rates. Although OA is a global phenomenon that will likely have far-reaching implications for many marine organisms, some areas will be affected sooner and to a greater degree.

Recent observations show that one such area in particular is the cold, highly productive region of the sub-arctic Pacific and western Arctic Ocean, where unique biogeochemical processes create an environment that is both sensitive and particularly susceptible to accelerated reductions in pH and carbonate mineral concentrations. The OA phenomenon can cause waters to become undersaturated in carbonate minerals and thereby affect extensive and diverse populations of marine calcifiers.

4.4 *The CO₂ problem*

Based on the most recent data of 2004 to 2013, 35.7 GtCO₂ (1 GtCO₂ = 10⁹ tons of carbon dioxide) of anthropogenic CO₂ are released into the atmosphere every year (Le Quéré et al. 2014). Of this, approximately 32.4 GtCO₂ come directly from the burning of fossil fuels and other industrial processes that emit CO₂. The remaining 3.3 GtCO₂ are due to changes in land-use practices, such as deforestation and urbanization. Of this 35.7 GtCO₂ of anthropogenically produced CO₂ emitted annually, approximately 10.6 GtCO₂ (or 29 per cent) are incorporated into terrestrial plant matter. Another 15.8 GtCO₂ (or 46 per cent) are retained in the atmosphere, which has led to some planetary warming. The remaining 9.5 GtCO₂ (or 26 per cent) are absorbed by the world’s oceans (Le Quéré et al. 2014).

As the hydrogen ions produced by the increased CO₂ dissolution take carbonate ions out of seawater, the rate of calcification of shell-building organisms is affected; they are confronted with additional physiological challenges to maintain their shells. Although alteration of the carbonate equilibrium system in the ocean reducing carbonate ion concentration, and saturation states of calcium carbonate minerals will play a role imposing an additional energy cost to calcifier organisms, such as corals and shellbearing plankton, this is by no means the sole impact of OA.

4.5 *What are the impacts of a more acidic ocean?*

Throughout the last 25 million years, the average pH of the ocean has remained fairly constant between 8.0 and 8.2. However, in the last three decades, a fast drop has begun to occur, and if CO₂ emissions are left unchecked, the average pH could fall below 7.8 by the end of this century (Rhein, et al. 2013).

This is well outside the range of pH change of any other time in recent geological history. Calcifying organisms in particular, such as corals, crabs, clams, oysters and the tiny free-swimming pteropods that form calcium carbonate shells, could be particularly vulnerable, especially during the larval stage. Many of the processes that cause OA have

long been recognized, but the ecological implications of the associated chemical changes have only recently been investigated. OA may have important ecological and socioeconomic consequences by impacting directly the physiology of all organisms in the ocean.

The altered environment is imposing an extra energy cost for the acid-base regulation of their internal body milieu. Through biological and evolutionary adaptation this process might have a huge variation of expression among different types of organisms, a subject that only recently has become the focus of intense scientific research.

Calcification is an internal process that in its vast majority does not depend directly on seawater carbonate content, since most organisms use bicarbonate, that is increasing under acidification scenarios, or CO₂ originating in their internal metabolism. It has been demonstrated in the laboratory and in the field that some calcifiers can compensate and thrive in acidification conditions.

OA is not a simple phenomenon nor will it have a simple unidirectional effect on organisms. The abundance and composition of species may be changed, due to OA with the potential to affect ecosystem function at all trophic levels, and consequential changes in ocean chemistry could occur as well. Some species may also be better able than others to adapt to changing pH levels due to their exposure to environments where pH naturally varies over a wide range. However, at this point, it is still very uncertain what the ecological and societal consequences will be from any potential losses of keystone species.

4.6 Socioeconomic impacts of ocean acidification

Some examples of economic disruptions due to OA have been reported. The most visible case is the harvest failure in the oyster hatcheries along the Pacific Northwest coast of the USA. Hatcheries that supply the majority of the oyster spat to farms nearly went out of business as they unknowingly pumped low pH water, apparently corrosive to oyster larvae, into their operation. Although intense upwelling that could have brought low oxygen water to hatcheries might also be a factor in these massive mortalities, low pH, “corrosive water” tends to recur seasonally in this region. Innovations and interactions with scientists allowed these hatcheries to monitor the presence of corrosive incoming waters and adopt preventive measures.

Economic studies have shown that potential losses at local and regional scales may have negative impacts for communities and national economies that depend on fisheries. For example, Cooley and Doney (2009) using data from 2007, found that of the 4 billion dollars in annual domestic sales, Alaska and the New England states likely to be affected by hotspots of OA, contributed the most at 1.5 billion dollars and 750 million dollars, respectively. These numbers clearly show that any disruption in the commercial fisheries in these regions due to OA could have a cascading effect on the local as well as on the national economy.

References

- Abraham, J.P., Baringer, M., Bindoff, N.L., Boyer, T., Cheng, L.J., Church, J.A., Conroy, J.L., Domingues, C.M., Fasullo, J.T., Gilson, J., Goni, G., Good, S.A., Gorman, J.M., Gouretski, V., Ishii, M., Johnson, G.C., Kizu, S., Lyman, J.M., Macdonald, A.M., Minkowycz, W.J., Moffitt, S. E., Palmer, M.D., Piola, A.R., Reseghetti, F., Schuckmann, K., Trenberth, K.E., Velicogna, I., & Willis, J.K. (2013). A review of global ocean temperature observations: Implications for ocean heat content estimates and climate change, *Review of Geophysics*, 51, 450–483, doi:10.1002/rog.20022.
- Ashok K., Guan, Z., Yamagata, T. (2001), Impact of the Indian Ocean dipole on the relationship between the Indian monsoon rainfall and ENSO, *Geophysical Research Letters*, 28, 4499–4502.
- Behera S. K., Luo, J.–J., Masson, S., Delecluse, P., Gualdi, S., Navarra, A., Yamagata, T. (2005), Paramount impact of the Indian Ocean dipole on the east African short rains: a CGCM study. *Journal of Climate*, 18, 4514–4530.
- Bresnan, E., Davidson, K., Edwards, M., Fernand, L., Gowen, R., Hall, A., Kennington, K., McKinney, A., Milligan, S., Raine, R., Silke, J. (2013). Impacts of climate change on harmful algal blooms, *MCCIP Science Review*, 236-243, doi:10.14465/2013.arc24.236-243.
- Buchan, J., Hirschi, J. J.-M., Blaker, A.T., and Sinha, B. (2014). North Atlantic SST anomalies and the cold north European weather events of winter 2009/10 and December 2010. *Monthly Weather Review*, 142, 922–932, doi: 10.1175/MWR-D-13-00104.1.
- Cai W., van Rensch, P., Cowan, T., Hendon, H.H. (2011), Teleconnection pathways of ENSO and the IOD and the mechanisms for impacts on Australian rainfall, *Journal of Climate*, 24, 3910–3923.
- Capotondi, A., Alexander, M.A., Bond, N.A., Curchitser, E.N., and Scott, J.D. (2012): Enhanced upper ocean stratification with climate change in the CMIP3 models. *Journal of Geophysical Research*, 117, C04031, doi:10.1029/2011JC007409.
- Chen, X. and Tung, K.-K. (2014). Varying planetary heat sink led to global-warming slowdown and acceleration. *Science* 345, 897-903.
- Ciais, P., Sabine, C., Govindasamy, B., Bopp, L., Brovkin, V., Canadell, J., Chhabra, A., DeFries, R., Galloway, J., Heimann, M., Jones, C., Le Quéré, C., Myneni, R., Piao, S., and Thornton, P.(2013). Chapter 6: Carbon and Other Biogeochemical Cycles, in: *Climate Change 2013 The Physical Science Basis*, edited by: Stocker, T., Qin, D., and Plattner, G.-K., Cambridge University Press, Cambridge.
- Cooley. S. R., and Doney, S.C. (2009). Anticipating ocean acidification's economic consequences for commercial fisheries. *Environmental Research Letters* 4.

- Cunningham, S. A., Kanzow, T., Rayner, D., Baringer, M.O., Johns, W.E., Marotzke, J., Longworth, H.R., Grant, E.M., Hirschi, J.J.-M., Beal, L.M., Meinen, C.S., Bryden, H.L. (2007), Temporal variability of the Atlantic meridional overturning circulation at 26.5 °N, *Science*, 317, 935–938, doi: 10.1126/science.1141304.
- Cunningham, S. A., Roberts, C.D., Frajka-Williams, E., Johns, W.E., Hobbs, W., Palmer, M.D., Rayner, D., Smeed, D.A., and McCarthy, G. (2013), Atlantic Meridional Overturning Circulation slowdown cooled the subtropical ocean, *Geophysical Research Letters*, 40, 6202–6207, doi:10.1002/2013GL058464.
- Doney S.C., Ruckelshaus M., Duffy J.E., Barry J.P., Chan F., English C.A., Galindo H.M., Grebmeier J.M., Hollowed, A.B., Knowlton, N., Polovina, J., Rabalais, N.N., Sydeman, W.J., Talley, L.D. (2012). Climate change impacts on marine ecosystems. *Annual Review of Marine Science* 4:11-37.
- Doney, S.C., Mahowald, N., Lima, I., Feely, R.A., Mackenzie, F.T., Lamarque, J.-F. and Rasch, P.J. (2007). Impact of anthropogenic atmospheric nitrogen and sulfur deposition on ocean acidification and the inorganic carbon system. *Proceedings of the National Academy of Sciences of the United States of America*, 104, 14580-14585.
- Domingues, C.M., Church, J. A., White, N.J., Gleckler, P.J., Wijffels, S.E., Barker, P.M. and Dunn, J.A. (2008). Improved estimates of upper-ocean warming and multi-decadal sea-level rise. *Nature*, 453, 1090–1093.
- Dlugokencky, E. and Tans, P. (2014). <http://www.esrl.noaa.gov/gmd/ccgg/trends>, last access: 25 May 2015.
- Drijfhout, S.S., Blaker, A.T., Josey, S.A., Nurser, A.J.G., Sinha, B. and Balmaseda, M.A. (2014), Surface warming hiatus caused by increased heat uptake across multiple ocean basins, *Geophysical Research Letters*, 41, 7868–7874.
- England, M.H., McGregor, S., Spence, P., Meehl, G.A., Timmermann, A., Cai, W., Gupta, A.S., McPhaden, M.J., Purich, A., Santoso, A. (2014). Recent intensification of wind-driven circulation in the Pacific and the ongoing warming hiatus. *Nature Climate Change*, 4, 403–407.
- Fabry, V.J., McClintock, J.B., Mathis, J.T., and Grebmeier, J.M., (2009). Ocean Acidification at high latitudes: the Bellwether. *Oceanography*, 22(4), 160–171.
- Falkowski, P.G. and Raven, J.A. (1997). *Aquatic photosynthesis*. Blackwell Science, Malden, MA.
- Fasullo, J.T., and Trenberth, K.E. (2008). The annual cycle of the energy budget. Part II: Meridional structures and poleward transports, *Journal of Climate*, 21, 2313–2325, doi:10.1175/2007JCLI1936.1.
- Feng, M., Meyers, G. and Wijffels, S. (2001). Interannual upper ocean variability in the tropical Indian Ocean *Geophysical Research Letters*, 28(21), 4151-4154.

- Field, C.B., Behrenfeld, M.J., Randerson, J.T. and Falkowski, P. (1998). Primary production of the biosphere: integrating terrestrial and oceanic components, *Science* 281, 237–240.
- Gadgil, S., Vinayachandran, P.N., Francis, P.A., and Gadgil, S. (2004). Extremes of the Indian summer monsoon rainfall, ENSO and equatorial Indian Ocean oscillation, *Geophysical Research Letters*, 31, L12213, doi: 10.1029/2004GL019733.
- Grist, J.P., Josey, S.A., Marsh, R., Good, S.A., Coward, A.C., de Cuevas, B.A., Alderson, S.G., New, A.L., and Madec, G. (2010). The roles of surface heat flux and ocean heat transport convergence in determining Atlantic Ocean temperature variability, *Ocean Dynamics*, 60, 771–790, doi:10.1007/s10236-010-0292-4.
- Halpern, B.S., Walbridge, S., Selkoe, K.A., Kappel, C.V., Micheli, F., D'Agrosa, C., Bruno, J.F., Casey, K.S., Ebert, C., Fox, H.E., Fujita, R., Heinemann, D., Lenihan, H.S., Madin, E.M.P., Perry, M.T., Selig, E.R., Spalding, M., Steneck, R., and Watson, R. (2008). A global map of human impact on marine ecosystems, *Science*, 319, 948-952.
- Harley, C.D.G., Hughes, R.A., Hultgren, K.M., Miner, B.G., Sorte, C.J., Thornber, C.S., Rodriguez, L.F., Tomanek, L., Williams, S.L. (2006). The impacts of climate change in coastal marine systems, *Ecology Letters*, 9, 228–241.
- Hartmann, D.L., Klein Tank, A.M.G., Rusticucci, M., Alexander, L.V., Brönnimann, S., Charabi, Y., Dentener, F.J., Dlugokencky, E.J., Easterling, D.R., Kaplan, A., Soden, B.J., Thorne, P.W., Wild, M. and Zhai, P.M. (2013). Observations: Atmosphere and Surface. In: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* [Stocker, T.F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V. and Midgley, P.M. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.
- Hays, G.C., Richardson, A.J., and Robinson, C. (2005). *Trends in Ecology and Evolution*. 20, 337-344.
- IOC-UNESCO, (2011). *Methodology for the GEF Transboundary Waters Assessment Programme*. Volume 6. Methodology for the Assessment of the Open Ocean, UNEP, vi + 71 pp.
- IPCC (2013), *Climate Change 2013: The Physical Science Basis*. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. Stocker, T.F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V., and Midgley, P.M. (eds.). Cambridge University Press, Cambridge, UK, 1535 pp.
- Ishii, M., and Kimoto, M. (2009). Reevaluation of Historical Ocean Heat Content Variations with Time-Varying XBT and MBT Depth Bias Corrections. *Journal of Oceanography*, Vol. 65: pp. 287-299.

- Johns, W.E., Baringer, M.O. Beal, L.M., Cunningham, S.A., Kanzow, T., Bryden, H.L., Hirschi, J.J.M., Marotzke, J., Meinen, C.S., Shaw, B. and Curry, R. (2011). Continuous, array-based estimates of Atlantic Ocean heat transport at 26.5°N. *Journal of Climate*, 24, 2429–2449, doi: 10.1175/2010JCLI3997.1.
- Kanzow, T., Cunningham, S.A., Rayner, D., Hirschi, J.J.-M., Johns, W.E., Baringer, M.O., Bryden, H.L., Beal, L.M., Meinen, Ch. S., Marotzke J. (2007). Observed flow compensation associated with the MOC at 26.5 °N in the Atlantic, *Science*, 317,938–941, doi: 10.1126/science.1141304.
- Keenlyside, N.S., Ding, H. and Latif, M. (2013). Potential of equatorial Atlantic variability to enhance El Niño prediction, *Geophysical Research Letters*, 40, 2278–2283.
- Kelly, K.A., Thompson, L-A. and Lyman, J. (2014). The Coherence and Impact of Meridional Heat Transport Anomalies in the Atlantic Ocean Inferred from Observations. *Journal of Climate*, 27, 1469–1487.
- Kiene, R.P., Visscher, P.T., Keller, M.D. and Kirst, G.O. (eds.). (1996). *Biological and environmental chemistry of DMSP and related sulfonium compounds*. Plenum Press, New York (ISBN 0-306-45306-1).
- Knutson, T.R., McBride, J.L., Chan, J., Emanuel, K., Holland, G., Landsea, C., Held, I., Sugi, M. (2010). Tropical cyclones and climate change. *Nature Geoscience*, 3(3), 157-163.
- Kosaka, Y., and Xie, S.-P. (2013). Recent global-warming hiatus tied to equatorial Pacific surface cooling. *Nature*, 501, 403–407.
- Kossin, J.P., Emanuel, K.A., Vecchi, G.A. (2014).The poleward migration of the location of tropical cyclone maximum intensity. *Nature*, 509(7500), 349-352.
- Kumar, K.K., Rajagopalan, B., Hoerling, M., Bates, G. and Cane, M. (2006). Unraveling the mystery of Indian monsoon failure during El Niño, *Science*, 314, 115-119.
- Le Quéré, C., Moriarty, R. Andrew, R.M., Peters, G.P., Ciais, P., Friedlingstein, P., Jones, S.D., Sitch, S., Tans, P., Arneeth, A., Boden,T.A., Bopp, L., Bozec, Y., Canadell, J.G., Chevallier, F., Cosca, C.E., Harris, I., Hoppema, M., Houghton, R.A., House, J.I., Johannessen, T., Kato, E., Jain, A.K., Keeling, R.F., Kitidis, V., Klein Goldewijk, K., Koven, C., Landa, C., Landschützer, P., Lenton, A., Lima, I., Marland, G., Mathis, J.T., Metzl, N., Nojiri, Y., Olsen, A., Peters, W., Ono, T., Pfeil, B., Poulter, B., Raupach, M.R., Regnier, P., Rödenbeck, C., Saito, S., Salisbury, J.E., Schuster, U., Schwinger, J., Séférian, R., Segsneider, J., Steinhoff, T., Stocker, B.D., Sutton, A.J., Takahashi, T., Tilbrook, B., Viovy,N., Wang, Y.-P., Wanninkhof, R., Van der Werf, G., Wiltshire, A. and Zeng, N. (2014). Global Carbon Budget 2014. *Earth System Science Data Discuss.*, 7, 521-610. DOI:10.5194/essdd-7-521-2014.
- Lee, S.-K., W. Park, W., van Seville, E., Baringer, M.O., Wang, C., Enfield, D.B., Yeager, S.G., and B.P. Kirtman (2011). What caused the significant increase in Atlantic Ocean heat content since the mid-20th century? *Geophysical Research Letters*, 38, L17607, doi:10.1029/2011GL048856.

- Levermann, A., Bamber, J.L., Drijfhout, S., Ganopolski, A., Haeberli, W., Harris, N.R.P., Huss, M., Krüger, K., Lenton, T.M., Lindsay, R.W., Notz, D., Wadhams, P. and Weber, S. (2012). Potential climatic transitions with profound impact on Europe: Review of the current state of six ‘tipping elements of the climate system’. *Climatic Change*, 110, 845–878, DOI 10.1007/s10584-011-0126-5.
- Levitus, S., Antonov, J.I., Boyer, T.P., Locarnini, R.A., Garcia, H.E., and Mishonov, A.V. (2009). Global ocean heat content 1955–2008 in light of recently revealed instrumentation problems, *Geophysical Research Letters*, 36, L07608, doi:10.1029/2008GL037155.
- Lima, F.P., and Wethey, D.S. (2012). Three decades of high-resolution coastal sea surface temperatures reveal more than warming, *Nature communications*, 3.
- Marengo, J. A., Tomasella, J., Alves, L.M., Soares, W.R., and Rodriguez, D.A. (2011), The drought of 2010 in the context of historical droughts in the Amazon region, *Geophysical Research Letters*, 38, L12703, doi:10.1029/2011GL047436.
- McCarthy, G., Frajka-Williams, E., Johns, W.E., Baringer, M.O., Meinen, C.S., Bryden, H.L., Rayner, D., Duchez, A., Roberts, C., and Cunningham, S.A. (2012). Observed interannual variability of the Atlantic meridional overturning circulation at 26.5°N, *Geophysical Research Letters*, 39, L19609, doi:10.1029/2012GL052933.
- McPhaden, M. J., (2012). A 21st century shift in the relationship between ENSO SST and warm water volume anomalies, *Geophysical Research Letters*, 39, L09706, doi:10.1029/2012GL051826.
- McPhaden, M.J., and Nagura, M. (2014). Indian Ocean dipole interpreted in terms of recharge oscillator theory. *Climate Dynamics*, 42, 1569-1586.
- Meehl, G. A., Hu, A., Arblaster, J.M., Fasullo, J.Y., and Trenberth, K.E. (2013). Externally forced and internally generated decadal climate variability associated with the Interdecadal Pacific Oscillation, *Journal of Climate*, 26, 7298–7310.
- Nagura, M., and McPhaden, M.J. (2010). Dynamics of zonal current variations associated with the Indian Ocean dipole. *Journal of Physical Research*, 115, C 11026,
- Occhipinti-Ambrogi, A. (2007). Global change and marine communities: Alien species and climate change, *Marine Pollution Bulletin.*, 55, 342-52.
- Palmer, M. D., and Haines, K. (2009). Estimating oceanic heat content change using isotherms, *Journal of Climate*, 22, 4953–4969, doi: 10.1175/2009JCLI2823.1.
- Perry, A., Low, P. J., Ellis, J.R., Reynolds, J.D. (2005). Climate change and distribution shifts in marine fishes, *Science*, 308, 1912–15.
- Ren, H.-L., and Jin, F.-F. (2013). Recharge oscillator mechanisms in two types of ENSO. *Journal of Climate*, 26, 6506–6523. doi: <http://dx.doi.org/10.1175/JCLI-D-12-00601.1>.
- Rhein, M., Rintoul, S.R., Aoki, S., Campos, E., Chambers, D., Feely, R.A., Gulev, S., Johnson, G.C., Josey, S.A., Kostianoy, A., Mauritzen, C., Roemmich, D., Talley, L.D.

- and Wang, F. (2013). Observations: Ocean. In: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change* [Stocker, T.F., Qin, D., Plattner, G.-K, Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex V. and Midgley, P.M. (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.
- Rodrigues, R. R., Haarsma, R.J., Campos, E.J.D., and Ambrizzi, T. (2011). The impacts of inter-El Niño variability on the Tropical Atlantic and Northeast Brazil climate, *Journal of Climate*, 24, 3402–22.
- Rodriguez-Fonseca, B., Polo, I., Garcia-Serrano, J., Losada, T., Mohino, E., Mechoso, C.R., and Kucharski, F. (2009). Are Atlantic Niños enhancing Pacific ENSO events in recent decades? *Geophysical Research Letters*, 36, L20705, doi:10.1029/2009GL040048.
- Roemmich, D., and Gilson, J. (2011), The global ocean imprint of ENSO, *Geophysical Research Letters*, 38, L13606
- Sabine C. L., Feely, R. A., Gruber, N., Key, R. M., Lee, K., Bullister, J. L., Wanninkhof, R., Wong, C. S., Wallace, D. W. R., Tilbrook, B., Millero, F. J., Peng, T. H., Kozyr, A., Ono, T., and Rios, A. F., (2004). The Oceanic sink for anthropogenic CO₂. *Science* 305: 367-371.
- Sabine, C.L. and Feely, R.A. (2007). The oceanic sink for carbon dioxide. pp. 31–49. In Reay, D., Hewitt, N., Grace, J., and Smith, K. (Eds.), *Greenhouse Gas Sinks*, CABI Publishing, Oxfordshire, UK.
- Saji, N.H., Goswami, B.N., Vinayachandran, P.N., and Yamagata, T. (1999). A dipole mode in the tropical Indian Ocean. *Nature*, 401, 360–63.
- Seidel, D.J., Fu, Q., Randel, W.J., Reichler, T.J. (2008), Widening of the tropical belt in a changing climate. *Nature Geoscience*, 1(1), 21-24.
- Smith, D.M., Murphy, J.M. (2007). An objective ocean temperature and salinity analysis using covariances from a global climate model. *Journal of Geophysical Research*. Vol. 112, Issue C2.
- Stocker, T.F., Qin, D., Plattner, G.-K., Alexander, L.V., Allen, S.K., Bindoff, N.L., Bréon, F.-M., Church, J.A., Cubasch, U., Emori, S., Forster, P., Friedlingstein, P., Gillett, N., Gregory, J.M., Hartmann, D.L., Jansen, E., Kirtman, B., Knutti, R., Krishna Kumar, K., Lemke, P., Marotzke, J., Masson-Delmotte, V., Meehl, G.A., Mokhov, I.I., Piao, S., Ramaswamy, V., Randall, D., Rhein, M., Rojas, M., Sabine, C., Shindell, D., Talley, L.D., Vaughan, D.G., and Xie, S.-P. (2013). Technical Summary. In: *Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change*. Stocker, T.F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V. and Midgley, P.M. (eds.). Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

- Suthers, I.M., Young, J.W., Baird, M.E., Roughan, M., Everett, J.D., Brassington, G.B., Byrne, M., Condie, S.A., Hartog, J.R., Hassler, C.S., Hobday, A.J., Holbrook, N.J., Malcolm, H.A., Oke, P.R., Thompson, P.A., Ridgway, K. (2011). The strengthening East Australian Current, its eddies and biological effects - An introduction and overview. *Deep Sea Research Part II: Topical Studies in Oceanography*, 58 (2011) 538–546.
- Takahashi, T., Sutherland, S.C., Wanninkhof, R., Sweeney, C., Feely, R.A., Chipman, D.W., Hales, B., Friederich, G., Chavez, F., Watson, A., Bakker, D.C.E., Schuster, U., Metzl, N., Yoshikawa-Inoue, H., Ishii, M., Midorikawa, T., Nojiri, Y., Sabine, C., Olafsson, J., Arnarson, Th. S., Tilbrook, B., Johannessen, T., Olsen, A., Richard Bellerby, Körtzinger, A., Steinhoff, T., Hoppema, M., de Baar, H.J.W., Wong, C.S., Bruno Delille and Bates N.R. (2009). Climatological mean and decadal changes in surface ocean pCO₂, and net sea-air CO₂ flux over the global oceans. *Deep-Sea Research II*, 56, 554-57.
- Taschetto, A.S., and England, M.H. (2009). El Niño Modoki impacts on Australian rainfall, *Journal of Climate*, 22, 3167–3174, doi:10.1175/2008JCLI2589.1.
- Tokinaga, H., and Xie, S.-P. (2011). Weakening of the equatorial Atlantic cold tongue over the past six decades. *Nature Geosciences*, 4, 222–226.
- Ummenhofer, C.C., Sen Gupta, A., England, M.H., and Reason, C.J.C. (2009). Contributions of Indian Ocean sea surface temperatures to enhanced East African rainfall. *Journal of Climate*, 22, 993-1013.
- von Schuckmann, K. and Le Traon, P.-Y. (2011). How well can we derive Global Ocean Indicators from Argo data? *Ocean Science*, 7, 783-91, doi:10.5194/os-7-783-2011.
- Woodruff, J. D., Irish, J.L., and Camargo, S.J. (2013). Coastal flooding by tropical cyclones and sea-level rise. *Nature*, 504(7478), 44-52.
- Yamagata T., Behera, S.K., Luo, J.-J., Masson, S., Jury, M., Rao, S.A. (2004). *Coupled ocean–atmosphere variability in the tropical Indian Ocean*. *Earth Climate: Ocean-Atmosphere Interaction*, Geophysical Monograph Series, 147, 189–211.
- Yeh, S.-W., Kug, J.-S., Dewitte, B., Kwon, M.-H., Kirtman, B., and Jin, F.-F. (2009). El Niño in a changing climate, *Nature*, 461, 511–514, doi:10.1038/nature08316.
- Yeh, S.-W., Kirtman, B.P. Kug, J.-S. Park, W. and Latif, M. (2011). Natural variability of the central Pacific El Niño event on multi-centennial timescales, *Geophysical Research Letters*, 38, L02704, doi:10.1029/2010GL045886.