Chapter 4. The Ocean's Role in the Hydrological Cycle

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1. The interactions between the seawater and freshwater segments of the hydrological cycle

The global ocean covers 71 per cent of the Earth's surface, and contains 97 per cent of all the surface water on Earth (Costello et al., 2010). Freshwater fluxes into the ocean include: direct runoff from continental rivers and lakes; seepage from groundwater; runoff, submarine melting and iceberg calving from the polar ice sheets; melting of sea ice; and direct precipitation that is mostly rainfall but also includes snowfall. Evaporation removes freshwater from the ocean. Of these processes, evaporation, precipitation and runoff are the most significant at the present time.

Using current best estimates, 85 per cent of surface evaporation and 77 per cent of surface rainfall occur over the oceans (Trenberth et al., 2007; Schanze et al., 2010). Consequently, the ocean dominates the global hydrological cycle. Water leaving the ocean by evaporation condenses in the atmosphere and falls as precipitation, completing the cycle. Hydrological processes can also vary in time, and these temporal variations can manifest themselves as changes in global sea level if the net freshwater content of the ocean is altered.

Precipitation results from the condensation of atmospheric water vapour, and is the single largest source of freshwater entering the ocean (~530,000 km^3/yr). The source of water vapour is surface evaporation, which has a maximum over the subtropical oceans in the trade wind regions (Yu, 2007). The equatorward trade winds carry the water vapour evaporated in the subtropics to the Intertropical Convergence Zone (ITCZ) near the equator, where the intense surface heating by the sun causes the warm moist air to rise, producing frequent convective thunderstorms and copious rain (Xie and Arkin, 1997). The high rainfall and the high temperature support and affect life in the tropical rainforest (Malhi and Wright, 2011).

Evaporation is enhanced as global mean temperature rises (Yu, 2007). The water-holding capacity of the atmosphere increases by 7 per cent for every degree Celsius of warming, as per the Clausius-Clapeyron relationship. The increased atmospheric moisture content causes precipitation events to change in intensity, frequency, and duration (Trenberth, 1999) and causes the global precipitation to increase by 2-3 per cent for every degree Celsius of warming (Held and Soden, 2006).

Direct runoff from the continents supplies about 40,000 km^3/yr of freshwater to the ocean. Runoff is the sum of all upstream sources of water, including continental precipitation, fluxes from lakes and aquifers, seasonal snow melt, and melting of mountain glaciers and ice caps. River discharge also carries a tremendous amount of solid sediments and dissolved nutrients to the continental shelves.

The polar ice sheets of Greenland and Antarctica are the largest reservoirs of freshwater on the planet, holding 7 m and 58 m of the sea-level equivalent, respectively (Vaughan et al., 2013). The net growth or shrinkage of such an ice sheet is a balance between the net accumulation of snow at the surface, the loss from meltwater runoff, and the calving of icebergs and submarine melting at tidewater margins, collectively known as marine ice loss. There is some debate about the relative importance of these in the case of Greenland. Van den Broeke et al. (2009), show the volume transport to the ocean is almost evenly split between runoff of surface meltwater and marine ice loss. In a more recent work, Box and Colgan (2013) estimate marine ice loss at about twice the volume of meltwater (see Figure 5 in that article), with both marine ice loss and particularly runoff increasing rapidly since the late 1990s. According to the Arctic Monitoring and Assessment Programme (AMAP, 2011), the annual mass of freshwater being added at the surface of the Greenland Ice Sheet (the surface mass balance) has decreased since 1990. Model reconstructions suggest a 40% decrease from 350 Gt/y (1970 - 2000) to 200 Gt/y in 2007. Accelerating ice discharge from outlet glaciers since 1995 - 2002 is widespread and has gradually moved further northward along the west coast of Greenland with global warming. According to AMAP (2011), the ice discharge has increased from the pre-1990 value of 300 Gt/y to 400 Gt/y in 2005.

Antarctica's climate is much colder, hence surface meltwater contributions are negligible and mass loss is dominated by submarine melting and ice flow across the grounding line where this ice meets the ocean floor (Rignot and Thomas, 2002). Freshwater fluxes from ice sheets differ from continental river runoff in two important respects. First, large fractions of both Antarctic ice sheets are grounded well below sea level in deep fjords or continental shelf embayments; therefore freshwater is injected not at the surface of the ocean but at several hundred meters water depth. This deep injection of freshwater enhances ocean stratification which, in turn, plays a role in ecosystem structure. Second, unlike rivers, which act as a point source for freshwater entering the ocean, icebergs calved at the grounding line constitute a distributed source of freshwater as they drift and melt in adjacent ocean basins (Bigg et al., 1997; Enderlin and Hamilton, 2014).

Sea ice is one of the smallest reservoirs of freshwater by volume, but it exhibits enormous seasonal variability in spatial extent as it waxes and wanes over the polar oceans. By acting as a rigid cap, sea ice modulates the fluxes of heat, moisture and momentum between the atmosphere and the ocean. Summertime melting of Arctic sea ice is an important source of freshwater flux into the North Atlantic, and episodes of enhanced sea ice export to warmer latitudes farther south give rise to rapid freshening episodes, such as the Great Salinity Anomaly of the late 1960s (Gelderloos et al., 2012).

The spatial distributions of these freshwater fluxes drive important patterns in regional and global ocean circulation, which are discussed in Chapter 5.

The Southern Ocean (defined as all ocean area south of 60°S) deserves special mention due to its role in the storage of heat (and carbon) for the entire planet. The Antarctic Circumpolar Current (ACC) connects the three major southern ocean basins (South Atlantic, South Pacific and Indian) and is the largest current by volume in the world. The ACC flows eastward, circling the globe in a clockwise direction as viewed from the South Pole. In addition to providing a lateral connection between the major ocean basins (Atlantic, Indian, Pacific), the Southern Ocean also connects the shallow and deep parts of the ocean through a mechanism known as the meridional overturning circulation (MOC) (Gordon, 1986; Schmitz, 1996, see Figures I-90 and I-91). Because of its capacity to bring deep water closer to the surface, and surface water to depths, the Southern Ocean forms an important pathway in the global transport of heat. Although there is no observational evidence at present, (WG II AR5, 30.3.1, Hoegh-Guldberg, 2014) model studies indicate with a high degree of confidence that the Southern Ocean will become more stratified, weakening the surface-to-bottom connection that is the hallmark of present-day Southern Ocean circulation (WG I AR5 12.7.4.3, Collins et al., 2013). A similar change is anticipated in the Arctic Ocean and subarctic seas (WG I AR5 12.7.4.3, Collins et al., 2013), another region with this type of vertical connection between ocean levels (Wüst, 1928). These changes will result in fresher, warmer surface ocean waters in the polar and subpolar regions (WGII AR5 30.3.1, Hoegh-Guldberg, 2014; WG I AR5 12.7.4.3, Collins et al., 2013), significantly altering their chemistry and ecosystems.

Imbalances in the freshwater cycle manifest themselves as changes in global sea level. Changes in global mean sea level are largely caused by a combination of changes in ocean heat content and exchanges of freshwater between the ocean and continents. When water is added to the ocean, global sea level adjusts, rapidly resulting in a relatively uniform spatial pattern for the seasonal ocean mass balance, as compared to the seasonal steric signal, which has very large regional amplitudes (Chambers, 2006). 'Steric' refers to density changes in seawater due to changes in heat content and salinity. On annual scales, the maximum exchange of freshwater from land to ocean occurs in the late Northern Hemisphere summer, and therefore the seasonal ocean mass signal is in phase with total sea level with an amplitude of about 7 mm (Chambers et al., 2004). Because most of the ocean is in the Southern Hemisphere, the seasonal maximum in the steric component occurs in the late Southern Hemisphere summer, when heat storage in the majority of the ocean peaks (Leuliette and Willis, 2011). Because globally averaged sea level variations due to heat content changes largely cancel out between the Northern and Southern Hemispheres, the size of the steric signal, globally averaged, is only 4 mm.

Globally averaged sea level has risen at 3.2 mm/yr for the past two decades (Church et al., 2011), of which about a third comes from thermal expansion. The remainder is due to fluxes of freshwater from the continents, which have increased as the melting of continental glaciers and ice sheets responds to higher temperatures. Multi-decadal fluctuations in equatorial and mid-latitude winds (Merrifield et al., 2012; Moon et al.,

2013) cause regional patterns in sea-level trends which are reflected in the El Niño/Southern Oscillation (ENSO) and the Pacific decadal oscillation (PDO) indices in the Pacific (Merrifield et al., 2012; Zhang and Church, 2012) and northern Australia (White et al., 2014). Interannual changes in global mean sea level relative to the observed trend are largely linked to exchanges of water with the continents due to changes in precipitation patterns associated largely with the ENSO; this includes a drop of 5 mm during 2010-11 and rapid rebound in 2012-13 (Boening et al., 2012; Fasullo et al., 2013).

Some key alterations are anticipated in the hydrological cycle due to global warming and climate change. Changes that have been identified include shifts in the seasonal distribution and amount of precipitation, an increase in extreme precipitation events, changes in the balance between snow and rain, accelerated melting of glacial ice, and of course sea-level rise. Although a global phenomenon, it is the impact of sea-level rise along the world's coastlines that has major societal implications. The impacts of these changes are discussed in the next Section.

Changes in the rates of freshwater exchange between the ocean, atmosphere and continents have additional significant impacts. For example, spatial variations in the distribution of evaporation and precipitation create gradients in salinity and heat that in turn drive ocean circulation; ocean freshening also affects ecosystem structure. These aspects and their impacts are discussed in Sections 3 and 4.

Another factor potentially contributing to regional changes in the hydrological cycle are changes in ocean surface currents. For example, the warm surface temperatures of the large surface currents flowing at the western boundaries of the ocean basins (the Agulhas, Brazil, East Australian, Gulf Stream, and Kuroshio Currents) provide significant amounts of heat and moisture to the atmosphere, with a profound impact on the regional hydrological cycle (e.g., Rouault et al., 2002). Ocean surface currents like these are forced by atmospheric winds and sensitive to changes in them - stronger winds can mean stronger currents and an intensification of their effects (WGII AR5 30.3.1, Hoegh-Guldberg, 2014), as well as faster evaporation rates. Shifts in the location of winds can also alter these currents, for example causing the transport of anomalously warm waters (e.g., Rouault, 2009). However, despite a well-documented increase in global wind speeds in the 1990s (Yu, 2007), the overall effect of climate change on winds is complex, and difficult to differentiate observationally from decadal-scale variability, and thus the ultimate effects of these currents on the hydrological cycle are difficult to predict with any high degree of confidence (WGII AR5 30.3.1, Hoegh-Guldberg, 2014).

2. Environmental, economic and social implications of ocean warming

As a consequence of changes in the hydrological cycle, increases in runoff, flooding, and sea-level rise are expected, and their potential impacts on society and natural environment are among the most serious issues confronting humankind, according to the Fifth Assessment Report (AR5) of the United Nations Intergovernmental Panel on

Climate Change (IPCC). This report indicates that it is very likely that extreme sea levels have increased globally since the 1970s, mainly as a result of global mean sea-level rise due in part to anthropogenic warming causing ocean thermal expansion and glacier melting (WGI AR5 3.7.5, 3.7.6; WGI AR5 10.4.3). In addition, local sea-level changes are also influenced by several natural factors, such as regional variability in oceanic and atmospheric circulation, subsidence, isostatic adjustment, and coastal erosion, among others; combined with human perturbations by land-use change and coastal development (WGI AR5 5.3.2). A 4°C warming by 2100 (Betts et al., 2011; predicted by the high-end emissions scenario RPC8.5 in WGI AR5 FAQ12.1) leads to a median sea-level rise of nearly 1 m above 1980-1999 levels (Schaeffer et al., 2012).

The vulnerability of human systems to sea-level rise is strongly influenced by economic, social, political, environmental, institutional and cultural factors; such factors in turn will vary significantly in each specific region of the world, making quantification a challenging task (Nicholls et al., 2007; 2009; Mimura, 2013). Three classes of vulnerability are identified: (i) early impacts (low-lying island states, e.g., Kiribati, Maldives, Tuvalu, etc.); (ii) physically and economically vulnerable coastal communities (e.g., Bangladesh); and (iii) physically vulnerable but economically "rich" coastal communities (e.g., Sydney, New York). Table 1 outlines the main effects of relative sea-level rise on the natural system and provides examples of socio-economic system adaptations.

It is widely accepted that relative trends in sea-level rise pose a significant threat to coastal systems and low-lying areas around the world, due to inundation and erosion of coastlines and contamination of freshwater reserves and food crops (Nicholls, 2010); it is also likely that sea-level effects will be most pronounced during extreme episodes, such as coastal flooding arising from severe storm-induced surges, wave overtopping and rainfall runoff, and increases in sea level during ENSO events. An increase in global temperature of 4°C is anticipated to have significant socio-economic effects as sea-level rise, in combination with increasingly frequent severe storms, will displace populations (Field et al., 2012). These processes will also place pressure on existing freshwater resources through saltwater contamination (Nicholls and Cazenave, 2010). Figure 1 outlines in more detail the effects of sea-level rise on water resources of low-lying coastal areas.

Small island countries, such as Kiribati, Maldives and Tuvalu, are particularly vulnerable. Beyond this, entire identifiable coherent communities also face risk (e.g., Torres Strait Islanders; Green, 2006). These populations have nowhere to retreat to within their country and thus have no alternative other than to abandon their country entirely. The low level of economic activity also makes it difficult for these communities to bear the costs of adaptation. A shortage of data and local expertise required to assess risks related to sea-level rise further complicate their situation. Indeed the response of the island structure to sea-level rise is likely to be complex (Webb and Kench, 2010). Traditional customs are likely to be at risk and poorly understood by outside agencies. Yet traditional knowledge is an additional resource that may aid adaptation in such settings and should be carefully evaluated within adaptation planning. A significant part of the economy of many island nations is based on tourism; this too will be affected by sea-level rise through its direct effects on infrastructure and possibly also indirectly by the reduced availability of financial resources in the market (Scott et al., 2012).

Coastal regions, particularly some low-lying river deltas, have very high population densities. It is estimated that over 150 million people live within 1 metre of the high-tide level, and 250 million within 5 metres of high tide. Because of these high population densities (often combined with a lack of long-range urban planning), coastal cities in developing regions are particularly vulnerable to sea-level rise in concert with other effects of climate change (World Bank, 2012).

Table 1. The main effects of relative sea-level rise on the natural system, interacting factors, and examples of socio-economic system adaptations. Some interacting factors (for example, sediment supply) appear twice as they can be influenced both by climate and non-climate factors. Adaptation strategies: P = Protection; A = Accommodation; R = Retreat. Source: based on Nicholls and Tol, 2006.

Natural System Effects		Interacting Factors		Socio-economic System Adaptations
		Climate	Non-climate	
1. Inundation, flood and storm damage	a. Surge (sea)	– wave/storm climate – erosion – sediment supply	– sediment supply – flood management – erosion – land use	– dykes/surge barriers [P] – building codes/floodwise buildings [A] – land use planning/hazard delineation [A/R]
	b. Backwater effect (river)	– run-off	– catchment management – land use	
2. Wetland loss (and change)		– CO_2 fertilization – sediment supply	– sediment supply – migration space – direct destruction	– land-use planning [A/R] – managed realignment/forbid hard defences [R] – nourishment/sediment management [P]
3. Erosion (direct and indirect morphological change)		– sediment supply – wave/storm climate	– sediment supply	– coast defences [P] – nourishment [P] – building setbacks [R]
4. Saltwater Intrusion	a. Surface Waters	– run-off	– catchment management – land use	– saltwater intrusion barriers [P] – change water abstraction [A/R]
	b. Ground-water	– rainfall	– land use – aquifer use	– freshwater injection [P] – change water abstraction [A/R]
5. Rising water tables/impeded drainage		– rainfall – run-off	– land use – aquifer use – catchment management	– upgrade drainage systems [P] – polders [P] – change land use [A] – land use planning/hazard delineation [A/R]

Effects of sea-level rise are projected to be asymmetrical even within regions and countries. Nicholls and Tol (2006), extending the global vulnerability analysis of Hoozemans et al. (1993) on the impacts of and responses to sea-level rise with storm surges over the 21st century, show East Africa (including small island States and countries with extensive coastal deltas) as one of the problematic regions that could experience major land loss. Dasgupta et al. (2009) undertook a comparative study on the impacts of sea-level rise with intensified storm surges on developing countries globally in terms of its impacts on land area, population, agriculture, urban extent, major cities, wetlands, and local economies. They based their work on a 10 per cent future intensification of storm surges with respect to current 1-in-100-year storm-surge predictions. They found that Sub-Saharan African countries will suffer considerably from the impacts. The study estimated that Mozambique, along with Madagascar, Mauritania and Nigeria account for more than half (9,600 km^2) of the total increase in the region's storm-surge zones.

Of the impacts projected for 31 developing countries, just ten cities account for two-thirds of the total exposure to extreme floods. Highly vulnerable cities are found in Bangladesh, India, Indonesia, Madagascar, Mexico, Mozambique, the Philippines, Venezuela and Viet Nam (Brecht et al., 2012). Because of the small population of small islands and potential problems with implementing adaptations, Nicholls et al. (2011) conclude that forced abandonment of these islands seems to be a possible outcome even for small changes in sea level. Similarly, Barnett and Adger (2003) point out that physical impact might breach a threshold that pushes social systems into complete abandonment, as institutions that could facilitate adaptation collapse.

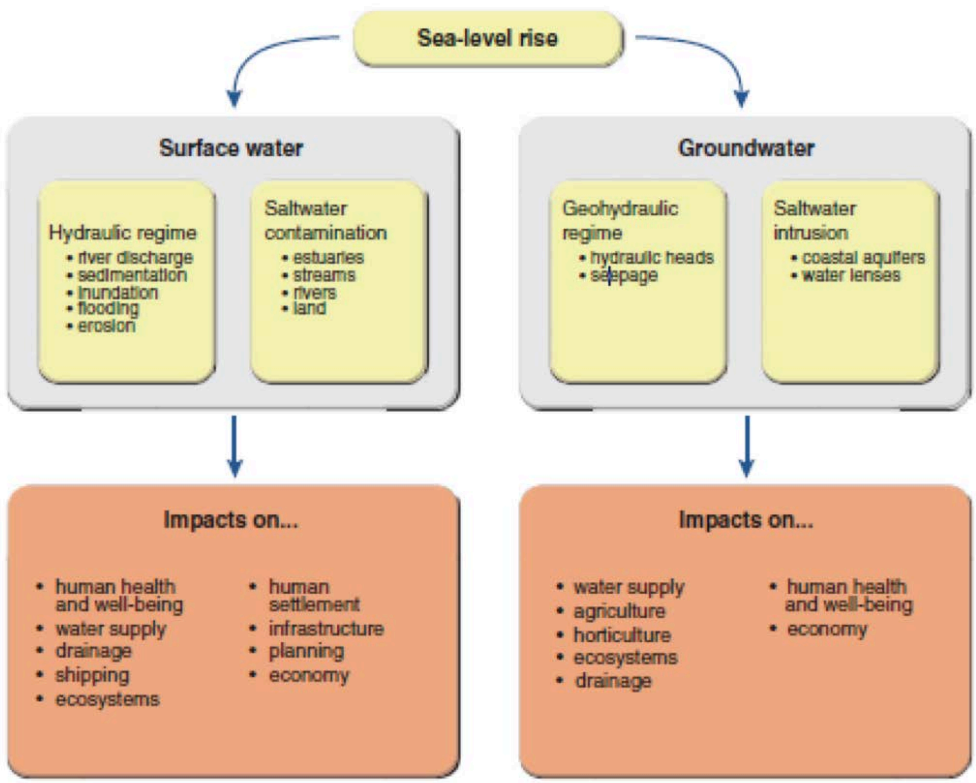

Figure 1. Effects of sea-level rise on water resources of small islands and low-lying coastal areas. Source: Based on Oude Essink et al. (1993); Hay and Mimura (2006).

Impacts of climate change on the hydrological cycle, and notably on the availability of freshwater resources, have been observed on all continents and many islands. Glaciers continue to shrink worldwide, affecting runoff and water resources downstream. Figure 2 shows the changes anticipated by the late 21st century in water runoff into rivers and streams. Climate change is the main driver of permafrost warming and thawing in both high-latitude and high-elevation mountain regions (IPCC WGII AR518.3.1, 18.5). This thawing has negative implications for the stability of infrastructure in areas now covered with permafrost.

Projected heat extremes and changes in the hydrological cycle will in turn affect ecosystems and agriculture (World Bank, 2012). Tropical and subtropical ecoregions in Sub-Saharan Africa are particularly vulnerable to ecosystem damage (Beaumont et al., 2011). For example, with global warming of 4°C (predicted by the high-end emissions scenario RPC8.5 in WGI AR5 FAQ 12.1), between 25 per cent and 42 per cent of 5,197 African plant species studied are projected to lose all their suitable range by 2085 (Midgley and Thuiller, 2011). Ecosystem damage would have the follow-on effect of reducing the ecosystem services available to human populations.

The Mediterranean basin is another area that has received a lot of attention in regard to the potential impacts of climate change on it. Several modelling groups are taking part in the MedCORDEX (www.medcordex.eu) international effort, in order to better simulate the Mediterranean hydrological cycle, to improve the modelling tools available, and to produce new climate impact scenarios. Hydrological model schemes must be improved to meet the specific requirements of semi-arid climates, accounting in particular for the related seasonal soil water dynamics and the complex surface-subsurface interactions in such regions (European Climate Research Alliance, 2011).

Even the most economically resilient of States will be affected by sea-level rise, as adaptation measures will need to keep pace with ongoing sea-level rise (Kates et al., 2012). As a consequence, the impacts of sea-level rise will also be redistributed through the global economic markets as insurance rates increase or become unviable and these costs are passed on to other sectors of the economy (Abel et al., 2011).

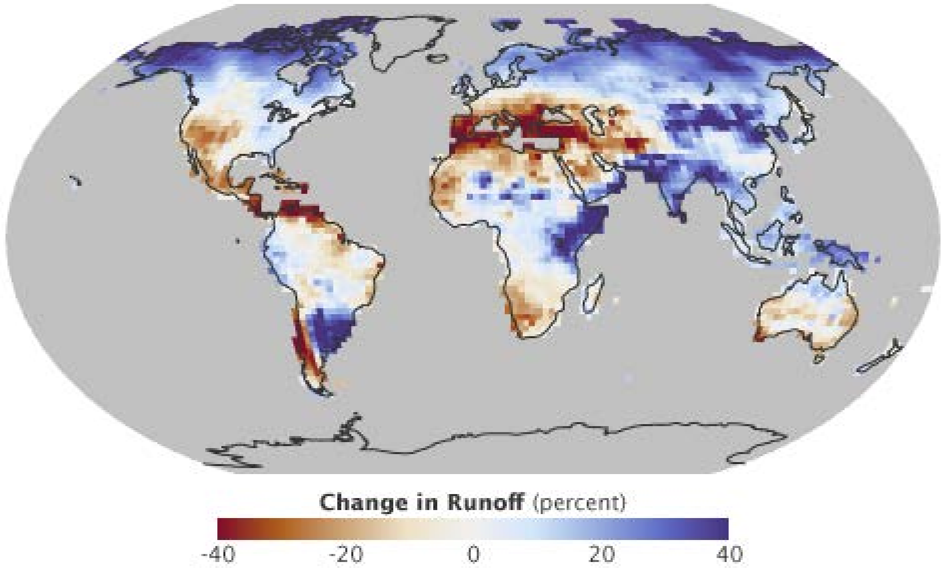

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Figure 2. Changes in water runoff into rivers and streams are another anticipated consequence of climate change by the late 21st Century. This map shows predicted increases in runoff in blue, and decreases in brown and red. (Map by Robert Simmon, using data from Milly et al., 2005; Graham et al., 2010; NASA Geophysical Fluid Dynamics Laboratory.)

3. Chemical composition of seawater

3.1 Salinity

Surface salinity integrates the signals of freshwater sources and sinks for the ocean, and if long-term (decadal to centennial) changes in salinity are considered, this provides a way to investigate associated changes in the hydrological cycle. Many studies have assessed changes to ocean salinity over the long term; of these, four have considered changes on a global scale from the near-surface to the sub-surface ocean (Boyer et al., 2005; Hosoda et al., 2009; Durack and Wijffels, 2010; Good et al., 2013). These studies independently concluded that alongside broad-scale ocean warming associated with climate change, shifts in ocean salinities have also occurred. These shifts, which are calculated using methods such as objective analysis from the sparse historical observing system, suggest that at the surface, high-salinity subtropical ocean regions and the entire Atlantic basin have become more saline, and low-salinity regions, such as the western Pacific Warm Pool, and high-latitude regions have become even fresher over the period of analysis (Figure 3). Significant regional-scale differences may be ascribed to the paucity of observational data, particularly in the pre-Argo era, the difference in temporal period over which each analysis was conducted, and differences in methodology and data selection criteria.

Despite regional differences, the broad-scale patterns of change suggest that long-term, coherent changes in salinity have occurred over the observed record, and this conclusion is also supported by shifts in salinity apparent in the subsurface ocean (Figure 4). These subsurface changes also show that spatial gradients of salinity within the ocean interior have intensified, and that at depth, salinity-minimum (intermediate) waters have become fresher, and salinity-maximum waters have become saltier (Durack and Wijffels, 2010; Helm et al., 2010; Skliris et al., 2014). Taken together, this evidence suggests intensification of the global hydrological cycle; this is consistent with what is expected from global warming (see Section 1). Actual changes in the hydrological cycle may be even more intense than indicated by patterns of surface salinity anomalies, as these may be spread out and reduced in intensity by being transported (advected) by ocean currents. For example, the work of Hosoda et al. (2009) and Nagano et al. (2014) indicates that large (ENSO-scale) salinity anomalies are rapidly transported from the central Pacific to the northwestern North Pacific (the Kuroshio Extension region).

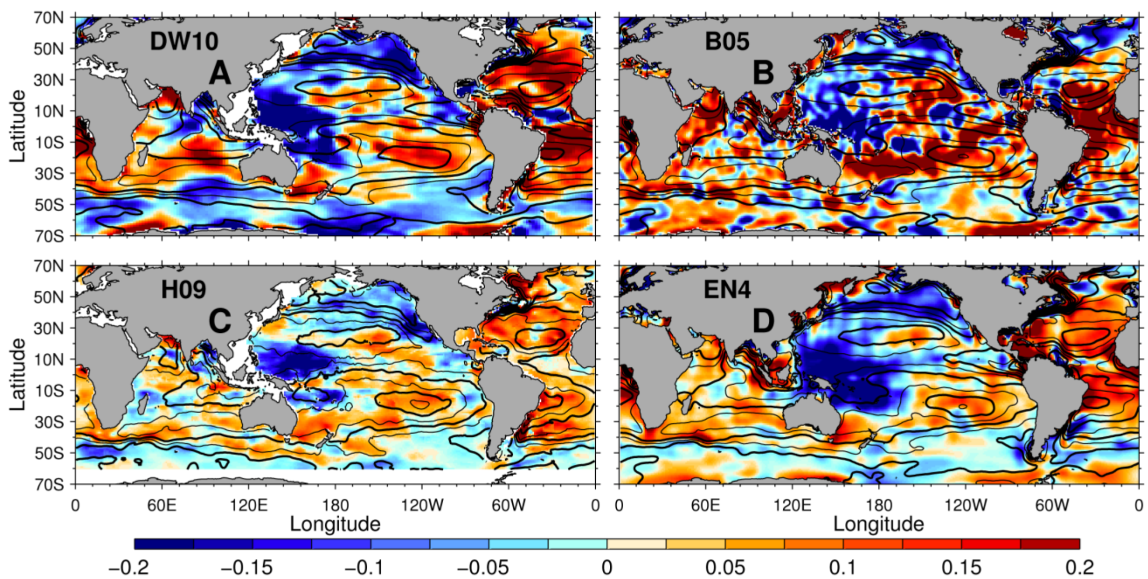

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Figure 3. Four long-term estimates of global sea-surface salinity (SSS) change according to (A) Durack and Wijffels (2010; ©American Meteorological Society. Used with permission.), analysis period 1950-2008; (B) Boyer et al. (2005), analysis period 1955-1998; (C) Hosoda et al. (2009), analysis period 1975-2005; and (D) Good et al. (2013), analysis period 1950-2012; all are scaled to represent equivalent magnitude changes over a 50-year period (PSS-78 50-year^{-1}). Black contours show the associated climatological mean SSS for the analysis period. Broad-scale similarities exist between each independent analysis of long-term change, and suggest an increase in spatial gradients of salinity has occurred over the period of analysis. However, regional-scale differences are due to differences in data sources, temporal periods of analysis, and analytical methodologies.

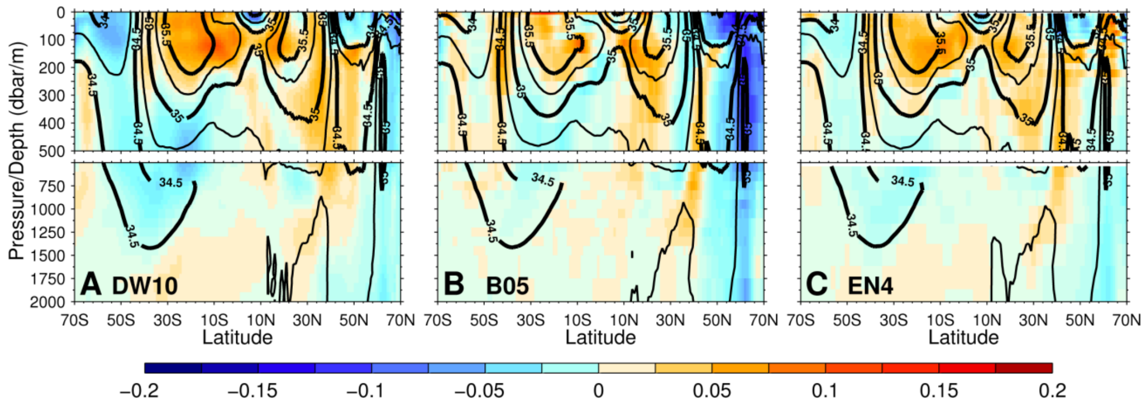

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Figure 4. Three long-term estimates of global zonal mean subsurface salinity changes according to (A) Durack and Wijffels (2010; ©American Meteorological Society. Used with permission.), analysis period 1950-2008; (B) Boyer et al. (2005), analysis period 1955-1998; and (C) Good et al. (2013),analysis period 1950-2012; all scaled to represent equivalent magnitude changes over a 50-year period (PSS-78 50-year^{-1}). Black contours show the associated climatological mean subsurface salinity for the analysis period. Broad-scale similarities also exist in the subsurface salinity changes, which suggest a decreasing salinity in ocean waters fresher than the global average, and an increasing salinity in waters saltier than the global average. However, regional differences, particularly in the high-latitude regions, are due to limited data sources, different temporal periods of analysis and different analytical methodologies.

3.2 Nutrients

Many different nutrients are required as essential chemical elements that organisms need to survive and reproduce in the ocean. Macronutrients, needed in large quantities, include calcium, carbon, nitrogen, magnesium, phosphorus, potassium, silicon and sulphur; micronutrients like iron, copper and zinc are needed in lesser quantities (Smith and Smith, 1998). Macronutrients provide the bulk energy for an organism's metabolic system to function, and micronutrients provide the necessary co-factors for metabolism to be carried out. In aquatic systems, nitrogen and phosphorus are the two nutrients that most commonly limit the maximum biomass, or growth, of algae and aquatic plants (United Nations Environment Programme (UNEP) Global Environment Monitoring System (GEMS) Water Programme, 2008). Nitrate is the most common form of nitrogen and phosphate is the most common form of phosphorus found in natural waters. On the other hand, one of arguably the most important groups of marine phytoplankton is the diatom. Recent studies, for example, Brzezinski et al., (2011), show that marine diatoms are significantly limited by iron and silicic acid.

About 40 per cent of the world's population lives within a narrow fringe of coastal land (about 7.6 per cent of the Earth's total land area; United Nations Environment Programme, 2006). Land-based activities are the dominant source of marine nutrients, especially for fixed nitrogen, and include: agricultural runoff (fertilizer), atmospheric releases from fossil-fuel combustion, and, to a lesser extent, from agricultural fertilizers, manure, sewage and industrial discharges (Group of Experts on the Scientific Aspects of Marine Environmental Protection, 2001; Figure 5).

An imbalance in the nutrient input and uptake of an aquatic ecosystem changes its structure and functions (e.g., Arrigo, 2005). Excessive nutrient input can seriously impact the productivity and biodiversity of a marine area (e.g., Tilman et al., 2001); conversely, a large reduction in natural inputs of nutrients (caused by, e.g., damming rivers) can also adversely affect the productivity of coastal waters. Nutrient enrichment between 1960-1980 in the developed regions of Europe, North America, Asia and Oceania has resulted in major changes in adjacent coastal ecosystems.

Nitrogen flow into the ocean is a good illustration of the magnitude of changes in anthropogenic nutrient inputs since the industrial revolution. These flows have increased 15-fold in North Sea watersheds, 11-fold in the North Eastern USA, 10-fold in the Yellow River basin, 5.7-fold in the Mississippi River basin, 5-fold in the Baltic Sea watersheds, 4.1-fold in the Great Lakes/St Lawrence River basin, and 3.7-fold in South-Western Europe (Millennium Ecosystem Assessment, 2005). It is expected that global nitrogen exports by rivers to the oceans will continue to rise. Projections for 2030 show an increase of 14 per cent compared to 1995. By 2030, global nitrogen exports by rivers are projected to be 49.7 Tg/yr; natural sources will contribute 57 per cent of the total, agriculture 34 per cent, and sewage 9 per cent (Bouwman et al., 2005). An example of this is discussed in Box 1.

Box 1: Example – Nutrients in the Pacific region

The Pacific Ocean basins form the largest of the mid-latitude oceans. In addition, the subarctic North Pacific Ocean is one of the most nutrient-rich areas of the world ocean; in 2013, the most recent year for which statistics have been compiled, the North Pacific (north of 40° N) provided 30% of the world's capture, by weight, of ocean fish (FAO, 2015). Many oceanographic experiments have been carried out over the last half century in the North Pacific Ocean; studies based on these datasets reveal the decadal-scale variation of nutrient concentrations in the surface and subsurface (intermediate) layers, as seen in Figure 6.

A linearly increasing trend of nutrient concentrations (nitrate and phosphate) has been observed in the intermediate waters in a broad area of the North Pacific (Figure 6b); Ono et al., 2001; Watanabe et al., 2003; 2008; Tadokoro et al., 2009; Guo et al., 2012; Whitney et al., 2013). Conversely, the concentration of nutrients in the surface layer has decreased (Figure 6a; Freeland, 1997; Ono et al., 2002; 2008; Watanabe et al., 2005; 2008; Aoyama et al., 2008, Tadokoro et al., 2009; Whitney, 2011). Surface nutrients are primarily supplied by the subsurface ocean through a process known as "vertical mixing", an exchange between surface and subsurface waters. Vertical mixing is partly dependent on the differences in density between adjacent ocean layers: layers closer to one another in density mix more easily.

A significant increase in temperature and a corresponding decrease in salinity (see above) have been observed during the last half-century in the upper layer of the North Pacific (IPCC, 2013, WG1 AR5). These changes are in the direction of increased stratification in the upper ocean and thus it is possible that this increased stratification has caused a corresponding decrease in the vertical mixing rate.

Superimposed on the linear trends, nutrient concentrations in the ocean have also exhibited decadal-scale variability, which is evident in both surface and subsurface waters (Figure 6c). Unlike the linear trends, the decadal-scale variability appeared synchronized between the surface and subsurface layers in the western North Pacific (Tadokoro et al., 2009). These relationships suggest that the mechanisms producing the trends and more cyclical variability are different.

4. Environmental, economic and social implications of changes in salinity and nutrient content

4.1 Salinity

Although changes to ocean salinity do not directly affect humanity, changes in the hydrological cycle that are recorded in the changing patterns of ocean salinity certainly do. Due to the scarcity of hydrological cycle observations over the ocean, and the uncertainties associated with these measurements, numerous studies have linked salinity changes to the global hydrological cycle by using climate models (Durack et al., 2012; 2013; Terray et al., 2012) or reanalysis products (Skliris et al., 2014). However, these studies only considered long-term salinity changes, and not changes that occur on interannual to decadal time-scales. These latter scales are strongly affected by climatic variability (Yu, 2011; Vinogradova and Ponte, 2013). As mentioned in Section 3, these studies collectively conclude that changes to the patterns of ocean salinity are likely due to the intensification of the hydrological cycle, in particular patterns of evaporation and rainfall at the ocean surface. This result concurs well with the “rich-get-richer” mechanism proposed in earlier studies, suggesting that terrestrial “dry” zones will become dryer and terrestrial “wet” zones will become wetter due to ongoing climate change (Chou and Neelin, 2004; Held and Soden, 2006).

4.2 Nutrients

Marine environments are unsteady systems, whose response to climate-induced or anthropogenic changes is difficult to predict. As a result, no published studies quantify long-term trends in ocean nutrient concentrations. However, it is well understood that imbalances in nutrient concentration cause widespread changes in the structure and functioning of ecosystems, which, in turn, have generally negative impacts on habitats, food webs and species diversity, including economically important ones; such adverse effects include: general degradation of habitats, destruction of coral reefs and sea-grass beds; alteration of marine food-webs, including damage to larval or other life stages; mass mortality of wild and/or farmed fish and shellfish, and of mammals, seabirds and other organisms.

Among the effects of nutrient inputs into the marine environment it is important to mention the link with marine pH. The production of excess algae from increased nutrients has the effect, *inter alia*, to release CO_2 from decaying organic matter deriving from eutrophication (Hutchins et al., 2009; Sunda and Cai, 2012). The effects of these acidification processes, combined with those deriving from increasing atmospheric CO_2, can reduce the time available to coastal managers to adopt approaches to avoid or minimize harmful effects on critical ecosystem services, such as fisheries and tourism. Globally, the manufacture of nitrogen fertilizers has continued to increase (Galloway et al., 2008) accompanied by increasing eutrophication of coastal waters and degradation of coastal ecosystems (Diaz and Rosenberg, 2008; Seitzinger et al., 2010; Kim et al., 2011), and amplification of CO_2 drawdown (Borges and Gypens, 2010; Provoost et al., 2010). In addition, atmospheric deposition of anthropogenic fixed nitrogen may now account for up to about 3 per cent of oceanic new production, and this nutrient source is projected to increase (Duce et al., 2008).

Figure 5 (a)

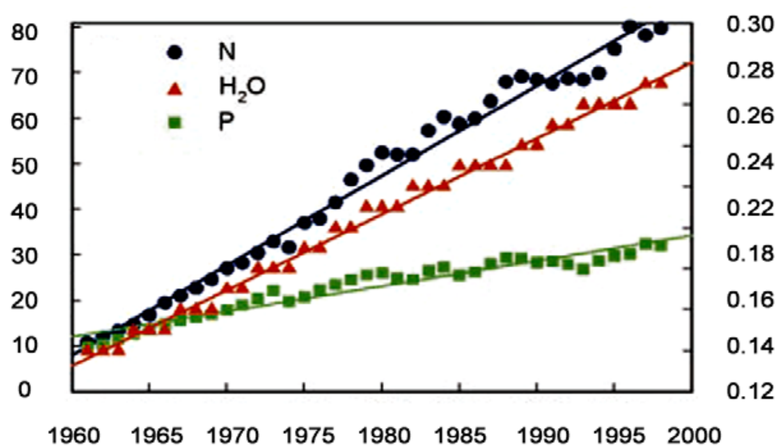


Figure 5 (b)

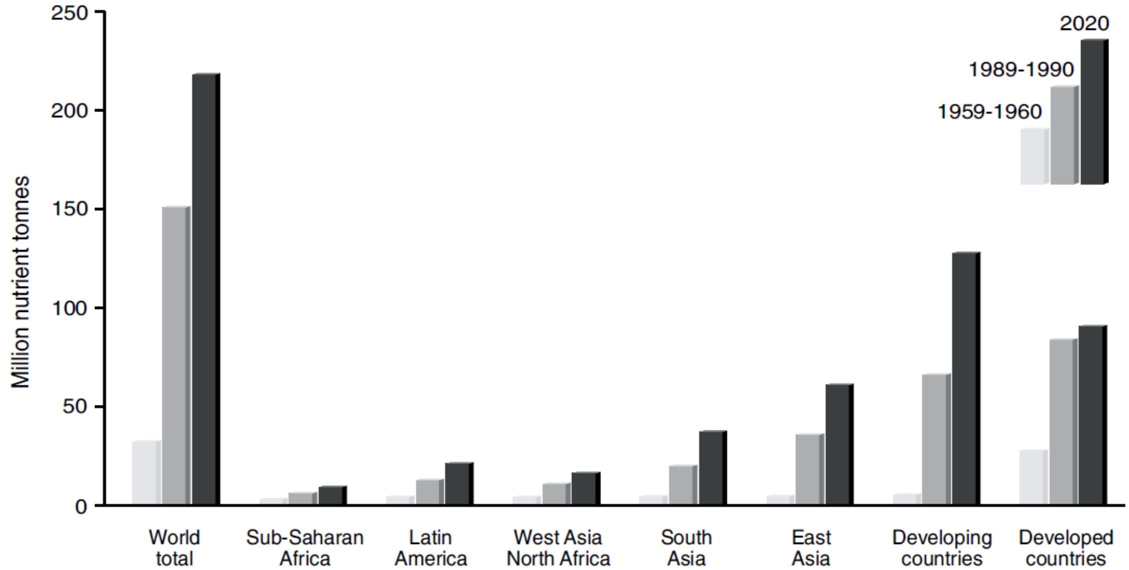
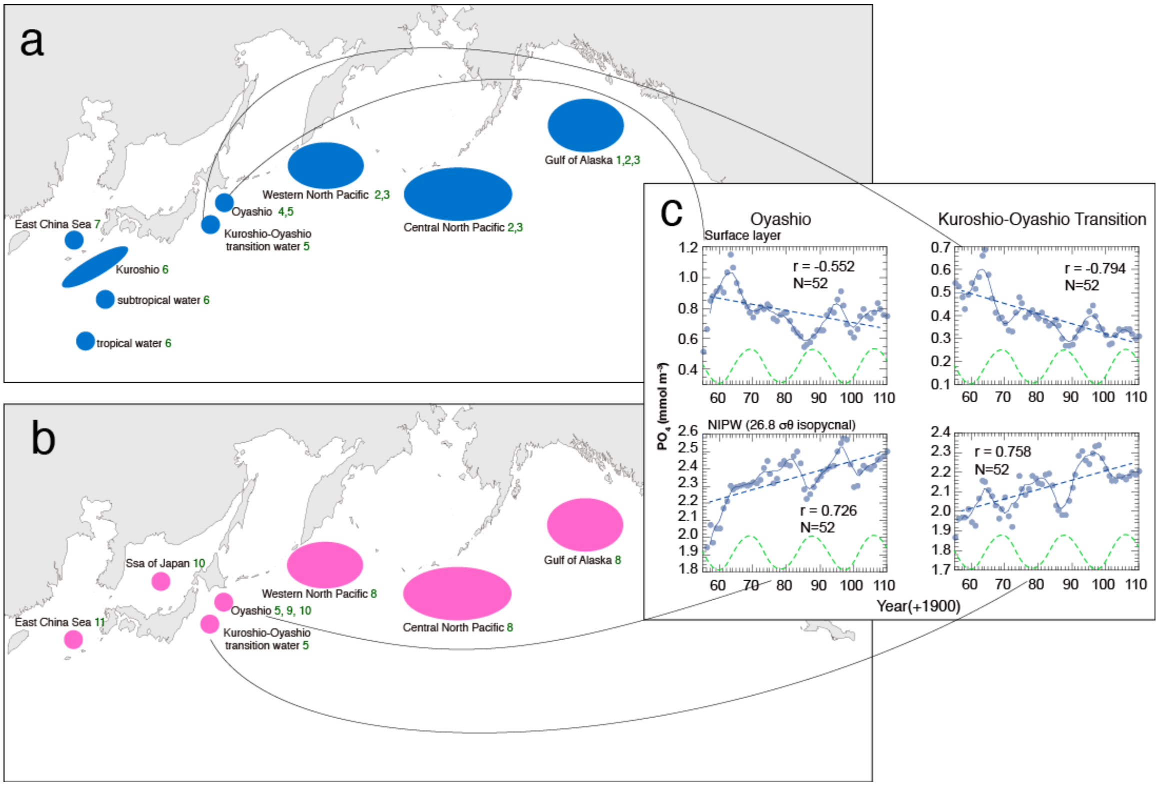


Figure 5 (a) Trends in annual rates of application of nitrogenous fertilizer (N) expressed as mass of N, and of phosphate fertilizer (P) expressed as mass of P_2O_5, for all States of the world except for many of the countries belonging to the United Nations regional group of Eastern European States and the former USSR (*scale on the left in 10^6 metric tons*), and trends in global total area of irrigated crop land (H2O) (*scale on the right in 109 hectares*). Source: Tilman et al., 2001. Figure 5 (b) Estimated growth in fertilizer use, 1960-2020. From GESAMP (2001). Source: Bumb and Baanante, 1996.



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. Synthesis of the decadal-scale change in nutrient concentrations in the North Pacific Ocean in the last fifty years. (a) The blue area shows the waters for which a decreasing trend in nutrient concentrations was reported in the surface layer. (b) The pink area shows the waters for which an increasing trend in nutrient concentrations was reported in the subsurface. (c) Example of the nutrient change in the North Pacific Ocean. Five-year running mean of the annual mean concentration (mmol m^{-3}) of Phosphate concentration in the surface and North Pacific Intermediate Water (NPIW) of the Oyashio and Kuroshio-Oyashio transition waters from the mid-1950s to early 2010. (Time series from Tadokoro et al., 2009). Blue broken lines indicate statistically significant trends of PO_4. Thin green broken lines represent the index of diurnal tidal strength represented by the sine curve of the 18.6-yr cycle.* The numbers following each area name indicate the referenced literature: (1) Freeland et al., (1997); (2) Ono et al., (2008); (3) Whitney (2011); (4) Ono et al., (2002); (5) Tadokoro et al., (2009); (6) Watanabe et al., (2005); (7) Aoyama et al., (2008); (8) Watanabe et al., (2008); (9) Ono et al., (2001); (10) Watanabe et al., (2003); (11) Guo et al., (2012).

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