

Chapter 1

Introduction

Many characteristics of climate change have been identified and attributed to global warming. Changing rainfall patterns and an increase in the intensity and frequency of extreme climate events i.e. heat waves and cyclones, are just two examples of how weather patterns are thought to respond to anthropogenic warming. While some evidence exists for such changes, it is not enough, however, to compile conclusions solely based on data that does not extend past the last century (generated by satellites, weather stations and ships), and therefore a widespread need for spatiotemporally resolved geological records is required. Moreover, it is essential to recognise how climate has evolved naturally throughout different eras in order to comprehend the extent to which modern day climate is changing and at what rate. Substantial uncertainties still exist in determining the natural feedback mechanisms and teleconnections within the earth's climate system, which have been continuously evolving to reach equilibrium since the earth's genesis.

Global Warming and Climate Change

'Global warming' is the most important environmental threat our world currently faces (Forster et al., 2007). The last century has seen a rapidly growing global population and more intensive exploitation of resources, leading to greatly increased emissions of gases, such as carbon dioxide and methane. Such gases are commonly referred to as greenhouse gases, and result from the burning of fossil fuels (oil, gas and coal), agriculture, cement production and deforestation (Houghton, 2005). Greenhouse gases absorb infrared radiation emitted by the Earth's surface and act as blankets keeping it warmer than it would otherwise be. Most greenhouse gases are naturally occurring (CO_2 , CH_4 , N_2O), however, as human population levels have risen and the need to exploit resources has grown, the accelerated release of greenhouse gases has triggered global warming (Cox et al., 2000). Associated with this warming is climate change.

The basic science behind the 'greenhouse effect' that leads to warming is well understood (Lashof and Ahuja, 1990); yet, to comprehend how climate is changing in its response is less clear. A more detailed understanding therefore relies on numerical models of climate that integrate the basic dynamical and physical equations describing the complete climate system (Cox et al., 2000).

The Earth's climate currently appears out of balance and is warming at an unnatural rate (Forster et al., 2007; Meehl et al. 2004). Global average surface temperatures have increased on average by about 0.6°C over the period 1956–2006 (Meehl et al., 2005). Many components of the climate system, including atmospheric, land and ocean temperatures, sea ice and mountain glaciers, sea level, precipitation, and the length of seasons are now changing at rates and patterns out of range of what is thought to be the natural variation. These are best explained by the increased atmospheric concentrations of greenhouse gases and aerosols generated by human activity during the 20th century, yet the interaction with natural climate variability needs to be further addressed in order to fully comprehend the ‘change’ (Cane, 2010).

There is now widespread concern that the Earth's climate will warm further; not only because of the lingering effects of the added carbon already in the Earth's system, but also because of further carbon additions as the human population continues to grow (Forster et al., 2007; Meehl et al., 2005). In the next 50 years, even the lower limit of impending climate change (an additional global mean warming of 1°C above the last decade) is far beyond the range of climate variability experienced during the past thousand years (Meehl et al. 2005). Adaptation to a rapidly changing environment is therefore essential to sustain species survival, while mitigation to reduce its impacts seems inevitable to combat the abrupt rate of change. The extent to which global warming has shifted the natural climate state will be fundamental in determining how the human population will manage food security, natural resources and indeed, survive. This is particularly evident for regions across eastern and southern Africa, whereby a combination of increasing population size and more frequent and intense flooding and drought events have already had a severe impact. This thesis provides a stepping stone towards unravelling the Earth's complex climate system and the implications global warming will have on it. With the help of geological archives we can begin to scrape the surface of this synergy.

Climate across eastern and southern Africa is largely heterogeneous due to topographic regional differences, and strong (weak) coastal (inland) links to the Indian Ocean (Funk et al., 2008). Warming of the south-central Indian Ocean (0-15°S, 60-90°E) is thought to reduce the onshore moisture

flux towards southern Africa in austral summer and fall, facilitating drought-like conditions (Goddard and Graham, 1999; Hoerling et al., 2006; Richard et al., 2000). Eastern Africa rainfall trends are less clear as both increasing (Goddard and Graham, 1999; Hoerling et al., 2006) and decreasing (Funk et al., 2008) patterns have been reported in response to Indian Ocean warming. Madagascar rainfall appears anti-correlated with southern African rainfall on interannual to interdecadal time-scales due to the shifting tropical temperature troughs in response to large-scale changes in SST and sea level pressure (Fauchereau et al., 2008; Reason and Rouault, 2002).

Unfortunately, long-term climate trends or multidecadal oscillations are poorly understood as weather station data is scarce and satellite data are often incomplete or too short to make a proper assessment, particularly in Madagascar (Dewar and Richard, 2007; Dewar and Wallis, 1999). Nevertheless, Pacific decadal variability is known to affect Indian Ocean temperatures and African climate (Charles et al., 1997; Cole et al., 2000; Crueger et al., 2009; Damassa et al., 2006; Zinke et al., 2009), and suggested to influence Indian Ocean and continental rainfall as well (Deser et al., 2004; Zinke et al., 2008). Madagascar's geographical position makes it particularly susceptible to a high variability in temperature, precipitation and cyclone activity (Dewar and Richard, 2007; Jury et al., 1995; Kuleshov et al., 2008), all of which pose a severe threat to Madagascar's rich marine and terrestrial biodiversity.

The Indian Ocean and Madagascar

The Indian Ocean is land locked in the northern hemisphere by the Eurasian continent, preventing northward heat export and allowing only weak ventilation of the thermocline waters from the north. It is the third largest oceanic body covering approximately 20% of the Earth's surface (water volume of 292,131,000 km³), surrounded by the three continents Africa, Asia and Australasia. The Asian continent drives the strongest monsoon on Earth, and the monsoonal winds generate large seasonal variations in ocean currents, many of which display annual reversals. The switching trade winds have historically facilitated early trade, also allowing the settlement of Indonesians in Madagascar.

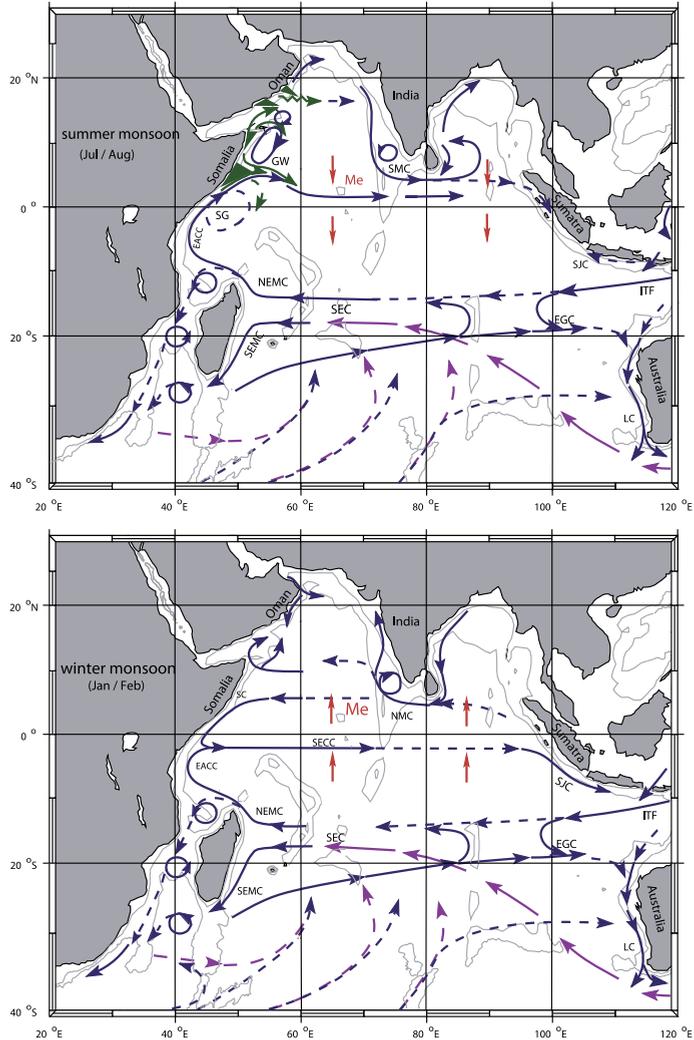


Figure 1.1. Schematic representation of current branches during the summer (southwest) monsoon (upper panel), and winter (northeast) monsoon (lower panel). Current branches indicated are the South Equatorial Current (SEC), South Equatorial Countercurrent (SECC), Northeast and Southeast Madagascar Current (NEMC and SEMC), East African Coastal Current (EACC), Somali Current (SC), Southern Gyre (SG) and Great Whirl (GW) and associated upwelling wedges (green shades), Southwest and Northeast Monsoon Currents (SMC and NMC), South Java Current (SJC), East Gyral Current (EGC), and Leeuwin Current (LC). The subsurface return flow of the supergyre is shown in magenta. Depth contours shown are for 1000 m and 3000 m (grey). Red vectors (Me) show directions of meridional Ekman transports. ITF indicates Indonesian Throughflow (from Schott et al., 2009).

The Indonesian Throughflow, which feeds the Southern Equatorial Current (SEC), provides the Indian Ocean with a low-latitude exchange with the Pacific Ocean. The SEC runs from the eastern side of the Indian Ocean westwards, to form the western limb of two subtropical gyres (Schott et al., 2009). As it hits the eastern side of Madagascar (near 17°S) it forms a clockwise current flowing north (NEMC) and an anticlockwise current flowing south (SEMC) (Schott et al., 2009; Fig. 1.1). The positioning of these gyre limbs vary depending on the season as surface currents are primarily driven by the monsoonal trade winds. The easterly trade winds, which prevail over the equatorial Pacific and Atlantic, cause shoaling of the thermocline in the eastern equatorial ocean and hence promote upwelling of cool subsurface waters. In contrast, the annual mean winds over the equatorial Indian Ocean are weak and westerly, therefore the thermocline is flat, with little or no upwelling in the eastern equatorial Indian Ocean. Precipitation patterns of the Indian Ocean are remarkably similar to those for sea surface temperatures (SST), indicative of the strong coupling between these two fields (SST and precipitation).

SW Indian Ocean Natural Climate Variability

Seasonal Variability

The Asian monsoon is a major manifestation of the seasonal cycle in the tropical-subtropical regions, driven by the pressure gradients resulting from differential heating over the Eurasian continent and the Indian Ocean, modified by the rotation of the Earth (Wang et al., 2005; Webster et al., 1998). The inherent seasonality of the Asian monsoon circulation affects the Indian Ocean, leading to strong seasonal variability in atmospheric circulation, SST and salinity (Wang et al., 2005). The Intertropical Convergence Zone (ITCZ) is described as the maximum line of solar heating, marked by a belt of high cloud cover and precipitation around the Earth that shifts back and forth across the equator in response to the sun's zenith point. During austral summer (DJF) it is located as far south as 20°S, where it is positioned over northern Madagascar (Fig. 1.2). This region of convection is essentially the rising limb of the Hadley cell and is responsible for Madagascar's heavy rains (Hu et al., 2007). The trade winds connecting the descending limb are located

approximately 30° north and south of the ITCZ. The descending limb of the Hadley cell is associated with higher air pressure and responsible for dryer conditions in the southern hemisphere during austral winter.

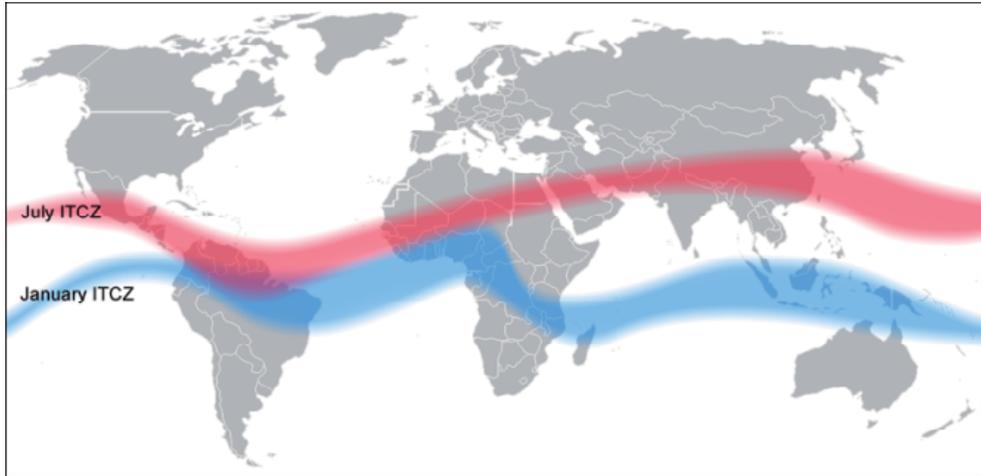


Figure 1.2. Location of the Inter Tropical Convergence Zone (ITCZ) in austral summer (blue) and austral winter (red).

North of 10°S the trade winds seasonally reverse with the monsoon. South of 10°S , the southeast trade winds, associated with Madagascar, differ as they are relatively steady in their direction. However, as the ITCZ moves north (south) during the boreal summer and fall (winter and spring) the trade winds strengthen (weaken) and their northern edge shifts northward (southward) (Schott et al., 2009). It is the strength of the trade winds feeding the ITCZ, coupled with the SW Indian Ocean SST, which determines the amount of regional precipitation Madagascar experiences. High regional SST and strong winds (strong SST-gradient with adjacent regions) drive maximum ITCZ convection, effectively drawing in air and moisture (Yasunari, 1998). In the Indian Ocean the ITCZ spawns on average ten tropical cyclones per year during austral summer through easterly waves and surges of the monsoon (Jury et al., 1994). The NE Madagascar region is vulnerable to these cyclones, which travel across the Indian Ocean in a westerly direction.

Interannual Variability

There are two prominent modes of interannual variability that impact the Indian Ocean; the El Niño Southern Oscillation (ENSO) (McPhaden et al., 2006) and the Indian Ocean Dipole (IOD) (Saji et al., 1999), which have frequencies of 2 - 7 years and 2 – 5 years, respectively. There are two other modes of variability that exist, namely the subtropical dipole (Behera and Yamagata, 2001; Reason, 2002) and the Tropical Biennial Oscillation (TBO) (Meehl and Alabaster, 2002); however, the impact of these on the Indian Ocean is much weaker than ENSO or the IOD.

The most dominant mode of natural climate variability in recent (instrumented) times is ENSO, characterised by recurrent warming (El Niño) and cooling (La Niña) of the eastern equatorial Pacific (McPhaden et al., 2006). These anomalies arise from changes in the Walker Circulation pattern. Normally, the tropical western Pacific is warm and wet under a low pressure system, and the cool and dry eastern Pacific lies under a high pressure system. This creates a pressure gradient from east to west and causes surface air to move in the same direction, from the high pressure in the eastern Pacific to the low pressure in the western Pacific. The ENSO induced SST anomalies of the eastern Pacific develop during boreal summer, peaking in boreal winter, and decaying in the following spring. There is an eastward (westward) shift in atmospheric convection over the Pacific during an El Niño (La Niña) event (see McPhaden et al., 2006). The convective currents then intensify over the central eastern (western) Pacific, resulting in slow anomalous subsidence over the western (eastern) Pacific and maritime continent (Fig. 1.3). These large-scale shifts in convection alter the atmospheric circulation remotely both in the tropics and extratropics via atmospheric wave adjustments (atmospheric bridge). These alterations have a global ‘knock on effect’ and are often referred to as global teleconnections. During a positive ENSO event (El Niño) there is a corresponding warming of the Indian Ocean, which peaks during the boreal spring (March – May). Maximum Indian Ocean SST therefore occurs approximately one season after the equatorial eastern Pacific (Klein et al., 1999; Nigam and Shen, 1993). Much of the basin wide Indian Ocean warming is caused by ENSO-induced surface changes in surface heat fluxes, particularly the wind-induced latent heat and cloud-induced solar radiation fluxes (Fig. 1.4; Klein et al., 1999).

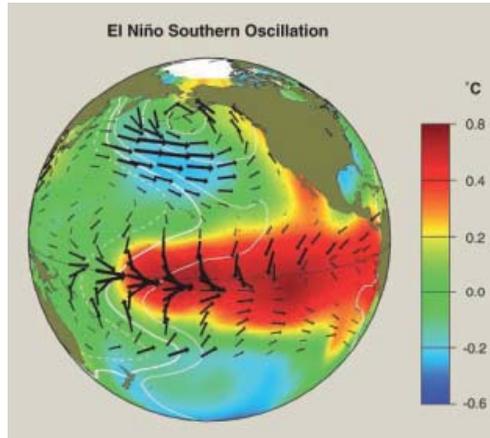


Figure 1.3. El Niño anomalies in SST (colour shading and scale in °C), surface atmospheric pressure (contours), and surface wind stress (vectors) in the Pacific basin. Pressure contour interval is 0.5 mb, with solid contours positive and dashed contours negative. Wind stress vectors indicate direction and intensity, with the longest vector equivalent to $\sim 1 \text{ N m}^{-2}$. The patterns in this graphic are derived from a linear regression against SST anomalies averaged over 6°N – 6°S , 90°W – 180° in the eastern and central equatorial Pacific. All quantities scale up or down with the intensity of anomalies in this index region, that is, higher for strong El Niños and lower for weak El Niños. Anomalies of opposite sign apply to La Niña events, although there are some differences in the spatial patterns of El Niño and La Niña. From McPhaden et al. (2006).

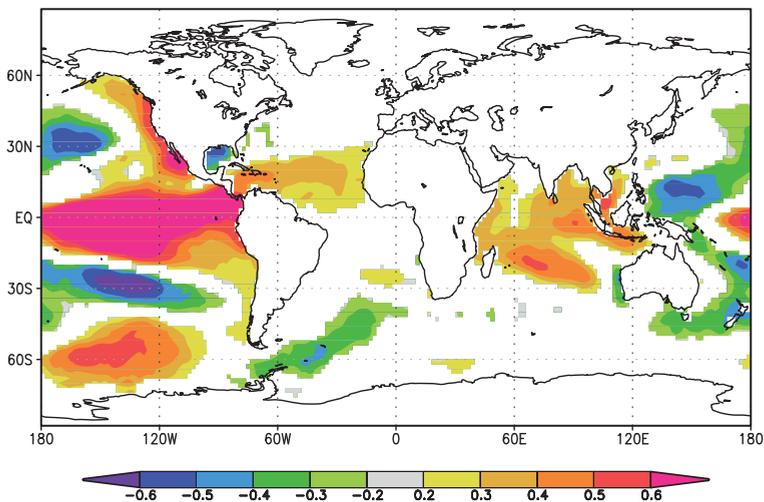


Figure 1.4. The spatial correlation of global SST (ERSSTv.3) with the Nino 3.4 index for Madagascar wet season (JFMA).

While it is evident that positive ENSO events are characterised by basin wide Indian Ocean warming (Fig. 1.4), the cause of warming in the SW Indian Ocean cannot solely be explained by surface fluxes (Klein et al., 1999). There is a thermocline ridge from 5°S to 10°S in the SW Indian Ocean where cool isotherms lie near the surface. It is here that thermocline variability strongly influences SST (Xie et al., 2002). During the mature phase of El Niño (December), anomalous anticyclonic wind stress forms in the tropical southeast Indian Ocean, caused by an atmospheric bridge from the Pacific, which is an anomaly of the Walker Circulation. This forces downwelling Rossby waves in the southeast Indian Ocean, which after they arrive in the SW Indian Ocean several months later, deepen the thermocline ridge and warm SST there (Fig. 1.5; Xie et al., 2002). SW Indian Ocean SST is therefore largely determined by remotely forced changes in the depth of this ridge.

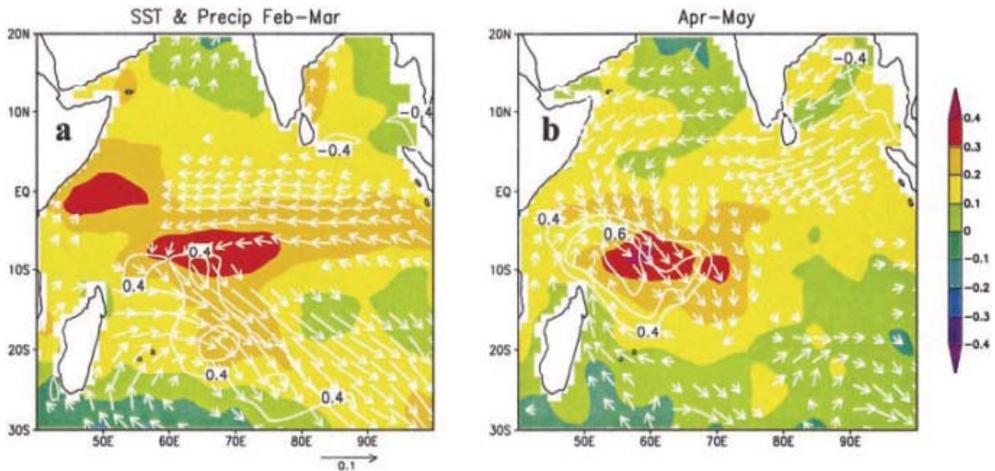


Figure 1.5. Regression coefficients of SST (colour shade in °C) and surface wind stress (vectors in 10^{-1} N m^{-2}) with the ENSO index in (a) Feb–Mar and (b) Apr–May. The precipitation correlation is plotted in contours. From Xie et al. (2002).

Local maximum SST forms over the thermocline ridge during boreal spring (April – July), where a mean standard deviation of 0.4°C - 0.5°C is reported (Xie et al., 2002). Moreover, embedded in the basin wide warming of the Indian Ocean that peaks between February – April following El Niño, SST anomalies display a positive core that co-propagates westward with the Rossby waves (Fig. 1.5; Xie et al., 2002). This core is the deepening

of the thermocline in the SW Indian Ocean, which together with SST and precipitation, shows positive correlations with ENSO (Fig. 1.5; Xie et al., 2002). As the ITCZ is displaced south of the equator during boreal winter, covering the 5°S to 10°S thermocline ridge, atmospheric convection and thus precipitation the SW Indian Ocean also increases in the region, with significant impacts on both local and remote climate (Fig. 1.5). Increased precipitation following positive ENSO events in the SW Indian Ocean is associated with an anomalous cyclonic circulation in the lower troposphere, as well as an increase in the number of tropical cyclones (Xie et al., 2002). Furthermore, the persistent SW Indian Ocean warming anchors convection south of the equator, delaying the onset of the south Asian monsoon in May (Joseph et al., 1994; Kawamura et al., 2001).

There is evidence to suggest that ENSO is correlated with the Indian Ocean Dipole, although from 1876 to 1999 about half of the IOD events occurred independently of ENSO (Meyers et al. 2007). Triggering of the IOD seems to therefore be by either ENSO, or by self-generation, provided the thermocline off Sumatra is shallow enough to support a type of feedback, namely Bjerknes feedback (Meyers et al., 2007). The IOD develops in June and peaks in October (Saji et al., 1999). Due to the strong seasonal variability of the monsoon winds, sometimes conditions prevail that allow a shoaling of the eastern equatorial thermocline, similar to La Niña-like ocean-atmosphere anomalies in the Pacific. This phenomenon is characteristic of an IOD event. South easterly winds, associated with the monsoon, form off the Sumatra coast, reaching a peak during the summer at which point the winds have an easterly component all along the equator. This forcing favours a shoaling of the thermocline and upwelling in the eastern equatorial Indian Ocean. Through the Bjerknes feedback mechanism, there is then a zonal gradient of tropical SST, with cooling off the coast of Sumatra and warming in the western ocean (Saji et al., 1999). As the IOD develops, an east–west dipole of anomalous rainfall is established over the tropical Indian Ocean, with precipitation increasing in the west because of the low-level convergence associated with the anomalous equatorial easterlies, and vice versa in the east (Saji et al., 1999; Schott et al., 2009). The intense winter monsoonal forcing eventually weakens the southeasterly winds off Sumatra, leading to the rapid termination of IOD events. The IOD is weak in most years, likely because the Sumatra

thermocline is too deep. It should be noted that the main impact region of the SW Indian Ocean is above 10°S, and therefore the impact on Madagascar is likely to be minimal.

Interdecadal and Multidecadal Variability

Both the IOD and ENSO display decadal variability. Further, the correlation between ENSO and the IOD varies on decadal timescales. As there is decadal variability in the depth of the eastern equatorial thermocline, the correlation between ENSO and the IOD increases when the thermocline is deep, as it takes a positive ENSO event to force an IOD event. It may well be that the decadal variability associated with the eastern equatorial thermocline is related to the Indonesian Throughflow (Santoso et al., 2011), which in turn might be related to another external decadal/multidecadal force. Indeed, IOD events occurred more independently from ENSO in the 1950's and 1960's, and then from the mid-1970's onwards became highly correlated (Schott et al., 2009). Interestingly, in 1976 the Pacific Decadal Oscillation (PDO) switched from a negative to a positive phase, and according to Williams and Grottooli (2010), the PDO can influence the depth of the thermocline, whereby a positive PDO phase (1976 – 1998) will deepen the thermocline in eastern equatorial Indian Ocean (Williams and Grottooli, 2010).

The PDO is a major internal mode of ocean-atmosphere variability (Mantua et al., 1997). Positive PDO phases are characterised by low SST in the central midlatitude Pacific and warm anomalies along the northern and eastern margins, and south of 30°N (Fig. 1.6). The PDO is remotely forced from the Tropics in part (Schneider and Cornuelle, 2005), and responsible for strong multidecadal (50 – 70 years; Minobe, 1997) and interdecadal Pacific oscillations in SST (IPO; 17 – 28 years) (Meehl and Hu, 2006). It is considered the leading mode of North Pacific SST defined by instrumental data for the past 120 years (Mantua et al., 1997). Mounting evidence indicates that the PDO has teleconnections extending over thousands of kilometers to the Indian Ocean (Cole et al., 2000; Crueger et al., 2009). The positive PDO phase corresponds to warm Indian Ocean SST anomalies (Deser et al., 2004), thought to exceed anomalies associated with ENSO (Krishnan and Sugi, 2003), particularly in the southwestern Indian Ocean (Fig. 1.6; Meehl and

Hu, 2006). While it is evident that changing rainfall patterns over Australia respond to the PDO (Arblaster et al., 2002; Lough, 2007), links to rainfall in southeastern Africa and the western Indian Ocean have also been suggested (Deser et al., 2004; Zinke et al., 2008).

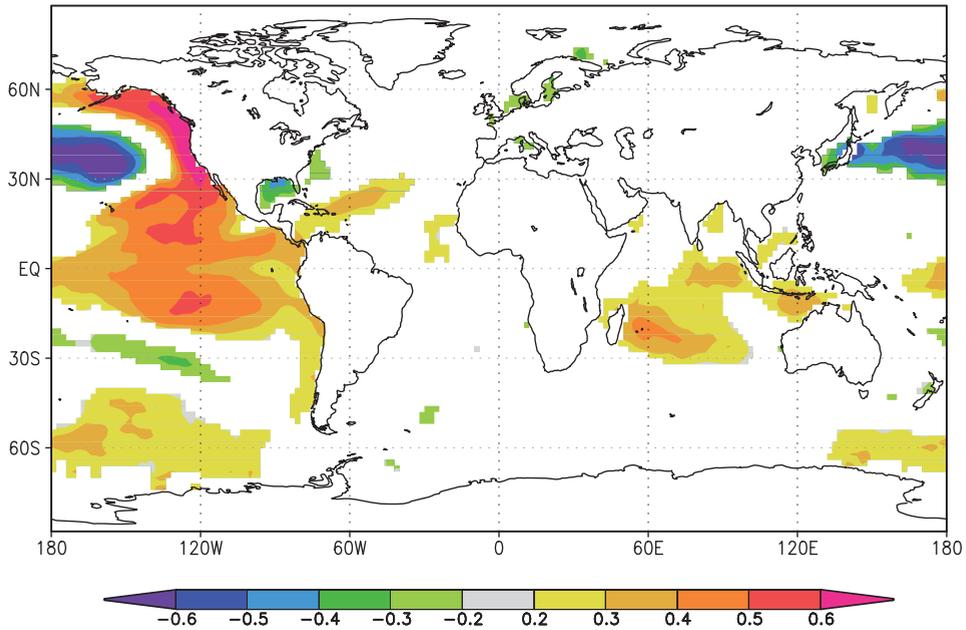


Figure 1.6. The spatial correlation of global SST (ERSSTv3;) with the PDO index for Madagascar wet season (JFMA).

Coral Reefs

“The organic forces separate the atoms of carbonate of lime, one by one, from the foaming breakers, and unite them into a symmetrical structure. Let the hurricane tear up its thousand huge fragments; yet what will that tell against the accumulated labour of myriads of architects at work night and day, month after month. Thus do we see the soft and gelatinous body of polypus, through the agency of the vital laws, conquering the great mechanical power of the waves of an ocean, which neither the art of man nor the inanimate works of nature could successfully resist.” Charles Darwin, in: *Journal of researches into the natural history and geology of the countries visited during*

the voyage round the world of H.M.S. Beagle under command of Captain Fitz Roy London, John Murray Eds, Albemarle Street, 1845.

Coral reef ecosystems are among the most biologically diverse ecosystems on the planet (Hallock, 2001). Corals have evolved to dominate well lit, shallow tropical waters, producing a diverse assortment of colourful colonies and morphologies. It is their ability to secrete an aragonite skeleton through extracting calcium (Ca^{2+}), carbonate (CO_3^{2-}) and bicarbonate (HCO_3^-) ions from seawater that allows symbiotic corals to construct giant reef frameworks upon which coral reef ecosystems thrive (Stanley and Schootbrugge, 2009). These living carbonate structures provide habitats for over a million species of plant, animal, fungi and bacteria, of which as much as 90% are yet to be discovered (Reaka-Kudla, 1996). In addition to the significant biodiversity, coral reefs provide food and resources for over 500 million people (Hoegh-Guldberg, 2011). Ecosystem services provided by coral reefs include food generated by small-scale fisheries, income from commercial fishing and tourism, and coastal protection against storms and hurricanes. Altogether, coral reefs play pivotal roles, bringing social, economic and cultural benefits which are crucially important to tropical island societies and nations (Spalding et al., 2001).

Symbiotic relationship

Scleractinian (stony) corals are the result of mutualistic symbiotic associations between cnidarian invertebrates and intracellular dinoflagellate algae of the genus *Symbiodinium*. This symbiosis underpins the success of coral reef ecosystems. Physiologically, the endosymbiont algae play a crucial role in host sustenance and survival in nutrient-poor tropical waters (Muscatine and Porter, 1977). The photosynthetically fixed carbon translocated from the algae can cover most or the entire energetic demands of the host (Muscatine et al., 1984) and powers the metabolically expensive process of calcification, allowing corals to deposit huge amounts of calcium carbonate (Hoegh-Guldberg, 1999). In return for the energy given to the coral host, the symbiotic algae receive access to inorganic nutrients resulting from animal metabolism, and are sheltered from threats such as ultraviolet (UV) radiation damage (Muller-Parker and D'Elia, 1997; Yellowlees et al., 2008). The close proximity of the autotrophic and heterotrophic components

facilitates molecular exchange and prevents nutrient losses (Tanaka et al., 2006), another adaptation to the oligotrophic conditions. Due to its crucial nutritional function, the phototrophic component of the association plays a major role in coral niche occupation. Specifically, algal symbiont functional diversity has been hypothesised to regulate the vertical distribution of corals (Iglesias-Prieto et al., 2004). Divergent genetic lineages of these algal symbionts appear to be adapted to particular environmental niches (Frade et al., 2008a; Frade et al., 2008b) and to confer varying levels of tolerance to increased seawater temperatures (Sampayo et al., 2008). The close relationship between corals and symbiotic dinoflagellates has been in existence for at least 220 million years (Muscatine et al., 2005), and is largely responsible for the huge reserves of limestone found in the upper layers of the earth's crust. The oldest fossil records of scleractinian corals date back 245 million year ago (Hoegh-Guldberg, 2011).

Threats

The current global distribution of coral reefs provides insight into the environmental conditions needed for maintaining corals and the reefs that they build. Because corals need abundant light for their photosynthetic symbionts, their distribution is limited to shallow seas (not deeper than 100 m) in regions within 30° north and 30° south of the equator. Coral reefs are also dependent on warm and thermally stable oceans, where temperatures do not fall below 18°C in the winter (Kleypas et al., 1999). However, one of the major contemporary threats to coral reefs comes in the form of widespread temperature rises associated with both ENSO and anthropogenic warming. These have been shown to cause coral bleaching; a loss or reduction of the symbiont populations or their photosynthetic pigments which often results in host mortality (Carpenter et al., 2008; Lesser, 2007). Bleaching most often results from physiological stresses on the algal symbionts caused by the synergistic effect of elevated seawater temperature and high light intensity (Douglas, 2003). Intact symbioses vary in their bleaching susceptibility and much of this variation has been attributed to differences in the tolerance of symbiont types towards temperature (Sampayo et al., 2008; Thornhill et al., 2008). Their narrow range of temperature tolerance puts coral reefs among the most threatened ecosystems on the planet (Carpenter et al., 2008; Lesser, 2007).

Coral bleaching is one of the most important causes for coral reef decline worldwide. This phenomenon has occurred with increasing frequency over the last twenty years, primarily due to the more frequent occurrence of strong ENSO events (Hoegh-Guldberg, 2004). Regional patterns in coral cover distribution in the Indian Ocean are driven mainly by episodic and acute environmental stress (Ateweberhan et al., 2011). In fact, coral reefs face widespread degradation, seen for instance in major losses of coral cover worldwide (Bak et al., 2005; Wilkinson 2008). In the Western Pacific the abundance of coral reefs has declined by as much as 50% and is continuing to decline at the rate of 1-2% per year (Bruno and Selig, 2007). At the local level, coral reefs are affected by declining water quality (resulting from increasing nutrient and sediment load from disturbed coastlines), over-exploitation of key marine species, destructive fishing practices and pollution from expanding urban areas (Hughes et al., 2003). Climate change is dramatically altering the circumstances under which corals have prospered for many millions of years, therefore exacerbating changes caused by local anthropogenic influence. Rapidly changing global climate is predicted to be shifting the face of coral reefs as we know them, carrying along important ecological and socioeconomic alterations (Hoegh-Guldberg, 2011).

Coral paleoclimatology

Massive tropical corals

Massive tropical corals continuously secrete a carbonate skeleton that may accumulate for centuries, creating giant structures. The most common genus of these scleractinian corals in the Indo-Pacific region is *Porites* spp. (Fig. 1.7). In the southwest Indian Ocean particularly, the primary reef-building species of *Porites* are *P. solida*, *P. lutea* and *P. lobata*. Massive colonies are spherical or hemispherical when small, and helmet or dome-shaped when large, which may be over 5 meters across (Fig. 1.7). Their corallites are small, immersed, and filled with septa, and the tentacles of most, but not all species, are extended only at night when they feed on zooplankton. Colours of the colonies mainly consist of cream, yellow or pale brown, however, brighter colours such as greens, blues and purples are found in shallower waters (Fig. 1.7). Differences in colours are largely related to the coral symbiont type.



Figure 1.7. A large *Porites* species photographed in eastern Madagascar. Photograph taken by Dr. Jens Zinke.

Calcification Processes

The coral skeleton is extracellular, located at the base of coral tissue, like a finger covered by a glove. Composition of the biomineral aragonite includes two fractions, one mineral and one organic, called the organic matrix. This organic framework is secreted by the coral periodically, on which mineral ions fix onto (Allemand et al., 2004; Cuif and Dauphin, 2005; Ingalls et al., 2003). Epithelial transport of molecules from seawater to the calcification site can be achieved by two mechanisms: a paracellular pathway, between cells (in this case, transport of the molecule is only driven by its chemical gradient, transport is diffusional and passive), or a transcellular pathway through cells (Allemand et al., 2004). In this last case, the molecule needs to enter the cell, cross the cell and then exit. Thus the transport is active, against a gradient with energy supply (Fig. 1.8). Transport of charged species across the hydrophobic

cell membrane requires specific carrier proteins (Allemand et al., 2004; Furla et al., 2000; Ingalls et al., 2003).

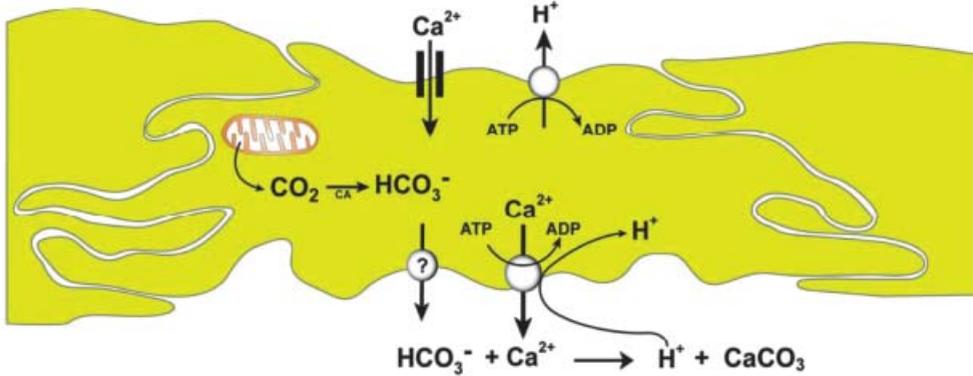


Figure 1.8. Schematic diagram showing the pathway of entry and exit of calcium through the calcicoblastic cells during calcification in hermatypic corals. From Allemand et al., 2004.

The accretion of calcium carbonate by reef building corals depends on a number of environmental factors, such as sun hour, light transparency, nutrient exposure, partial pressures and temperature. These environmental factors affect the accretion of high and low density bands within the skeletal matrix of coral colonies. The high density and low density bands are revealed by X-ray radiographic techniques (Carricart-Ganivet and Barnes, 2007; Chalker and Barnes, 1990). Generally one year of growth consist of two density bands, one high density band and the other low density band, providing a means to age the coral by counting the paired bands. The precipitation of these bands is mostly related to seasonal variability. Coral colony growth rates of *Porites* spp. can vary between 5 and 20 mm per year depending on environmental conditions (see above).

Calcification is a product of both coral extension rates and density (Cantin et al., 2010; De'ath et al., 2009; Lough and Barnes, 2000). By assessing changes in calcification over time there is great potential to link growth variability to environmental conditions (Cantin et al., 2010; De'ath et al., 2009; Scoffin et al., 1992). Physical characteristics of coral bands, such as their skeletal density, linear extension rate, tissue thickness and calcification rate, provide time-series data on the environmental conditions controlling coral growth, such as sea surface temperature (SST) (Cantin et

al., 2010; Lough and Barnes, 2000). As coral reefs around the world are becoming increasingly threatened by bleaching, pollution, ocean acidification and sedimentation, the coral skeleton itself therefore provides an indicator of a changing environment (Helmle et al., 2011). Furthermore, locked within the coral skeleton, geochemical proxies can identify and record these past environmental conditions, providing important information on both the coral reefs environmental response and global climate itself (Felis and Paetzold, 2003; Lough, 2004).

Coral Proxies

Massive corals growing in shallow tropical and sub-tropical oceans have micro-laminations (density banding) that contain physical, isotopic and geochemical evidence of past environments at annual and sub-annual resolution (Gagan et al., 2000; Lough, 2004). The stable oxygen isotope ratios ($\delta^{18}\text{O}$) of coral carbonate provide information on past SST, sea surface salinity and the hydrological balance of oceans (Guilderson and Schrag, 1999; McConnaughey, 1989; McConnaughey, 2003; Zinke et al., 2004). When the $\delta^{18}\text{O}$ composition of seawater is more strongly governed by the interaction of precipitation, evaporation and water advection, the coral carbonate provides a record of changes in the hydrologic balance (Pfeiffer et al., 2004). However, the $\delta^{18}\text{O}$ composition of seawater may or may not correlate with precipitation, depending on SST variability (Charles et al., 1997). Moreover, if the coral records are positively correlated with precipitation, $\delta^{18}\text{O}$ can be used to reconstruct past rainfall (see Gagan et al. 2000 for a review). Alternatively, pairing $\delta^{18}\text{O}$ with geochemical SST proxies (e.g. Sr/Ca) gives a more accurate indication of past hydrological changes (Pfeiffer et al., 2006; Zinke et al., 2004).

In addition to $\delta^{18}\text{O}$, the stable carbon isotope ratios ($\delta^{13}\text{C}$) of the coral skeleton may show strong seasonal annual cycles. However, the $\delta^{13}\text{C}$ of coral carbonate is confounded by many effects, other than the $\delta^{13}\text{C}$ of seawater (Omata et al., 2008; Reynaud et al., 2007; Reynaud et al., 2002; Rodrigues and Grottoli, 2006; Suzuki et al., 2005), meaning surface productivity reconstructions are severely constrained. Coral autotrophy (Omata et al., 2008; Reynaud et al., 2007), heterotrophy (Reynaud et al., 2002), riverine

DIC (Moyer and Grottoli, 2011) and productivity of the water column all influence the skeletal $\delta^{13}\text{C}$ signal, complicating the signal.

The geochemistry of coral bands have also been used to reconstruct past SSTs, with the most robust being the strontium-calcium ratio (Sr/Ca) (Alibert and McCulloch, 1997; Gagan et al., 1998). A negative relationship exists with SST, whereby as temperatures increase, less Sr is incorporated into the aragonite lattice relative to Ca. Since Sr has a long oceanic residence time, heterogeneity in skeletal Sr/Ca is assumed to mainly reflect SST variability, therefore down-core sampling of massive corals yields an *in situ* SST time-series in which the resolution is only limited by the coral growth rate (Sun et al., 2005). However, recent studies suggest that the Sr/Ca signal is also confounded by vital effects, meaning coral growth parameters play an important role in determining skeletal Sr/Ca ratios (Cohen and Gaetani, 2010; Gaetani et al., 2011). A compilation of recent coral Sr/Ca-SST calibrations revealed that the mean Sr/Ca response to a 1°C increase in SST is -0.061 mmol/mol, with a range of -0.04 mmol/mol and -0.08 mmol/mol (Correge, 2006). Other coralline geochemical proxies for temperature have been explored (Fallon et al., 2003), including magnesium-calcium ratios (Mg/Ca), uranium-calcium ratios (U/Ca), boron (B) and fluorine (F) (Hart and Cohen, 1996; Min et al., 1995; Mitsuguchi et al., 1996). Nevertheless, the Sr/Ca ratio remains to this date the most robust.

As well as assessing the stable isotope $\delta^{18}\text{O}$ signal in coral skeletons to reconstruct rainfall, other proxies exist that are related to terrestrial erosion products, indirectly linked to river runoff and precipitation. Luminescent banding found in coral skeletons has been used to improve dating of coral records as well as provide a proxy for precipitation and river runoff from adjacent land masses (e.g. Hendy et al. 2003). Although the exact nature and cause of luminescent banding is still debated, there are positive correlations with river runoff variability (Isdale et al., 1998; Lough, 2011; Lough, 2007). The incorporation of terrestrial humic acids was first proposed as the likely cause of luminescence, meaning increasing runoff would increase humic acid erosion and therefore the incorporation of the fluorescent compound into the skeleton (Isdale, 1984). However, coral skeletal density was later proposed as the likely source, as aragonite itself is fluorescent, and banding was observed in corals far from terrestrial influence (Barnes and Taylor, 2001, 2005). A

detailed assessment of luminescent banding in corals can be found in Chapter 2 (Grove et al., 2010).

The barium-calcium ratio (Ba/Ca) is used to reconstruct sediment runoff (McCulloch et al., 2003; Prouty et al., 2010; Sinclair and McCulloch, 2004). Ba is desorbed from sediment particles in the low-salinity range, mixed into the water column and incorporated into the coral skeleton (Coffey et al., 1997). Elevated skeletal Ba/Ca concentrations therefore indicate increased sediment runoff conditions. Other coralline geochemical proxies applied to reconstruct runoff include yttrium (Y) and manganese (Mn), both of which are associated with the erosion of bedrock (Lewis et al., 2007).

Similar to all other palaeoclimate proxy records, corals are biased by factors unrelated to climate. For example, the upward growth of a coral can lead to the exposure of its surface to shallower water depths, and in turn slightly different temperature, salinity and light intensity levels. This has potential ramifications for the isotopic and geochemical records obtained (Felis et al., 2003; Gagan et al., 2000; Inoue et al., 2007; Omata et al., 2008; Reynaud et al., 2007; Suzuki et al., 2005). One way to identify and adequately take such effects into account is to cross match coral records and/or create composite records (Cahyarini et al., 2008; Pfeiffer et al., 2009). Indeed, some studies have done this, attempting to quantify the reliability of climate proxies contained within coral records through local, regional and global cross matching as well as comparing single and composite records with instrumental records (Cahyarini et al., 2008; Evans et al., 2002; Guilderson and Schrag, 1999; Hendy and Gagan, 2003; Pfeiffer et al., 2009). However, many studies continue to rely on the implicit assumption that variations identified in a single coral record are attributable to one or more climatic variables (Lough, 2004). Some caution therefore needs to be applied in utilising coral records. If the various potential influences on proxies contained within corals are well understood, and the coral records obtained are well calibrated and validated, they provide an extremely powerful tool for understanding climate change and variability from seasonal through to millennial time scales. However, it is important to remember that a coral is an animal, not a piece of equipment.

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