14
The Oceans' Role in the Hydrological Cycle

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Opposite page: Splashing waves at Inhaca Island, Mozambique. © José Paula.

INTRODUCTION

More than 70 per cent of the earth's surface is covered by the oceans (Stewart 2008). Because of the water's high heat capacity, the oceans absorb and retain a greater amount of solar energy, far more than the land and atmosphere (Linacre and Geerts 1997). Almost half of the absorbed solar energy at the sea surface is released back to the atmosphere in form of outgoing longwave radiation (OLR) and latent heat flux. The latter produces atmospheric water vapour (Linacre and Geerts 1997). Water vapour plays a key role in the Earth's energy balance and drives important processes within the hydrological cycle (Linacre and Geerts 1997, Talley and others, 2011), upon which human existence and permanence on Earth's surface depend. In this chapter, the hydrological cycle is the process of water circulation and exchange through the hydrosphere - atmosphere - lithosphere systems.

Bengtsson (2010) indicates that the total amount of available water on Earth's surface is about 1.5×10^9 km^3. When distributed through the components of the hydrological cycle, the oceans retain about 1.4x109 km^3, inland waters, ice and glaciers about 29x106 km^3, and ground-water systems 15x106 km^3 (Bengtsson 2010). These estimates indicate that the oceans retain nearly 97 per cent of the available water.

The different freshwater systems interact with the seawater in diverse and complex ways, causing both short and long-term impacts on the compartments of the Earth's environment, with significant social, economic and ecological implications. For example, large amounts of inland sediments transported by the rivers, when deposited in the coastal ocean systems can lead to high levels of ocean turbidity that can reduce penetration of solar radiation in the water column. Deficiency of solar radiation can negatively affect photosynthetic processes, thus reducing primary production. Another example is the effect of anomalous freshening in the ocean caused by inputs of freshwater that can affect seawater stratification and water mass formation. This in turn can influence the strength of the general thermohaline oceanic circulation, potentially modifying the Earth's climate.

According to Lagerloef and others (2010), the global evaporation process is estimated to be about 13 Sv (1 Sv = 10^6 m^3 s^{-1}), precipitation about 12.2 Sv, river discharge about 1.2 Sv and melting glaciers only about 0.01 Sv. As these estimates suggest, among such processes, evaporation and precipitation greatly dominate the hydrological cycle. However, each oceanic basin exhibits its particular influence on the hydrological cycle, associated with morphologic and hydrographic characteristics. For the purpose of this review, in this chapter only the role of the western Indian Ocean on the hydrological cycle is assessed.

INTERACTIONS BETWEEN SEAWATER AND FRESHWATER SEGMENTS OF THE HYDROLOGICAL CYCLE

The Indian Ocean is the smallest ocean after the Pacific and Atlantic, exhibiting distinct characteristics, which affect its hydrology. Among many, the most important characteristics are: (1) the shape of the Indian Ocean, ending in the northern subtropics, (2) seasonally reversing monsoons, and (3) the influence of equatorial current system in blocking the spread of thermocline waters (Schott and others, 2009, Talley and others, 2011). The shape of the Indian Ocean with the subtropical northern limit prevents a northward heat export, permitting only a weak ventilation of the thermocline from the north (Schott and others, 2009). The seasonal reversal of the monsoonal winds cause reversal of the ocean currents such as the Somali Current in the western Indian Ocean, and the southwest and northeast Monsoon currents south of India and Sri Lanka (Schott and others, 2009). The prevention of spread of thermocline waters by the equatorial current systems results in the coastal portion of the western Indian Ocean being the only gateway for advective transfer of thermocline water between the southern and northern Indian Ocean sectors (Talley and others, 2011).

Rivers runoff

The total volumes involved in the interaction between the Indian Ocean and the surrounding continental freshwater segments is small when compared with the Atlantic and Pacific Oceans. However, estimates of the net transport of Indian Ocean waters moving to inland freshwater systems is about one-third of the volume, compared to the Atlantic Ocean where an estimated two-thirds of the volume moves inland (Bengtsson 2010).

Dai and Trenberth (2002) compiled information on the fifty-one largest rivers flowing into the world's ocean. For each river, the information consisted of observed and modelled annual mean river discharge both at the river mouth and at a given station, and its respective drainage area. Many rivers flowing into the Indian Ocean also have a strong impact on the hydrology of the region. The ten largest rivers flow, as estimated in the river's mouth, are presented in Table 14.1. These include: Brahmaputra River in Bangladesh, the Ganges River in India and the Irrawady River in Burma, all of which flow into the Gulf of Bengal. Into the Arabian Sea flows the Indus River in Pakistan. Into the Somali Basin flow the Juba River in Somalia and the Rufiji River in Tanzania; while into the Comoros Basin flows the Tsiribihina River from the northwest of Madagascar. Finally, in the eastern side of the Mozambique Channel flows the Mangoky River from central Madagascar, while into the western side flows the Limpopo River at Chokwe and the Zambezi River at the Sofala Bank (Dai and Trenberth 2002).

The Zambezi River is the fourth largest in Africa, and the largest in east Africa. It transports more than 75 per cent of the annual mean runoff of the region's interior and discharges more than 40 per cent (Mukosa and Mwiinga, 2008) through its delta on the Mozambique Channel. These discharges greatly impact the coastal water properties on the continental shelf, significantly enhancing the productivity of the Sofala Bank (Hoguane 2007).

The regime of each of the above-mentioned rivers is strongly modulated by rainfall. Studies on the effect of climate change on water resources indicate that in tropical regions rivers runoff regime and water resources depend entirely on changes in annual precipitation and its distribution during a year. Global warming is expected to result in more changes in extreme minimum and maximum discharges of rivers. Some studies have suggested, that the hydrology of the northern and Western Indian Ocean are

Table 14.1. Main rivers runoff in the north and western Indian Ocean, showing annual mean flow (km^3) and drainage area (km^2), based on model results from Dai and Trenberth (2002).

River Name	Annual Flow (km^3)	Drainage Area (10^3 km^2)
Brahmaputra	628	583
Ganges	404	956
Irrawady	393	406
Indus	104	1143
Juba	7,5	234
Rufiji	40	187
Tsiribihina	31	49
Mangoky	19	60
Limpopo	14	420
Zambezi	117	1989

likely to be very sensitive to such global warming scenarios (eg Shiklomanov 1998).

Precipitation and evaporation rates

Lagerloef and others (2010) computed and mapped global annual average of the dominant processes of the water cycle: evaporation, precipitation, and the difference between evaporation and precipitation (P-E), for the period spanning from 2050 to 2008 (see Figure 14.1). While it is well known that all subtropical oceanic gyres and subtropical margin seas are dominated by excess evaporation, a remarkable feature to observe is a maximum evaporative imbalance in the western Indian Ocean of about 2.2 m per year occurring within the Red Sea, outside the strict WIO region (Lagerloef and others, 2010).

In the northwest Indian Ocean, two contrasting oceanographic regimes dominate, to the west and east of India (see Figure 14.1). The western region, between India and Africa is defined by the Red Sea and Arabian Sea, which are evaporative basins (Beal and others, 2006). As a result of high evaporation rates they have very saline waters, eg Red Sea waters, with observed highest salinity range between 38 and 42 PSU (Talley and others, 2011).

The rainfall distribution is characterized mainly by a large annual mean difference between the western and eastern Indian Ocean. Such a pattern is completely different to what is usually known to occur in subtropical conditions. The precipitation in the former region was estimated at 10 cm per year occurring in the Arabian Sea, and the latter 300 cm per year occurring near Sumatra and Andaman Sea (Tomczak and Godfrey 1994).

In the south Indian Ocean, the rainfall distribution is characterized by small variations of about 200 cm per year in the southwest, near Madagascar, and 50 cm per year in the southeast, near Australia. As a result, the isolines of precipitation are nearly zonal, and the P-E follows closely the rainfall distribution (Tomczak and Godfrey 1994, Schott and others, 2009).

As shown by Talley (2008), the Indian Ocean is a net evaporative basin. This implies that it must import freshwater via ocean circulation (Talley 2008). In order to balance the system with a required volume of about 0.37 ± 0.10 Sv, the Indian Ocean receives tributaries from the Pacific Ocean through the Indonesian Through-Flow (ITF) of about 0.23 ± 0.05 Sv, from the Southern Ocean through a shallow gyre circulation, 0.18 ± 0.02 Sv, and a small southward export of about 0.04 ± 0.03 Sv due to freshening of bottom waters while upwelling into deep and intermediate layers (Talley 2008).

Chemical composition of seawater

Continental freshwater runoff and estuarine systems are linked to regional and large-scale hydrology through interactions among soil, water and evaporation. These systems and the whole water cycle control the movement of major nutrients and trace elements over long distances from the

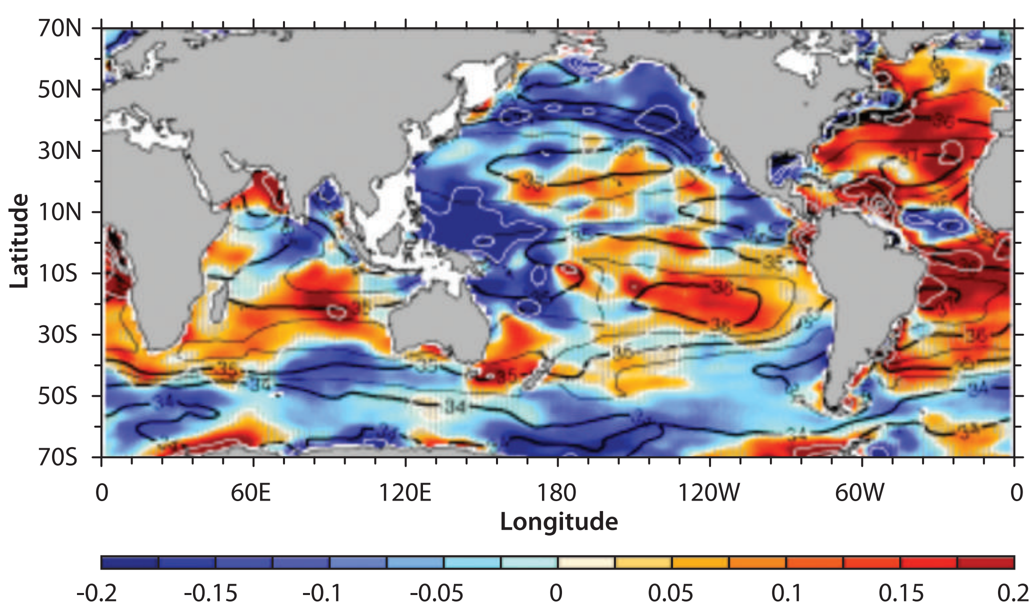

Figure 14.1. Long-term estimate of global sea surface salinity change for period covering 1950 – 2008. Notice the contrast between the northwest and northeast Indian Ocean. (Figure source: Durack and Wijffels, 2010; © American Meteorological Society, used with permission).

land to the oceans. Major nutrients in the ocean are nitrogen, phosphorus and silicon.

The physical and chemical properties of freshwater from the rivers are considerably different from the saline seawater in the oceans. A list of major constituents of the seawater is given by Stewart (2008). It involves the following ions: chlorine (by 55.3 per cent), sodium (30.8 per cent), sulphate (7.7 per cent), magnesium (3.7 per cent), calcium (1.2 per cent) and potassium (1.1 per cent). On average, the Indian Ocean surface water salinity ranges from 32 to 37 PSU, being highest in the Arabian Sea portion.

In-situ observations have shown high salinity content in the Red Sea, Arabian Sea (35.5–36.8 PSU) and Persian Gulf (34.8–35.4 PSU). Relatively low salinity waters from the ITF (34.4–35.0 PSU) are transported into the western Indian Ocean by the South Equatorial Current that feeds the northward East African Coastal Current passing along Tanzanian coast. Low salinity waters from rain-dominated areas such as the Bay of Bengal (28.0–35.0 PSU) can be advected toward the Western Indian Ocean region, where they eventually mix with waters from Indonesian origin. Studies have shown that in the Western Indian Ocean, the salinity maximum linked to Red Sea water can be traced all the way to the coast of South Africa. The salinity core decreases from 38 PSU in the Red Sea to about 34.7 PSU in the Mozambique Channel (Emery 2003).

Among four main studies addressing the observed long-term changes to ocean salinity, the study by Durack and Wijffels (2010) appears to provide a more robust measure of absolute temporal trends (Siedler and others, 2013). Durack and Wijffels (2010) suggest that in general, areas dominated by precipitation have undergone freshening, while evaporation-dominated areas have become saltier (Figure 14.1). Strong amplification of salinity contrast is indicative of an intensification of the hydrological cycle in response to a warmer climate (Held and Soden 2006).

ENVIRONMENTAL, ECONOMIC AND SOCIAL IMPLICATIONS OF OCEAN WARMING AND SEA LEVEL CHANGE

Ocean warming

The tropical Western Indian Ocean encompasses the largest warm sea surface temperature (SST) pool of the world's oceans, hence it has the potential to influence both regional and global climate (Schott and others, 2009, Talley and others, 2011). The 1997-1998 El Niño Southern Oscillation (ENSO), linking anomalous warm SST accompanied by extreme rainfall and drought, with strong social and economic implications is worthy of mention. ENSO is an anomalous warming of the ocean and atmosphere in the tropical Pacific that occurs roughly every three to seven years and lasts for 12 to 18 months. It is the dominant mode of natural climate variability.

The anomalous warming in the tropical eastern Pacific reduces the atmospheric pressure causing weakening of the trade winds. This sets the stage for an evolution of an equatorial rain band stretching from Indonesia to the central Pacific (Siedler and others, 2013). The surface oscillations in the central Pacific induces an eastward propagation of Kelvin waves that on reaching the eastern boundary of the equatorial Pacific trigger a chain of westward propagating Rossby waves. Studies have shown that these waves cross the entire Pacific basin, entering into the tropical Indian Ocean, with far reaching impact in the western region.

The ENSO causes significant climatic disturbances in most parts of Africa, either inducing drought or flooding, or increasing sea temperatures leading to cyclones. The 1997-1998 ENSO devastated both northeastern and western Indian Ocean countries, with unusual torrential convective rains that flooded Somalia, Ethiopia, Kenya, Sudan and Uganda, and severe droughts in Papua New Guinea and Indonesia (FAO 2001). These had severe social consequences: extensive crop failures and livestock losses, food and drink water shortage (McPhaden 1999). Over a thousand people died and hundreds of thousands misplaced (Schott and others, 2009).

It is important to mention that while ENSO induced warming covers the whole tropical Indian Ocean, the cause for the warming in the southwest Indian Ocean is different from that in the rest of the basin (Schott and others, 2009). This indicates that the warming in this oceanic sector cannot be explained only by surface fluxes. The ocean interior dynamics are crucial. The role of the thermocline ridge between 5-10°S has been shown to play an important role on the variability of SST (Xie and others, 2002).

Washington and Preston (2006) have investigated the role of the Indian Ocean SST on extreme wet occurrences over southern Africa during the 19th century. Their results, based on the two most extreme rainfall events in 1974 and 1976, indicate that while ENSO serves as an important control of rainfall variability, other processes that enhance warm SST anomalies in the subtropics and cool anomalies in the northern southwest Indian Ocean are the main drivers of extreme rainfall events. One such critical process was found to be associated with cold SST anomalies in the Mascarene region, which influenced anomalous low-level easterly moisture flux along 10-20ºS to the east of southern Africa. This had a direct impact on moist convective uplift leading to enhanced precipitation. Economic and social implications of changes on the hydrological water cycle tend to be large.

According to official retrospectives relating to the year 2000 floods in Mozambique, the costs were estimated at US$ 273 million in physical damage, US$ 247 million in lost production, US$ 48 million in lost exports and US$ 31 million in increased imports. During mid to late January 2001, heavy rains over Zambezia Province caused the Licungo River to flood in Mozambique, with nearly 500 000 people affected by the floods (UNEP 2002), 650 people killed and more than half a million left homeless (UNEP 2002).

Increased warm SST leads to increased cyclone activity. UNEP (2002) reports that "the Western Indian Ocean islands typically experience ten cyclones a year, between November and May, which bring strong winds and heavy rainfall. This causes destruction of infrastructure, particularly in low-lying areas and where settlements have encroached into flood-prone areas". This natural phenomenon also has strong social impacts. For example, cyclones Eline (mid-February) and Gloria (early-March) in the year 2000 in Madagascar left 184 000 people in need of immediate relief support of the total 737 000 affected (UNEP 2002).

Sea level change

Warming SST affects the low atmospheric pressure systems with implication throughout the oceanic water column. Sea level oscillations reflect the vertical movements of the thermocline. Investigating sea level rise around Pacific and Indian Ocean islands, Church and others (2006) have shown large sea surface height (SSH) variability being caused by ENSO. They also suggested that fewer individual tide-gauge records contribute to uncertainty of historical rates of sea level rise. For the period from 1993 to 2001, both satellite and tide-gauge data show large rates of sea level fall in the eastern Pacific and western Indian Ocean (approaching – 10 mm per year), which reflects weak ENSO conditions. Over the region 40°S to 40°N, 30°E to 120°W, the average rise was estimated at about 4 mm per year (Church and others, 2006). Their analysis suggests a continued and increasing rate of sea level rise, causing serious problems for inhabitants of some of these islands during the 21st century (Church and others, 2006). According to recent projections (IPCC 2013), it is very likely that in this century and beyond, sea level change will show strong regional patterns (Figure 14.2), with some places experiencing significant deviations from the global mean change (Church and others, 2013).

Natural variabilities appear to be the main drivers of the regional patterns of the observed sea level change (Siedler and others, 2013). In the Indian Ocean, minimal sea level rise pattern has been observed in the south tropical band (Figure 14.2). While Han and others (2010) attributed such a pattern to a strengthening of the Indian Ocean Walker and Hadley cells, Schwarzkopf and Boning (2011) attributed such a pattern to the change of winds in the western equatorial Pacific that transmit thermocline anomalies through the Indonesian Archipelago, and their subsequent westward propagation by baroclinic Rossby waves (Church and others, 2011).

Information compiled in the IPCC (2007) report in connection with the sea level changes in east Africa, indicated that western Indian Ocean countries, from Sudan in the Red Sea to South Africa, including coastal wetlands and some near-shore islands off the coast of Tanzania and Mozambique, as well as oceanic islands of Madagascar, Seychelles, Comoros, Mauritius, and Reunion, are highly vulnerable. In fact, simulations of sea level rise in Tanzania have shown that for a rise of 0.5 and 1 m per century, about 2 090 and 2 117 km^2 of land would be inundated, respectively, while the latter would erode an additional area of about 9 km^2 (Mwaipopo 2000). Looking at the economic implications of such rising sea level, the projected damage estimated as exceeding US$ 544 million for a 0.5 m rise and US$ 86 000 million for 1 m rise.

A recent study focused on Africa by Brown and others (2011), for period from 2000 to 2100, based on an integrated biophysical and socio-economic impact model, has shown that for the Western Indian Ocean, Mozambique and Tanzania stand out as countries with high people-based impacts associated with flooding and forced migration. For South Africa higher economic damage from sea level rise was foreseen. Mozambique and South Africa were identified as countries in the region with the highest adaption costs (Brown and others, 2011). Nevertheless, more studies in the region are needed, to improve such estimates and examine other aspects of impact and adaptation in more detail.

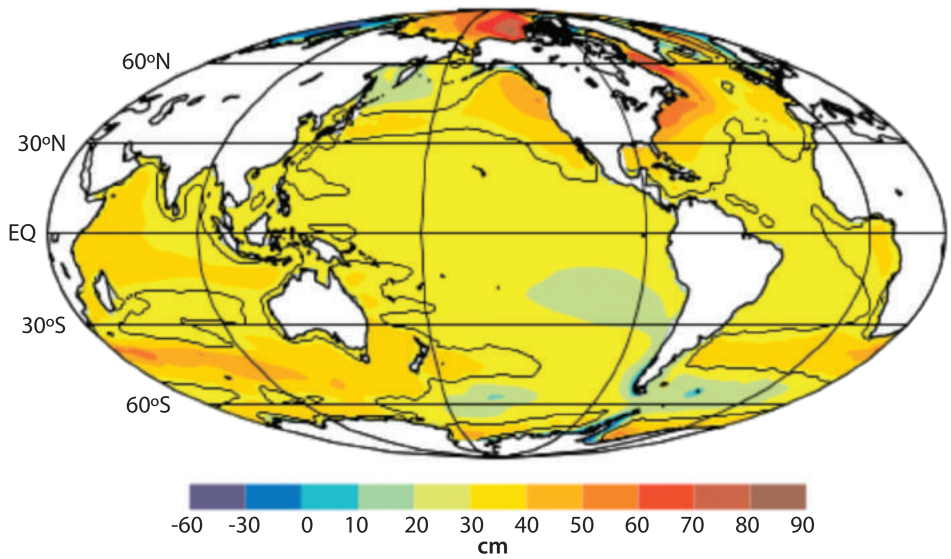

Figure 14.2. The regional distribution of the projections of sea level change for 2090 compared to 1990, combining global average sea level projections, dynamic ocean departure from the global average, and regional changes associated with the changing mass distribution in the cryosphere. The black contour is the "average" value at 2090 of 38 cm, dividing those regions with above and below-average sea level rise. (Figure source and caption taken from Church and others, 2011).

For Kenya, Awuor and others (2008) indicate that the impacts of sea level rise are likely to be felt beyond coastal and national boundaries. Activities such as infrastructure, tourism, aquaculture and agriculture are likely to be negatively affected due to rising sea level. For instance, with 0.3 m rise and without adaptation, an estimated 17 per cent of the Mombasa District will be submerged (Oyieke 2000). This rise would also affect 7 per cent of the population in the Tana Delta, and an area of about 481 km^2 could be lost between 2000 and 2050 (Ericson and others, 2006).

For Mauritius, the coastal zone is degrading at an accelerated rate due to sea level rise. The review by Brown and others (2011) mentions that with 1 m of rise, about 26 km of beach would disappear on the west coast, including flooding of local housing, tourism and infrastructure facilities. Inundation will also affect plantations and major coastal roads.

For the Seychelles islands, the impact of sea level rise will have also strong negative impacts on fishing and tourism, which are main sectors of socio-economic importance (Brown and others, 2011). The ports and airports which have been built on reclaimed low ground would also be severely affected (Brown and others, 2011). Furthermore the increasing sea level rise will lead to greater erosion which potentially will lead to an increased landslides, and beaches would be inundated, resulting in a severe damage to freshwater aquifers systems.

Anthropogenic and other changes on freshwater fluxes into the sea

Both natural and anthropogenic activities affect climate variability, with anthropogenic activities affecting the pace of their occurrence and severity. The African continent has been regarded as a region where human activities, associated with the use of freshwater systems (for construction of reservoirs, canals for irrigation and navigation and other uses), may cause severe changes to the hydrological system and river flow regimes (Shiklomanov 1998).

Indicators of climate change, such as unprecedented ocean warming, sea level rise, and changes on freshwater fluxes into the sea, will occur more frequently in future, and the largest number of people affected will be those in low-lying deltas in Asia and Africa (IPCC 2007, IPCC 2013). As an example, the UNEP (2002) report states that in Kenya, low reservoir levels resulting from drought and siltation have been linked to human deforestation and inappropriate management of land and water resources. This has led to reductions in hydropower generation, necessitating water and power rationing with severe social and economic consequences, which devastated the country's economy in 1999 and 2000. Losses from power rationing alone were estimated at US$ 2 million per day, and the cost of unmet electricity demand was estimated at US$ 400-630 million, equal to 3.8-6.5 per cent of GDP (UNEP 2002).

THE OCEANS' ROLE IN HEAT TRANSPORTATION

On the basis of a study of Mayotte Island corals (recent and fossil) over a long time series (1881 - 1994), Zinke and others (2008) showed evidence for hydrological changes that have taken place in the Western Indian Ocean. The balance P-E in the region has natural variations on timescales of 5 to 6 years and 18 to 25 years. The temporal changes on the P-E balance as well as on OLR in the region were anti-correlated with the SST, thus higher rainfall events induced cooler mean SST.

Recent analysis on the interannual variability of SST and circulation in the tropical Western Indian Ocean for the period 1980 to 2007 by Manyilizu and others (2014), in the oceanic environment around the Tanzanian coast, suggests that changes in the thermocline (which induces SST variability) are linked with anomalous short wave radiation, plus heat anomalies from the oceanic currents, namely the South Equatorial Current. In the offshore environment, the changes in the thermocline appeared to be linked to local Ekman pumping from the wind stress curl and Indian Ocean Dipole (IOD) and ENSO (Manyilizu and others, 2014). The IOD is a basin-wide mode of variability in the tropical Indian Ocean involving a strong ocean-atmosphere interaction. In its positive phase it is characterized by an anomalous cold temperature to the west of Sumatra, and warm temperatures in the tropical Western Indian Ocean.

The Indian Ocean gains heat to the north of 20°S and imports warm ITF water into the upper ocean near 12°S. For this ocean to find its equilibrium, a net of about 0.8 PW of this heat is transported southward, with less than half 0.3 PW, being mostly in the upper Indian Ocean subtropical gyre, and half 0.4 PW by the overturning cell from deep to intermediate (or thermocline), and 0.1 by the ITF (Talley 2008). As stated by Talley (2008), the deep upwelling is significant for the global heat budget, but insignificant for the freshwater balance. Important for the freshwater balance of the Indian Ocean is the upper branch of the subtropical gyre and the ITF that carries most of the northward transport of freshwater required balancing the net evaporation (Talley 2008).

In the northern Indian Ocean, the net heat gain is observed in the region of the warm pool tongue. This requires a balance, which is made through the southward export of heat across the equator and out of the tropics. Upwelling of deep to thermocline depth waters is thought to be responsible for a significant part of the heat export (Talley 2008). The coral study from Mayotte Island by Zinke and others (2008), referred to above, also concluded that warm ENSO events are associated with a negative freshwater balance in the southwest Indian Ocean. Therefore, the region south of 10°S exhibits an opposite ENSO response compared to the equatorial Indian Ocean region (Zinke and others, 2008).

ENVIRONMENTAL, ECONOMIC AND SOCIAL IMPLICATIONS OF CHANGES IN SALINITY AND NUTRIENTS

There is evidence for water vapour volumes increasing faster than precipitation, which will have far-reaching consequences (Held and Soden 2006). The residence time of water vapour in the atmosphere is also increasing, which implies that the exchange of mass between the boundary layer must decrease. This could result in fewer tropical cyclones in a warmer climate. In regions of net evaporation (eg Western Indian Ocean) salinity is increasing (Figure 14.1), leading to increased ocean vertical mixing. In the Intertropical Convergence Zone (ITCZ) and extra-tropical storm tracks, ocean salinity is decreasing due to high precipitation thus reducing ocean vertical mixing. The interaction between river discharges in the western Indian Ocean during the summer monsoon season is known to influence the surface salinity distribution over thousand of kilometres offshore.

Results from the Argo float observations from 2004 to 2013 indicate that seawater is becoming increasingly salty in the Western Indian Ocean and near the equator in the western and central Pacific Ocean. Such a change is indicative of changes in the hydrological cycle.

Amplification of strong contrast of surface salinity between the evaporative Arabian Sea and precipitation and runoff-dominated basin in the Bay of Bengal has been observed in a fifty-year linear trend data record from 1950 to 2008 (Durack and Wijffels 2010) (see Figure 14.1). The study suggests that in the Arabian Sea, a maximum salinity increase is found along the northeast with a magnitude of about 0.52 (+/- 0.26) at 13°N and 74°E (Figure 14.2). In contrast, the Bay of Bengal shows a strong surface freshening signal of -0.40 (+/-0.31) found in the northeastern regions of the Bay (19°N, 92°E). This freshening is constrained to the shallowest waters of the Bay (Figure 14.1) and extends southward and crosses the equator with maximum surface amplitude south of India. In general, the global averaged surface salinity change is small, however at basin scale such change is large for Atlantic and Pacific Oceans, while near neutral for the Indian Ocean (-0.001+/- 0.061). Strong amplification of salinity contrast is indicative of an intensification of the hydrological cycle in response to a warmer climate (Held and Soden 2006).

Álvarez and others (2011) have investigated temporal changes in salinity and nutrients in the eastern and western subtropical gyre of the Indian Ocean, along a repeated hydrographic zonal transect that stretches from Australia to Africa, along 32ºS, for a period ranging from 1987 to 2002. With the aid of a numerical model running backwards, a simulation was made for the period from 1987 to 1960. During this period, salinity has shown an increasing trend, contrary to the period of 1987 – 2002. In the western Indian Ocean a decreasing trend in the transit time distribution indicates a faster delivery of Sub Antarctic Mode Water, thus less biogenic remineralisation. This explains the observed oxygen increase and nutrient decrease.

ENVIRONMENTAL, ECONOMIC AND SOCIAL IMPACTS OF CHANGES IN OCEAN TEMPERATURE

A direct link exists between changes to environmental oceanic conditions and economic and social impacts. Among many environmental parameters that could be inspected, here the influence of ocean temperature in the western Indian Ocean was considered. Major changes in ocean temperature can be associated with climate modes of variability such as El-Niño (La-Niña) phenomena for warmer (cooler) SST.

At great length, the impact of the sea level changes in the western Indian Ocean has been presented in this chapter. One must bear in mind that one important parameter associated with such changes is temperature, especially at the sea surface. While the general picture of the Indian Ocean's warm pool in the eastern and central portions is known to be increasing at a moderate rate in the last half-century, a more recent study based on both observational data and outputs from an ocean-atmosphere couple model by Roxy and others (2014), suggests a more drastic warming. Their study revealed a warming rate greater than was thought to be the case for the tropical western Indian Ocean, and one that has been developing for over a century (Figure 14.3). The rate of this warming is now classified as the fastest of any tropical global ocean regions, and is accountable for the largest contribution of the overall trend in the global mean sea surface temperature (Roxy and others, 2014), hence having a direct contribution to greenhouse warming.

For the period considered by Roxy and others (2014) from 1901 to 2012, it revealed an increasing rate of 0.7 ºC in the warm pool region, while the tropical western Indian Ocean has witnessed an anomalous sea surface temperature increase of 1.2 ºC during the summer season (Figure 14.3). Such significant changes on zonal SST gradients has the potential to change the Asian monsoonal circulation and rainfall patterns, as well as the productivity of the marine ecosystem in the western Indian Ocean (Roxy and others, 2014). Another important result from this study suggests that this long-term warming trend over the western Indian Ocean during summer is strongly dependent on the asymmetry in the ENSO teleconnections and positive SST skewness associated with ENSO during recent decades. Simply put, it is inferred that while major warming events such as ENSO induces a significant warming in the western Indian Ocean, a major cooling event such as La-Nina on the other hand fails to reverse such change accordingly (Roxy and others, 2014).

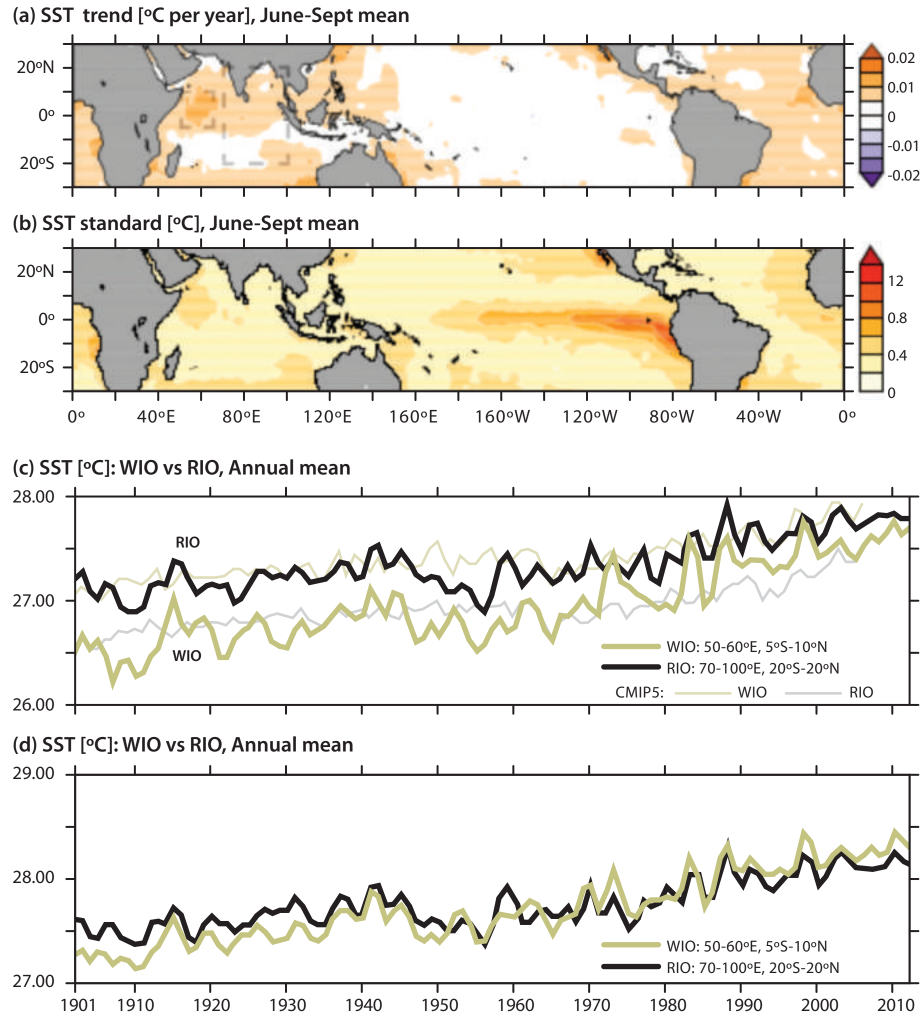

Figure 14.3. **(a)** Observed trend in mean summer (June to September) sea surface temperature (SST), units in °C year^{-1} over the global tropics from 1901 to 2012; **(b)** Interannual standard deviation of SST in °C for the same domain and time period; **(c)** Time series of mean summer SST in °C; **(d)** annual mean SST in °C over the western Indian Ocean (WIO [red: 5°S-10°N, 50-65°E]), with remaining Indian Ocean (RIO [black: 20°S-20°N, 70-100°E]). WIO and RIO are marked with inserted boxes in panel **(a)**. The CMIP5 ensemble means based on 25 climate models averaged over the WIO (light red) and RIO (light grey) are shown in panel **(c)**. (Figure source: Roxy and others, 2014; © American Meteorological Society, used with permission).

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