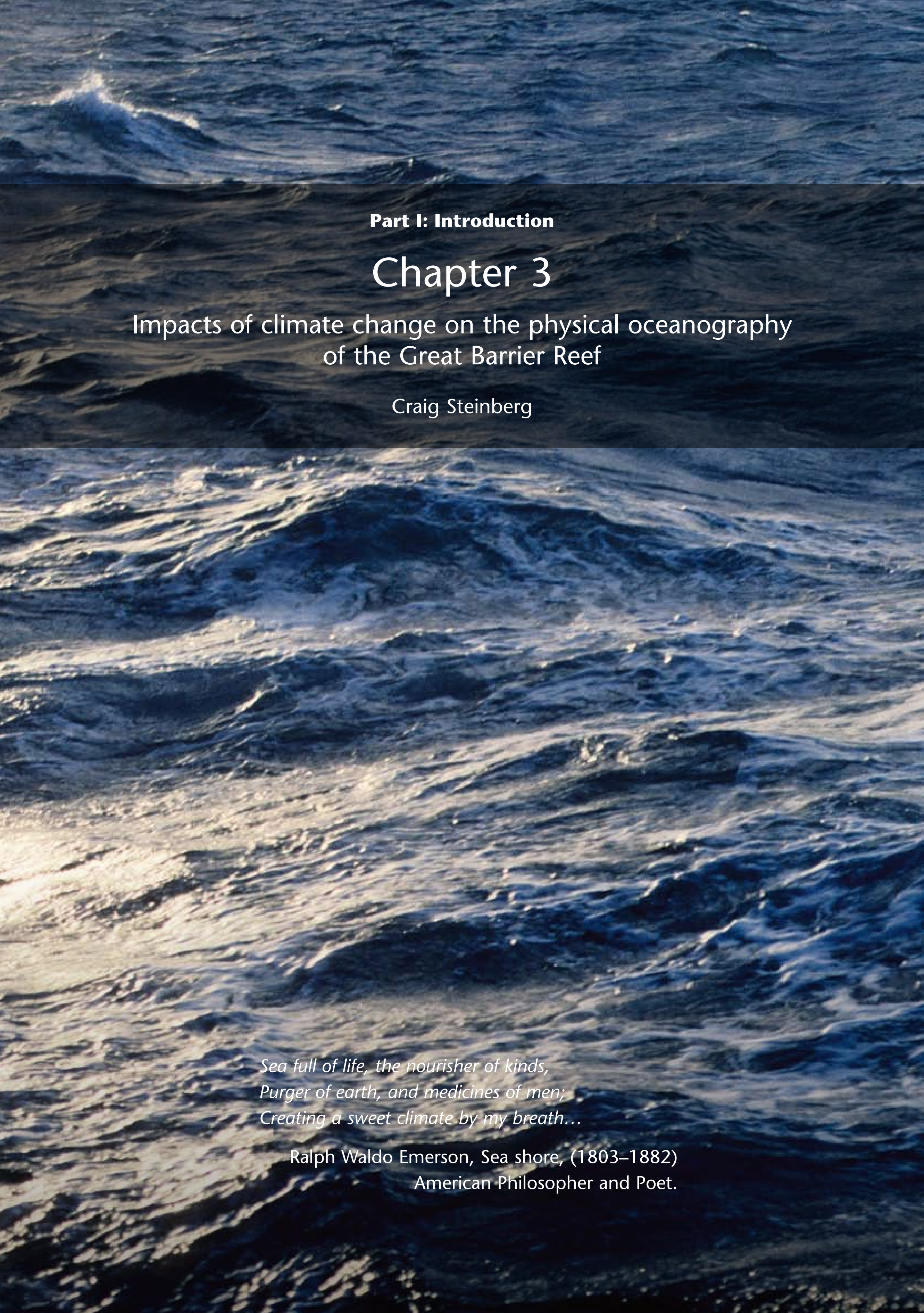


Part I: Introduction

Chapter 3

Impacts of climate change on the physical oceanography of the Great Barrier Reef

Craig Steinberg

Sea full of life, the nourisher of kinds,
Purger of earth, and medicines of men;
Creating a sweet climate by my breath…

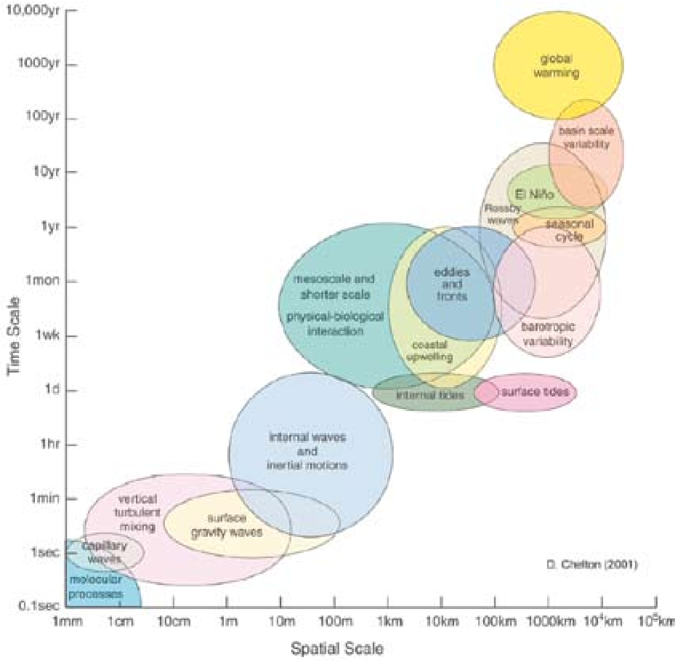
Ralph Waldo Emerson, Sea shore, (1803–1882)
American Philosopher and Poet.

3.1 Introduction

The oceans function as vast reservoirs of heat, the top three metres of the ocean alone stores all the equivalent heat energy contained within the atmosphere[29]. This is due to the high specific heat of water, which is a measure of the ability of matter to absorb heat. The ocean therefore has by far the largest heat capacity and hence energy retention capability of any other climate system component. Surface ocean currents (significantly forced by large scale winds) play a major role in redistributing the earth's heat energy around the globe by transporting it from the tropical regions poleward principally via western boundary currents such as the East Australian Current (EAC). These currents therefore have a major affect on maritime and continental weather and climate.

It is important to understand the temporal and spatial scales that influence ocean processes. Energy is imparted to the ocean by sun, wind and gravitational tides. The energy of the resulting large-scale motions is transmitted progressively to smaller and smaller scales of motion through to molecular vibrations where energy is finally dissipated as heat[42]. The oceans therefore play an important role in climate control and change, and Figure 3.1 shows the ranges of time and space, which characterise physical processes in the ocean and their hierarchical nature. Within this scheme, global warming occurs over different temporal (centuries to millennia) and spatial (global to hundreds of kilometres) scales. Through the energy cascade, climate change will affect all the other oceanic processes at smaller scales (summarised in Figure 3.1) and may alter their range, intensity and frequency and so strong regional variations in response are expected.

Figure 3.1 *Domain of space-time scales of physical processes (Reproduced courtesy of Chelton[17])*



Prior reviews of regional physical oceanography include Australian oceanographic processes in Church and Craig[21], Coral Sea circulation in Burrage[8] and the GBR in Pickard et al.[48]. For a more detailed review of physical processes on the GBR, Wolanski[62] is recommended. This chapter reviews the expected key climate-influenced oceanographic processes that affect the Great Barrier Reef (GBR). Section 3.2 explains how sea level variations are used to observe longer-term effects of climate change. Section 3.3 discusses processes involved in the air-sea heat budget that may result in a warming surface layer and the mixing mechanisms that are available in the water column to dissipate it. Section 3.4 provides a review of the current understanding of Coral Sea circulation. Section 3.5 discusses currents in the GBR. Finally, Section 3.6 discusses conclusions and recommendations. At the end of each sub section some effort is made to identify projected changes on the identified processes, assess their certainty and any regional detail if not already given in the previous chapter.

3.2 Observing long term climate changes: sea level and the El Niño-Southern Oscillation (ENSO)

The oceans remain the least understood and most sparsely sampled regions of the world due to the expense of sampling in marine environments and the historical reliance on slow ship based observations. Recent technological advances in observing networks, such as satellite remote sensing and Argo profiling drifters, is leading to a more global coverage and more frequently updated picture emerging. Long-term trends, however, remain difficult to recover due to the shortage of long-term records and the difficulty in separating out different signals from other processes. For example, one of the longest reliable instrument records available is sea level derived from coastal tide gauges. Sea level can be a good integrator of large-scale currents, temperature variability due to the expansion of seawater during warming and changes in meteorological forcing (eg setup due to wind stress).

Sea levels vary temporally and spatially over a wide range of scales. Surface gravity waves generated by storms attain heights on the order of 10 metres and storm surges can pile water up on the coast and may subsequently propagate along the coast as trapped waves. Tide fluctuations in sea level can range up to 10 metres depending on location. Large-scale currents can cause sea level fluctuations of up to one metre. Annual variability in sea level is principally due to seasonal warming and thermal expansion of the water column, variations in prevailing wind strength and direction and changes in the strength and timing of the gravitational tide.

To effectively reveal long-term (eg inter-annual and inter-decadal) sea level changes associated with climate change, long-term records are assessed after shorter term seasonal and shorter period variability is removed. This is achieved by subtracting the predicted tide and smoothing out these higher frequency processes by filtering. Correction must also be made for tectonic movement and crustal deformation caused by the loading of flooding tides on continental shelves. The National Tide Centre at the Bureau of Meteorology is responsible for observing long-term sea levels in Australia through the Australian Baseline Sea Level Monitoring Project, and throughout the South Pacific, through the South Pacific Sea Level and Climate Monitoring Project. Long-term sea level trends on the GBR show an increase of 2.9 mm per year since 1991 at Cape Ferguson (central GBR) and a 2.6 mm per year increase since 1992 at Rosslyn Bay (southern GBR). Based on satellite altimetry, Church et al.[23] calculated an average sea level rise in the Indian Pacific region for the period 1993 to 2001 to be 4 mm per year. Uncertainty in the accuracy of these trends is caused by variability associated with global-scale phenomena like ENSO, and the relative shortness of high precision records from satellite altimetry and the tide gauge network.

Sea level anomalies with the predicted tides, seasonal cycles and linear trend removed are shown in Figure 3.2 for Rosslyn Bay and Cape Ferguson. Both show that during the 1997–1998 El Niño event, sea levels rose approximately 18 cm over a 12 month period revealing a regional-scale variation two orders of magnitude larger than the long term sea level rise. The West Pacific Warm Pool (WPWP), in the seas around northern Papua New Guinea and the Solomon Islands, exhibited a change in sea level of more than twice that observed in the GBR at 40 cm (Figure 3.3).

Figure 3.2 *Sea level anomalies for Cape Ferguson and Rosslyn Bay. Units are in metres (Adapted from the Australian Baseline Sea Level Monitoring Project, Annual Sea Level Data Summary Report July 2004 to June 2005*[a]*)*

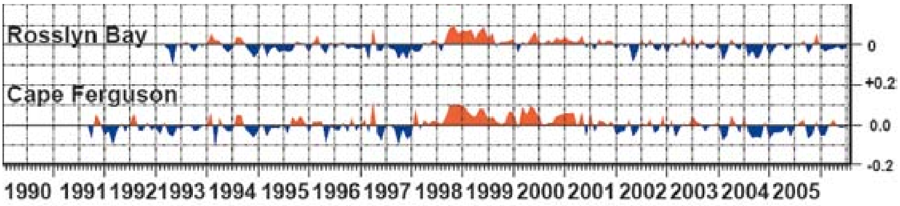
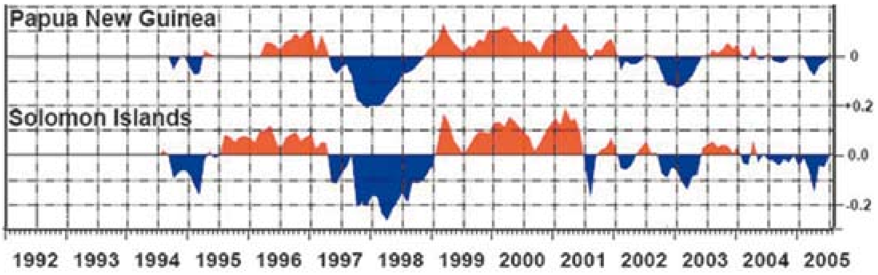


Figure 3.3: *Sea level anomalies observed by the Manus Island, PNG and Honiara, Solomon Islands sea level gauge (Adapted from Pacific Country Report, Sea Level and Climate: Their present state, Papua New Guinea, June 2005*[b]*)*



The three to seven year ENSO cycle affects sea level through a complex interaction between atmospheric and oceanographic processes. ENSO was originally observed as a change in the difference in atmospheric pressure between Darwin and Tahiti[5]. This difference provided a simple indicator of the shifting atmospheric Walker circulation in which lower pressures occur where air rises over warm ocean waters and higher pressures occur where air descends over cooler waters. Figure 3.4 shows that during the opposite ENSO phase, La Niña, the WPWP (centred north of Papua New

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[a] http://www.bom.gov.au/oceanography/projects/abslmp/abslmp.shtml

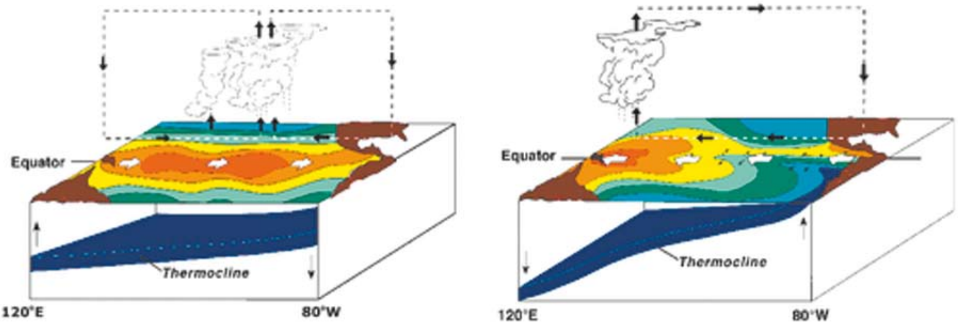
[b] http://www.bom.gov.au/oceanography/projects/spslcmp/spslcmp.shtml

Guinea) warms the equatorial air causing it to rise, and thereby lowering the atmospheric pressure at Darwin. At the same time, the eastern Pacific experiences characteristically cooler upwelled waters from the divergence of surface waters away from the South American continent due to the local winds and the equatorward transport of cooler waters by the Peru Current from the south. Air subsides in the eastern Pacific to complete the convection cell and increases the air pressure at Tahiti. During the El Niño phase of the Southern Oscillation the reverse occurs. As the heat anomaly moves eastward the warmer waters form a low-density cap, which prevents upwelling and the cooling of waters off the coast of Peru. Sea levels in the warmer regions are higher due to a combination of thermal expansion and a convergence of currents causing the ocean surface to dome up.

There is also a deepening of the surface warm, low-density layer. The region of transition between the warmer surface waters and colder deep oceanic water is known as the thermocline. This is displaced lower due to the build up of the warmer waters[70]. Thus during La Niña the eastern Pacific experiences cold water upwelling (with the thermocline breaking the surface), and the Western Equatorial Pacific experiences a depressed thermocline that hinders any upwelling due to the cooler water being much deeper. In contrast, during El Niño, the thermocline is much closer to the surface in the Western Pacific making the cooler waters potentially more available to continental shelves. It is important to note that whilst the sea level may vary by tens of centimetres, thermocline depth can vary over a range of hundreds of metres, typically occurring between 50 and 200 metres depth during ENSO. This effect arises because of the relatively small density contrast between the internal layers that make up the thermocline and the huge density difference between the ocean and the atmosphere. Actual thermocline response along the GBR by ENSO is unknown and may not necessarily occur south of the WPWP, however it can be inferred from the sea level anomaly data that there should be a response. There is an El Niño related sea level increase of 10 cm that can be seen as far south as South Australia. Here the thermocline has been observed to shallow to 60 to 120 metres, 150 metres above the mean thermocline depth[46]. This signal however is thought to propagate from the Indo-Pacific WPWP via the Indo-Pacific throughflow and along the shelf edge wave guide from Western Australia to South Australia, not along the east coast[25].

The extreme 1997–1998 ENSO event (the largest on record) is likely to have been exacerbated by an in-phase Pacific Decadal Oscillation (PDO) (see Lough chapter 2) that behaves in a similar manner to ENSO in the Western Pacific. Over the last few decades, considerable effort has been put into explaining and predicting regional ENSO affects. With each event there are significant variations in behaviour and

Figure 3.4 *Schematic of the two phases of the El Niño Southern Oscillation during a La Niña (left) and El Niño (right) phase (Schematic courtesy of NOAA/ PMEL/TAO)*



hypotheses are refined. Whilst progress has been made in predicting the onset of El Niño, more recent studies are looking at the longer-term modulation of ENSO and different triggering mechanisms observed by ocean observing systems[59].

The 1997–1998 ENSO event shows that changes in currents, transport of warmer waters and thermocline displacement are likely to cause significant impacts on the GBR. Currents are highly variable on a whole range of scales and accurate long-term measurements are limited. Sea level has therefore been the traditional indicator of large-scale oceanographic changes. However the sea level signal can also be from thermal expansion (if warming) and not just from changes associated with changes in the strength and direction of major currents.

3.3 Thermal stratification and mixing of the water column

Section 3.2 discussed the importance of thermocline depth changes for regulating the appearance of cool, nutrient rich oceanic waters at the surface. This section will explore how ocean warming forms a buoyant surface layer and how mixing can disperse the heat throughout the water column.

Incoming solar radiation varies naturally by a few tenths of a percent due to dark sunspots and the 11-year solar cycle. Whilst these changes are small, they do influence climate variability. Increases in greenhouse gas concentrations are changing the Earth's atmospheric composition and radiative balance. In response, the Earth system is absorbing excess heat and the global oceans in particular are taking up much of this excess. The amount of radiation incident at the ocean surface depends upon the amount of cloud cover, aerosols, water vapour, angle of incidence, reflection and scattering. Short-wave radiation spectra are comprised of far infrared, visible and ultra-violet radiation. Absorption varies according to the wavelength, with the longer infrared being absorbed in the first few centimetres of surface waters and ultra-violet radiation penetrating to hundreds of metres into the ocean. However, the greatest amount of warming occurs near the surface because 75 percent of the total short-wave energy is absorbed in the top 5 metres[42].

Fluxes across the air-sea interface include the incoming and outgoing short and long wave radiation, sensible and latent heat fluxes. The tropical oceans tend to have a net gain of heat over the year although cold air temperatures, wind, evaporation and night time back radiation can cause periods of heat loss. The surface heating stabilises the top layer due to thermal expansion of the water reducing its density. Winds are the major source of turbulence that can mix these waters further down into the water column. The depth to which they can mix is dependent on the strength of wind, so over a warming period numerous thermal layers can develop that are progressively mixed deeper in the water column by successively stronger wind events.

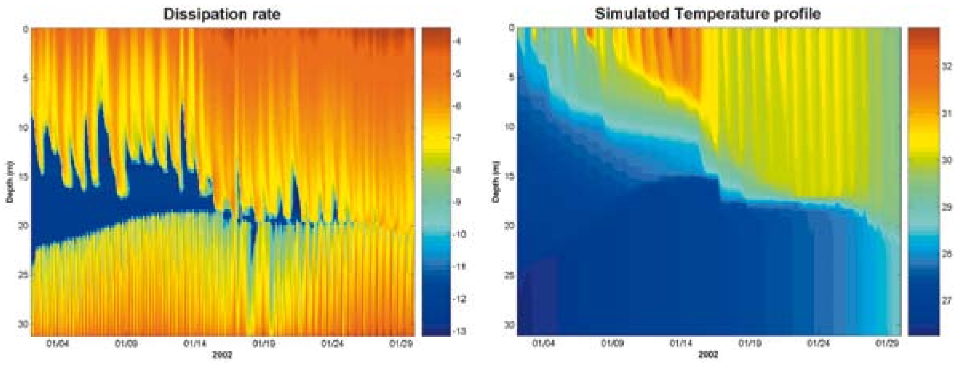
In the tropical ocean this well-mixed surface layer can extend down to depths exceeding 100 metres and this depth usually defines the location of the main thermocline for surface waters. Deeper mixing or additional thermocline deepening occurs through surface water convergence, entrainment driven by larger turbulent eddies or current shear instabilities[36] and basin-scale tilting. The GBR shelf is considered to be well mixed during most of the year assisted by the strong southeast trades[21] however an important exception occurs during summer warming events. Whilst the summer stratification may be considered weak compared with more temperate regions, the fact that corals are living at the limit of their thermal tolerances means that these episodes have a higher ecological importance than would otherwise be the case.

Figure 3.5 illustrates the wind-generated turbulence through the water column over the course of one month during summer as indicated by the turbulent energy dissipation rate in the panel. The winds vary in strength diurnally but mix to a depth of between 10 and 20 metres over the course of the month. The panel shows how temperature response to the mixing is controlled by daily heating and a prolonged warming period from 9 to 15 January 2002. This warm surface layer is progressively mixed down to about eight metres, eventually overcoming the increased stability of the buoyancy from the lower density warmer waters. When the winds strengthen mid-month, the heat extends deeper to over 20 metres depth during the remainder of the month. The result is that the high surface temperatures are gradually redistributed into deeper layers. In tandem, the winds also cool the surface waters by releasing latent heat to the atmosphere through evaporation. Other factors contributing to cooling surface waters include cloud cover, which reduces direct heating by insolation during the day and clear skies at night enabling long wave radiation to escape the atmosphere.

On the GBR shelf, additional energy for vertical mixing from the sea floor up through the water column is sourced from currents that are dominated by the tides (Figure 3.5). Tidal currents generate turbulence from the shear produced by friction at the sea floor. A persistent cool bottom boundary layer exists in this case. Where tides are stronger and/or the water depth is shallower, they can mix all the way to the surface and this is a common feature along coasts, in channels between reefs and in macro-tidal areas. In deeper regions, where the surface wind mixing doesn't overlap with the tidal mixing from the sea floor, a central core can result, exhibiting a reduced dissipation rate where there is negligible turbulence available for mixing, as seen in the first half of the month in Figure 3.5.

Figure 3.6 shows two satellite images of sea surface temperatures for the GBR and Coral Sea. Persistent summer cooler waters are found along the outer far northern GBR during the austral summer in December and January. Hotter waters are apparent on the reef tops of large mid-shelf reefs (especially off Princess Charlotte Bay located at 14° S) and in the shallow waters along the coast. In contrast, waters are 2°C cooler along the outer edge of the continental shelf where the Ribbon Reefs occupy over 90 percent of the shelf break. It is thought that intrusions from a variety of processes (see section 3.4) cause cooler, deeper water to encroach onto the outer shelf, mixing waters around

Figure 3.5 *Simulated profile of the time history of turbulent energy dissipation rate (top) and temperature (bottom) for a midshelf location on the central GBR in January 2002*



the coral reefs[53]. These mechanisms effectively provide a microclimate for the outer reef corals keeping them cooler and less susceptible to heat stress than their mid-shelf counterparts. Figure 3.7 shows average December sea surface temperatures for the central and southern GBR. The dense outer reef matrix between latitudes 19° and 23° experiences consistently lower temperatures than the coastal and largely reef-free lagoon that extends to the middle of the shelf.

Figure 3.6 *December (left) and January (right) sea surface temperature climatology of the northern and far northern GBR, averaged from 1990 to 2000 (Source: Skirving et al.[54])*

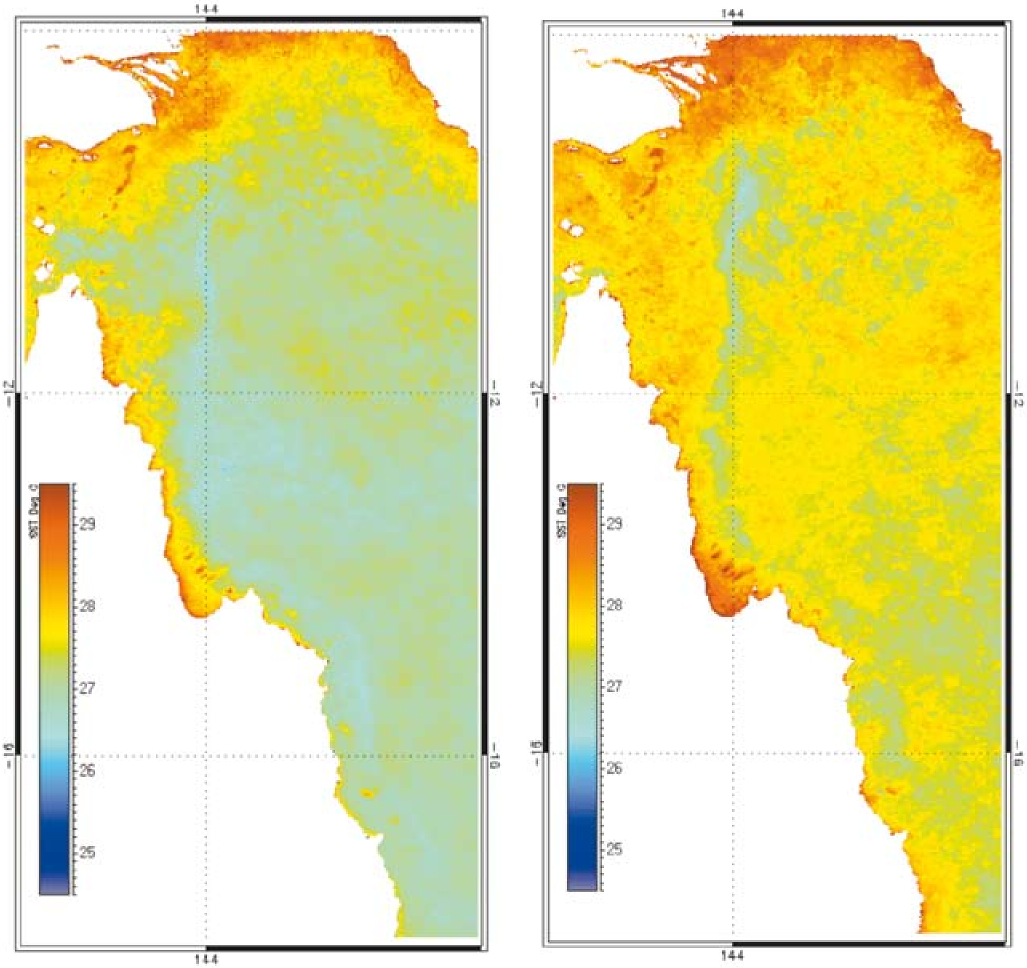
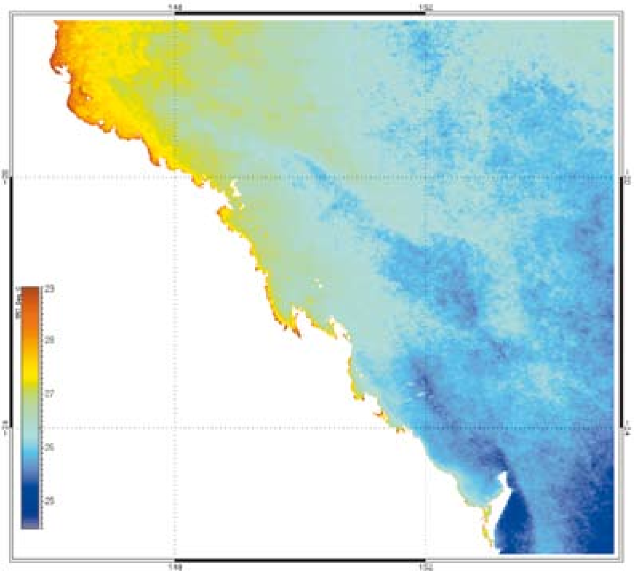


Figure 3.7 *December sea surface temperature climatology of the central and southern GBR averaged from 1990 to 2000 (Source: Skirving et al.[54])*

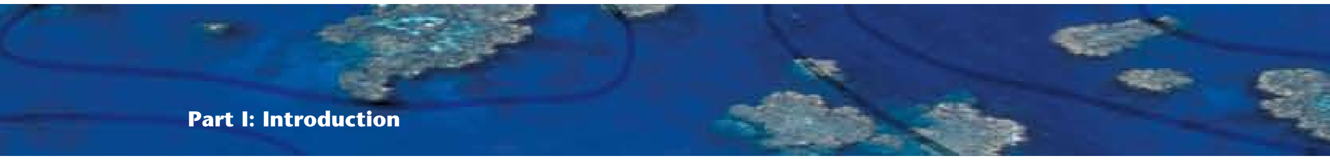


The oceanic surface mixed layer mediates the exchange of mass, momentum, energy, heat and dissolved gases between the atmosphere and the ocean, and hence plays a central role in determining long-term climate response[36]. In tropical waters the surface layer tends to be oligotrophic and so any deepening of the mixed layer can potentially provide a source of nutrients to the photic zone, enabling an increase in primary productivity and cooling of surface waters. Thus deep chlorophyll maxima are widespread in the open ocean near the thermocline. Since heating is stabilising, it tends to suppress the penetration of turbulence down into the water column so mixing will be confined to a shallower surface layer. Climate modelling predicts that this may suppress the upward flux of nutrients reducing oceanic primary productivity, but also induce oscillations and increased variability in the amount of phytoplankton in the deep chlorophyll maximum and export of carbon into the ocean interior[34]. Momentum imparted to the more stable, shallower surface layer by the wind may speed up surface currents as the majority of the energy transfer will be confined to this layer.

Projected:
Thermal stratification is to increase. Depth of thermocline to rise and surface layer currents to increase.

Certainty:
High-significant warming already observed and projected to continue. Medium – Thermocline depth response

Regional detail:
Oceanic thermocline likely to shallow, stratified regions on the GBR shelf to increase in areas with less energetic tides.



3.4 Coral Sea inflows

3.4.1 South Equatorial Current

The South Equatorial Current (SEC) primarily drives Coral Sea circulation. It is the northern arm of the ocean basin scale South Pacific Gyre. The gyre is driven by the latitudinal contrast of the strong westerly winds forming the eastward flowing circumpolar current in the Southern Ocean and the south-easterly trade winds in the lower latitudes forcing the westward flow of the SEC. The surface waters of the SEC are warmed by several degrees as they traverse the equatorial Pacific and form a well-mixed surface layer of around 150 metres in thickness[16].

The classical view of Coral Sea circulation is derived from ship based hydrographic observations[52,19,2] and early, low resolution numerical modelling[33]. This view has the broad SEC entering the Coral Sea from the east and bifurcating at the GBR into a northern arm, the North Queensland Current or Hiri Current and the poleward flowing East Australian Current (EAC). The location of the bifurcation varies seasonally between 14° S and 20° S and lies at the southern end of this range during the southeast trade wind season (April to November). Underlying the EAC is a permanent undercurrent that flows northwards and eventually joins up with the Hiri Current[20,19,33,8].

Over the last two decades technological advances in computing, satellite and acoustic remote sensing and ship positioning has revealed significant complexity and detail in ocean circulation. Webb[60] used a numerical model to suggest that the broad westward SEC inflow is broken up into a number of zonal jets by shallow bathymetry associated with island archipelagos. The reef systems effectively impede the flow and force the waters around them. Figure 3.8 is a schematic showing that jets form north and south of Fiji[56], Vanuatu and New Caledonia. More recently, Ridgway and Dunn[49] have been able to discern these features in climatological data with recent increased resolution and after allowing for bathymetric control. Once in the Coral Sea, currents deviate around the reef systems on the Bellona (west of New Caledonia), Queensland and Marion Plateaus. This topography produces multiple pathways for the SEC to reach the GBR. Once the jets encounter the Australian continental shelf they form the EAC and Hiri currents flowing along the western boundaries. Kessler and Gourdeau[38] found evidence that jets can also be caused by quasi-permanent structures in the wind field, independent of the island and reef systems.

Figure 3.9 shows a snapshot of an eddy resolving ocean circulation model[51] showing the complexity of the Coral Sea circulation. It contains most of the features mentioned above as well as significant meanderings of the current flow. A large recirculation of the Hiri Current, known as the Papuan Gyre, provides a pathway from Papua New Guinea waters back to the far northern GBR[8]. A smaller recirculation is seen off the southern GBR, south of the Swain Reefs and east of the Capricorn Bunker Group. The model also reproduces the transient zonal jets extending eastward from southern end of the GBR (off Fraser Island), which were first detected using sea surface temperature imagery[11,12].

The SEC and EAC strengthen during an El Niño when the southern WPWP moves eastwards along the equator to the central Pacific. This is due to the SEC being displaced south[69,43,35], which favours EAC flow rather than contributing to the Hiri current. Burrage et al.[10] found a strengthening of the EAC in the central GBR and Wolanski and Pickard[66] speculate that ENSO may account for anomalous currents in their data during the 1982–1983 El Niño.

Figure 3.8 *Southern Equatorial Current (SEC) pathways to the Coral Sea: North Fiji Jet (NFJ), South Fiji Jet (SFJ), North Vanuatu Jet (NVJ), South Vanuatu Jet (SVJ), North Caledonia Jet (NCJ), South Caledonia Jet (SCJ). Once the streams approach the GBR the flow bifurcates to form the East Australian Current (EAC) and a northern arm, called the Hiri Current or North Queensland Current. This branch then feeds the New Guinea Coastal Current (NGCC) that feeds the West Pacific warm pool and is a source for the Equatorial Under Current (EUC). The currents in orange indicate major seasonal changes during the NW Monsoon. The NGCC reverses and the Southern Equatorial Counter Current (SECC) reverses the SEC nearest the equator. (Adapted figure prepared by SPICE[c], with alterations by the author for flows in the Coral, Solomon and Bismarck Seas)*

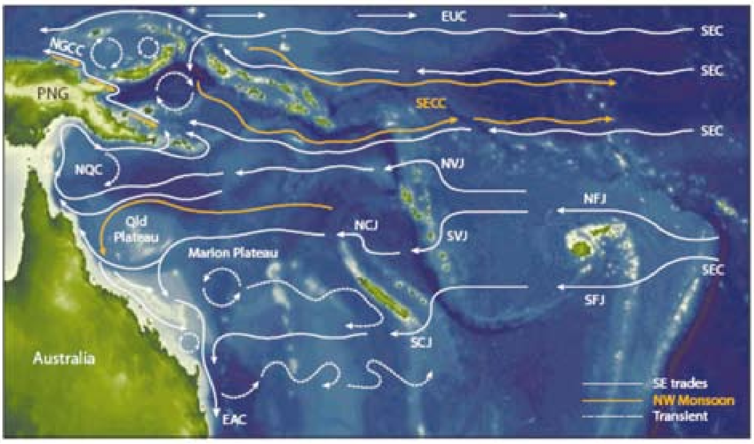
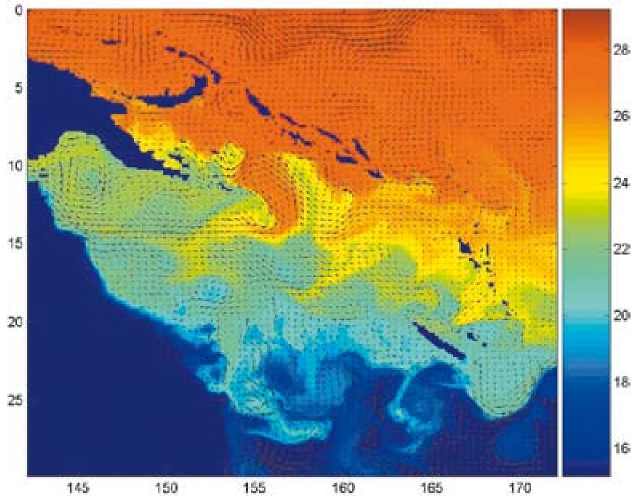


Figure 3.9 *Snapshot of the surface circulation of the Coral Sea from the OFAM model (run spinup4) for 27 February 2002. Arrows indicate current strength and direction, the background colour indicates model temperatures (model data courtesy Bluelink[d])*



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c http://www.ird.nc/UR65/SPICE/

d http://www.cmar.csiro.au/bluelink

Projected:
Climate modelling by Cai et al.[15] found that the southern EAC will strengthen in the Tasman Sea due to the Southern Annular Mode causing lighter mid latitude winds and stronger southern ocean westerlies. Observations by Roemmich et al.[50], who analysed 10 years of Argo floats and satellite altimetry, have found that the South Pacific Gyre has spun up over the last decade due to increased southeast trade winds, although it appears to be subsiding in recent years.

Certainty:
Low – Medium

Regional detail:
The variations in the strength and breadth of the SEC are critical to understand given that it is the main driver of the Coral Sea circulation. Whilst the SEC shows only a small seasonal variation, the relative contributions of the various zonal jets entering the Coral Sea will vary the location of the bifurcation and hence the relative strengths of the EAC and Hiri current.

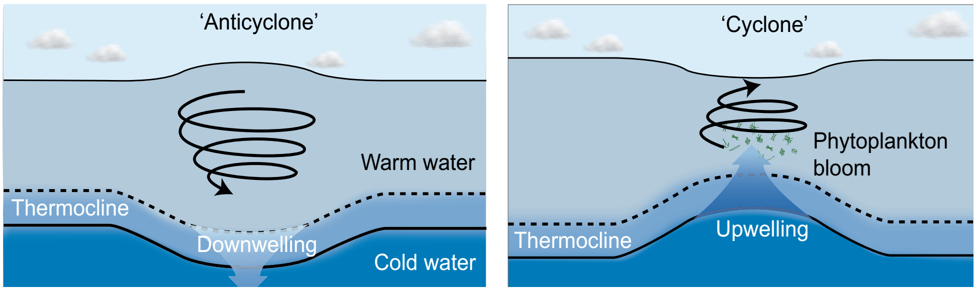
3.4.2 Eddies

The majority of kinetic energy in the ocean resides not in the steady ocean basin gyres but in eddies. These eddies are embedded in the larger scale currents and are therefore a large source of current variability in the ocean[37]. Eddies can be formed by a number of mechanisms, including baroclinic flow instabilities in the wake of islands or variations in wind fields. Energetic currents often become hydrodynamically unstable, creating meanders, which eventually shed eddies. These isolated eddies can last for years and can be transported with the prevailing flow of the major current systems. For example, in the Tasman Sea the EAC forms a meander and anticyclonic eddies pinch off from the main current moving south past Tasmania[24]. Stammer et al.[55] have mapped the global occurrence of eddies from over a decade of satellite altimeter data and have found that there has been a general decline in activity in the Western Pacific Ocean but an increase to the east of Australia. This coincides with the area of the most rapid warming in the ocean[61]. Thus any change in the intensity of the EAC upstream, in the Coral Sea is considered to be a factor leading to the warming. Eddies in the Coral Sea are found to have an annual cycle with a maximum in the austral summer and minimum in winter, with a period of 70 to 80 days.

Eddies have cyclonic (clockwise in the Southern Hemisphere) or anticyclonic (anticlockwise) senses of rotation. An analogous system can be found in atmospheric circulation around high and low air pressure systems. For example, weather charts show anticlockwise movement of air around a central high pressure. In both cases, higher pressure or sea level lies to the left of the direction of flow in the Southern Hemisphere due to the pressure gradient forces balancing the Coriolis force. Figure 3.10 shows a cross-section of oceanic eddies. The anticyclonic eddy has a convex surface with higher sea level at the centre whereas a cyclonic eddy is characterised by a concave surface with lower sea level at the centre. The anticyclonic eddies have warm cores due to the convergence of warmer surface waters at the centre which deepens the thermocline. In contrast, cyclonic eddies are characterised by a cold core due to the divergence of surface water allowing the thermocline to dome upwards. This can allow increased local productivity by bringing nutrient rich water into the euphotic layer. These eddies can often be seen in sea surface temperature[12] and ocean colour imagery of the Coral Sea.

Figure 3.11 shows a cyclonic eddy (with a chlorophyll-a signature in the centre) on the shelf edge near Hydrographers Passage, east of Mackay. The ocean circulation model (Figure 3.9) also shows a cold core eddy embedded in the flow east of the GBR at about 18° S.

Figure 3.10 *Cross section of an anticyclonic (left) and cyclonic (right) rotating eddy showing the respective convex and concave displacements of the sea surface and thermocline*



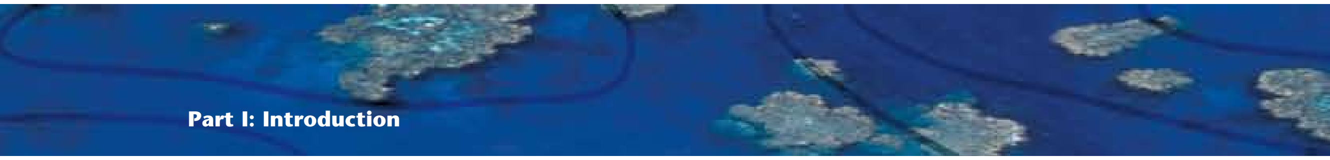
Projected:
With any increase in the SEC, eddy activity is also expected to increase. Perturbations to thermocline are likely to increase in magnitude.

Certainty:
Medium

Regional detail:
Eddies expected to form with increasing frequency from the same topographic features that generate them however other mechanisms are also at play. Eddies impacting on the GBR can also affect the location of the bifurcation between the EAC and Hiri currents.

3.4.3 Rossby waves

Another source of variability are Rossby waves. These are formed from the transient adjustment of ocean circulation to changes in wind and thermal forcing at the sea surface[18]. They can propagate westwards at less than 10 cm/s and can take months to decades to traverse the Pacific Ocean. They propagate fastest near the equator and are slower at higher latitudes. Whilst they have small surface amplitudes of about 5 cm, the thermocline can be displaced by over 50 metres, and impact shelf edge mixing and transport along the GBR. Due to their small amplitudes, they have only recently been able to be detected from satellite altimetry although the theory of their existence is well established[29].



Projected:
Rossby wave activity is expected to increase, perturbations to thermocline are likely to increase in magnitude.

Certainty:
Low – Medium

Regional detail:
Westward propagating waves are not expected to speed up, however, their amplitudes may increase. Their propagation in to the Coral Sea is complicated by topographic barriers and so regional effects are uncertain.

3.5 Great Barrier Reef currents

Inside the reef matrix there is a complex circulation due to the interaction of currents from tide, wind, continental shelf waves, inflows from the SEC and the physical barrier of the reefs themselves. The following sections describe the major characteristics of these flows along the GBR.

3.5.1 Bathymetry

The topographic complexity of the GBR significantly influences circulation and mixing on the shelf. The continental shelf in the far northern and northern GBR is a relatively narrow 50 to 70 km for most of its length with the exception of Princess Charlotte Bay located at latitude 14° S and toward Torres Strait at 10° S where it widens to over 150 km. The shelf is relatively shallow, gradually deepening to 40 to 60 metres toward the shelf edge. In the central region the shelf gradually widens from 18° S to about 110 km and sloping to a depth of around 100 metres. The GBR is widest in the southern region at latitude 21° S, near Broad Sound at around 250 km. There is a sudden narrowing south of the Swains in the southern region to approximately 60 km width where the outer shelf reverts to a relatively shallow 40 to 50 metres. The Capricorn channel at a depth of 90 metres extends from the southeast where the shelf narrows just south of the Swains, forming a trough in the lagoon separating the inner and the outer half of the shelf where there Swains and Pompey reef complexes are situated.

The Great Barrier Reef lives up to its name in the far northern and northern regions where long 'ribbon' reefs are oriented along the shelf, and in the southern region where the outer half of the wide shelf is occupied by a 'barrier' reef matrix. The reefs cover over 90 percent of the outer shelf leaving only narrow channels for oceanic and tidal flows to pass[48,67]. The central region is characterised by a more open reef matrix and together with a change in orientation of the shelf from north to south to a south-easterly orientation at a latitude around 19° S, allows the southward flowing EAC to penetrate directly into the GBR lagoon[7]. To a lesser extent, passages in the northern GBR allow Coral Sea water to flow through Trinity and Grafton passages, offshore from Port Douglas and Cairns. In this region the Ribbon Reefs give way to a less dense reef matrix to the south.

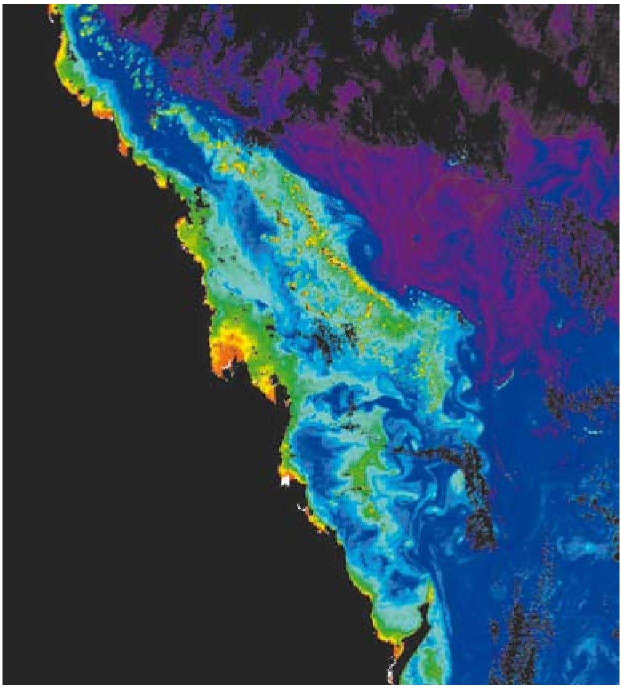
3.5.2 Western Boundary Currents: EAC and Hiri Current

The southward surface flow of the EAC in the central GBR peaks in November to December and is at a minimum in April to May due to the opposing southeast trade winds[9]. The EAC however is not considered fully developed until it reaches a latitude of 26° S where the rest of the branches of the SEC, such as the South Caledonia Jet converge at the Queensland continental shelf and contribute to the EAC.

The Hiri current[33] is an equatorward low-latitude western boundary current and is fully developed north of 14° S. It is guided by topography around the perimeter of the Gulf of Papua and the majority eventually flows around the Louisiade Archipelago to the Papua New Guinea northern coast in the Solomon Sea. The Louisiade Archipelago is an extension of the shelf islands and reefs from the southeast of Papua New Guinea. Some of the Hiri current recirculates as the Papua Gyre back to the far northern GBR[57,2,8].

These major current systems also drive along-shelf flows on the continental shelf that remain in geostrophic balance and so are principally driven by cross-shelf sea level gradients setup by the currents[9,10,62]. The lagoonal branch of the EAC can often be tracked as a southeastward extending low chlorophyll tongue of oceanic water in satellite ocean colour imagery (Figure 3.11) eventually moving out through the Capricorn Channel in the southern GBR.

Figure 3.11 *MODIS Chlorophyll-a image of the central and southern GBR showing oceanic (blue) water intruding through Palm Passage and toward Capricorn Channel. Red is high Chlorophyll-a or turbid water; blue is low Chlorophyll -a water; and black is land and cloud (Image courtesy of AIMS Remote Sensing)*



The sea level and geostrophic pressure gradients set up by the currents also cause the thermocline to mirror these movements. As higher sea levels are found to the left of the geostrophic current flow in the Southern Hemisphere, the thermocline will fall. Thus a strong Hiri current depresses the thermocline and suppresses the ability of cooler deep waters to access the continental shelf, whereas the opposite is true for the poleward flowing EAC. The thermocline rises and sea level lowers along the continental shelf. Furnas and Mitchell{27} found that primary productivity is at a maximum at 21° S off the Swains and Pompey group of reefs where the EAC is well formed and large tides can force a current across the continental slope and shelf (Figure 3.12) assisting the delivery of deeper oceanic waters to the shelf and mixing to the surface. The pulsing of these currents therefore generates significant variations in the thermocline depth and control over any shelf margin intrusions into the GBR.

In the Capricorn Bunker Group of the southern GBR, the circulation along the continental shelf margin is dominated by the meandering EAC. The sudden narrowing of the shelf south of the Swains reefs allows the current to meander and regularly produces a clockwise gyre with a mean northwest flow on the outer shelf[31,12]. Figure 3.13 shows the warm EAC seaward of the Swains reefs heading south and turning northwest just north of Fraser Island. It can be seen entraining Capricorn Channel waters from the lagoonal branch of the EAC into the cyclonic gyre. A shelf break front can be seen which separates the oceanic and shelf waters. When the recirculation is strong, the thermocline will be lowered, suppressing upwelling along the shelf break and forcing warmer surface Coral Sea waters past the Capricorn Bunker Group of reefs.

Projected:
The EAC and Hiri currents are expected to increase in strength due to direct forcing from the SEC. Central GBR currents may weaken and reverse if the bifurcation moves south.

Certainty:
Medium – Modulations from ENSO and PDO will be significant.

Regional detail:
Location of bifurcation is likely to move according to the large-scale wind stress curl driving the SEC. Whether one current strengthens at the expense of the other remains uncertain.

Figure 3.12 *Cross section, looking north, of the boundary currents showing thermocline adjustment for accelerating currents: the northward flowing Hiri current (left) has sea levels rising along the GBR and coast and the thermocline deepening. The EAC (right) shows sea levels dropping at the coast and the thermocline rises allowing waters to upwell onto the shelf.*

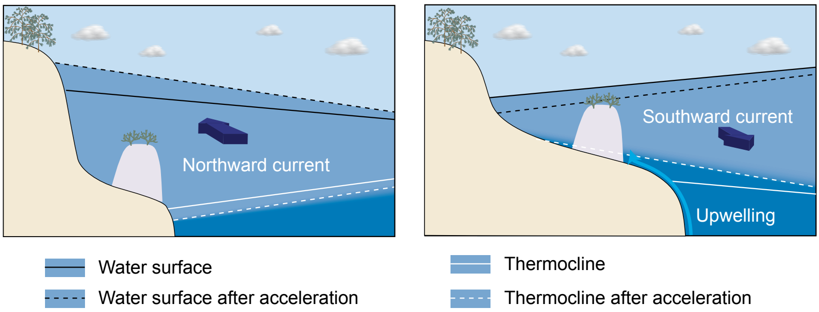
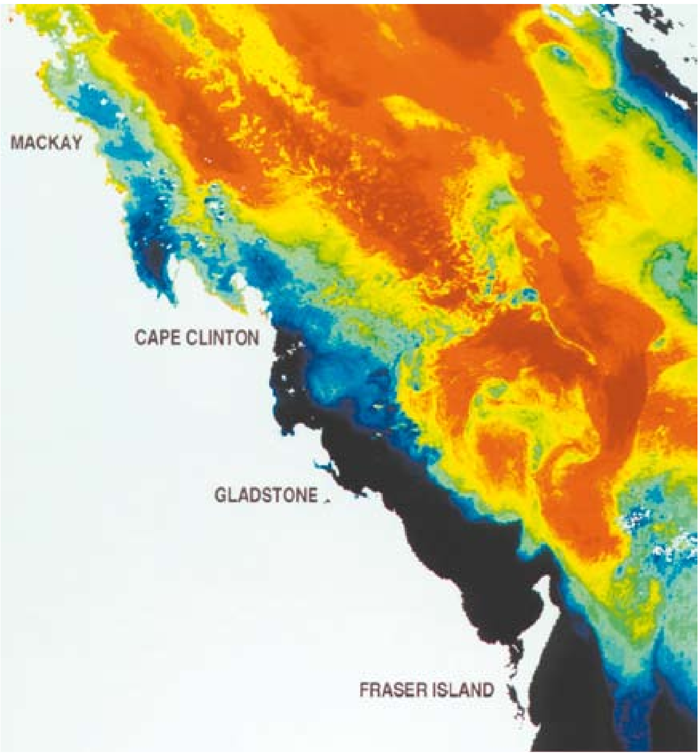


Figure 3.13: *Sea surface temperature image of the southern GBR showing the EAC re-circulation off the Capricorn Bunker Group*



3.5.3 Wind generated currents

There are two main seasons in the sub tropical GBR and they are characterised by the south-east trade winds, that are prevalent from April to November encompassing the austral winter and the variable south-west monsoon, usually active from December to March during the summer. Low frequency winds are highly correlated spatially over the majority of the GBR[62], especially during the south-east trades which commonly reach 25 to 30 knots. The winds reinforce the northward flowing Hiri current and oppose the poleward flowing EAC causing a seasonal minimum in transport and occasional surface flow reversals. Closer to the coast the south-east trade winds dominate the inner shelf resulting in well mixed northward coastal currents[14]. Modelling studies by King and Wolanski[39] and Brinkman et al.[7] in the central GBR suggest there is a disconnect between well mixed, coastal wind forced currents moving north and mid-lagoon EAC driven currents moving toward the south. Overall, southeast trade winds force surface waters on-shore and will suppress upwelling along the coast and shelf edge. During the north-west monsoon, winds tend to be less consistent and lower in strength. Episodic coastal upwelling can occur to replace the surface waters transported offshore by the wind. Any change in the strength of the Asian Monsoon (see Lough chapter 2) will be a major determinant on the relative roles of the seasonal winds experienced in the GBR.

Projected:
Climate models predict stronger southeast winds are expected in the Coral Sea.

Certainty:
Medium – Modulations from the Asian monsoon and ENSO will dominate variability.

Regional detail:
Coastal boundary layer to extend seaward. The Hiri current will strengthen with the local south-east trades. The EAC will exhibit more pulsing due to these winds impeding the poleward flow.

3.5.4 Continental shelf waves

Continental shelf waves propagate freely into the GBR from distant meteorological forcing in the south[32,22]. They are subject to scattering into higher order modes as they reach Fraser Island and travel past the Capricorn Bunker group in the southern GBR[30]. Continental shelf waves are also generated locally along the GBR from atmospheric pressure and wind forcing. They produce relatively small currents of 10 to 20 cm per second, but can transport water large distances and are a major source of current variability on the continental shelf[63,44,9,14]. Shelf waves also displace the thermocline vertically along the shelf break and can lead to intermittent cold water intrusions as they propagate past reef passages[1].

Projected:
Changes to continental waves propagating from the south into the GBR will be due to any change in occurrence and intensity of weather systems in the south. The observed increase in the Southern Annular Mode affects those weather systems. How far these waves propagate into the central GBR remains unclear.

Certainty:
Low

Regional detail:
GBR generated waves are likely to increase in strength assuming local atmospheric perturbations in the wind and pressure fields increase.

3.5.5 Tide

Tides dominate the sea level variations of GBR waters and are a major source of energy for mixing in the GBR. They are a mixture of semi-diurnal and diurnal tides along most of the shelf with the exception of the region around Broad Sound in the southern GBR. Here there is a significant topographic amplification of the semi-diurnal constituents (Table 3.1).

The sea level height and the strength of currents along the GBR vary according to shelf width and the degree to which the reef presents a barrier to tidal flow. Tides are generally small in the deep ocean but are amplified over the wide and shallow continental shelves. For example, the tidal range at Elusive Reef on the seaward edge of the Swains reefs (offshore from Broad Sound) is only 1.7 metres.

Table 3.1 *Mean spring ranges (defined as mean high water springs minus mean low water springs) along the GBR from north to south*

Location	Latitude	Range (m)
Harrington Reef	11° S	2.7
Cairns	16° S	2.1
Townsville	19° S	2.3
Mackay	22° S	4.6
Broad Sound	22° S	6.7
Gladstone	23° S	3.9

The exceptionally large tides of Broad Sound in the southern GBR are due to the wide shelf (250 km) and the barrier effect of the offshore reef complex[45,6]. This barrier forces tidal flows to go around the reefs, both from the south, up Capricorn Channel and from the north, through the sparse reef matrix off Townsville in the central GBR, where they superimpose to form a macro-tidal standing wave.

A feature of the tidal current is that it tends to have a significant cross-shelf component near the shelf edge and is an important factor in regulating and mixing any slope water intrusions and upwelling processes[58,69,31]. The horizontal excursion of tidal waters at the shelf edge is limited to only a few kilometres however the tides play an important role mixing the cooler waters towards the surface. Satellite sea surface temperature imagery shows these thermal signatures (Figures 3.6 and 3.7).

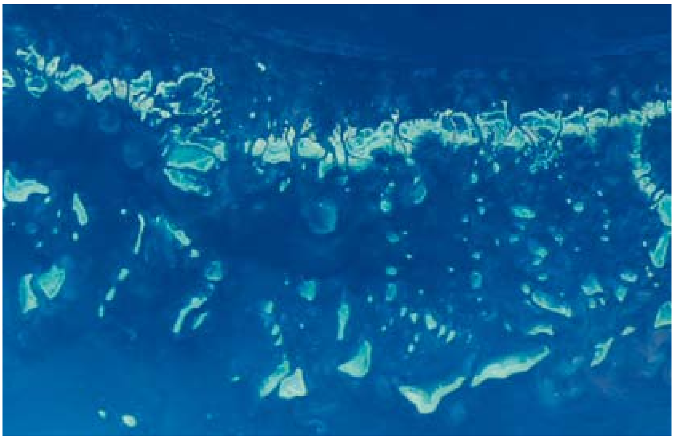
Tidal currents within the reef matrix also provide a source of enhanced mixing in the vicinity of the complex topography of the reefs. This causes flow acceleration and formation of tidal phase eddies in the lee of the reefs due to channelling of the flow (Figure 3.14). Eddies can then separate from the source reef and flow downstream influencing others in their path[62]. Predictable tidal currents are critical sources of mixing vertically around reefs especially during the summer monsoon season. Coral bleaching events are characterised by periods of high insolation and low wind speeds with alternative sources of surface mixing, to break down the stable surface layer, lacking.

Projected:
Gravitational tidal currents are not expected to increase significantly, however, as sea levels rise the tidal range will increase according to local shelf and coastal topography. Where waters can encroach on land, this effectively increases the shelf width, resulting in an amplification of the tidal range.

Certainty:
High

Regional detail:
Tides centred on Broad Sound in the southern GBR are likely to show the largest increase in tides along the GBR.

Figure 3.14 *Tidal phase eddies forming during a flooding tide after traversing narrow reef channels near Hydrographers Passage on the outer southern GBR*



3.5.6 River plumes

Consistent rainfall occurs in the wet tropics in the northern region where coastal mountain ranges provide the necessary uplift of the humid trade winds to produce rainfall, feeding the local rivers during the south-east trades. Two major catchments provide significant seasonal flows to the GBR during the monsoon and cyclone season (December to April) and are the main source of river plumes. The two major rivers are the Burdekin River in the central region and the Fitzroy River in the southern region. Freshwater plume dynamics control the direction of flow, with plumes turning northward at the river mouths (in the Southern Hemisphere) and following the coastline northward where the plumes are subject to mixing by wind and tide forcing[13]. Large flood events can bring flood waters to the outer reef in the narrow northern and far northern regions, and to a lesser extent in the central and southern regions where the shelf is wider[68,64,40]. Oceanic inflows through the Palm and Magnetic passages around 19° S inhibit cross-shelf surface flows of the plumes reaching the outer shelf reefs. A detailed review of riverine impacts on the GBR can be found in Furnas[26].

3.5.7 Upwelling

Throughout this chapter a recurring theme has been to identify processes that cause thermocline displacements along the GBR shelf edge. These displacements occur as a result of a large range of oceanographic processes: basin scale ENSO and PDO, EAC and Hiri current variability, impinging eddies and Rossby waves, tides, wind forced continental shelf waves and internal waves and tides on the thermocline itself[65,62].

Mechanisms that enhance the delivery of sub thermocline waters to the shelf include bottom generated Ekman layer currents[28], tidal induction[58], geostrophic pumping[47] and favourable winds and currents. The intruded waters may penetrate to the GBR lagoon but remain subsurface, such as found in the central GBR[3]. Waters however can mix upwards around the coral reef fringes within the reef matrix assisted by vertical mixing provided by the tides, wind and wave activity.

These processes affecting the depth of the thermocline are important to the health of the GBR as they control the relative amount of warm oligotrophic surface waters or cool nutrient rich waters that reach the continental shelf from the Coral Sea[4,41]. The intrusions from below the thermocline enhance primary productivity and alleviate heat stress experienced during coral bleaching events. Andrews[1] found that shelf break waters can be 1 to 4.5°C cooler than the surface lagoonal waters.

Projected:
Highly variable and episodic. Shallowing of the thermocline due to increased stability of the surface may allow the thermocline to lift above the shelf edge. Increased southeast trades will be less favourable. ENSO and PDO effects remain unobserved.

Certainty:
Low

Regional detail:
If the SEC bifurcation moves southward, the northward flowing Hiri current will deepen the thermocline resulting in the central and southern section reefs experiencing a reduction in nutrients and warmer waters. A waxing Hiri current and waning EAC result in a deepening thermocline along the entire GBR.

3.6 Conclusions

Climate change will affect GBR circulation patterns, the stability and depth of the surface mixed layer and the depth of the main thermocline. All these processes play an important part in regulating heat, connectivity, productivity and exchanges with the atmosphere. The heat content of the ocean is a fundamental environmental variable and influences the health of the GBR ecosystem. The Coral Sea also plays an important role in determining regional climate systems beyond the GBR. Northern Coral Sea waters through the Hiri current feed the WPWP and in turn the equatorial current systems that determine ENSO. The southern branch, EAC, sends warm tropical waters poleward affecting the climate along the eastern seaboard of Australia.

Given the sparse number of observations of the GBR and Coral Sea, it is therefore important to encourage initiatives such as CLIVAR's Southwest Pacific Circulation and Climate Experiment (SPICE) and dedicated regional array on the GBR that can monitor the EAC variability and structure over the longer term. The recent Australian Integrated Marine Observing System initiative goes some way toward achieving these goals for the GBR. Without these dedicated systems for long-term accurate measurements, detection of climate related change in oceanographic processes will remain unresolved or uncertain. Further modelling studies are needed to provide hypothesis testing on local affects of climate change through downscaling from global predictions.

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