The Dynamics of the Leeuwin Current during the middle and late Quaternary

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Except where otherwise acknowledged in the text, this thesis represents original research by the author

Michelle Ianthe Spooner
Acknowledgements

"We know more about the 'dead seas' of Mars than our own ocean." I was reminded of Jean-Michel Cousteau’s famous quote in a lecture by Patrick De Deckker as an undergraduate at ANU. Patrick grabbed my attention and I wanted to know more about the Earth’s oceans and their influence on our climate, so began my love and appreciation of Marine Geology. I have a huge amount of gratitude for my supervisor Professor Patrick De Deckker for his continual patience, guidance and support, not just through my PhD but over many years at ANU. Because of your belief in me and your vision, I have experienced many wonderful opportunities over the last few years. Thank you for plucking me from obscurity when I was an undergraduate. I hope we will always be able to share a good bottle of red together in the future.

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Azreal, 2005

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Abstract

This thesis details the dynamics of the Leeuwin Current, specifically trying to determine the nature and timing of oceanic events that prevailed off the western and southern coastlines of Australia during the last 550,000 years. The Leeuwin Current is an anomalous eastern boundary current transporting warm, low salinity water formed within the Indonesian Throughflow. The influence of this current extends from North West Cape in Western Australia to (at times) the southern tip of Tasmania. The strength of the Leeuwin Current is seasonal and has a temporal variability due to variation in the along-shore pressure gradient and prevailing equatorward winds. It is believed the Leeuwin Current operated differently during glacial and interglacial periods due to an alteration in the forcing mechanisms; this has implication for the regional oceanography and climate.

Data were obtained from marine cores located below the present-day pathway of the Leeuwin Current, core MD61 located on the shelf edge offshore Western Australia (113°28.63'E, 22°04.92'S) and core MD2607 (137°24.39'E, 36°57.64'S) located on the upper continental slope offshore South Australia. An additional core, located offshore Sumatra (BAR9403, 5°49.20'S 103°61.90'E) was also examined to understand possible forcing mechanisms that may alter the past dynamics of the Leeuwin Current. Planktonic foraminifera assemblages, the δ18O and δ13C of near-surface dwelling foraminifera, and sea-surface temperature estimates were used to reconstruct the vertical structure of the water column through the past ∼550,000 years. The forcing mechanisms behind the strength of the Leeuwin Current can also be inferred from these data.

These findings indicate a weak Leeuwin Current was present on the western coastline of Australia during glacial periods but it did not reach the southern core site during Marine Isotope Stage 6. However, there is some evidence of the Leeuwin Current reaching the southern core site during the last glacial maximum due to the presence of transitional water.

There was a greater influence of South Indian Central Water at the site of core MD61 offshore Western Australia due to the northward migration of the Indonesian Throughflow Water /South Indian Central Water frontal system by 3-4° which also resulted in a 6-9°C decrease in SST, a thickening of the mixed layer and the dominance of transitional species during glacial periods. The dominance of South Indian Central Water also suggests that the West Australian Current, which presently sits below the Leeuwin Current, was strengthened during the glacial periods and aided in weakening...
the Leeuwin Current. At core MD2607, offshore South Australia, there was an increase in Subantarctic Mode Water and upwelling of nutrient-rich water due to the reduction/absence of the Leeuwin Current during glacial periods.

There is an upwelling signal in core BAR9403 (offshore Sumatra), at 14 000 yrs BP that disrupts the warm stratified water column, which was a feature between MIS 3 to present. The upwelling offshore Sumatra is timed to a regional pattern of an intensified Australasian Monsoon. Findings within cores MD61 and MD2607 also suggest a stronger Leeuwin Current at ~14 000 yrs BP; it was also stronger during Marine Isotope Stage 5, 7, and 11 due to a thicker component of Indonesian Throughflow Water sourced from the Warm Pool, located in the equatorial Indian-Pacific region. This may also suggest an intensified monsoonal system during interglacial periods. Conversely, it appears that the Australasian Monsoon was generally weakened during the glacial periods which aided in reducing the flow of the Leeuwin Current and strengthening the West Australian Current.
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Chapter 1

Introduction

In recent years, the term ‘enhanced greenhouse effect’ has become well known in modern society and yet aspects of this possible ‘future climate’ are poorly understood. The characteristics of the geological record provide a basis to understand this possible future climate through the examination of alternating glacial (cold) and interglacial (warm) stages. We must understand the natural cycles of the oceans to understand the evolution of our climate through geological time. It is only through investigations into past oceanographic and climatic systems that we will be able to predict the likelihood of future climates. Recent scientific projects such as the World Ocean Circulatory Experiment (WOCE), the Tropical Ocean-Global Atmosphere (TOGA) study and the multiproxy approach for the reconstruction of the glacial ocean surface (MARGO) have identified the importance of palaeoclimatic and palaeoceanographic reconstructions within different climatic states. In addition, this information can be used to inform and guide long term environmental policies and planning and to predict the impact of climate change on land and our habitat (Kucera et al., 2005).

Understanding the variability of sea-surface temperature and biological processes in the water column provides insight into general circulation of present and palaeo-ocean currents. Sea-surface temperature can indicate if a current has been displaced by a colder or warmer current. A reduction of sea-surface temperature (SST) may also indicate the movement of deep, cold and nutrient-rich water to the sea-surface (upwelling), providing sites for increased primary productivity.

As the ocean and the atmosphere interact at the sea-surface, the palaeoclimate can also be inferred due to the coupled system of the ocean and the atmosphere. This is demonstrated in the equatorial region of the Warm Pool (WP), north of Australia, which is the largest single expanse of warm water on our planet with a sea-surface temperature (SST) consistently higher than 28°C (Yan et al., 1992). The high atmospheric temperatures and the high SST of the WP induce evaporation of oceanic water at the sea-surface and combined with the dynamics of the Australasian Monsoon causes increased precipitation in northern Australia. The water vapour and heat from these equatorial oceans is then transported to the middle and high latitudes generally affecting
the global climate (Broecker, 1994). Any changes to the SST or biological processes in the upper water column may suggest that this coupled system has changed in the past.

This thesis will focus on sea-surface processes, specifically trying to determine the nature and timing of oceanic events that prevailed off the western and southern coastlines of Australia during the last 550 000 yrs BP. The investigations are centred on the occurrence of the Leeuwin Current during glacial and interglacial cycles. The Leeuwin Current is a warm surface current of low salinity that flows over a distance of 5500km, from its origin at North West Cape (Western Australia) to the southern tip of Tasmania (Ridgeway and Condie, 2004). This is one of the very few extended zonal coastal boundaries found anywhere in the world and possibly the longest (Ridgeway and Condie, 2004).

The Leeuwin Current is part of the global circulation system transporting warm tropical water into higher latitudes, so understanding the dynamics of this system is important in terms of global responses to climate change. Past investigations on the Leeuwin Current have been inconclusive, especially in regard to the occurrence of this warm surface current during glacial periods. There is some evidence that the Leeuwin Current either operated differently or was absent during glacial periods due to a strengthening of the opposing West Australian Current and/or reduction of the along shore pressure gradient. Alternatively, during interglacial periods it is believed the Leeuwin Current may have been stronger transporting a greater amount of warm, low salinity water into the Southern Ocean. Either of these situations would have significantly changed the oceanography of the region and this in turn would affect the climate of the region. Understanding how the Leeuwin Current responded during extreme interglacials will help us understand what may happen in the region during an enhanced greenhouse climate.

The advent of deep-sea drilling in the 1950’s prompted the use of planktonic foraminifera (zooplankton) as palaeoceanographic indicators (Haslett, 2002). Planktonic foraminifera are protists and one of the most common groups of pelagic organisms in the open ocean today (Bé and Tolderlund, 1971). Planktonic foraminifera provide a natural archive of past environmental changes (Haslett, 2002), due to their global distribution through passive transport by ocean currents, their prolific productivity and sensitivity to environmental variations (Bé and Tolderlund, 1971). Therefore, species abundances will change in accordance with physical-chemical parameters such as SST, nutrients and salinity. Our knowledge of the ecology and present-day distribution of
planktonic foraminifera is the basis of palaeoceanographic reconstructions and a main experimental technique used in this study.

This study is based around the relative abundance of planktonic foraminifera, isotopic results from species living at the sea-surface and sea-surface temperature estimates using a modern analogue technique. This will provide information into the dynamics of the sea-surface layers and the presence or absence of the Leeuwin Current through time. This study is significant as the dynamics of the Leeuwin Current is not well understood and the data provided by this project will gain more insight into sea-surface interactions at a local and regional level. The Leeuwin Current is an anomalous eastern boundary current and we do not really know is how important it is in terms of climate change. Whether the Leeuwin Current has a significant impact on our climate in Australia is something that will be addressed in this study. Recent evidence during times of El Niño indicates that the Leeuwin Current does affect our climate in Australia (Meyers, 1996). Therefore, what are the effects, if any, of a reduced or absent Leeuwin Current?

Data obtained during the cruises of the *RV Franklin* in 1995 and 1996 and the *RV Baruna Jaya* in 1994 are the major sources of information in the study area. A study by Wells and Wells (1994) used planktonic foraminifera for sea-surface temperature reconstructions for the past 130,000 yrs B.P along the pathway of the Leeuwin current. However, this study provided no insight into the abundance changes of planktonic foraminifera. More extensive research has been conducted by Martinez et al. (1998) and (1999) in the Throughflow, the area of the ‘Java upwelling system’ and along the flow of the Leeuwin Current to 32°S. However, Martinez et al. (1998) study concentrated on present day assemblages while Martinez et al. (1999) study only covered the last 50,000 yrs BP. Further south, recent foraminiferal studies have been conducted in the Great Australian Bight by Li et al. (1996) and Li et al. (1999). A palaeoceanographic study using planktonic and benthic foraminifera has been conducted by Almond et al. (1993) and indicated the removal of the Leeuwin Current during glacial periods but the study lacked a good age model so the timing of these events are uncertain.

This project’s advantage over preceding investigations is that cores will be sampled at much higher resolutions and over a longer time scale using cores from the most recent cruises in the area: TIP 2000 and AUSCAN. With the aid of the modern analogue technique I will also estimate the thickness of the mixed layer which has not been previously estimated in the region, this will indicate if upwelling has occurred during glacial periods. In addition, two cores are located at the Leeuwin Current’s
various latitudinal extremes. This study will be the first attempt to analyse the existence of the Leeuwin Current during glacial and interglacial periods along both the western and southern coastline of Australia. The hope was to determine synchronous data between the two cores and to indicate whether the Leeuwin Current was absent or reduced during glacial periods and enhanced or unchanged during interglacial periods. Therefore, this PhD will be the most comprehensive study of the Leeuwin Current to date and will provide new insights into the dynamics of this current during climate extremes.

1.1 Aims and Objectives of this Thesis

To verify the presence and behaviour of the Leeuwin Current, deep-sea cores were obtained from beneath the present-day pathway of the Leeuwin Current (see Fig. 1). These cores were taken during two expeditions with the Marion Dufresne. One core MD002361 (MD61) was taken during the TIP 2000 expedition of the Marion Dufresne in 2000, it is located at 113°28.63'E, 22°04.92'S, 1805 mbsl. The other Marion Dufresne core MD032607, 137°24.39'E, 36°57.64'S, 865mbsl (MD2607) was taken during the AUSCAN cruise in 2003 and is located in the region of the Murray Canyon system. Both cores represent the various latitudinal extremes of the present-day Leeuwin Current. Core MD61 is located at a latitude where the Leeuwin Current can be distinguished from the Indonesian Throughflow while core MD2607 represents the terminal end of the Leeuwin Current's pathway or end member where it turns into the South Australian Current. Both Marion Dufresne cores are long, high-resolution cores and it was hoped that they would capture several Marine Isotopic Stages (MIS) preceding the last interglacial. Isotopic results indicated that core MD61 capture rare isotopic events such as MIS 11 and 12 that have not been obtained in deep-sea cores in the area, while core MD2607 captures events up to MIS 7.

Another core BAR9403 (5°49.20’S 103°61.90’E, 2034mbsl) was taken off the southwest coast of Sumatra. Investigating this core may provide insight into general palaeoceanographic conditions in the north-eastern Indian Ocean that may link to patterns to be found within core MD61.

To define the history of the Leeuwin Current, foraminifera and geochemical analyses were performed to determine past ecological conditions in the water column and to gain insight into the interactions within the water column especially near the sea
surface. Isotopic analyses were also conducted to provide a chronology and possible changes to the nutrient cycles of the core locations.

1.2 Experimental Analysis

1) Radiocarbon dating ($^{14}$C) will be utilised to provide a good age model for core MD61.

2) The $\delta^{18}$O isotopic results from planktonic foraminifera will provide an ice volume effect and an age model beyond the limits of radiocarbon dating.

3) The $\delta^{13}$C of planktonic foraminifera (Globigerinoides ruber and Globigerina bulloides) will also be investigated in all cores for possible indication of productivity at the sea-surface.

4) The relative percentage abundance of planktonic foraminifera analysed through time will provide insight into the dynamics of the water column.

5) The relationship and ecological groups between planktonic foraminifera will be analysed via principal component analysis and factor analysis.

6) The relative abundances of planktonic foraminifera will be used to reconstruct past sea-surface temperature using a modern analogue technique developed by Dr. T.T Barrows.
Chapter 2
Regional Oceanography

The cores in this study were selected under the pathways of the Leeuwin and South Java Currents to understand the dynamics of these currents through time. This chapter defines the present day characteristics of the currents mentioned above. In addition, this chapter also explores the important features of other regional oceanic currents that feed, drive and thereby influence the dynamics of the Leeuwin and South Java Currents.

In the region between Indonesia and northwest Australia, a complex set of surface currents are mainly driven by seasonally changing monsoonal winds. Therefore, these surface currents vary in strength and direction seasonally and interannually (Wijffels et al., 2001). The development of large scale eddies have also been identified and affect the circulation within this region (Wijffels et al., 2002; Sprintall et al., 2002). Generally, the water that passes through the Indonesian Archipelago eventually spreads out into the Indian Ocean as the west-flowing South Java and South Equatorial Currents, and as the south-flowing Leeuwin Current (Godfrey and Ridgway, 1985; see Fig 1.). The currents within this study area will be discussed in sequence of occurrence from low to high latitudes, due to the inter-relationship of these currents.

2.1 Indonesian Throughflow

The flow of water from the Pacific Ocean into the Indian Ocean, called the Indonesian Throughflow (ITF), is an important oceanographic feature as it provides the thermodynamic link between these two major oceans. The ITF has a major impact on the mass, heat and freshwater budgets of the South Pacific and the Indian Oceans (Sprintall et al., 2002). The heat and freshwater carried by the ITF into the Indian Ocean is believed to aid in driving global circulation as it represents one of three global centers of tropical convection where warm, low salinity water is injected into the global circulation system (Kinkade et al., 1997). Locally, the ITF strengthens the Indian Ocean’s Leeuwin Current, the South Equatorial Current (SEC) and the South Java Current (SJC), while affecting the heat budget of the Indian Ocean (Godfrey, 1996; see Fig. 1).
Figure 1.
The location of cores and the surface-sub surface currents in the study area. The blue arrows represent cold water currents; conversely the red arrows indicate warm water currents. The green arrows represent currents that have some influence in upwelling areas such as off the coast of Java and the southern coastline of Australia. The West Australian Current is located offshore and flows northward under the Leeuwin Current. The small green arrows near the coast locate the Ningaloo (north) and Capes Current (south).

The hydrological characteristics of the Throughflow are due to the dynamics of the Warm Pool (WP), which is the largest single expanse of warm water on our planet with a sea-surface temperature (SST) consistently higher than 28°C (Yan et al., 1992; see Fig. 2). The high atmospheric temperatures in the region combined with the high SST of the WP induce evaporation of oceanic water at the sea-surface. This transfer of latent heat, combined with the dynamics of the Asian Monsoon and the Trade wind
system over the Pacific Ocean result in low-pressure cells and rain; the overall balance is a gain of freshwater from precipitation at the surface of the ocean. Consequently, SSTs are high and sea-surface salinities (SSS's) low (35.5‰) within the ITF (Tomczak and Godfrey, 1994).

In addition, the low-salinity ITF varies in strength with the Australasian monsoon. Maximum strength of the ITF is observed during the austral winter in August/September (Wyrtki, 1987; Molcard et al., 1996).

![Figure 2](image.png)

The annual global mean sea-surface temperature. Note the warm pool around Indonesia with temperature close to 30°C. Also note the cooler water off the western coastlines of southern Africa and South America compared to the warmer downward dipping contour off the coast of Western Australia due to the presence of the Leeuwin Current. Data from World Ocean Atlas 2001 (Conkright et al., 2001) illustrated with Ocean Data View (ODV; Schlitzer, 2005).

### 2.2 South Equatorial Current

The South Equatorial Current (SEC) flows strongly westward, advecting the relatively pure Indonesian Throughflow water (ITW; 34.4‰) from the internal Indonesian seas (Bray et al., 1997; Sprintall et al., 2002; see Fig. 1). The SEC can be traced as a tongue of low salinity water extending westward, that occurs only in the upper 150m of the water column (Rochford, 1966; Wijffels et al., 1996; Hautala et al 1996), but varies seasonally in strength, as it incorporates the seasonally varying...
Throughflow transport (Wijffels et al., 1996). However a recent study by Wijffels et al. (2002) has shown that the SEC is not only fed by the Throughflow, but also receives waters via recirculation from the eastward flowing South Java Current (SJC), the South Indian Central Water (SICW) re-circulating from the eastward-flowing Eastern Gyral Current (EGC) and an un-named eastward flow of mixed intermediate waters also flowing eastward between 16°S and 20°S (Wijffels et al., 2002).

The union of a strongly westward flowing SEC with an intensified eastward flow within the EGC was observed during the World Ocean Circulation Experiment (WOCE) IR6 cruise, September 1995 and maximum flow was identified during the SE Monsoon (Meyers et al., 1995; Bray et al., 1997; Sprintall et al., 2002). The work by Wijffels et al. (1996) showed that the SEC was absent in April, strong in September and weakened in November (being the transition between the monsoon ‘seasons’). In addition, coastal upwelling south of Java contributes ~2.4 Sv of water to the South Equatorial Current, during the SE Monsoon season (August–September; Wyrtki, 1962).

### 2.3 South Java Current

The South Java Current (SJC) flows along the Java shelf break and slope (Wijffels et al., 1996; see Fig. 1), and appears to be composed of two layers due to different source water and density differences. The warm, shallow SJC (above the thermocline) carries water that is fresher than the Throughflow (32%) that results from runoff in the Java Sea or along the west coast of Sumatra (Wijffels et al., 1996; Bray et al., 1997). The very high temperatures and low salinities of the waters found in the surface flow of the SJC have a significant impact on the total net transport. During all five transects conducted in the area, very fast surface velocities (16.7 Sv) were associated with the upper part of the SJC (Sprintall et al., 2002). The relatively saline deeper component (Bray et al., 1997) of the SJC, extending to at least 1000m, carries relatively saline North Indian Central Water (NIW) into the Throughflow region (Wijffels et al., 1996).

The Indian Ocean thermocline waters enter via the SJC to the north of the SEC and the subtropical Eastern Gyral Current (EGC) to the south. The SJC is believed to join the Throughflow within the Indo-Australian Bight (IAB) and exit in the SEC (see Fig. 1). Similarly the shallow eastward flowing EGC also must recirculate within the IAB and most likely exits westward in the SEC, with a small contribution to the southward flowing Leeuwin Current found at the West Australian shelf break (Sprintall et al., 2002).
While flow in the upper SJC varies in direction approximately every 3 months, mostly in accordance with the expected semi-annual signal (Quadfasel and Cresswell, 1992), a subsurface core of eastward flow near 300-400m along the coast of Java was always present (Sprintall et al., 2002). Due to the impact of the strong, eastward surface SJC and its deeper extensions, the transport in this boundary current largely balances the offshore westward flow in the SEC. The banded flow structure in the geostrophic currents along the southern end of the transect means that the cumulative transport is essentially zero at ~ 14°S (Sprintall et al., 2002).

Observations by various cruises have revealed the dynamic nature of the SJC. Generally, the shallow SJC is strongest eastward in November, weaker in September and possibly westward in April (Wijffels et al., 1996). During the SE Monsoon season (austral winter), the SJC flows from the southeast to the northwest and is primarily fed by the Indonesian Throughflow (Takahashi and Okada 2000; see Fig. 1). The Leeuwin Current, by this time of the year, is weak and the West Australian Current dominates circulation along Western Australia. Simultaneously, the sea-level difference between Java and Australia becomes large, and the Indonesian Throughflow is at its maximum (Tomczak and Godfrey, 1994). November marks a transition between the monsoons, and coincides with one of two semi-annual maxima in eastward transport of the SJC, the other being in May (Wijffels et al., 1996).

Another consequence of the SE Monsoon is the southward Ekman transport offshore the south of Java. This phenomenon causes coastal upwelling off the southern Java coast (Takahashi and Okada 2000). Low SST's (<15°C), and high inorganic phosphate (~0.5 mg atm/l) were recorded at 100 and 200m water depths in this area (Wyrtki, 1962). However, the upwelled deeper water does not seem to reach the surface even south of Java. Instead, the Indonesian Throughflow is intensified during the SE Monsoon season (Tomczak and Godfrey, 1994), and the capping effect of the Indonesian Throughflow Water (ITW) prevents upwelling from reaching the surface. Consequently, the phytoplankton concentration is low in both monsoon seasons (Takahashi and Okada, 2000) within the SJC.

During the late NW monsoon season (March–April), the SJC (derived from the Equatorial Counter Current) moves towards the southeast to meet the Leeuwin Current (Tomczak and Godfrey, 1994), strengthening this current. The semi-annually reversing SJC plays an important role in distributing freshwater into and out of the southeast Indian Ocean (Sprintall et al., 2002).
2.3.1 Eddy development

The development of cyclonic and anti-cyclonic eddy systems within the IAB has been captured by satellite imagery and measured surface dynamic heights between Indonesia and Australia. Low salinity of the ITW appears to enter the Indian Ocean in a series of eddies and bullets, interspersed by saltier water masses native to the Indian Ocean (Gordon et al., 1997; Bray et al., 1997). Some eddies were analysed during each of the World Ocean Circulation Experiment surveys (WOCE I10/IR6) in the SEC between 14-12°S. During both the April 1995 and September 1995 WOCE surveys the eddie fields were seen to be responsible for much of the velocity structure and mass transport of the SEC (Sprintall et al., 2002).

The direct current estimates support the findings that the strongest flow across all sections is found within the eddy structures. SADCP and LADCP velocity data suggests a strongly cohesive circulation within eddies from top to bottom (Sprintall et al., 2002). SADCP data show strong flow >80 cm/s on the western edge of anticyclonic eddy centered at ~10°S, advecting the very fresh cap found in the surface layer of the SJC southwards to 13°S (Wijffels et al., 2002; Sprintall et al., 2002).

Feng and Wijffels (2001) found that eddies within the SEC are seasonally modulated, occurring more frequently during July to September. Chlorophyll responses measured by Asanuma et al. (2003) also supports Feng and Wijffels (2001) model with synchronous blooming occurring at the onset of southwestery wind in March. Chlorophyll responses measured by Asanuma et al. (2003) also support the findings of Wijffels et al. (2002) and Sprintall et al. (2002) that velocity flow and mixing are important components of the eddy systems and current transport in the eastern Indian Ocean. For example, cyclonic eddy (A) measured by Asanuma et al. (2003) had a chlorophyll-α concentration of ~3.0mg m⁻³, diameter of approximately 30km and SST of ~27°C indicating the movement of deeper water up to the sea-surface and the eddy moved ~80 km from the exit of the Lombok Strait within four days (Asanuma et al., 2003).

2.4 Leeuwin Current off Western Australia

The Leeuwin Current, off the coast of Western Australia, is unique when compared to eastern boundary current systems of continents in the Southern Hemisphere. Most eastern boundary regions of southern oceans are synonymous with cold, nutrient enriched (upwelling), equatorward surface flow, and are large heat sinks
from the atmosphere (Tomczak and Godfrey, 1994). This, for example, is seen along the coastlines of Chile-Peru (Humboldt Current) and southwest African coast (Benguela Current; see Fig. 2). Instead, the Leeuwin Current transports warm, "fresh" tropical waters poleward from near the Indonesian Archipelago and from the Indian Ocean northwest of Exmouth, to south of Cape Leeuwin, where the current changes direction, due to Coriolis force, and moves across the Great Australian Bight (Cresswell, 1991; Wells and Wells, 1994; Murray-Wallace et al., 2000).

The SST varies from ~29°C in the north to ~21°C in the south in February, and ~27°C in the north to ~16°C in the south in August. This gradient steepens south of ~22°S at the southern boundary of the Warm Pool and the ITF–SICW front. Sea-surface salinity (SSS) has a similar distributional pattern (Martinez et al., 1998). Importantly, it is the only eastern boundary region with a large heat transfer from the ocean to the atmosphere, with values equivalent to a western boundaries current (Josey et al., 1999). The ocean–atmosphere heat flux is largest along the southwest Australian coastline where the annual heat loss exceeds 50Wm\(^{-2}\) (Morrow et al., 2003).

In the northern sector, the Leeuwin Current is wide, up to 440 m deep and sluggish, whereas south of the Northwest Cape the flow is narrower, shallower and stronger. The main core of the Leeuwin Current is typically 100–200 m deep, flows parallel to the coast at speeds up to 1ms\(^{-1}\) and usually tracks along the continental shelf edge (Cresswell and Peterson, 1993). On the northwest shelf at 19° 5'S, 116° 5'E, at ~100 m water depth, Holloway and Nye (1985) observed a southwest current pulse that was strongest from February and June. They linked this to the onset of the Leeuwin Current, which flows southward along the northwest shelf (Church et al., 1989; Creswell and Peterson, 1993). Creswell and Peterson (1993) also observed this signal in currents near 12°S, 120°E, in the Timor Sea, and indicated that the Indonesian Throughflow fed the Leeuwin Current. Furthermore, the strength of the current and its position along the continental shelf can also vary between years (Gaughan and Fletcher, 1997).

More recent studies such as Wijffels et al. (2002) and Sprintall et al. (2002) regard the Leeuwin Current as being indirectly affected by the ITF and the variations in its strength are complicated by the interactions with the SJC and SEC as explored above. However, the influence of the Indonesian Throughflow on the Leeuwin Current is conspicuous during non-El Niño Southern Oscillation years when additional warm water passes through the Pacific-Indian Ocean gateway in Indonesia (Wells and Wells, 1994; Hantoro, 1997; Murray-Wallace et al., 2000).
There is some disagreement on what causes the variability in the Leeuwin Current. Some studies suggest the northward wind stress during the austral summer causes the most variation (McCreary et al., 1986; Kundu and McCreary, 1986; Smith et al., 1991). Other studies promote the meridional pressure gradient at the shelf break as the main forcing mechanism (Thompson, 1984, 1987; Godfrey and Ridgway, 1985; Weaver and Middleton, 1989; Batteen and Rutherford, 1990; Godfrey and Weaver, 1991; Morrow and Birol, 1998; Feng et al., 2003). Recently, Ridgway and Condie (2004) confirmed that the along shore pressure gradient was the dominant force driving the Leeuwin Current along the western coastline of Australia but also acknowledged that the seasonality in the strength of the Leeuwin Current is due to variations in both along shore pressure gradient and the coastal wind stress (south-westerly winds).

The cause of the anomalously large along-shore pressure gradient is probably due to the passage from the western equatorial Pacific Ocean to the eastern equatorial Indian Ocean through the Indonesian Archipelago (Godfrey and Ridgway, 1985; Godfrey and Weaver, 1991; Smith, 1992). This mechanism is supported by Godfrey (1996) who observed "A low sea-level difference between the Pacific and Indian Oceans reduces the strength of the Indonesian Throughflow and consequently the strength of the Leeuwin Current." The alongshore pressure gradient is a gradient of geopotential anomaly at the sea-surface. This gradient has a value greater than 2 x 10^{-6} m s^{-2}, which means an alongshore slope upward toward the equator of about 20 cm per 1000 km, which is anomalously large compared to other eastern boundary regions (Smith, 1992). The model study of Godfrey and Ridgway (1985) showed (1) that the alongshore pressure gradient had a larger seasonal amplitude than the wind stress; (2) that the two terms have similar phases so their seasonal cycles reinforce one another (3) that in April-July, the pressure gradient is much stronger than the opposing wind stress and coincides with the time of the year the Leeuwin Current is strongest. Morrow and Birol (1998) have also suggested a secondary maximum of alongshore pressure gradient in November. All of these factors result in the Leeuwin Current having a maximum southward geostrophic speed of 45 cm s^{-1} during April-May, while its maximum southward volume transport of 5 Sv occurs in June-July (Feng et al., 2003).

Generally, the flow of the Leeuwin Current is strong enough to overwhelm the upwelling tendencies of the coastal wind stress; therefore, upwelling via Ekman transport does not occur (Reason et al. 1999). However, due to seasonal differences in equatorward wind stress along the west coast of WA, the Leeuwin Current flow is weaker between October and March (Pearce and Pattiaratchi, 1999). The general pattern
of wind change in relation of the strength of the Leeuwin Current is described in detail in Godfrey and Ridgway (1985) in the following paragraph. “In winter (July), strong westerlies are found south of 30°S; while the region north of 20°S is dominated by vigorous Southeast Trades and moderate southerly winds occur near the shelf edge in 20-30°S. From July to September wind stress generally veer northward and northward coastal winds increase, this tendency is repeated from September to November. In November the monsoon changes from Southeast to Northwest and by January the NW Monsoon is well established out to sea north of 20°S and longshore winds are very strong south of 20°S. In March the monsoonal winds again reverse; onshore coastal southerly winds remain strong between 20-35°S. However, by May the winter wind pattern is re-established and there is a strong decrease in equatorward wind stress occurs (Godfrey and Ridgway, 1985).

The Leeuwin Current restricts the eastern arm of the Indian Ocean gyre to offshore regions (the West Australian Current), generating large-scale downwelling as it travels along the continental shelf break (Pearce, 1991; Smith et al., 1991). Suppression of upwelling due to an alongshore pressure gradient is unusual in other eastern boundary regions, but has been observed off Peru during El Niño phases (Huyer et al., 1987; Smith, 1992). Alternately, the Leeuwin Current has been observed to be weakened during an El Niño year due to coastaly trapped waves (Clarke and Lui, 1994; Meyers, 1996; Wijffels and Meyers, 2003; Feng et al., 2003). It is postulated that without the Leeuwin Current, the prevailing equatorward wind, driving the northwards Capes Current along the Western Australian coastline, would almost certainly produce strong coastal upwelling due to offshore Ekman transport. This would be analogous to the present-day currents off the coasts of Chile and Peru and south-eastern Africa, and would be associated with enhanced productivity (Suess and Thiede, 1983; Veeh et al., 2000).

High-resolution satellite imagery shows the surface structure and variability of the Leeuwin Current and its associated meanders and eddies in its passage southwards down the coast and then eastwards towards the Great Australian Bight (see Fig. 3). Most long-lived anticyclonic eddies are generated in May–June, when the Leeuwin Current is most intense, and at three preferred locations: 20–21.5°S, 24–25°S and 28°–31°S (Fang and Morrow, 2003). Eddies were observed some 350km offshore from the Leeuwin Current in September 2000, around 25–30°S, and their vertical structure and temporal evolution show they were formed at the coast in May, when the Leeuwin Current is strongest. After their separation from the current, the warm-core eddies drifted WNW
following isopycnal contours, and was strongly steered by bathymetry. These eddies penetrated down to at least 1500m depths; the strongest eddy, Eddy B, influenced isopycnals to 2500 m (Morrow et al., 2003).

Figure 3

2.5 Ningaloo Current

Inshore of the Leeuwin Current, an equatorward coastal counter current is driven by upwelling-favourable southerly winds which prevail during the austral summer (December–March). A combination of wind-forced shelf currents and localized Ekman-
driven upwelling (Gersbach et al., 1999; Woo et al., in review) generates the Ningaloo Current (NC) which extends along the Gascoyne continental shelf, from North West Cape (21°.31'S) to Shark Bay (26°.51'S, Hanson et al., 2005) and seasonally restricts the inshore extent of the Leeuwin Current (Pearce and Pattiaratchi, 1999; Taylor and Pearce, 1999).

Generally, the prevalence of downwelling along the coast of WA is thought to prevent deep nutrient concentrations from reaching surface waters (Pearce, 1991). However, it is believed that nutrients within the Ningaloo Current are most likely sourced from the nutricline at the base of the Leeuwin Current (Woo et al., in review; Hanson et al., 2005). The nutricline within the LC occurs between 70 and 200m deep (Pearce, 1997; Hanson et al., 2005). The depth of the Leeuwin Current's nutrient-depleted mixed layer governs nutrient concentrations within upwelled water, as suggested by Gersbach et al. (1999) in relation to Capes Current upwelling off southwestern WA (Hanson et al., 2005).

Hanson et al. (2005) hypothesize that the biological impact of any upwelling on the inner shelf would be a function of: (a) the depth of the Leeuwin Current's nutrient-depleted mixed layer, (b) the strength and duration of upwelling-favourable winds (i.e. the intensity of upwelling), and (c) geographical location, primarily with respect to the width of the continental shelf and resultant proximity of upwelling flows to deep nutrient pools. Hanson et al. (2005) study of productivity levels within the Ningaloo Current showed that even though the productivity levels within the Ningaloo Current were five times greater than offshore waters of the Leeuwin Current the productivity during the early summer of 2000/01 was considerably less than upwelling regimes along other eastern ocean boundaries.

### 2.6 West Australian Current

The West Australian Current (WAC) represents the eastern boundary current off Australia and consists mainly of salty South Indian Central Water (SICW) (Wijffels et al., 1996). The West Australian Current flows northward and anticlockwise beneath the Leeuwin Current (Tomczak and Godfrey, 1994; Murray-Wallace et al., 2000; see Fig. 1). At a 200 to 2000 m water depth, the SICW flows northeast and returns west beneath the SEC (Wijffels et al., 1996) This current in turn influences water masses as deep as 2000 m (Tchernia, 1980), and is part of a major southern hemisphere subtropical gyre,
moving anticlockwise in the Indian Ocean. The WAC is strongest in September–November (Wijffels et al., 1996), when the Leeuwin Current is weakened.

Figure 4
A set of sea-surface temperature satellite images of the Leeuwin Current showing the strength of the Leeuwin Current during April as it enters the Great Australian Bight along the shelf break.

2.7 Capes Current

Satellite imagery and water sampling methods have provided evidence for a cool, seasonal, inner shelf current which flows northwards against the alongshore steric
height gradient in the summer months when the equatorial wind stress is at maximum strength (Gersbach et al., 1999). The strongest flow of this seasonal current is between Cape Leeuwin (34°20'S) and Cape Naturaliste (33°30'S) and is therefore named the Capes Current. The Leeuwin Current tends to flow closer inshore between the Capes in the winter months, and then moves offshore in summer as the southerly wind stress drives the Capes Current northward along the coast (Pearce and Pattiaratchi, 1997).

When the Leeuwin Current is weaker, the localized upwelling seems to occur in the southwest coast of Australia (Pearce and Pattiaratchi, 1997), due to the Capes Current and as with the Ningaloo Current it is believed that the source off upwelled nutrients is from the base of the Leeuwin Current’s nutricline (Gersbach et al., 1999). The low nitrate/high productivity signature associated with this flow is consistent with an aging upwelled water mass (Dugdale et al., 1990). This is supported by observations of low silicate levels and a high proportion of centric diatoms within Capes Current surface waters (Kudela et al., 1997; Tilstone et al., 2000; Hanson et al., 2005).

2.8 Leeuwin Current off the Southern Coastline

On the south coast the Leeuwin Current is driven by westerlies in winter (Ridgeway and Condie, 2004). As stated previously satellite and ocean buoy data reveals that the Leeuwin Current moves eastward at Cape Leeuwin, along the southern margin of Australia into the Great Australian Bight (GAB; see Fig. 4). The current mixes with warmer and more saline waters from the Bight to form a modified water system, the Great Australian Bight Current (Rochford 1986; James et al. 1997), and continues flowing eastward. Downwelling of warmer and more saline waters into the deeper ocean has been observed especially from the Bight and off the Spencer Gulf mouth (James et al. 1997; Li and McGowran, 1998).

Together with the warm waters of the GAB, the current extends its flow southeasterly, across the Lincoln Shelf, until being diffused into offshoots near Kangaroo Island (see Fig. 5). As a result the southern shelf water is relatively warm about 21°C (19°C in winter) in the west and 17°C (15°C in winter) in the east (Legeckis and Cresswell, 1981).

Analysis of a series of SST images representing the seasonal cycle of SST within the GAB reveal that this band is the combination of two water features; the Leeuwin Current which flows along the shelf edge in a narrow band to approximately 130°E, and a warm water mass within the GAB which is found over a broad region of
the shelf eastward of this longitude (Herzfield, 1997). This W-E temperature gradient was confirmed during a RV Franklin Cruise in July 1995 (Li et al., 1999). A warm pool develops in the shallow western GAB due to surface heating (Herzfeld and Tomczak, 1997), appearing in October and intensifying throughout the summer and autumn (Herzfield, 1998). The maximum eastward extent of this GAB water is greater than 136°E in autumn, and the maximum SST is 2-3°C higher than the surrounding water (Herzfield, 1998). The vertical temperature gradient varies little in summer but increases significantly during winter months due to much cooler to cold bottom water (<10°C is constant at >400m depths), particularly offshore from Albany and Esperance and in the central part of the Bight (Li et al., 1999).

The warm saline waters of the eastern extend of the Leeuwin Current contrasts with both the cool and less saline oceanic water and the warmer and more saline gulf water (Li et al., 1996b). Gulf waters of the Spencer and adjacent St Vincent Gulf are much warmer and more saline (Nunes and Lennon, 1986; Li et al., 1996b; see Fig. 5). Gulf-shelf water exchange is weaker in summer because of the existence of SST fronts at gulf entrances (Petrusevics, 1993).

Prevailing swells flow northeast towards the coast and mixing of water masses occurs on the Neptune Sector of the shelf (Li et al., 1996b). Rochford (1984, 1986) described the interchange of waters in the region, but to what degree these two water systems interact, as well as the potential impact on biotic distribution and sediment accumulation, is poorly known (Li et al., 1999).

### 2.9 Flinders Current

As stated above, the circulation along the southern shelf is mainly wind driven. In the winter, the flow is towards the East, and in summer, for the most part, it is towards the West (Rochford, 1957; Bye, 1986). Thus it is apparent that the Flinders Current flows as a countercurrent to the poleward circulation of the Leeuwin Current (Bye, 1998; see Fig.1).

Throughout the year, a positive wind stress curl is maintained over the South Australian Basin which drives a permanent deep sea anticyclonic gyre (Bye, 1998). The positive wind stress curl south of Australia leads to an equatorward Sverdrup transport in the deep ocean and the development of the Flinders Current which flows from east to west along the slope (Middleton and Platov, 2004). The flow of the Flinders Current is estimated at 15 Sv (Bye, 1972).
2.9.1 Bonney Upwelling

The southern margin of Australia is generally considered a downwelling coast with GAB waters having very low nutrient concentrations (Motoda et al., 1978). This is due to a weak shelf-break jet which is driven, especially in winter, by a combination of Leeuwin current inflow, onshore Ekman flux, alongshore winds and evaporation. The strong downwelling and consequent low surface nutrient levels off Western Australia result in a very oligotrophic ecosystem (Griffin, 1998). The GAB is sandwiched between this and the productive regions to the east, and also productive regions south of the subtropical front. Limited, transient upwelling does occur, however, but this is probably as variable as the wind, unlike the more continuous processes to the east (Griffin, 1998).

However, the development of large seasonal coastal upwelling that establishes in austral summer (December-April) along Australian southern shelves, between Cape Jaffa and Portland along the Bonney Coast (~800km) has recently been established (Kämpf et al., 2004). Longshore variation in the transport of the shelf circulation,
possibly due to changes in longshore wind forcing brought about by changes in coastal orientation (Bye, 1983) gives rise to a series of onshelf and offshelf transports in the eastern GAB and Bonney Coast (Bye, 1998).

Coastal upwelling occurs simultaneously in three upwelling centres: off southern Eyre Peninsula, off south western Kangaroo Island, and along the Bonney Coast (Hahn, 1986; Schahinger, 1989; Griffin et. al., 1997; Kämpf et al., 2004; see Fig. 6). The Bonnie upwelling system has 2-3 major upwelling events, each lasting ~1 week (Kämpf et al., 2004).

The shelf width varies greatly; ranges from narrow (~20km) shelf along the Bonney Coast to broader (100km) shelf of the GAB. It has been suggested by Kämpf et al. (2004) that the unison of SST response of similar magnitude in all region indicates a pre-existing larger scale process which lifts cold (<12°C) water onto the shelf. It has also been suggested by Herzfeld and Tomczak (1997) that the slope of the GAB contributes to shelf anticyclonic circulation that occurs in the central GAB when the Bonney Upwelling occurs.

Exploration of a major upwelling event in March 1988 shows the evolution of peak surface chlorophyll-a concentrations of >4µg/L lagging the onset of upwelling by ~1 week. During later measurements distinct sub-surface chlorophyll-a maximum at a depth of 50m was found along the upwelling front (Kämpf et al., 2004). In addition, model studies of Middleton and Platov, (2004) show upwelling from the 600 m deep permanent thermocline.
Figure 6
The three upwelling centres in the eastern Great Australian Bight observed via satellite. Note the larger centre along the Bonnie coast near Cape Jaffa and smaller centers off Kangaroo Island and the Eyre Peninsular. Core MD2607 of this study is indicated in the upper left image along with location markers, KI: Kangaroo Island, SG: Spencer Gulf, GSV: Gulf of St Vincent. Satellite data from NOAA compiled at the CSIRO Marine Laboratories Remote Sensing Facility 1996, Image from David Griffin, (CSIRO Marine) abstracts of The South East Indian Ocean and Great Australian Bight USA/Australia Bilateral Workshop 1998, Port Lincoln.
2.10 General Water Masses and Frontal boundaries of the Study Areas

This section will discuss the characteristics of water masses in the study areas, from source at high latitude to low latitude, and the location of frontal systems which determine the location of the water masses.

2.10.1 Southern Ocean

The surface of the Southern Ocean between Antarctica and Australia displays a strong latitudinal, zonal structure of oceanographic fronts and boundaries. These include the Subtropical Front (SF), Subantarctic Front (SAF), Polar Front (PF) and the Antarctic Divergence, which are arranged in a poleward fashion following definite sea-surface temperature gradients (Nees et al., 1999). The STF, the SAF and the PF are associated with the filaments of the Antarctic Circumpolar Current (Nowlin et al., 1977; Joyce et al., 1978; Belkin and Gordon, 1996; see Fig. 7).

The Subtropical Convergence, like other frontal structures, is formed in response to high-latitude wind fields and represents the boundary between the salty, warm water of the southeast Indian Ocean to the north and the Southern Ocean to the south (Belkin and Gordon, 1996). Fronts are comparatively narrow zones and can be recognised by sharp changes in vertical structure, temperature, salinity and nutrients (Belkin and Gordon, 1996). Significantly increased sea-surface productivity and subsequent downward flux of organic matter into the deep-sea are observed at oceanographic fronts (Yoder et al., 1994).

The North Subtropical Front (NSTF) is distinguished in the Indian Ocean between 31-38°S (see Fig. 7). The NSTF location is coincident with the position of the wind convergence between westerlies and easterlies (Belkin and Gordon, 1996), as generally the location of frontal zones are set by the interplay of wind stress fields, and bathymetric features that tend to steer the wind-driven current jets (King and Howard, 2000).

Deep water within the South Australian Basin consists of Circumpolar Deep Water (CPDW) and is generally characterised by a salinity minimum core (Osborne et al., 1983). However, CPDW can be divided into upper and lower CPDW due to dissimilar characteristic’s derived from different source waters. Lower CPDW is characterised with a salinity maximum and a nutrient minimum and believed to be derived from mixed North Atlantic Deep Water (Park et al., 1993). Upper CPDW contains the oxygen minimum layer entering the South Australian Basin from the west.
and derived from North Indian Deep Water (Rochford, 1965; Park et al., 1993; see Fig. 8).

Figure 7
The locations of fronts in the Southern Ocean according to Belkin and Gordon (1996). In the Australian sector, the South Subtropical Fronts confluence with the North Subtropical Fronts at 38°S to become the Subtropical Front (STF). The Subantarctic Mode Water (SAMW) is indicated, and formed, between the STF and Subantarctic Front.

Antarctic Intermediate Water (AAIW), formed by sinking along the Antarctic Polar Front between 48°S and 50°S (Meinardus, 1923; Bohnecke, 1936; Deacon 1933, 1937; Sverdrup et al., 1942; Mackintosh, 1946; Houtman, 1964; Gordon, 1967; Gordon and Goldberg, 1970; Gordon, 1971; Belkin and Gordon, 1996), is the most extensive and important water mass at intermediate depths in the world oceans (Emery and Meinke, 1986). The northward spreading AAIW (600-1400m) expands into the North Atlantic, northern Indian Ocean and equatorial Pacific and can be distinguished from
other water masses by a salinity minimum (<34.5‰; see Fig.8). It is further characterized by a temperature range 3.2-7°C and a high content of dissolved oxygen (Sturm, 2003).

Subantarctic Mode Water (SAMW) is closely related to the AAIW and occurs at depths between 400 and 600 m north of the Subantarctic Front (McCartney, 1977; Heath, 1985). The SAMW is a thick well oxygenated subsurface layer, that is almost isothermal (8-10°C; see Fig. 8), and originate in the region between the Subtropical Convergence and the Subantarctic Front (Martinez, 1997). The cool surface layer sinks in the summer and is isolated by warmer surface waters, forming a distinct thermostad (a layer characterised by a minimal vertical temperature gradient (McCartney, 1977; Passlow et al., 1997).

Core MD2607, lies near the shelf edge of the eastern GAB (865m) and is influenced by deep, intermediate and surface water masses (see Fig. 8). At the core location, CPDW is situated between 4000m to 1200m (Emery and Meincke, 1986; Gingele and De Deckker, 2005) and AAIW lies above the CPDW between 850-1100m (Passlow et al., 1997), In the Australian sector, SAMW is circulated by an anticyclonic gyre which brings it to the continental margin at depths 450-850m (McCartney, 1977; Edwards and Emery, 1982; Passlow et al., 1997) and isolated from the sea surface by warmer surface waters (Passlow et al., 1997).
Figure 8
The surface and intermediate-deep water masses over the core site of MD2607 as described in section 2.10.1 above. LC= Leeuwin Current, SAMW=Subantarctic Mode Water, AAIW= Antarctic Intermediate Water, CPDW= Circumpolar Deep Water. Note the influence of the Leeuwin Current at the shelf break with increased SST and SSS compared to the offshore water. Images constructed in Ocean Data View (ODV; Schlitzer, 2005), with data from Conkright (2001).
2.10.2 Eastern Indian Ocean

Wijffels et al. (2002) and Fieux et al. (2005) identified four major thermocline waters in the southeast Indian Ocean, (1) South Indian Subtropical Water (STW), (2) South Indian Central Water (SICW), (3) North Indian Water (NIW) and (4) the Indonesian Throughflow Water (ITW). When exiting into the Indian Ocean, ITW is fresh compared to the local Indian Ocean water masses. In the eastern Indian Ocean, the ITF meets the high-saline SICW that deviates westward with the West Australian Current. The contact between the ITF and the SICW occurs at ~23°S at the sea-surface and ~17°S at 500 m water depth. This oceanographic front is regarded as one of the most important of the world ocean (Tomczak and Godfrey, 1994) and must exert a strong control on the distribution of plankton in the eastern Indian Ocean (Martinez et al., 1998). Wijffels et al. (2002) suggested that the convergence of thermocline waters, associated with the ITW/ SICW-STW front, occurs at approximately 14°S at the sub surface. The other major thermocline water mass fronts occur at 5-10°S between ITW and NIW (Wijffels et al. 2002; see Fig. 9).

Relatively salty North Indian Water (NIW) is advected eastward south of Java within and below the South Java Current. While the upper portion of this flow appears ‘blocked’ at Lombok Strait by the Throughflow, the deeper portions penetrate further east below the Lombok sill depth (350 m; Wijffels et al., 2002). A narrow core of eastward flow is also found between 700 and 2000m, and strongly trapped against the Australian shelf. The eastward flow was suggested in the water mass analysis of Wijffels et al. (2002) to consist of a mixture of relatively salty NIW and the fresher Indonesian Intermediate Water (IIW) (see Fig. 9).

The South Indian Subtropical Water (STW) is capped under the fresher ITW water in the middle of the thermocline with a salinity maximum near the upper thermocline (18°C isotherm) and 25.8 isopycnal (Fieux et al., 2005; see Fig. 10). Subduction around 30°S under the high evaporation regime of the subtropical South Indian Ocean produces STW (the saltiest waters found in the region) (Wijffels et al. 2002; Fieux et al., 2005). STW is carried north by the large-scale interior anticyclonic circulation (Wijffels et al. 2002).

The high salinity SICW is formed along the Subtropical Front in response to negative Ekman pumping or subduction (Tomczak and Godfrey, 1994). Water moving downwards is trapped along isopycnal surfaces in late autumn and winter to travel north affecting the uppermost 800-900m of the upper water column (Tomczak and Godfrey,
South Indian Central Water (SICW) is subducted further south in the subtropical Indian Ocean than STW, and thus is denser, fresher and is marked by high oxygen concentrations. At lower latitudes SICW lies below the STW salinity maximum (Wijffels et al. 2002). The recently ventilated SICW water mass is clearly defined as a region of high oxygen and low silica concentrations between depths of 200 and 500 m (Wijffels et al. 2002; Fieux et al., 2005). Near the bottom of the SICW lies an oxygen maximum centered along 400 m, which is due to Subantarctic Mode Water (SAMW) formed by subduction in the subantarctic zone (McCartney, 1977; Toole and Warren, 1993; Wijffels et al. 2002).

Figure 9.
The three major thermocline water masses from Wijffels et al. (2002). Properties on neutral surface $\gamma = 24.5$ based on the raw observed data compiled in Levitus et al. (1994), averaged as described by Lozier et al. (1994). (a) Salinity; (b) thermocline depth (m) and (c) oxygen concentrations (mmol/kg). The three major thermocline water masses are marked in (a): North Indian Water (NIW), South Indian Subtropical Water (STW) and Indonesian Throughflow Water (ITW).
Antarctic Intermediate Water (AAIW) is formed in the Antarctic Polar Zone (Fine, 1993; Toole and Warren, 1993; Sloyan and Rintoul, 2001). The thickness of the minimum salinity layer decreases toward lower latitudes, eroded by mixing with higher salinity water masses above and below (Fieux et al., 2005; see Fig. 10). Antarctic Intermediate Water (AAIW) marked by a salinity minimum near 6–7°C lies below the SAMW, and has relatively high oxygen and low silica concentrations compared to the waters to the north (Wijffels et al. 2002; see Fig. 10).
The temperature and salinity of surface, thermocline and intermediate water masses of the Southeast Indian Ocean from 12°S to 36°S. Note the low salinity warm water of the ITW carried by the Leeuwin Current at the sea-surface and the high salinity tongue of the STW that interact near the core of MD61 (22°S). ITW= Indonesian Throughflow Water, STW= South Indian Subtropical Water, SICW= South Indian Central Water, SAMW=Subantarctic Mode Water, AAIW= Antarctic Intermediate Water. Images constructed in Ocean Data View (ODV; Schlitzer, 2005), with data from Conkright (2001).
Chapter 3
Regional ecology

3.1 Eastern Indian Ocean

The present-day ecology of marine organisms along the coastline of Western Australia is highly influenced by the location and occurrence of the Leeuwin Current. The Leeuwin Current is responsible for the presence of tropical marine organisms off the west and south coasts of the continent (Maxwell and Cresswell, 1981; Pearce and Walker, 1991; Hutchins and Pearce, 1994; Pearce and Pattiaratchi, 1999). For example, the Leeuwin Current has a major influence on the life histories of the southern bluefin tuna *Thunnus maccoyii* (Davis and Lyne, 1994; Pearce and Pattiaratchi, 1999), the western rock lobster *Panulirus Cygnus* (Pearce and Phillips, 1988; Pearce and Pattiaratchi, 1999), coastal fisheries (Lenanton et al., 1991; Pearce and Pattiaratchi, 1999) and the distribution of seabirds (Wooller et al., 1991; Pearce and Pattiaratchi, 1999).

A majority of investigations into the distribution of planktonic foraminifera, in the Indian Ocean, have been plankton tow studies which identify present day assemblages; these are by: Ujiie (1968). Ujiie and Nagase (1971), Bé and Tolderlund (1971), Bé and Hutson (1977) and Bé (1977, see Appendix 1). At the same geographic location, different assemblages can be represented in plankton tow and core top material (Bé and Hutson, 1977). This is mostly due to dissolution in the water column and within the sediment. Post-mortem transport and winnowing by ocean currents can also effect the distribution of planktonic foraminifera within the sediment by preferentially removing juvenile species. In regards to this study, more accurate comparisons can be made to core top studies based on fossil assemblages that represent ‘present day’ assemblages and more accurate interpretation of palaeoceanographic conditions can be applied (Martinez et al., 1998). However, it has to be considered that core-tops integrate sedimentation over decades and centuries.

Extensive analysis of fossil planktonic foraminifers and their geographical provinces was conducted by Bé and Hutson (1977) within the Indian Ocean. They identified five main groupings of planktonic foraminifers, due to their distribution on the sea-floor, and linked this to environmental parameters (see Fig 11). Martinez et al (1998) redefined the planktonic foraminiferal provinces of the eastern Indian Ocean by
considering the dynamics of the Warm Pool and the Java Upwelling System (see Fig. 12). In addition, Martinez et al. (1998) used principal component, canonical correspondence and factor analysis on the core top material to define faunal assemblages. Another study by Ding et al. (2006) used cluster analyses on core-tops to define five provinces: 1. the Banda/Java region (I); 2. the Timor region (II); 3. the Java upwelling region (III); 4. the Indian monsoon Sumatra region (IV) and 5. the NW Australia margin region (V) (see Fig. 13).

Differences in species assemblage for the similar geographic locations are to be noted between the three studies. The Upwelling (I) and Warm Pool (II) assemblages of Martinez et al. (1998) are positioned between 0°S to 18°S (see Fig. 12). This is the region identified by Bé and Hutson (1977) to include tropical (S4) and subtropical-tropical (S3) assemblages. This area included provinces I to III and V of Ding et al. (2006; see Fig. 13).

Ding et al. (2006) province IV, the Indian monsoon Sumatra region, contains the location of BAR9403 (see Fig. 13). The species assemblages of Ding et al. (2006) contain high abundance of Gs. ruber, N. dutertrei and P. obliquiloculata. This indicated an oligotrophic, low salinity environment with small seasonal temperature differences and a shallow thermocline.

Martinez et al. (1998) identified the Upwelling (I) assemblage as being dominated by Gr. menardii, P. obliquiloculata, Gr. tumida, S. dehiscens and N. pachyderma (dextral). These species are classed in this group because they are deep-dwelling, symbiotic-barren and heterotrophic consumers that are abundant when the thermocline is shallow and within the photic zone, such as upwelling regions (Thunell and Reynolds, 1984; Hemleben et al., 1989; Ravelo et al., 1990; Andreasen and Ravelo, 1997; Martinez et al., 1998; see Fig. 11). Tropical assemblages (S4) of Bé and Hutson (1977) consisted of Gr. menardii, Gs. quadrilobatus (Gs. sacculifer), N. dutertrei, P. obliquiloculata, Gr. tumida and Gl. aequilateralis (see Fig. 12). In Ding et al. (2006) this area coincided with the Java Upwelling province which is characterised with high abundances of Gr. menardii, N. dutertrei, and P. obliquiloculata believed to be due to the ITF limiting the upwelling of cold, nutrient-rich subsurface waters.

The Warm Pool species (II) of Martinez et al. (1998) consisted of Gs. sacculifer, Gs. ruber, Gl. aequilateralis, O. universa and Gs. conglobatus supported by the fact these species generally are shallow-dwelling, symbiont-bearing species commonly found in oligotrophic regions where the thermocline extends below the photic zone (Thunell and Reynolds, 1984; Hemleben et al., 1989; Martinez et al., 1998). While the
subtropical and tropical species (S3 group) of Bé and Hutson (1998) consisted of Gs. ruber, Gn. glutinata, Gs. quadrilobatus (Gs. sacculifer), Gl. aequilateralis and Gt. rubescens. Province V of Ding et al. (2006) had high percentage abundance of Gs. ruber and Gs. sacculifer also indicating oligotrophic, low salinity conditions.

Transitional species (III/IV) combined with the Southern species (V) cover the latitudinal area defined by the S5 group of Bé and Hutson (18°S to 35°S). Transitional species of Martinez et al. (1998) consists of Gn. glutinata, Gt. rubescens, Gt. tenella and Gr. crassaformis. This assemblage includes shallow to deep dwelling species and symbiont-bearing to symbiont-barren species reflecting a changing structure of the upper-water column and mark seasonal variations in temperature and nutrients (Martinez et al., 1998).

Below this group, the Southern assemblage (V) consists of Gr. inflata, Gr. truncatulinoides (sinistral), Ga. bulloides and N. dutertrei; these species are intermediate to very deep dwelling, symbiont-barren and typically found at high latitudes (Thunell and Reynolds, 1984; Hemleben et al., 1989; Martinez et al, 1989). The tropical-subtropical boundary current assemblage (S5) group of Bé and Hutson (1977) includes species Gn. glutinata, Ga. bulloides and Gr. menardii. It is possible that S5 is not a true representation of the present day assemblages off Western Australia as Bé and Hutson (1977) study was done before the presence of the Leeuwin Current was acknowledged (Cresswell and Golding, 1980). It may have been presumed that a typical eastern boundary current existed in this location. It is possible that the abundance of Ga. bulloides in this group represent the upwelling of the Capes Current as most of the cores are located between 30-35°S (see Fig. 11).
Figure 11
Foraminiferal geographical provinces in the Indian Ocean (from Bé & Hutson 1977). Foram species in order left to right

S1 a) *N. pachyderma*, sinistral b) *N. pachyderma*, dextral c) *Ga. bulloides* d) *T. quinqueloba*. 
S2 a) *Gr. inflata* b) *Gr. truncatulinoides*. 
S3 a) *Gs. ruber* b) *Gn. glutinata* c) *Gs. quadrilobatus* (*Gs. sacculifer*) d) *Gl. aequilateralis* e) *Gt. rubescens*. 
S4 a) *Gr. menardii* b) *Gs. quadrilobatus* (*Gs. sacculifer*) c) *N. dutertrei* d) *P. obliquiloculata* e) *Gr. tumida* f) *Gl. aequilateralis*. 
S5 a) *Gn. glutinata* b) *Ga. bulloides* c) *Gr. menardii*.
Figure 12
The foraminiferal assemblages of Martinez et al. (1998) in the eastern Indian Ocean, note stars representing the core tops analysed by Martinez et al. (1998)

The southern groups of Bé and Hutson (1977) extend beyond the sample area of Martinez et al. (1998). The polar and subpolar assemblages (S1 group) consisted of sinistral and dextral *N. pachyderma*, *G. bulloides* and *T. quinqueloba*. The transitional species (S2 group) were represented by *G. inflata* and *G. truncatulinoides* (see Fig. 11).

It is shown that species can be placed into quite different groups for example, *N. pachyderma* (dextral) reflecting nutrients in the Upwelling group of Martinez et al. (1998) but is classed as a polar species in Bé and Hutson’s (1977) groups. Also *N. dutertrei* is placed in the tropical group of Bé and Hutson (1977) and Ding et al. (2006) cluster III and IV while it has some representation in the warm pool group (II) and,
placed in the Southern (V) assemblage in Martinez et al. (1998). It has to be acknowledged that all studies are over different size areas (see Fig 11, 12 and 13) so species may be averaged out in Bé and Hutson (1977) while, in Martinez et al (1998), the study may undervalue the diversity of the southern regions as no core tops were analysed beyond 32°S. Regardless of these differences, all these studies provide a good comparison on the present day fossil assemblage in the Indian Ocean to which my samples can be compared.

Figure 13
The core-top locations of Li et al (2006) and five provinces defined by cluster analysis on foraminiferal assemblages. The five provinces are: I Banda-Java region, II Timor region, III Java Upwelling region, IV Indian Monsoon Sumatra region, V Australia margin region.

3.2 South Australian coastline
The southern Australian coastline hosts one of the world’s largest cool water carbonate provinces on Earth, it is located in a high-energy, swell-dominated
oceanographic setting (James et al., 1992). The Holocene sediments are enriched in coralline algal particles and conspicuous large tropical foraminifers (cf. Marginopora) and depleted in bryozoans, as compared to coeval deposits on the Lacepede and Otway shelves off south-eastern Australia. These differences are interpreted to reflect warmer waters of the Leeuwin Current and prevalent downwelling in this area as opposed to the general upwelling and colder waters in the east (James et al., 1992).

The benthic foraminifera species Marginopora bears algal symbionts and requires warm, oligotrophic waters. The distribution of Marginopora signals the flow of the Leeuwin Current as Marginopora is abundant in Western Australian waters and is restricted to the Lincoln Shelf and farther west. Its absence from the neighbouring Lacepede Shelf suggests that the influence of the Leeuwin Current is weaker there (Li et al., 1996). But, in the past, the flow of the Leeuwin Current is expected to have been different especially at glacial and interglacial extremes (see Chapter 5). The difference between the Lincoln and the Lacepede Shelves is due to Kangaroo Island, which acts as an oceanographic and biogeographic barrier (Li et al., 1996a; see Fig. 5).

The present-day faunal community structure and distribution of planktonic foraminifera on the Lincoln Shelf is outlined in Li et al. (1996a). Temperate species Gr. inflata dominates the assemblages, while Gs. ruber, Ga. bulloides, Ga. falconensis and O. universa were also common. Rare but consistently occurring species include Gr. scitula, G. trilobus (sacculifer), N. dutertrei and Gn. glutinata. The dominance of Gr. inflata is alternated by fluctuations of other species, particularly Gs. ruber, Ga. bulloides-Ga. falconensis and (in deeper sites) Gr. truncatulinoides. Li et al. (1996a) considered that the assemblages were related to depth, shifting dominances were attributed to a wandering of currents and the greater influence of deeper waters as the Leeuwin Current is moved to inner shelf areas. For example, the high in G. ruber/Gr. inflata ratio may signal the existence of a warm shelf current as it shifts to inner and outer shelf locations (Li et al., 1996a).

Not all species common to the tropics and subtropics occurs in the southern shelves of Australia as Gs. conglobatus and P. obliquiloculata are not represented (see Li et al., 1996) in the Lincoln and Lacepede Shelves. On the Lincoln and Lacepede Shelves, Gs. sacculifer, Gr. menardii and other typical (sub) tropical forms are rarer than to the west or absent and the planktonic fauna is dominated by Gr. inflata, Gs. ruber, Ga. bulloides and, in deeper sites, Gr. truncatulinoides (Li et al., 1996; Li et al., 1999).
Identification of Holocene planktonic assemblages along the southern shelf of Western Australia by Li and McGowran (1998) and Li et al. (1999) have revealed that assemblages were dominated in the west by subtropical forms (*Globigerinoides trilobatus* (*Gs*. *sacculifer*), *Gr*. *menardii* and *N*. *dutertrei* and in the east by the temperate species *Gr*. *inflata*. Species *Gs*. *ruber* is abundant in the east and west of the Bight but less so in the central part; further east, its abundance falls from 20–40% on the Lincoln Shelf to <20% on the Lacepede Shelf (Li and McGowran, 1998). In contrast, the temperate index taxon *Gr*. *inflata* increases eastwards from <10% offshore of Bunbury, 10–30% offshore of Albany, 20–50% offshore of Esperance, and 30–70% in the Bight and on the Lincoln and Lacepede Shelves. This is accompanied by an increase in *Gr*. *truncatulinoides* from <5% from the far west to 5–20% in the Bight and farther east. Species *Ga*. *bulloides* and *Ga*. *falconensis* show a similar pattern, increasing from <5% in the Western Australian sector to 15–30% in the Bight. Their abundance remains at 10–20% on the Lincoln and Lacepede shelves, where *Gr*. *inflata* is predominant (Li and McGowran, 1998). This West to East gradation indicates that the Leeuwin Current has played a key role in influencing the distribution of foraminifera at least from the last interglacial, because this pattern exists in the Holocene and also in the relict foraminiferal biofacies (Li et al., 1999).

In the deeper part of the Bight a high number of *Ga*. *bulloides* and *Ga*. *falconensis* but many specimens of other species with a small size, including *Gt*. *rubescens*, *T*. *quinqueloba*, *N*. *pachyderma*, *Tenuitellinata juvenilis* and *Tenuitella* sp, is observed indicating a high productivity (Li and McGowran, 1998). Li nd McGowran (1998) based on the presence of numerous small species in the assemblage from the deeper part of the Bight suggests (1) a fertile water mixed with water from the head of the Bight and Leeuwin Current from the west and upwelling water from the south, and (2) a modification of the surface circulation from a domination by the Leeuwin Current to a flow influenced more intensively by the warm and arid local climate (Herzfeld 1997; Li and McGowran, 1998).
Chapter 4

Regional Palaeoceanography

Previous studies along the western and southern coastlines of Australia provide an insight into the adjustments and interactions of the main current systems of the region, such as the interrelation of the Leeuwin Current and the Indonesian Throughflow and the opposing West Australian Current during glacial and interglacial phases (see Fig. 1).

The study by Martinez et al. (1999) extended the core-top analysis of Martinez et al. (1998) and found groups of planktonic foraminifera that characterised the Last Glacial Maximum time-slice. Martinez et al. (1999) found the percentage of shallow-dwelling species (above 50m; e.g. *G. sacculifer*) was reduced by 10% during the LGM compared to the modern day. In contrast, the percentage abundance of intermediate (50-100m; e.g *N. dutertrei*) and deep-dwelling species (below 100m; e.g *G. inflata*) increases significantly during the LGM (see Fig. 14). In addition, symbiont-bearing species decreased during the LGM by ~10%, whereas symbiont-facultative species increased by 10% relative to modern percentages. In contrast, the symbiont-barren species show an inverse pattern, with maximum values in the north during the LGM as compared to the minimum observed today.

Past investigations (see Fig. 15 and 16), into the occurrence of the Leeuwin Current during glacial and interglacial periods have established two main hypotheses for the position of the Leeuwin Current during glacial cycles while the main argument for the last interglacial (MIS 5; ~125, 000 yrs B.P.) is that the Leeuwin Current was enhanced. The results and hypotheses of past studies are presented below.
Figure 14

The comparison of present day and LGM distribution of planktonic foraminifera (%) grouped according to their depth habitat from Martinez et al. (1999). Martinez et al. (1999) study was based on 10 gravity cores taken within major oceanographic features of the eastern Indian Ocean such as the Warm Pool, the SICW/ITW front, the Java Upwelling System and the offshore transitional waters for comparison, between ~12°S to 32°S. The present day abundances are based on Martinez et al. (1998) study of 57 core top in the same region.
Figure 15
The other cores studied by Wells and Wells (1994), Wells et al. (1994; blue dots) and Martinez et al. (1999; green dots). Note the location of core MD61 that is very close to the location of Fr10/95-17 which has been analysed by numerous studies as shown in Figure 15.
4.1. Absence of the Leeuwin Current

Sea-surface temperature reconstructions off Western Australia by Wells and Wells (1994) and Wells et al. (1994), indicate a major change in surface-water circulation during glacial times in the Late Quaternary. Applying a transfer function to planktonic foraminiferal assemblages, Wells and Wells (1994) established SST gradient residuals for several climate extremes in the Late Quaternary. They indicated that areas north of ~18°S changed very little over the last ~ 130 000 yrs BP, while south of ~18°S there were significant changes in surface-water temperatures during the LGM and at the end of the penultimate glaciation (MIS 6; ~ 130 000 yrs B.P.) (Wells and Wells, 1994). These authors believed the Leeuwin Current did not flow during glacial periods, and large areas of anomalously low sea-surface temperatures were established off-shore at mid-latitudes (~19° S- 25° S), and further south (Wells and Wells, 1994) resulting in a SST 6°-7°C cooler than today (Wells et al., 1994). This statement will be investigated further with core MD61.

The absence of the Leeuwin Current during glacial periods is also supported by De Deckker (1997) and Martinez et al. (1999), who identified a possible mechanism behind the Leeuwin Current shifting further offshore and migrating westward into the Indian Ocean. They suggested the reduced low salinity water input from the Indonesian Throughflow and sea levels, would have resulted in more saline Throughflow water and encouraged its incorporation into the SEC or SJC enhancing these systems. The dominance of the SEC over the Leeuwin Current was also the findings of Takahashi and Okada., (2000). They found a high abundance of calcareous nanoplankton offshore southern Java, during glacial periods, indicating low SST's, the reduction of the Leeuwin Current and the dominance of the SEC during this period.

Analysis of corals from the Houtman Abrolhos (28.5°S) and Ningaloo coral reefs (21-23°S) also support the removal of the Leeuwin Current from shelf areas (Collins et al., 1993; 2003). The fringing reef systems existence is due to the presence of the Leeuwin Current at a latitude usually dominated by more temperate communities. Studies such as Collins et al. (1993, 2003) indicated two periods of reef development during the last interglacial (~125 000 yrs B.P; MIS 5) and during the Holocene (12 000 yrs B.P-present). This may suggest the Leeuwin Current was only present near the coast of Western Australia during these two time periods and, therefore, movement of the Leeuwin Current occurred away from the coast during glacial periods.
Figure 17 (a,b)
Figure from Barrows and Juggins (2005) showing the modern (World Ocean Atlas, 1998) and the LGM mean annual SST estimates. Note in the modern frame the downward movement of the isotherms near the Western Australian coastline implying the presence of the Leeuwin Current. However, during the LGM the isotherms near the Western Australian coastline are not depressed as far south, compared to the modern isotherms, implying the influence of the Leeuwin Current on the South Western coastline is reduced.
4.2 Reduction of the Leeuwin Current

It has been suggested that the West Australian Current (see Fig. 1) had greater influence during glacial periods thus reducing the influence of the Leeuwin Current. Barrows and Juggins (2005) updated previously mapped SST of the LGM by Prell et al. (1980), Wells and Wells (1994) and by Barrows et al. (1996), and indicated the cooling of the tropics during the LGM (see Fig. 17a and b). Barrows and Juggins (2005) suggested that reduced SST off Western Australia may be due to cooler waters entering the area via a strengthened West Australian Current, the South Equatorial Pacific current and upwelling of colder waters of the Java Upwelling System.

Calcareous nannofossil studies (e.g. Okada and Wells, 1997) also suggested that SICW (driven by the WAC) increased in dominance during Marine Isotope Stage 2 and 6 (MIS 2 and 6) and this resulted in a weaker Leeuwin Current. Takahashi and Okada (2000) also working with calcareous nannofossils similarly indicated that the Leeuwin Current was reduced during the LGM and importantly did not reach their southern core site (Fr 10/95 GC.20, 111°49.75'E, 24°44.67'S; see Fig. 15).

Sedimentological studies suggest that the Leeuwin Current was reduced during glacial periods. The Gingele et al. (2001) analysis of clay minerals in the Indonesian Throughflow region and on the Northwest Shelf has provided insight into the pathways of the Indonesian Throughflow (ITF) and the Leeuwin Current during the Last Glacial Maximum (LGM). This study found that the volume of the ITF decreased during the LGM as less kaolinite and chlorite reached the Timor Passage. In addition, offshore the Northwest Cape, it was suggested that a reduction in chlorite may also indicate a decreased Leeuwin Current because of the apparent reduction in the flow of the ITF.

4.3 Enhanced Leeuwin Current

Tropical and subtropical faunas have been used to indicate an enhanced Leeuwin Current at higher latitudes. There appears to be evidence for a stronger-than-present Leeuwin Current resulting in warmer seas around the western and southern coasts of Australia during the last interglacial (MIS 5; ~125,000 yrs B.P). This is also timed to enhanced monsoonal activity across tropical and subtropical parts of the continent, including northern Western Australia, during the early part of MIS 5 (Murray-Wallace et al., 2000).
Wells and Wells (1994) also suggested an enhanced Leeuwin Current along the western coastline during interglacial periods, based on planktonic foraminiferal assemblages. They indicate that sea-surface temperatures during Isotope Stage 5e is similar to today (±1.3°C error range) in summer and up to 3°C warmer in winter. While coral reef development of the Ningaloo Reef system indicates that the last interglacial reef growth, toward the end of the sea-level high stand (120 000–115 000 yrs BP), was more extensive than the Holocene, perhaps as a result of a stronger Leeuwin Current (Collins et al., 2003). Perth Basin deposits of Substage 5e also record a southerly expansion of warm-water corals and other fauna consistent with shelf temperatures warmer than present (Kendrick et al., 1991).

The establishment of hyposaline-estuarine habitats favourable to arcoid bivalve *Anadara trapezia* (subtropical species) has been identified by Murray and Belperio (1991) and Murray-Wallace et al. (2000), along the southern coastline of Australia and suggests higher, more sustained levels of river discharge than at present. Murray-Wallace and Belperio (1991) suggest that inshore water temperatures during Isotope Stage 5e along the western and southern coasts of Australia were about 2-3°C warmer than present due to the location of coral and molluscan fossils on the Eyre Peninsula, South Australia. Using comparison with the modern *Marginopora* community in Esperance WA, Cann and Clarke (1993) suggest the presence of tropical benthic foraminifera *Marginopora* in the Spencer Gulf was due to a stronger Leeuwin Current during the last interglacial resulting in a 5°C increase in SST.

### 4.4 Was there upwelling along the Coastlines?

#### 4.4.1 West Australian Coastline

There has been some evidence of pulses of nutrient-rich water ‘upwelling’ along the western coastline of Australia during glacial periods. However, a fully developed upwelling system comparable to the eastern boundary currents of Peru and Namibia has not been identified. Different intensities of ‘upwelling’ have been described along the coasts of the Indonesian archipelago and the West Australian coastline. The suggested mechanism behind possible upwelling is attributed to major changes in oceanic circulation during the glacial intervals, involving the increased importance of the West Australian Current (CLIMAP, 1984; Prell and Hutson, 1979; Prell et al., 1979, 1980;
A strengthening of the westerly airstream onto the southern parts of the west coast of Australia during the LGM (Prell et al., 1980) was accompanied by an increase in the strength of the tropical easterlies blowing off the Australian continent (Kolla and Biscaye, 1977). The stronger offshore winds had the potential to cause upwelling of sub-surface water at low latitudes off the west coast of Western Australia (Wells et al., 1994). However, Martinez et al. (1999) suggested that a steeper latitudinal gradient, due to a weak Leeuwin Current prevented substantial upwelling along the western Australian coastline.

Wells et al. (1994) analysis of benthic foraminifera, $\delta^{13}$C trends of benthic foraminifera Cibicidoides wuellerstorfi, and planktonic foraminifera Globigerinoides sacculifer indicated that during MIS 6 bottom waters were significantly depleted in $\delta^{13}$C, and strong $\delta^{13}$C gradients were established in the water column, while during the LGM, $\delta^{13}$C trends did not differ greatly from that of the Holocene. In addition, abundance of uvigerinids (benthic foraminifera), and of taxa preferring an infaunal microhabitat, indicate high influx episodes of particulate organic matter in most sites during glacial episodes, particularly during stage 6, while this evidence for 'upwelling' during the LGM is not as strong. The penultimate glaciation (MIS 6) upwelling was established within the areas of low sea-surface palaeotemperature indicated by planktic foraminifera (Wells et al., 1994). McCorkle et al. (1994) used mass accumulation rates of benthic foraminifera to indicate increased productivity during the LGM and suggested the Leeuwin Current was removed during this period.

Also along the West Australian coastline Okada and Wells (1997) used nannofossils to indicate mild upwelling for MIS 7-5 in the upper-photic zone and intensified upwelling during the penultimate glaciation (MIS 6). The existence of upwelling was not clear during MIS 2 and not strong enough to produce blooms of Emiliania huxlei. Therefore, there was no significant increase in productivity at the LGM offshore Western Australia as found by Okada and Wells (1997). This seems to be supported by the benthic foraminifera analysis of Murgese (2003) in core Fr10/95 GC17, who showed high percentages of C. wuellerstorfi during 31-18 000 yrs BP, suggesting oligotrophic conditions and increased ventilation by lateral advection of active bottom currents.

Mass accumulation rates of biogenic (carbonate, organic carbon, Ba) and terrigenous (Al, Th) sediment components, together with excess $^{230}$Th in gravity cores from the Exmouth Plateau and the Perth Basin on the West Australian continental
margin were analysed for indication of enhanced productivity by Veeh et al. (2000; see Fig. 15). Veeh et al (2000) suggested that there was no compelling evidence in the sediment record for strong coastal upwelling comparable to that in the modern ocean off the west coasts of Africa and South America (Veeh et al., 2000). They dismissed that a major reorganization of ocean circulation involving the replacement of the Leeuwin Current with the north flowing West Australian Current occurred during the LGM.

Eastern Indian Ocean foraminiferal studies, such as that of Martinez et al. (1999), have revealed contrasting distribution patterns for the LGM when compared to present day assemblages resulting in a substantial difference in the depth of the thermocline and a slight decrease in SST. The foraminiferal record indicated an intensification of upwelling and/ or more influence of corrosive Antarctic Intermediate Water (AAIW) offshore southern Java during the LGM due to an increase of species G. cultrata (Gr. menardii) and N. dutertrei with increased foraminiferal fragmentation. Takahashi and Okada (2000) also identified there was an enhanced upwelling system of the South Java Current (SJC) due to the high abundance of small placoliths, which are indicative of eutrophic conditions. However, De Deckker and Gingele (2002) argue that, although there was a rise in nitrate levels near the sea-surface, the water column was actually stratified. This allowed blooms of the giant diatom Ethmodiscus rex to develop offshore southeast Sumatra, and therefore, upwelling to the sea-surface was not recorded during glacial times off the coast of Sumatra.

4.4.2 South Australian Coastline

There is also conflicting evidence of coastal upwelling along the southern Australian coastline during glacial periods. AMS $^{14}$C and U/Th dating, seismic and isotopic data evidence indicate that bryozoan mounds of Pleistocene age developed cyclically in response to glacial productivity cycles. Increased upwelling during sea level lowstands promoted active mound growth, in contrast to the thin mud accumulations that draped inactive mounds during highstands (Feary et al., 2004). During glacial periods, the lowered sea level, the weakened Leeuwin Current, and increased upwelling, provided enhanced carbon flux and nutrients, contributing to prolific bryozoan growth and mound development (James et al., 2000; Holbourn et al., 2002; Feary et al., 2004). Alternatively, increased nutrient supply and enhanced primary productivity may have been enhanced by northward movement of the Subtropical Convergence Zone (James et al., 2004).
Another causal influence of bryozoan growth may be due to the pockets of gas hydrates in the shelfal muds providing nutrients during low sea-level. During Leg 182 (DSDP) the presence of brine (106%) within the upper slope sediments was associated with the highest H₂S concentration (15%) ever measured during any Deep Sea Drilling Project (DSDP) or ODP leg (Swart et al., 2000).

Gingele and De Deckker (2005) showed evidence of increase productivity above core MD2607 (see Fig. 14). They used various proxies, including concentrations of aragonite, low and high magnesium calcite, total carbonate, total organic carbon, sulphur and δ¹³C of planktonic foraminifera Globigerina bulloides, which revealed cyclic increases of productivity during glacial periods.

4.5 Movement of the Frontal Systems

4.5.1 Indian Ocean

Late Quaternary fluctuations in the position of the STC have been analysed by several investigators using faunal and floral parameters (Williams, 1976; Bé and Duplessy, 1976; Prell et al., 1979; Prell et al., 1980; Morley, 1989; Howard and Prell, 1992; Passlow et al., 1997; Nees et al., 2000; Gersonde et al., 2005). There is evidence that the Subtropical Convergence (STC) shifted northward during the glacial periods and southward during interglacial periods. Analysing the relative abundances of planktonic foraminifera within cores MD61 and MD2607 will provide insight into the dynamics of the water column. This may provide additional evidence that there was a shift in the Subtropical Front during glacial periods with greater influence of the Indian Central Water driven by the West Australian Current, which was observed by Martinez et al. (1999). Martinez et al. (1999) indicated that stronger cooling occurred in areas 16-23°S off the western coastline of Australia and was accompanied with a dramatic increase in transitional species Gr. inflata.

4.5.2 Southern Ocean

During Late Quaternary glacial episodes, the Polar Front migrated from its present day position at 48°S to near 40°S expanding the distribution of polar waters and reducing the extent of the subpolar water mass. A northward shift also occurred in the STC of approximately 2° latitude but less than 5° latitude (Prell et al., 1979). However, the extent of this migration has been questioned by Armand (1997) Sturm (2003) and Gersonde et al. (2005).
Nees et al. (2000), benthic foraminiferal study of core MD88-779 on the South Tasman Rise (47°50.69S, 146°32.75E; 2260 m water depth), suggests significant increases in ocean surface productivity during glacial periods and, in particular, during isotopic stages 2, late 3 and 6. Increased productivity at the sea-surface is indicative that both the Subtropical Front and the Subantarctic Front may have shifted northward during glacial periods and that the Subantarctic Front was near the coring site on the South Tasman Rise for these periods (Nees et al., 2000).

The frontal migration of approximately 2-6° that has been identified in the southern Atlantic and Indian Oceans (Howard and Prell, 1992; Brathauer and Abelmann, 1999; Crosta et al., 1998) has shown to be restricted by the bottom topography of the South Tasman Rise and the Campbell Plateau (Moore et al., 1999) in the Australian and New Zealand components of the Southern Ocean. It has also been suggested that the East Australian Current which, like the Leeuwin Current, advects warm and comparatively saline water into higher latitudes, may affect the positioning of the frontal systems. It may be the frontal systems are compressed during glacial periods with stronger SST gradients, because the glacial Subtropical Front was locked to a position south of Tasmania (Armand, 1997; Sturm, 2003). Gersonde et al. (2005) observed the abundance of diatoms and radiolarian from 122 cores from the Atlantic, Indian and Pacific sectors of the Southern Ocean to reconstruct the LGM frontal systems. They suggest the northward expansion of Antarctic waters by 5–10° in latitude and only a relatively small displacement of the Subtropical Front. Therefore, the thermal gradients were steeper during the LGM in the northern zone of the Southern Ocean (Gersonde et al., 2005).

During warm phases the fronts are seen to move back towards the poles. Howard and Prell (1992) suggested poleward movement of the STC only four times during the last 500 000 yrs. Maximum poleward excursions were found during MIS 9 (330 000-340 000 yrs BP) and MIS 11 (390 000-425 000 yrs BP) to approximately 45°S (Howard and Prell, 1992). However, Morley’s (1989) radiolarian study did not record this poleward shift during MIS 9. Other excursions occurred during the Holocene, specifically between 7500 and 10 000 yrs BP (Francois et al., 1993), and MIS 5e (118 000-125 000 yrs BP; Howard and Prell, 1992).
Chapter 5

Modern Climate

Due to the interrelationship of the atmospheric and oceanic circulatory systems, it is important to understand the present day climatic conditions and how these systems may have been altered during glacial and interglacial cycles. The use of proxies to identify past oceanic conditions within these cores can point to alteration within the atmospheric systems such as changes to the dynamics of the Australasian Monsoon. This is exacerbated by the influence of the El Niño—Southern Oscillation (especially in northern and eastern Australia) which introduces pronounced interannual variability to the Australian climate (Allan, 1988; Allan et al., 1996; Hesse et al., 2004).

5.1 Effects of SST on Climate in the Australasian region.

There is an inter-relationship between high SST and high precipitation levels along the western and southern coastlines of Australia. High SST especially within the region of the Warm Pool, enhances the transfer of moisture from the sea-surface to the atmosphere. Current Warm Pool SSTs ensure that this region has the most extensive coverage of organised, deep convective clouds on Earth. Within the 26.5-29°C range of SST, organised convection is much increased (Waliser and Graham, 1993; De Deckker et al., 2003). This transfer of latent heat, combined with the dynamics of the Asian Monsoon, causes increased precipitation in northern Australia.

The average SST off Australia’s west coast near 32°S is 20.8°C, which at the same latitude is 5.6°C and 4.2°C higher than the eastern boundary current of the Pacific and Atlantic Ocean respectively (Table 1; Feng et al., 2003). Correspondingly, the precipitation rate in the coastal area of southwest Australia is more than twice the precipitation rates in southwest South America and Africa (Table 1; Feng et al., 2003) and suggests the warm water carried by the Leeuwin Current influences the climate of Australia. The long coastline and the lack of mountain barriers mean that areas of Australia are subject to oceanic influences, although major short-term effects of the surrounding oceans on temperatures, humidity and precipitation are confined to coastal regions (Hobbs, 1998). This is supported by Pearce and Pattiaratchi, (1999), who suggest that the higher winter air temperatures and rainfall in Western Australia compared with those in corresponding latitudes elsewhere may be linked with the warm
water of the Leeuwin Current. This demonstrates the impact of the Leeuwin Current on the coastal climate in Western Australia and the potential significance of alteration to this current in glacial and interglacial periods.

Table 1
This table shows the comparison of SST and precipitation along the western coastline of Australia, South America and Africa. Note the higher SST and precipitation rate along the Western coastline compared to other eastern boundary current regions (South America and Africa). aThe SST is taken from Levitus and Boyer (1994). bThe mean precipitation rate is taken from the Merged Analysis of Precipitation (CMAP) data averaged between 1978 and 2000 (Xie and Arkin, 1997). This table is from Feng et al. (2003)

<table>
<thead>
<tr>
<th></th>
<th>Australia</th>
<th>South America</th>
<th>Africa</th>
</tr>
</thead>
<tbody>
<tr>
<td>SST °C</td>
<td>20.8</td>
<td>15.2</td>
<td>16.6</td>
</tr>
<tr>
<td>Precipitation, (mm day⁻¹)</td>
<td>1.8</td>
<td>0.8</td>
<td>0.8</td>
</tr>
</tbody>
</table>

Some modelling studies failed to find a correlation of increased SST causing increased precipitation along the western coastline of Australia. Drosdowsky and Chambers (1998) analysed near-global SST principal component patterns as potential rainfall predictors but found little correlation with rainfall in the southwestern WA region. Model studies by Frederiksen et al. (1999) did not reveal any direct link between rainfall and SST anomalies of the Indian Dipole (cool SST of the Indonesian region occurring every ~ 5 years) unless anomalies were made relatively large. However, Allan and Haylock (1993) noted that rainfall decrease appeared to be linked with Indian Ocean SST. Drosdowsky (1993) identified a link between northwest to southwest winter rainfall pattern and SST of the southern Indian Ocean and Ansell et al. (2000) found that southwestern Western Australian rainfall in winter was significantly correlated with winter SST over Indian and Southern Ocean regions from 1907-1994.

As stated above (section 2.1) the Warm Pool is one of three global sites for tropical convection. The atmosphere above the warm pool has been called the ‘boiler box’, because of the massive convective cloud systems of the region (Sturman and Tapper, 1996). The variability of thermal structure in the Warm Pool region is
particularly important for the climate of Australia. A pattern of anomalously cool sea-surface temperatures in the Indonesian region, as seen in El Niño cycles, affects the rainfall in Victoria and central Australia. Drought occurs, due to the lack of warm water carried by the Leeuwin Current, and these are severe enough to have a damaging economic impact (Meyers, 1996).

A lack of cyclone activity due to reduced SST can also be observed during El Niño years (Bureau of Meteorology, 2003; see Fig. 18a and 18b). Considering monsoonal influence and tropical cyclones are important rain producers in low latitudes (Hobbs, 1998), this shows the impact of SST on the climate of the region. Studies by McGowran et al. (1997), De Deckker (1998), Martinez et al. (1999) and Pillans and Bourman, (2001) argue that the removal of the Leeuwin Current could have greatly affected the Australian climate in the past. One of the components of this project will be to investigate the effects of possible change in sea-surface temperature (SST) and precipitation during glacial periods.
Average annual number of tropical cyclones - La Nina years

Based on a 2 x 2 degree resolution gridded analysis using 30 years of data (1969/70 to 1998/99 tropical cyclone seasons).

Average annual number of tropical cyclones - El Nino years

Based on a 2 x 2 degree resolution gridded analysis using 30 years of data (1969/70 to 1998/99 tropical cyclone seasons).

Figure 18a and 18b
The average occurrence of tropical cyclones during (a) normal oceanic conditions (La Nina) and (b) El Niño. Lower occurrences of cyclones are associated with El Nino years when SST in the Warm Pool are lower. Images from Commonwealth Bureau of Meteorology, Australia.

5.2 Australasian Monsoon

The modern climate of the Indonesian Archipelago and northern Australia is dominated by biannual monsoonal circulation. Core BAR9403 is located within the Warm Pool which is both influenced and contributes to the dynamics of the biannual Australasian Monsoon, while, core MD61 is located beneath the pseudo-monsoon circulation near the southern limit of monsoonal activity (see Fig. 19). Heavy rain accompanies northwesterly winds between November and March (Austral summer), during the Northwest (NW) Monsoon. The dry season corresponds to the Southeast (SE) Monsoon period from May to September (Austral winter).

![Figure 19](image_url)

The late summer (February) patterns in the Australian region of (a) pressure; and (b) associated air flow (From Sturman and Tapper, 1996).
The intertropical convergence zone (ITCZ) is the driving force behind monsoonal variability. The ITCZ, a pressure trough where the southeast and northeast trade winds (Northern Hemisphere) meet, usually lies about 10°-15° north of the equator in the Austral winter and migrates south, close to or over northern Australia in summer (Hobbs, 1998). At present, the yearly movement of the ITCZ in Australasia is larger than anywhere else in the world (Singh and Agrawal, 1976).

An important feature of northern Australia that contributes to the southward movement of the monsoon trough is the development of two low pressure systems from intense heating of the lower atmosphere that persist through the warmer months of the year. These occur in the semi-arid Australian tropics in the Pilbara region of north-west Western Australia and in the Cloncurry region of north-west Queensland (Sturman and Tapper, 1996; see Fig. 19). The low-pressure system over Australia and a high-pressure system over Asia during summer (February), promote northeasterly winds over Southeast Asia (Tomczak and Godfrey, 1994) and northwesterly winds from 0°S to the ITCZ.

During the Austral summer, the Northwest Monsoon gathers large amounts of moisture while crossing the sea from the Asian high-pressure belt on its way to the Inter-Tropical Convergence Zone (ITCZ), which has shifted south to approximately 10-15°S (Hobbs, 1998). At the ITCZ, the moisture-laden air rises, resulting in heavy rains (van der Kaars et al., 2000). Therefore, Northern Australia is characterised by wet summers and dry winters. In northern Australia, the annual rainfall varies from around 500 (austral winter) to 1500 mm (austral summer) with the highest rainfall in coastal areas (van der Kaars et al., 2000; see Fig. 20). Vegetation cover also responds to the change in precipitation away from the coastline, as the amount of rainfall decreases from NE-SW, so does the vegetation cover (Gingele et al., 2001). Rainfall in the tropical north of Australia occurs regularly during the austral summer. In the more arid western areas of the continent precipitation is mostly associated with single, short events, such as cyclones (Gingele et al., 2001).

The monsoon season can last from two weeks to four months, with break periods during most monsoon seasons, when dry southeasterly flow becomes temporarily re-established as the equatorial trough retreaths northward. Break periods represent 20% of monsoon season (Sturman and Tapper, 1996). Some of the trade winds are deflected by the heat lows south of the ITCZ forming pseudo-monsoon in north-western Australia (Sturman and Tapper, 1996) in the region of core MD61.

In the Australasian region, the ITCZ lies 10-15°N of the equator in the austral winter largely in response to the seasonal shift in continental heating (Sturman and Tapper, 1996; see Fig. 21). During this time the interactions between the high-pressure system over Australia and a low-pressure system over Asia during winter (August) promote southeasterly trade winds over most of Australia (Tomczak and Godfrey, 1994) which are relatively dry and cool (van der Kaars et al., 2000) losing most of their moisture along and close to the east coast (Hobbs, 1998). The southeast trade wind flow is generated by the semi-permanent South Pacific high-pressure system centred between 85°-100°W and 27°-37°S (Hastenrath, 1988; Hobbs, 1998) and when the migratory subtropical ridge lies across the Australian continent (Sturman and Tapper, 1996). The flow is stronger in winter when the subtropical pressure cells are more intense and the equatorward pressure gradient is greatest (Sturman and Tapper, 1996; see Fig. 21).
The position of the subtropical high pressure belt varies both seasonally and annually with the Southern Oscillation Index (Drosdowsky and Chambers, 1988) and also with altitude, being more northerly at higher elevation. (Kawahata, 2002; see Fig. 21). Generally, the southern subtropical high pressure belt dividing the tropical south-easterly circulation (trade winds) from the mid-latitude westerlies lies at a latitude of approximately 30°S in winter and is located near 40°S in summer (Sturman and Tapper, 1996; see Fig. 21).

Figure 21
The summers and winters mean zonal wind and mass flux in the southern Hemsphere. Note the movement of equatorial trough (ITCZ) and the subtropical ridge during the summer and winter periods and the incursion of the northern Hadley Cell causing the southward movement of the ITCZ in summer. (Image from Sturman and Tapper, 1996).

5.3 Mid latitude Westerlies

In southern Australia, summers are dry and hot and rainfall is more effective and reliable in the cool winter months across the arid zone. During winter the sub-tropical
ridge moves north and cold fronts embedded in the westerly circulation bring moisture inland (Hesse et al., 2004). This region is dominated by the occurrence of fronts as subtropical and polar air masses interact (Sturman and Tapper, 1996).

Lying between the subtropical ridge (highs) and sub-Antarctic trough (lows in Southern Ocean) is the zone of the mid-latitude westerlies. This zone moves northwards to affect southern Australia in winter when the westerly flow is particularly strong (Sturman and Tapper, 1996; see Fig. 21). Mobile, high pressure systems moving east across the continent are the main controls on seasonal patterns of atmospheric circulation and rainfall. They cross the east coast at about 38°S in late summer and about 28°S in late winter (Hobbs, 1998). During winter (May-October), when the anticyclones track across Australia at relatively low latitudes, cool westerly winds pass over the southern part of the continent. Frequent cold frontal low pressure systems embedded in the westerlies then bring a regular alternation of winds from northwesterlies to south-westerlies (Hobbs, 1998).

In the summer half of the year (November-April), the mid-latitude anticyclones are usually considered further south, largely retreating from the continent (Sturman and Tapper, 1996). Summer heating of the northern part of the continent and the southward shift of the ITCZ combine to produce a broad area of low pressure across the north, particularly in the northwest and direct south-easterly winds over much of the continent (Hobbs, 1998).
Chapter 6
Regional Palaeoclimate

Investigations into Late Quaternary palaeoclimate provide an insight into the onset of drier and wetter climate conditions and suggest possible mechanisms behind regional climate change. Potentially important isotopic stages for understanding our future climate such as Marine Isotope Stages 11, which is captured in core MD61 of this study, are discussed below.

6.1 Marine Isotope Stage 11

A few studies (e.g. Loutre and Berger, 2003) suggest that the period from 405,000 to 340,000 yrs, including a large part of MIS 11, could be a good analogue for future climate. External climate forcing has been recognised to be the main driver of our climate and the time interval between 360,000 and 423,000 yrs BP (MIS 11) is characterized by a configuration of Earth’s orbit that is similar to that of the present (Berger, 1978; Poli et al., 2000). During MIS 11, and at present, small amplitude insolation changes coincided with the minima of 400,000-year eccentricity cycle. Eccentricity will reach almost zero within the next 25,000 yrs, damping the variations of precession considerably (Loutre and Berger, 2003).

The apparent high amplitude of the temperature change across the transition from MIS 12 to 11, from the most severe glacial of the past 0.5 million years to the longest and warmest interglacial interval of the past 500,000 yrs BP, has been recognized globally (Imbrie et al., 1993; Howard, 1997; King and Howard, 2000). Millennial-scale climate instability and associated changes in North Atlantic Deep Water are believed to occur during both MIS 11 and MIS 12. A series of interstadial events that occurred at a 5-6,000-year periodicity, and the occurrence of ice-rafted debris indicate icebergs were present at least as far south as 34°N during MIS 12 (Poli et al., 2000). By comparison, simulations using the climate model developed in Louvain-la-Neuve (LLN 2-D NH) show that both MIS 11 and the future are characterized by small amount (if any) of continental ice, with almost no variation during the whole interval (Loutre and Berger, 2003). The long ice free interval of MIS 11 was also
indicated in the North Atlantic by McManus et al. (1999). In contrast, MIS 5 is exhibiting larger variability in simulated ice volume (Loutre and Berger, 2003).

Terrestrial evidence from Europe (Rousseau et al., 1992; Loutre and Berger, 2003) and marine evidence from New Zealand (Howard, 1997; King and Howard, 2001) suggests that MIS 11 was warmer than the Holocene and MIS 5. The Chinese loess–soil sequence shows that the summer monsoon was particularly strengthened during MIS 11 and 5 (Guo et al., 2000; Loutre and Berger, 2003). Moreover, sea-level highstands have been reported from Alaska, England, Bermuda and the Bahamas (Hearty et al., 1999; Kindler and Hearty, 2000; Loutre and Berger, 2003).

However, there are discrepancies in the record such as the marine sequences from the North and South Atlantic (Oppo et al., 1998; Hodell et al., 2000). These studies indicate that MIS 11 was not warmer, but actually slightly cooler than the Holocene. Therefore, considerable uncertainties remain in regard to the characteristics of MIS 12 and 11. Little is known about these stages in the Indian Ocean and it is hoped that the study of core MD61 will provide some valuable data for the region.

### 6.2 Marine Isotope Stages 6-10

The Marine Isotope stages of 6-10 are not very well documented due to the sparseness of terrestrial records. There is evidence in Northwest Australia of enhanced runoff, and therefore a stronger monsoon, is inferred in interglacial stages to at least Stage 9 (English et al., 2001; Bowler et al., 2001; Hesse et al., 2004). In the Eyre Basin major episodes of fluvial deposits from Coopers Creek have been identified by Nanson et al., 1990, 1992b) from MIS 7 to late MIS 8. This is further indicated by fluvial sediments on the Lake Eyre western shore with a major wet period earlier than 170 000 yrs BP, possibly Stage 7 or early Stage 6 (Croke et al., 1996).

The fossil distribution of A. trapezia in Western Australia lends support to the proposition that MIS 5e and MIS 7 climates along what are now the semi-arid and Mediterranean-type coasts of south western and southern Australia were significantly more humid than at present, with more regularly sustained and less seasonally contrasting precipitation patterns influenced by monsoonal and epimonsoonal phenomena (Murray-Wallace et al., 2000)

Lake Wangoom, western Victoria has a record that covers the last two glacial cycles. Interglacials MIS 7, 5e and 1 show high values for forest or woodland taxa,
especially Eucalyptus, while dryer conditions during glacial periods are indicated by grasses and by herbaceous and shrubby members of the Asteraceae (Harle et al., 2002).

6.3 Last Interglacial (MIS 5)

Many terrestrial records and deep-sea cores within the Indian Ocean only capture more recent isotopic events, such as the last interglacial, and therefore more information is available from these time periods in Australia.

The last interglacial is generally considered to be a stage of high precipitation, with pollen records along the west coast of Australia indicating periods of maximum summer rain at 100 000, 80 000 and 70 000 yrs BP (van der Kaars and De Deckker, 2002). Runoff records (lakes and rivers) from monsoon dominated areas show alternating wet and dry phases superimposed on a clear drying trend over the last 130 000 yrs BP at all sites (Hesse et al., 2004).

Conditions during MIS 5 were wetter than at any subsequent time, resulting in larger than present Lake Eyre up to 25m deep over prolonged periods (Magee et al., 1995; Murray-Wallace et al., 2000). Most of this stream discharge originated from northern summer monsoonal precipitation. A high lake-level phase of Lake Eyre between 130 000 to 90 000 yrs BP is confirmed by both palaeo-environmental and computer simulated data. This coincides with a time of high sea level and strong orbital forcing of the Asian Monsoon, the primary forcing mechanism for the north Australian Monsoon (Croke et al., 1999). Magee et al., (1995) suggested that the NW Monsoon promoted wetter phases in the Lake Eyre catchment of central Australia at 125 000 yrs BP and 80 000 yrs BP.

Additional evidence for wetter conditions during MIS 5e is seen in the extensive river terraces about the margins of the Adelaide Hills in South Australia (Bourman, 1968; Bourman et al., 1997; Murray-Wallace et al., 2000). Similarly, pollen records from Southern Ocean marine core E55-6 (Murray-Wallace et al., 2000) and from Lake Wangoom in western Victoria (Harle et al., 2002) also indicate that for the Late Quaternary, at least, maximum precipitation occurred in MIS 5e (Murray-Wallace et al., 2000).

6.4 Marine Isotope Stage 3 and 4

Significant review studies by Bowler (1976), Rognon and Williams (1977), Chappell and Grindrod (1983), Wilford (1984), Wasson and Clark (1988), Wasson and
Donnelly (1991) and Colhoun (1991) indicate that the climate of this period was cool and moist in south-eastern Australia and New Guinea, while in north-western Australia, New South Wales and mid-northern South Australia the climate was wetter (Allan and Lindesay, 1998). Evidence of this wetter phase is supported by Torgersen et al. (1988). Beginning at ~35 000 yrs BP and continuing until ~26 000 yrs BP, the net water balance of the Gulf of Carpentaria (basin) became wetter with aquatic pollen and ostracod chemistry indicating intense wet-season precipitation (Torgersen et al. 1988). The study of Lake Mungo by Bowler (1998) also indicated the alteration of wet and dry phases throughout MIS 3 with an overall drying trend.

However, the dating of Lake Eyre sand dunes (Gardner et al., 1987; Nanson et al., 1988, 1990, 1992a, 1995; Croke et al., 1996) from all published dated linear dunes in the Lake Eyre Basin shows a pronounced increase in dune activity after 40,000 yrs BP, extending to the Holocene (Hesse et al., 2004). High lake levels in Lake Eyre during MIS 4 gave way to drying and deflation only in early MIS 3 (Magee et al., 1995; Hesse et al., 2004). The study by Kawahata (2002) of trace aluminium from in a Tasman Sea core shows a much clearer increase in dust flux in MIS 4, persisting to the last termination, around 15 000 yrs BP.

In addition, marine core Fr10/95-17 from off the coast of Western Australia, indicates significantly drier conditions and reduced summer rain after 46 000 yrs BP compared to 100 000-64 000 yrs BP (van der Kaars and De Deckker, 2002). Vegetation changed from open Eucalyptus woodlands rich in grasses to open Eucalyptus and Gyrotemon shrublands rich in Chenopodiaceae/Amaranthaceae and Asteraceae Tubuliflorae, in the period from 46 000-40 000 yrs BP (van der Kaars and De Deckker, 2002). This period was also characterised by significant base level lowering and channel incision between 50 000 yrs - 31 000 yrs BP. Lake lowering induced fluvial incision of up to 9m, scouring several meters into basal silicified Miocene sediments some time after 50 000 yrs BP(Croke et al., 1996).

During late MIS 3 there was a shift to more southern (winter rainfall), drier vegetation, in and following the LGM until wetter vegetation was re-established after 14 000 yrs BP (van der Kaars and De Deckker, 2002). In the Lake Eyre basin, linear sand dune building was widespread by 30 000 yrs BP. Detailed shoreline evidence points to a long period of dry lake conditions through this time (Magee and Miller, 1998). Evidence during the 27 000 to 24 000 yrs BP period, suggests that in central and northern Australia conditions were drier with reduced intertropical rainfall (Rognon and Williams, 1977). This indicates that the subtropical anticyclone belt and the monsoon
trough were displaced further to the north than they are at present, so that the SE-trade winds were strengthened (Allan and Lindesay, 1998). The Gulf of Carpentaria, (Lake Carpentaria), also responded to this drier phase between ~23 000 and ~11 800 yrs BP with increased dominance of terrestrial grasses in the pollen and high charcoal that suggest a cyclical wet-dry sequence which may have been seasonal (Torgersen et al., 1988). Bowler (1998) has shown the alternation of wet and dry phases at Lake Mungo throughout MIS 3 and 2, with an overall drying trend.

6.5 Last Glacial Maximum

Sedimentological studies indicate a significantly lower input of terrigenous components during the last glacial maximum than during the Holocene, due to reduced runoff in monsoon-controlled northwestern Australia (Veeh et al., 2000; Gingele et al., 2001). The results are consistent with expanded aridity during glacial times and a somewhat wetter climate in NW Australia during interglacials (Veeh et al., 2000). The final phase was a period of aeolian and ephemeral-fluvial deposition, which peaked between 20 000 and 18 000 yrs BP, coincident with the LGM (Croke et al., 1996).

As outlined below, proxies such as dust, pollen, dune behaviour, and wind modelling have also been used to determine the characteristics of the climate of the LGM.

6.5.1 Dust

During dry periods airbourne dust is deposited in two main regions due to the direction of wind circulation across Australia (Hesse and McTainsh, 1999). The two cores (MD61 and MD2607) of this study lie beneath these two sites. Australia is the largest contemporary dust source area (Rea, 1994) and during the LGM contributed around three times more dust to the southwestern Pacific Ocean (Hesse, 1994; Hesse and McTainsh, 1999).

However, the evidence provided by dust recorded in marine cores over the last glacial termination that the size of silt transported to the Tasman Sea did not change and therefore the westerly winds were no stronger in the LGM than in the Holocene (Hesse and McTainsh, 1999). However, the study of Gingele and De Deckker (2005) from Murray Canyons which laid directly under the “Eastern Australian Dust Plume” suggest the atmospheric conditions during the glacial periods were comparable to present winter
patterns, when strong westerly winds dominate over the area and bring dust from the central desert areas.

Increased glacial atmospheric dust loadings and dust deposition are due more to source area aridity, greater dust production, and less-efficient scavenging than to stronger atmospheric circulation over the mid-latitude source areas (Hesse et al. 2004). Elevated dust fluxes from Australia to the Tasman Sea were not driven by more erosive winds but by greater aridity in inland southeastern Australia. There is also evidence for weaker rain scavenging of dust over the Tasman Sea, suggestive of a drier atmosphere and a weaker hydrological cycle, at least regionally (Hesse and Metanish, 1999).

6.5.2 Dunes

The onset of cooler, windier and drier conditions is indicated by the beginning of the extensive inland dune activation and building phases that culminated in the LGM (Wilford, 1984; Gardner et al., 1987; Wasson and Clark, 1988; Wasson, 1989; Nanson et al., 1988, 1990, 1992, 1995; Harrison, 1993; Croke et al. 1996; Allan and Lindesay, 1998). Activation of desert dune fields covered over 40% of Australia during the LGM (Hesse et al., 2004). In addition, several sites around northern Australia, from the Kimberley to Cape York Peninsula, record sand dunes blown from the exposed continental shelf during the LGM (Lees, 1992).

However Hesse and Metanish (1999) suggest dune activation inland of Australia also appears to be a result of reduced stabilising vegetation cover rather than increased wind strength. The LGM climate of inland Australia must have been considerably more stressful for plants as a result of lower precipitation and/or carbon dioxide stress to achieve the implied levels of surface destabilization (Hesse and Metanish, 1999).

6.5.3 Pollen

A review by Hope et al. (2004) identifies that the pollen records from southeastern Australia are interpreted mainly in terms of changing precipitation and there is little indication of temperature variation. Likewise, van der Kaars and De Deckker (2002) observed this trend in north-western Australia with the pollen record showing more arid assemblages during the LGM, by comparison with modern plant distributions.

This situation is likely to be a result of a combination of factors including the limited representation and restricted distribution of temperature controlled rainforest
taxa, the dominance of sclerophyll and herb taxa containing a number of species with
different and often wide bioclimatic ranges, little topographic variation and
environments more exposed to incursions of subtropical high pressure systems and
therefore very sensitive to moisture variation (Hope et al. 2004).

6.5.4 Wind

Hesse (1994) reconstructs the Quaternary history of the southeast Australian
dust plume from Tasman Sea cores taken from 30°S to 45°S. These cores show a 3-6°
northward shift in the high pressure subtropical ridge (STR), dividing the tropical
easterly circulation from the midlatitude westerlies. Kawahata (2002) took four cores
further to the north of Hesse (1994) study between 25°S to 35°30'S to quantify past
changes in primary production and aeolian dust sedimentation. Kawahata (2002)
suggests a small northward shift of 2-3° of the high pressure subtropical ridge during
 glacial times.

From pollen and lake studies (Colhoun, 1991; Markgraf et al., 1992; Harrison,
1993) it has been proposed stronger anticyclonic winds around an expanded STR and
poleward migration of the westerlies. However, Wyrwoll et al. (2000) modelled the
position of the westerlies and found little change in the Australian region and
insufficient resolution to determine either poleward or equatorward movement of the
subtropical front (Hesse et al., 2004).

Hesse et al., (2004) suggest that the largest changes in circulation patterns over
the glacial cycle probably occurred in the location and/or intensity of summer tropical
convergence in northern Australia. Over southern Australia, changes to the temperature
and humidity of the westerly circulation have been more significant than the small
fluctuations in latitude of the sub-tropical high pressure ridge. This is supported by
Shulmeister et al. (2004) who concluded that westerly circulation was as strong now as
at any time in the LGM as the changes in latitudinal boundaries in the westerlies of (3–
4°) during glaciation-interglaciation movements may be nearly as large in inter-annual
zonal shifts (2°) presently.

6.6 Holocene (12ka- present)

There is general agreement that warmer, wetter conditions existed from around
14 000-5000 yrs BP, possibly due to the re-intensification of the Australasian Monsoon
(van der Kaars and De Deckker, 2002) or the re-initiation of the NW Monsoon
(Wyrwoll and Miller, 2001). The period from 14 000 to 3 000 yrs BP was wetter with heavier summer rain compared to today, probably as a result of higher sea-surface temperatures (van der Kaars and De Deckker, 2002). Dune field analysis by Nott et al. (1999) in tropical northern Australia found that although the climate was clearly becoming substantially wetter after 9 000 yrs BP, as evidenced by pollen records and increased stream activity, at least one phase of longitudinal dune development occurred between 8200 to 5900 yrs BP.

Clay analysis has been conducted on core MD2607 by Gingele and De Deckker (2005), provides an insight into the climatic conditions of the core area in the GAB. Accounting for sea level fluctuations and the position of the “palaeo-Murray mouth,” there is evidence for stronger river discharges and more humid climate during the early Holocene (11 000-6 000 yrs BP), interrupted by a short dry spell (9000-8000 yrs BP), before present arid conditions (aeolian deposition) became finally established at 4000 yrs BP (Gingele and De Deckker, 2005). However, Ogden et al. (2001) found the flow of the Murray Basin rivers declined between 14 000-7 000 yrs BP. Wyrwoll and Miller (2001) found drier, ephemeral conditions at Lake Gregory with a maximum water level around 6 000 yrs BP. Lees (1992) also identified increasing coastal dune activity after 5500 yrs BP which he attributed to the onset of a late Holocene drying trend (interrupted by brief wetter periods) and Lake Eyre experienced a minor lacustral episode through to 4000–3000 yrs BP (Magee and Miller, 1998). Lake Frome entered a drying trend after 7000 yrs BP, interrupted by a brief lacustral event at 4000–3000 yrs BP (Luly, 2001).

Dust transported to the Tasman Sea by the westerlies shows clearly the boundary of the spring/early summer westerly zonal wind belt. Over the LGM to Holocene transition, and previous terminations, the westerlies retreated south by up to 3° latitude (Hesse, 1994; Kawahata, 2002).


Chapter 7

Methods

7.1 Core description BAR 9403

During the 1994 cruise of the RV Baruna Jaya, core BAR9403 was taken at 2034 meters below sea-level offshore the south-western coastline of Sumatra (5°49.20'S 103°61.90'E). This is a short piston core, 275cm in length, which was sampled every 5cm. Based on the observations of Murgese (2003) the top 105cm consists of a fine grained, light yellow clay. Below 105cm, the colour of the clay changes to grey-olive for the remainder of the core. Minor bioturbation is evident between 130-120cm. Two lenses of medium to coarse foraminiferal sand are found between 156-154cm and between 255-245cm (see Fig. 22).

Figure 22
Stratigraphic log of piston core BAR9403 consisting of foraminifera sand and clays from 0cm to 275cm.
7.2 Core description MD61

Core MD002361 (MD61) was taken during the TIP 2000 expedition of the Marion Dufresne in 2000, at 113°28.63'E, 22°04.92'S and at a depth of 1805 mbsl. At the time of coring this 42m piston core broke the record for the longest core obtained in the region. However, only the upper 1360cm was analysed after the primary MIS’s were reached. Core MD61 was sampled every 4cm between 0-300cm primarily to obtain a detailed age model for this interval (see Section 8.5). The rest of core MD61 was sampled every 5cm between 300-1360cm.

The main features of core MD61 is the occurrence of 1-1.5m bands of red clay during interglacial periods which alternate with 1-1.5m thick bands of light grey foraminifera sand during most of the glacial periods (see Fig. 23). Manganese laminations and nodules are a feature of the core mostly during glacial periods but become more prominent towards the base of the core (920-1330cm). Clay lenses are found between 459-490cm and spheres of red clay are found around the LGM (160-215cm) and MIS 7 (710-750cm) possibly due to intensified bottom currents (see Fig. 23).

From 0-90cm a homogenous, fine grained mid brown-tan clay mergers to red brown clay from 90cm to 155cm. The lithology changes at 155cm to a medium to coarse foraminiferal sand in a light grey clay matrix to 305cm. A feature of this 155-305cm section is occasional red brown streak/layers and spheres which decrease downcore.

At 240cm there is a pulse of red brown clay which coincides with the MIS 2-MIS 3 transition. The foraminiferal sand and the clay matrix between 305-360cm changes colour to a green grey and black manganese nodules are also a feature of this interval. The colour of the foraminiferal sand and clay matrix returns to a light grey between 360 to 459cm.

From 459cm to 495cm the core becomes more clay dominated and there is a corresponding colour change to light to mid brown. At 459-490cm there are light grey foraminiferal sand and clay lenses. The sediment merges to fine red brown clay at 495cm which continues to 605cm, this coincides with the later stages of MIS 5. A dark grey manganese lense occurs at 510cm and manganese laminations between 560-600cm.
From 605-710 the lithology changes to a foraminiferal sand in a dark grey-brown clay matrix with dark grey manganese laminations and nodules. At 710cm the sediment merges back to fine, red brown clay with light-dark brown clay spheres. There is a gradual change at 750cm to light grey brown foraminifera sand with manganese nodules at 760cm.

At 790cm the foraminifera sand merges back to fine red brown clay to 840cm, which coincides with the MIS 7 period. This section of red brown clay is dissected at 815cm with 1-2cm bands of light brown foraminifera sand.

The next phase of the core has very distinct alternating lithology (see Fig. 23). Light grey foraminiferal sand with manganese nodules occur during MIS 8 and MIS 10 between 840-960cm and 1050-1125cm and red brown, manganese nodule clays occur between 960-1050cm and 1125-1233cm during MIS 9 and MIS 11.

The end of the analysed core section has tan grey foraminiferal sand between 1240-1278cm. A short phase of mid brown-tan clay with manganese nodules between 1278-1305cm merges to light grey foraminiferal sand in a clay matrix from 1305-1360cm.
Mid brown tan clay

-40

Merge red brown clay

-120

Foraminifera sand in light grey clay matrix, occ red/brown clay layers decreasing down core

-160

Foraminifera sand in grey clay matrix with red/brown clay spheres

-200

Foraminifera sand in light grey clay matrix

-240

red brown pulse of clay

-280

Foraminifera sand in light grey clay matrix

-320

Foraminifera sand in green grey clay with black manganese nodules

-360

Foraminifera sand in light grey clay matrix

-400

Foraminifera sand in light grey clay matrix

-440

Foraminifera sand in red clay

-480

Transition to red brown clay

-520

More clay dominated sediment, light mud brown with light grey clay lenses

-560

Manganese laminations in red brown clay

-600

Transition to red brown clay

-640

Manganese laminations in red brown clay

-680

Foraminifera sand in dark grey-brown clay matrix with dark grey manganese laminations and nodules

-720

Gradual transition from red brown clay to forams in a light grey brown clay manganese nodules

-760

Gradual transition from light grey brown forams sand to red brown clay

-800

Large bands of foraminiferal light brown sand in red brown clay

-840

MIS 6

LGM 180.6 Ka

MIS 2

-291

MIS 3

-243

MIS 4

-160

MIS 5

-40

Holocene
Figure 23
The log of MD61 (in cm) showing the change in sediment from light grey foraminiferal sands and dark grey brown clays during glacial periods to red brown clays during the interglacial periods. The sediment log is placed against the timing of the interglacial and glacial periods to show the timing of the distinct change in sediment types. Red= interglacial warm periods, blue= cold interglacial period. The less extreme interstadial periods MIS 3 and MIS 4 are represented by purple and light blue respectively. The timing of the interglacial and glacial cycles are based on the $\delta^{18}O$ record of Gs. ruber in MD61 (see Section 8.4 for details).
7.3 Core description MD2607

Core MD03-2607 was taken from the upper Sprigg Canyon at a water depth of 865m (36°57.64'S, 137°24.39'E) during the 2003 AUSCAN cruise by the RV Marion Dufresne. The following core description contains information from the AUSCAN final report (see Fig. 24). The upper 14.7m of this giant piston core consist of massive fine to very fine pale-dark olive grey sand. Upward fining sequences occur between 4-5m and 9-10m. Below 14.7m the sediments are dominantly clayey with silty clay layers between 14.7-15m and 25.8-28.5m becoming silty clay and sand from 31m to the base. Shelly lenses and shell-rich sand lenses consisting of large foraminifera and foraminifera fragments occur throughout the core. Tiny rock fragments are noted at 21.4m and wood fragments at 22m. Echinoid and pteropod fragments were observed in most samples during foraminiferal counts.

Initial onboard analysis indicated high sedimentation rates towards the base of the core and this was considered while sampling the core. The Holocene section was sampled every 5cm. Sample intervals in the sections with higher sedimentation rates were increased to 20cm (MIS 2–4), 30cm (MIS 5) and 60cm (MIS 6). A total of 116 samples were collected from MD2607
<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Lithology</th>
<th>Structure</th>
<th>Colour</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>Massive</td>
<td></td>
<td>Olive-grey</td>
<td>Mottled olive-grey fine calcareous foraminiferal sand. Rare micrit and water-worn lenses.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Colour change abundant black specks and streaks. Belge bioturbation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Belge bioturbation.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Olive-grey</td>
<td>Large shelly foraminifera. Abundant 2-5 mm black specks. Less black specks.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>More black specks. Black specks cease.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Pale olive</td>
<td>Lens of large foraminifera. Coarse sand layer with shells and pelecypods. Fining-up sequence. Echinoid spine.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Coarse gravelly sand with shells. Coarse lens with shells. Irregularly layered at cm scale. Colour gradation from pale to dark olive-grey.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Sandy shell lens. Occasional black specks throughout.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Shelly lens</td>
<td>Shell. Shark clast. Fining upward sequence. Fining upward sequence.</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Darker olive-grey mottle.</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>Olive-grey</td>
<td>Rare black specks. Shark clast. Wet, soupy. Very uniform massive olive-grey. Rare black specks.</td>
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<td></td>
<td></td>
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<td></td>
<td>Thin irregular darker band. Darker olive-grey layer.</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>Rare black specks. Irregular darker and paler olive-grey layers with mottling.</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Thin diffuse shelly layer. Shell. Occasional isolated shells. Irregular cm-scale pale or dark olive-grey layers with shelly lense.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Shelly lens</td>
<td>Motting. Abundant black specks.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Small paler &lt;1 cm bioturbation. Larger 1 cm black streaks. Irregular pale layer with black upper and lower boundaries.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Small paler &lt;1 cm bioturbation or motting. Abundant black specks.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Lithology</th>
<th>Structure</th>
<th>Colour</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Massive</td>
<td></td>
<td>Olive-grey</td>
<td>Olive grey-green clay with abundant, high-density black staining that may indicate bioturbation.</td>
</tr>
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<td></td>
<td></td>
<td>Rare black sand lenses about 1 cm circular diameter.</td>
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<td></td>
<td></td>
<td>Sand lenses.</td>
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<td></td>
<td></td>
<td>Sand lenses.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Minor black staining. Dense black staining.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Echinoid fragments. Minor black staining. Dense black staining.</td>
</tr>
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<td></td>
<td></td>
<td></td>
<td>Occasional sand lenses roughly circular about 0.5 cm diameter.</td>
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<td></td>
<td></td>
<td></td>
<td>Sand lens. Dense black staining. Rare black staining.</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Sand lens. Dense black staining.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Large and small sand lenses up to 3 cm long. Thin layer of very fine sand about 2 cm thick.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Large and small sand lenses.</td>
</tr>
</tbody>
</table>

74
Figure 24
The stratigraphic log of core MD2607 analysed and constructed onboard the AUSCAN Cruise 2003. The core has a massive structure of olive green grey sediment from the top to the base of the core. From Hill and De Deckker (2003)

7.4 Isotope analysis

The calcareous tests of specific species of planktonic foraminifera were analysed to quantify the isotopic composition of oxygen and carbon through time in each of the cores. The chemical composition of the calcareous shell (e.g. stable isotope and trace elements) holds clues to the chemical and physical state of the ambient seawater and is useful in the reconstruction of temperature, chemical state, and biological productivity of the ancient marine environment (Schiebel and Hemleben, 2005).

The δ¹⁸O results will provide a stratigraphy for the three cores. In the study region the decrease in sea-level from present day to Last Glacial Maximum is ~129m (Yokoyama et al., 2000, 2001). The effect of ice volume on the δ¹⁸O record is demonstrated by Schrag et al. (2002), who conducted high-resolution oxygen and hydrogen measurements on pore fluids from deep-sea sediments in the North and South Atlantic. Schrag et al. (2002) estimates the global change in δ¹⁸O of seawater as 1.0±0.1‰ during the LGM and Holocene transition, confirming previous estimates of a 1.0‰ ice volume effect from foraminifera (Schrag et al., 1996). Duplessy et al., (2002) has combined the pore water analysis with calculation of glacier volume and suggests 1.05‰ in δ¹⁸O is due to the global ice volume effect during the LGM. For the Marine
Isotope Stage (MIS) 11 to MIS 12, transition the $\delta^{18}$O ice volume effect is estimated by Chappell (1998) to be 1.5‰.

The estimates of $\delta^{13}$C were obtained concurrently with $\delta^{18}$O during the analysis of samples with the Mass Spectrometer. The $\delta^{13}$C signals in the three cores of this study could be used with other proxies, such as changes in planktonic foraminifer abundance, to highlight changes of circulation patterns in deeper water masses and productivity at the sea-surface.

Isotopic analysis for all cores was sampled every 4 to 6 cm. Each sample consisted of 3 cm$^3$ of sediment sampled at 4 to 6 cm intervals. The sediment was oven dried and then soaked in dilute (3%) hydrogen peroxide solution until the clays separated from the microfossils. The samples were then wet sieved using the 63µm mesh size to collect the planktonic foraminiferal tests, which were then allowed to dry in an oven at 60°C.

Average sample sizes of approximately 15 foraminifera were selected from approximately the 250µm-size fraction for isotopic analysis. Isotopic analyses were conducted on tests of Ga. bulloides from core MD2607 while, for cores BAR9403 and MD61, the tests ofGs. ruber were used. The foraminifera were cleaned in 100% ethanol to remove any adhering particles.

Samples from cores BAR9403 and MD61 were analysed by the Finnigan MAT 251 Mass Spectrometer at the Research School of Earth Sciences at the Australian National University. Approximately 26 samples were analysed each day and run concurrently with the international standards NBS-19 (7 samples per run) and NBS-18 (1 sample per run). Therefore, in core MD61, 365 samples were run concurrently with ~75 samples of NBS-19 and 11 samples of NBS-18. In core BAR9403, 56 isotopic samples were analysed. The analytical error for $\delta^{18}$O is better than ±0.08 per mil and for $\delta^{13}$C is better than ±0.05 per mil (Joe Cali, personal communication). Isotopic samples (115) for core MD 2607 were analysed by the Finnigan MAT251 at GEOMAR in Kiel, Germany by Dr A. Sturm. Analyses gave a precision of ±0.07 per mil for $\delta^{18}$O and ±0.03 per mil for $\delta^{13}$C.

7.5 Age Model

The chronology of deep-sea cores in this study was deduced from, (1) oxygen-isotope stratigraphy, (2) $^{14}$C accelerator mass spectrometer analyses and (3) biostratigraphic last appearance of pink Ga. ruber.
Initially radiocarbon dating was arranged with Dr Martine Paterne at Gif-sur Yvette, Paris, after receiving a FEAST grant for our collaboration. Unfortunately this could not occur due to problems with the Tandetron. Fourteen samples of 10 mg *Gs. ruber* were prepared while at LSCE. Fe-graphite targets were prepared by Dr M. Paterne after my departure. These samples were radiocarbon dated at the Department of Nuclear Physics, Research School of Physical Sciences and Engineering, The Australian National University, with Dr K Fifield.

The radiocarbon dates were used to construct an age model for the upper sections of MD61. Radiocarbon dates were calibrated using the CALIB 5.0 program (Stuiver et al., 2005), and the Marine04 radiocarbon age calibration curve (Hughen et al., 2004). For the oceanic reservoir correction the estimate of the regional ΔR mean for NW Australia and Java of 67±24 (Bowman, 1985; Southon et al., 2002) was obtained from the CALIB Marine Reservoir Correction Database. The age model for the remainder of the giant piston core MD61 is based on the chronology of Bassinot et al. (1994) as the core extends beyond the SPECMAP timescale. Bassinot et al. (1994) study captures isotopic events in the tropical Indian Ocean that were accurately tuned to the astronomical time scale. This age model is used in other studies capturing MIS 12 such as King and Howard (2000). The Bassinot et al. (1994) astronomical time scale is also ideally located within the Indian Ocean and, as it is based on planktonic foraminifera very similar patterns, were found between the $\delta ^{18}$O stack of Bassinot et al. (1994) and core MD61.

To obtain an age model for cores MD2607 and BAR9403, the $\delta ^{18}$O curve was correlated against the SPECMAP chronology of Martinson et al. (1987) using the Analyseries software (Paillard et al., 1996), in accordance with previous studies on these cores by Murgese (2003), Gingele et al. (2004) and Gingele et al. (2005). Likely errors for this age model range from ±1370 cal yrs BP at stage 2.2 to ±5560 cal yrs BP at stage 4.0 (Martinson et al., 1987).

Biostratigraphic chronology was utilised in cores MD61 and MD2607 with older marine sediment. The last occurrence in the Indian Ocean of the pink form of *Globigerinoides ruber* is dated at 120 000 yrs B.P by Thompson et al. (1979) and will be noted in cores MD61 and MD2607.
7.6 Planktonic Foraminifera

For each core over 400 planktonic foraminifera per sample were identified, at a species level, to reconstruct faunal assemblages through time. The species nomenclature used in this study follows the taxonomy of Saito et al. (1981). Counts of planktonic foraminifera were made on splits of the >150 µm fraction to provide a base level for the ecological counts, removing small juvenile and possibly unidentifiable foraminifera. Each sample was split by an Otto-micro splitter until ~ 400 species were present in the final split. The number of fragments was also recorded to give an indication of the preservation status of the sample.

7.7 Modern analogue technique

Sea-surface temperatures (SST) were estimated from the planktonic foraminifera assemblage data using the modern analogue technique (MAT), in conjunction with the AUSMAT-F4 database (Barrows and Juggins, 2005), consisting of 2619 core tops. Each SST estimate was calculated as the mean of the best 10 analogues from the global database, using the square chord distance as the dissimilarity coefficient. The mathematical square chord ‘distance’ to the nearest analogue (MAT) is a measure to how well the modern training set represents the fauna of a sample (Barrows and Juggins, 2005). The distance to the nearest analogue, the mean distance and the standard deviation were also calculated to assess the quality of the analogue. The most precise variable is mean annual temperature with a root mean squared error of prediction (RMSEP) of 0.84°C. Tmax also have a low RMSEP of 0.87°C whereas, Tmin SST is the least well predicted with a RMSEP of 0.98°C (Barrows and Juggins, 2005). The quality of SST estimates is measured by the square chord distance (dissimilarity coefficient). The analogue estimate is reduced with values greater than 0.2, conversely squared-chord distances of <0.2, show that the samples have good analogues in the AUSMAT-F4 database.

The annual SST (Tmean), and the temperature of the warmest and coolest months were estimated for all the cores. In a tropical monsoonal setting, the months with the most extreme temperatures are usually not the calendar months in the middle of summer and winter. The minimum monthly temperature along the western and southern coastline of Australia is during September, while the maximum monthly temperature occurs during March-April on the western coastline and February along the southern coastline (Conkright et al., 2001). In addition, a temperature minimum (Tmin) of the sea
surface could correspond to conditions during upwelling when cool water is brought to the sea surface (Troelstra and Kroon, 1989; van Iperen et al., 1993). This highlights the importance of reconstructing the temperature of these months whenever they occur throughout the year (Barrows and Juggins, 2005).

In addition to SST parameters, the temperature at 50 m, 100 m, and 150 m depth (using Levitus and Boyer, 1994) were estimated. These depth levels are important because the highest concentration of foraminifera live below the surface, in the mixed layer and at the top of the thermocline. The 100m and 150m depth levels are at the base of the photic zone where a DCM could develop in regions with a thin mixed layer. Therefore these temperature variables could be more representative of the changes influencing the assemblage (Barrows, personal communication).

To get an insight into the dynamics of the Leeuwin Current during the LGM and how the estimates of core MD61 relate to other regional studies, SST maps of the temperature extremes (Tmin and Tmax) were constructed. The other cores in the region were originally studied by Prell et al. (1980), Prell et al. (1985), Wells and Wells (1996) and Martinez et al. (1999). The SST estimates of Prell et al. (1980), Wells and Wells (1996) were estimated via a transfer function while a MAT was used by Prell et al. (1985) and Martinez et al. (1999). The original data from these studies have been recently re-analysed by Barrows and Juggins (2005) with the AUSMAT_F4 also used by this study and therefore, SST estimates can be compared between the cores. Unfortunately, only the LGM timeslice can be compared as this is the extent of Barrows and Juggins (2005) study but it will provide insight into how the Leeuwin Current responded to maximum global cooling which is comparable to the cooling by the other glacial periods recorded in core MD61.

From the original studies of Prell et al. (1980), Prell et al. (1985) and Wells and Wells (1996) the AUSMAT_F4 found warmer Tmax estimates of approximately 3-5.7°C ± 0.87°C between 20°S and 33.5°S whereas estimates between 14°S and 20°S were equivalent to those of the original studies. The Tmin estimates of the AUSMAT_F4 were cooler by 0.8-2.5°C ± 0.98°C between 14°S and 18°S whereas the estimates between 19°S and 33.5°S were 1.7-5.7°C ± 0.98°C warmer.

7.8 Statistical Analysis

Statistical analysis was performed on all of the cores to understand the variance within the abundance counts of foraminifera. Principal component analysis and Imbrie
and Kipp Factor analysis (Q-mode Factor analysis) is used to find ecological groups within the data. This technique finds variance within the raw data counts and does not correlate this to a separate parameter. The C2 program, version 1.4 (Juggins, 2003) was used to perform the statistical analysis. In accordance with other studies (Martinez et al., 1998) species abundance below 1% was excluded from analysis and raw counts were entered into the analysis to reduce error.

Imbrie and Kipp factor analysis (a Q-mode factor analysis) will also be used to identify assemblage groups within the data as this process involves oblique and orthogonal analysis. The factor analysis eliminates ecological redundancy and simplifies the data displayed but also offers insights into the underlying structure and complexity of the ecosystem as well (Imbrie and Kipp, 1971). Factor analysis compared to principal component analysis can oversimplify the data. However, assemblage groups can represent several ecological situations such as independent reactions to physical conditions, seasonal responses, and vertical stacked water masses to which species responses may overlap. The advantage of factor analysis is to make tendencies clear which may be obscured by random error and to focus attention on ecological or diagenetic questions worthy of further inquiry (Imbrie and Kipp, 1971).

Analysing the data with Principal Component Analysis may find more complex groups that may be more explanatory in terms of the variability in abundance groups. Principal component analysis involves mathematical transformations of the original data set to produce a set of orthogonal (i.e. uncorrelated) eigenvectors that describe the main modes of variance in the multiple parameters making up the data set (Grimmer, 1963; Stidd, 1967; Daultrey, 1976; Bradley, 1999). Each eigenvector is a variable that expresses part of the overall variance in the data set. Although there are as many eigenvectors as original variables, most of the original variance will be accounted for by only a few of the eigenvectors (Bradley, 1999). The first eigenvector represents the primary mode of distribution of the data set and accounts for the largest percentage of its variance (Mitchell et al., 1966; Bradley, 1999). Subsequent eigenvectors account for lesser and lesser amounts of the remaining variance (Bradley, 1999). The proportion of variance explained = Eigenvalue / sum of Eigenvalues (Darlington, 2005). A scree test (Cattell, 1966) will be used to determine how many factors to retain. A scree test is a graphical method of plotting the Eigenvalues. You can dismiss the additional Eigenvalues from where the Eigenvalues level out in the plot. (Statsoft, 2005).
Chapter 8

Results

Stable isotope and faunal analyses results from the three deep-sea cores analysed in this study are presented below. Comments referring to the variation in the relative abundance of planktonic foraminifera through the various glacial and interglacial cycles, relate to the ecological preferences of each planktonic foraminifera species outlined in Appendix 1. The initial results for each core are listed in table 2.

Table 2
The initial data of the three piston cores analysed in this study and the details of this table are discussed below.

<table>
<thead>
<tr>
<th>Core ID</th>
<th>Length of core (cm)</th>
<th>Latitude °S</th>
<th>Longitude °E</th>
<th>Depth (mbsl)</th>
<th>Ave Sediment Rate cm/1000yrs</th>
<th>Sample interval (cm)</th>
<th>Time between samples (yrs B.P)</th>
<th>Previous Work</th>
</tr>
</thead>
<tbody>
<tr>
<td>BAR9403</td>
<td>275</td>
<td>5°49.20'</td>
<td>103°61.90'</td>
<td>2034</td>
<td>8.3</td>
<td>5cm</td>
<td>399-894</td>
<td>Murgese and De Deckker (2005)</td>
</tr>
<tr>
<td>MD002361</td>
<td>1360</td>
<td>22°04.92'</td>
<td>113°28.63'</td>
<td>1805</td>
<td>2.5</td>
<td>2-10 cm</td>
<td>143-5000</td>
<td></td>
</tr>
<tr>
<td>(MD61)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MD03-2607</td>
<td>3270</td>
<td>36°57.64'</td>
<td>137°24.39'</td>
<td>865</td>
<td>18.7</td>
<td>Initially at 5cm Increased to 60cm due to high sedimentation rate</td>
<td>375-3259</td>
<td>Gingele et al. (2004); Gingele and De Deckker (2005)</td>
</tr>
<tr>
<td>(MD2607)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

8.1 Core BAR9403

8.1.1 Age Model

The age model of core BAR9403 is based on the $\delta^{18}$O record from planktonic foraminifera *Gs. ruber*. The following isotopic events 1.1, 2.0 and 3.0 were identified (see Fig. 25 and Appendix 2).
Figure 25
Age model for BAR9403 from species *Gs. ruber* with the isotopic events of Martinez et al. (1987) indicated by the MIS numbers on the δ¹⁸O curve.

8.1.2 Isotope Analysis

This short core (275cm) records the isotopic events from the LGM to the present day. The lightest δ¹⁸O value (-2.99‰) is recorded during the Holocene while the LGM is positioned at the heaviest δ¹⁸O value of -0.59‰ (see Fig. 26). The mean δ¹⁸O range from the glacial to interglacial extreme in BAR9403 is 2.08‰. With the current estimated ice volume effect at 1.05 ‰ (Duplessy et al., 2002) a significant change in sea-surface temperature (SST) and/or salinity must account for the remaining 1.03‰ Δ δ¹⁸O (see Section 8.1.5). The δ¹³C record ranges from a minimum at approximately 8500 yrs BP of 0.16‰ to a maximum value of 1.31‰ during MIS 3.
8.1.3 Age versus Depth

The age versus depth profile shows a fairly constant sedimentation rate in core BAR9403 with an average sedimentation rate of 8cm/1000 yrs (see Fig. 27). The sedimentation rate for the Holocene period is 6cm/1000 yrs, for MIS 2 is 6cm/1000 yrs, and for the late MIS 3 section is 12cm/1000 yrs. However, it should be noted that only a part of MIS 3 is represented in BAR9403 and may display a higher sedimentation rate.
Figure 27
Age versus depth profile for BAR9403 showing a near constant sedimentation rate which increases down the core.

8.1.4 Faunal Analysis

A total of 24,374 planktonic foraminiferal individuals were identified from core BAR9403. This is equivalent to approximately 435 individuals per sample, at an average sample interval of 5cm. The relative abundance of planktonic foraminifera shows an increase in the pre-upwelling or Deep Chlorophyll Maximum (DCM) species *N. pachyderma* (dextral) during MIS 3 with abundances around 6%. The relative abundance of *N. pachyderma* (dextral) reduces to <2% near the LGM and does not recover in the Holocene (see Fig. 28 and Appendix 3). Another DCM-dwelling, but warmer water species, *N. dutertrei* remains fairly stable throughout the entire core but has a 5% increase in abundance during MIS 3 and at the Holocene/ MIS 2 transition. Tropical species and sub-surface dweller *P. obliquiloculata* increases in abundance during MIS 3 and the Holocene with abundances >10%, compared to <10% during the LGM (see Fig. 28 and Appendix 3).

The most obvious change in relative abundance in core BAR9403 is shown by the sub-polar to transitional ‘upwelling’ species *Ga. bulloides* with abundances of 26% at approximately 14 000yrs BP and 22.7% during the Holocene (see Fig. 28 and
Appendix 3). This is compared to the periods from MIS 3 to the LGM where the relative abundance of *Ga. bulloides* is generally <10%. *Gr. menardii*, a tropical ‘upwelling’ species, also increases its relative abundance during MIS 2 from < 8% during MIS 3 to a peak abundance of 16% at ~ 17 000yrs B.P (see Fig. 28 and Appendix 3).

Tropical and subtropical species are most abundant during the Holocene in core BAR9403. Subtropical species *Gs. ruber* records its highest relative abundance of 22.8% during the Holocene but also records a high relative abundance of 20.5% around the LGM (see Fig. 28 and Appendix 3). Similarly, tropical species *Gs. sacculifer* records its highest relative abundance during the Holocene of 17% but also records a high abundance of 15% at ~19 000yrs B.P.

There is a reduction in the relative abundances of subtropical and tropical species between 35 000yrs B.P to 26 000yrs B.P. This time period corresponds to an increase in the relative abundance of *N. pachyderma* (dextral). The other period of low abundance for *Gs. ruber* and *Gs. sacculifer* is after the LGM with abundances not increasing until ~10 000yrs BP (see Fig. 28 and Appendix 3). This period of low abundance coincides with the increase in abundance of ‘upwelling’ species *Ga. bulloides*.

The relative abundance of *Gn. glutinata* is maintained around 10% and displays no obvious patterns of abundance, as this species is a globally ubiquitous species with a wide range of physical and chemical tolerances this is not unusual.

The relative abundances of the minor species within core BAR 9403 also show a greater influence of deeper waters in the water column during MIS 3. Polar to transitional species *T. quinqueloba* appears in the record with a peak in relative abundance during MIS 3 of 2%. Deep-water dwellers such as *Gr. truncatulinoides* and *Gr. crassaformis* also appear in the record during MIS 3 (see Fig. 29).
The variation in $\delta^{18}O$, $\delta^{13}C$, and the relative abundance of planktonic foraminifera in core BAR9403 versus Specmap Age. Species which prefer tropical-subtropical water conditions are coloured red. *Gr. menardii* is coloured deep red as it represents tropical, upwelling conditions. Ubiquitous species *Gn. glutinata* is indicated in mid green. *N. dutertrei* is coloured lime green as it represents nutrient-rich (DCM) water conditions. *N. pachyderma* is coloured deep blue as it is a cold subsurface water indicator. *Ga. bulloides* is coloured aqua blue for other species, which prefer cooler and nutrient-rich water conditions.
Figure 29
The variation in δ¹⁸O, δ¹³C, and the relative abundance of minor planktonic foraminifera in core BAR9403 versus the Specmap Age. Tropical and subtropical species are coloured red, transitional species purple and cooler water, sub-surface dwellers are coloured deep blue.
8.1.5 Sea-surface Temperature

Sea-surface temperature (SST) estimates were investigated by a modern analogue technique (MAT) and indicated that SST remained relatively stable from MIS 3 to present. The 10 best analogues were recorded for each sample within core BAR9403. The locations of these analogues are shown in figure 30, and are mainly equatorial cores. Many of the best analogues were found from the studies of Prell (1999).

![Figure 30](image)

The core sites from the AUSMAT-F4 database that provides an analogue and SST estimates for the samples in core BAR9403. Original data of analogues are from Prell (1985), Prell et al. (1999) and Martinez and Bedoya (2001).

The mean SST (Tmean) is generally above 28°C ± 0.84°C with the highest mean temperature of 28.6°C ± 0.84°C recorded during MIS 3, MIS 2 and the Holocene (Fig 31). The temperature estimate of the warmest month (Tmax) fluctuates between 29.8°C ± 0.87°C and 29.1°C ± 0.87°C and the temperature of the coolest month (Tmin) fluctuates from 24.7°C ± 0.98°C and 27.3°C ± 0.98°C during all of the isotopic stages. The warmest Tmax of 29.8 °C ± 0.84°C is recorded at numerous times during MIS 2 and the Holocene.
Importantly, at the LGM, the mean SST remained relatively unchanged at 28.5°C ± 0.84°C. The largest deviation in SST is recorded at approximately 14,000 yrs BP with a temperature of 26.1°C ± 0.84°C (Tmean). The time frame of 14,000 yrs BP also records the coldest measure of Tmax (28.7°C ± 0.87°C) and Tmin (23.5°C ± 0.98°C) respectively. Two other periods of lowered SST are 10,200 yrs BP with a Tmean of 26.4 °C± 0.84°C which may represent the Younger Dryas event (Fig. 31) and at ~8000 yrs BP with a Tmean 27.6°C± 0.84°C.

The mean annual depth of the mixed layer was also estimated from the MAT. This is an additional tool that can give us insight into the dynamics of the mixed layer. The present MAT mixed layer appears to be quite shallow at an annual thickness of 31m. Obviously, this is an average estimate from the ‘seasonal’ extremes of the Java ‘upwelling’ system present in this area. A reference line has been placed at the modern day MAT estimate of 31m. In figure 31, the mixed layer is shallower than at present during the later stages of MIS 2 and at the MIS 2-Holocene transition. The shallowest mixed layer of -29m occurs at ~11 200yrs BP, while the mixed layer is deeper during MIS 3, reaching a maximum depth of -41m.

Temperature estimates at the 50m 100m and 150 m water depths indicate that the water temperatures are pretty stable at these depths but do change during the 10 000-15 000yrs BP interval. The most dramatic temperature change is at 50m. Colder water also appears to overall reduce the temperature of the mixed layer during MIS 3 (see Fig. 31).

Most of the SST estimates have good squared-chord distances of <0.2 except towards the end of the core in MIS 3. Estimates of <0.2 indicate that the samples have good analogues in the AUSMAT-F4 database. The highest quality estimates are in the youngest interval (<5 000 yrs BP) where the distances are ~0.1 or less. Below this level, distances are mostly >0.1 (see Fig. 31). Higher square chord distances coincide with the occurrence of cold water species such as *N. pachyderma* (dextral), which is not surprising at a tropical core site.
Various temperature estimates of core BAR9403 at different depths (sea-surface, 50m, 100m and 150m) compared to the $\delta^{18}$O and $\delta^{13}$C record through time. The various extremes of SST are also calculated, Tmean represents the mean annual temperature, Tmax the warmest month and Tmin the coldest month at the sea-surface. Note the decrease in all temperature estimate for the 15 000-10 000yrs BP period. The highlighted section represents the timing of MIS 2 with the coldest stage at the LGM.
8.1.6 Factor Analysis

Q-mode Factor analysis identified two relevant factors in the species counts of BAR9403. The first factor accounts for ~ 90.6% of the variance in core BAR9403 (see Table 3), and is dominated by subtropical and DCM species *N. dutertrei* (see Table 4). The second and third highest species scores are recorded by subtropical species *Gs. ruber* and ‘upwelling’ species *Ga. bulloides*, respectively. *N. dutertrei* is associated with warm subtropical water and the increase of nutrients at the thermocline. The association between *N. dutertrei* and other species with high species scores in Factor 1 is not very clear due to the different habitat preferences of these species (see Table 4). This may be due to the relatively low sedimentation rates and, therefore, mixing of seasonal assemblages. Bi-plots (see Fig. 32) and principal component analysis (see Section 8.1.7) will be used to explore these relationships and provide more insight to whether it is temperature or nutrients which cause 90.6% of the variance in core BAR9403.

Table 3

The variance of the species counts described by the Factor value of each Factor axis for core BAR9403. Most of the variance in the species counts is explained by Factor 1. The amount of variance explained in the species counts decreases from Factor 1 to Factor 5.

<table>
<thead>
<tr>
<th>Factor axis</th>
<th>Factor value</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.906</td>
</tr>
<tr>
<td>2</td>
<td>0.031</td>
</tr>
<tr>
<td>3</td>
<td>0.025</td>
</tr>
<tr>
<td>4</td>
<td>0.012</td>
</tr>
<tr>
<td>5</td>
<td>0.007</td>
</tr>
</tbody>
</table>

Factor 2 accounts for 3.1% of the variance and is dominated by the positive scores of *Ga. bulloides* and *Gr. menardii* (see Table 3 and 4). Highly negative scores in Factor 2 are recorded by tropical-subtropical species (*Gs. ruber, Gs. sacculifer, P. obliquiloculata* and *Gn. glutinata*). Negative values are meaningful, both mathematically and ecologically (Imbrie and Kipp, 1971), and indicate the lack of covariance to the dominant species. The 3% variation displayed in Factor 2 appears to be
due to ‘upwelling’ as both of the dominant species (*Ga. bulloides* and *Gr. menardii*) are known to represent ‘upwelling’ systems in tropical regions.

Table 4
The Factor scores of species from Factor analysis. The dominant species which have positive scores in Factor 1 and Factor 2 are highlighted in bold text and account for 93.5% variance in core BAR9403.

<table>
<thead>
<tr>
<th>Species</th>
<th>Species score Component 1</th>
<th>Species score Component 2</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Ga. bulloides</em></td>
<td>0.31</td>
<td>0.51</td>
</tr>
<tr>
<td><em>Ga. falconensis</em></td>
<td>0.08</td>
<td>0.09</td>
</tr>
<tr>
<td><em>T. quinqueloba</em></td>
<td>0.02</td>
<td>0.00</td>
</tr>
<tr>
<td><em>Gt. rubescens</em></td>
<td>0.12</td>
<td>0.00</td>
</tr>
<tr>
<td><em>Gt. tenellus</em></td>
<td>0.02</td>
<td>-0.05</td>
</tr>
<tr>
<td><em>Gs. ruber</em></td>
<td><strong>0.40</strong></td>
<td>-0.41</td>
</tr>
<tr>
<td><em>Gs. sacculifer</em></td>
<td><strong>0.22</strong></td>
<td>-0.30</td>
</tr>
<tr>
<td><em>Gs. conglobatus</em></td>
<td>0.03</td>
<td>-0.05</td>
</tr>
<tr>
<td><em>O. universa</em></td>
<td>0.02</td>
<td>-0.01</td>
</tr>
<tr>
<td><em>Gl. aequilateralis</em></td>
<td>0.08</td>
<td>-0.03</td>
</tr>
<tr>
<td><em>Gl. calida</em></td>
<td>0.06</td>
<td>-0.05</td>
</tr>
<tr>
<td><em>Gn. glutinata</em></td>
<td><strong>0.29</strong></td>
<td>-0.14</td>
</tr>
<tr>
<td><em>Gr. inflata</em></td>
<td>0.01</td>
<td>0.00</td>
</tr>
<tr>
<td><em>Gr. hirsuta</em></td>
<td>0.02</td>
<td>0.04</td>
</tr>
<tr>
<td><em>Gr. scitula</em></td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td><em>Gr. truncatulinoides</em></td>
<td>0.00</td>
<td>0.00</td>
</tr>
<tr>
<td>(s)</td>
<td></td>
<td></td>
</tr>
<tr>
<td><em>Gr. truncatulinoides</em></td>
<td>0.03</td>
<td>-0.03</td>
</tr>
<tr>
<td>(d)</td>
<td></td>
<td></td>
</tr>
<tr>
<td><em>Gr. menardii</em></td>
<td><strong>0.20</strong></td>
<td>0.27</td>
</tr>
<tr>
<td><em>Gr. tumida</em></td>
<td>0.01</td>
<td>0.00</td>
</tr>
<tr>
<td><em>Gr. ungulata</em></td>
<td>0.02</td>
<td>0.00</td>
</tr>
<tr>
<td><em>N. pachyderma (d)</em></td>
<td>0.06</td>
<td>0.02</td>
</tr>
<tr>
<td><em>N. dutertrei</em></td>
<td><strong>0.66</strong></td>
<td>0.32</td>
</tr>
<tr>
<td><em>P. obliquiloculata</em></td>
<td><strong>0.31</strong></td>
<td>-0.51</td>
</tr>
<tr>
<td><em>G. hexagona</em></td>
<td>0.01</td>
<td>-0.01</td>
</tr>
</tbody>
</table>
Figure 32

The bi-plot of the species scores of Factor 1 and Factor 2 from core BAR9403 suggests two ecological groups. The length of the vectors represents the dominant species of each group, such as *N. dutertrei* (Group 1) and *Gs. ruber* (Group 2). Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin of the plot.

The ecological relationships are made clearer when the species scores of Factor 1 and Factor 2 are plotted in a bi-plot (see Fig. 32). The dominant species in Factor 1 and Factor 2 which have a preference for nutrients within the water column form Group 1. The vector lengths indicate that *N. dutertrei* is the most dominant species in Group 1. Group 2 is represented by oligotrophic, subtropical-tropical species and are characterised by the positive scores in Factor 1 and negative scores in Factor 2. The vector lengths reveal that subtropical and oligotrophic species *Gs. ruber* is the dominant species of this group (see Fig. 32).
The relative percentage abundance of the two ecological groups for core BAR9403, revealed by Q-mode factor analysis (from Table 4) compared to the $\delta^{18}O$ and $\delta^{13}C$ record through time. The highlighted section represents the timing of MIS 2 with the coldest stage at the LGM.
The cumulative species abundances of each group are plotted against time (see Fig. 33) and reveal different stages of dominance. The nutrient responsive species of Group 1 increase its abundance during MIS 3 and around the Holocene-MIS 2 transition. The subtropical-tropical oligotrophic species of Group 2 are more dominant during the initial stages of MIS 2, including the LGM, and the late Holocene.

8.1.7 Principal Component Analysis

![Scree Test BAR9403](image)

**Figure 34**
The scree plot of the first 10 Eigenvalues from the species abundance counts of core BAR9403, showing how less variance is explained for each consecutive Eigenvalue.

Analysis via the scree test (Cattell, 1966) was conducted on the species scores from core BAR9403 to show how many components should be retained to describe the variation of relative abundance within the core (see Fig. 34). The main variance is explained by Component 1 (Eigenvalue 1). However, relatively high variance is explained in the first four components, and according to the scree test 4 Eigenvalues (components) it should be analysed to understand the variance in core BAR9403 (see Fig. 34). Beyond Eigenvalue 4-5, the variance is significantly reduced and therefore can be disregarded.
In 56 observations, 28 variables were identified and analysed. Component 1, which explains 32.3% of variance, is dominated with positive scores of *Ga. bulloides*, *Gr. menardii* and *N. dutertrei* (see Table 5 and 6). Component 1 is also associated with a high negative score by *P. obliquiloculata* (see Table 6). Compared to the mixed signal of the first factor in the Q-mode analysis (temperature and nutrients), the first component appears to be solely related to the presence of nutrients due to the positive response from the ‘upwelling’ species *Ga. bulloides* and *Gr. menardii* and the DCM species *N. dutertrei*. The negative values from the tropical and subtropical species *P. obliquiloculata*, *Gs. sacculifer* and *Gs. ruber* confirm this observation (see Table 6).

**Table 5**
The Eigenvalues of the first five axes resulting from principal component analysis. The variance within the species counts of core BAR9403 is explained by each axis, which decreases for each successive axis.

<table>
<thead>
<tr>
<th>Axis</th>
<th>Eigenvalue</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.323</td>
</tr>
<tr>
<td>2</td>
<td>0.244</td>
</tr>
<tr>
<td>3</td>
<td>0.147</td>
</tr>
<tr>
<td>4</td>
<td>0.070</td>
</tr>
<tr>
<td>5</td>
<td>0.063</td>
</tr>
</tbody>
</table>

Component 2, which accounts for 24.4% of the variance in core BAR9403, is dominated by *N. pachyderma* (dextral), *N. dutertrei*, *P. obliquiloculata* and *Gn. glutinata* (see Table 5 and 6). This component is also associated with strong negative responses by *Ga. bulloides*, *Gs. ruber*, *Gs. sacculifer*. The dominant species are non-spinous and live in the sub-surface layers of the water column. *N. pachyderma* and *N. dutertrei* are associated with the development of DCM layer in the tropics. Whereas the species with negative responses are symbiont-bearing species that live close to the sea-surface and prefer oligotrophic conditions. Therefore, Component 2 seems to be associated with subsurface-dwelling species.
Table 6
The species scores for the first four components in core BAR9403. The species highlighted in bold are the dominant species of each component.

<table>
<thead>
<tr>
<th>Species</th>
<th>Component 1</th>
<th>Component 2</th>
<th>Component 3</th>
<th>Component 4</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Ga. bulloides</em></td>
<td>2.96</td>
<td>-2.03</td>
<td>1.00</td>
<td>0.31</td>
</tr>
<tr>
<td><em>Ga. falconensis</em></td>
<td>0.49</td>
<td>-0.10</td>
<td>0.79</td>
<td>-0.08</td>
</tr>
<tr>
<td><em>T. quinqueloba</em></td>
<td>-0.01</td>
<td>0.21</td>
<td>0.17</td>
<td>0.12</td>
</tr>
<tr>
<td><em>Gs. rubescens</em></td>
<td>-0.11</td>
<td>0.26</td>
<td>1.36</td>
<td>0.50</td>
</tr>
<tr>
<td><em>Gs. tenellus</em></td>
<td>-0.25</td>
<td>0.08</td>
<td>0.09</td>
<td>0.00</td>
</tr>
<tr>
<td><em>Gs. ruber</em></td>
<td>-1.50</td>
<td>-2.47</td>
<td>-0.75</td>
<td>-2.74</td>
</tr>
<tr>
<td><em>Gs. sacculifer</em></td>
<td>-1.06</td>
<td>-2.41</td>
<td>-3.09</td>
<td>1.92</td>
</tr>
<tr>
<td><em>Gs. conglobatus</em></td>
<td>-0.19</td>
<td>0.03</td>
<td>-0.39</td>
<td>0.18</td>
</tr>
<tr>
<td><em>O. universa</em></td>
<td>-0.01</td>
<td>-0.08</td>
<td>-0.16</td>
<td>0.19</td>
</tr>
<tr>
<td><em>S. dehiscens</em></td>
<td>-0.02</td>
<td>0.01</td>
<td>0.00</td>
<td>0.02</td>
</tr>
<tr>
<td><em>G. aequilateralis</em></td>
<td>0.03</td>
<td>-0.51</td>
<td>-0.42</td>
<td>0.38</td>
</tr>
<tr>
<td><em>G. calida</em></td>
<td>-0.17</td>
<td>-0.26</td>
<td>-0.11</td>
<td>0.23</td>
</tr>
<tr>
<td><em>G. digitata</em></td>
<td>0.00</td>
<td>-0.02</td>
<td>0.02</td>
<td>0.03</td>
</tr>
<tr>
<td><em>Gn. glutinata</em></td>
<td>-0.51</td>
<td>1.01</td>
<td>0.64</td>
<td>-1.11</td>
</tr>
<tr>
<td><em>C. nitida</em></td>
<td>-0.03</td>
<td>-0.03</td>
<td>-0.02</td>
<td>0.05</td>
</tr>
<tr>
<td><em>G. inflata</em></td>
<td>-0.02</td>
<td>0.13</td>
<td>0.00</td>
<td>-0.11</td>
</tr>
<tr>
<td><em>G. hirsuta</em></td>
<td>0.22</td>
<td>0.05</td>
<td>0.00</td>
<td>-0.15</td>
</tr>
<tr>
<td><em>G. scitula</em></td>
<td>-0.01</td>
<td>0.13</td>
<td>0.23</td>
<td>0.14</td>
</tr>
<tr>
<td><em>Gr. truncatulinoides (s)</em></td>
<td>-0.01</td>
<td>0.02</td>
<td>0.03</td>
<td>0.03</td>
</tr>
<tr>
<td><em>Gr. truncatulinoides (d)</em></td>
<td>-0.20</td>
<td>0.63</td>
<td>0.26</td>
<td>-0.06</td>
</tr>
<tr>
<td><em>Gr. menardii</em></td>
<td>1.95</td>
<td>-0.50</td>
<td>-0.89</td>
<td>-3.59</td>
</tr>
<tr>
<td><em>Gr. tumida</em></td>
<td>0.00</td>
<td>0.12</td>
<td>-0.12</td>
<td>-0.18</td>
</tr>
<tr>
<td><em>Gr. ungulata</em></td>
<td>0.00</td>
<td>0.15</td>
<td>0.06</td>
<td>0.09</td>
</tr>
<tr>
<td><em>N. pachyderma (s)</em></td>
<td>-0.02</td>
<td>0.00</td>
<td>0.00</td>
<td>-0.01</td>
</tr>
<tr>
<td><em>N. pachyderma (d)</em></td>
<td>0.05</td>
<td>1.68</td>
<td>-0.15</td>
<td>-0.26</td>
</tr>
<tr>
<td><em>N. dutertrei</em></td>
<td>2.46</td>
<td>2.07</td>
<td>-3.35</td>
<td>0.27</td>
</tr>
<tr>
<td><em>P. obliquiloculata</em></td>
<td>-2.29</td>
<td>1.63</td>
<td>-1.16</td>
<td>-1.33</td>
</tr>
<tr>
<td><em>G. hexagona</em></td>
<td>-0.01</td>
<td>0.06</td>
<td>-0.02</td>
<td>-0.17</td>
</tr>
</tbody>
</table>

Component 3 accounts for 14.7% of the variance and is dominated by the positive score of *Gt. rubescens* and the strongly negative responses by *Gs. sacculifer, N. dutertrei* and *P. obliquiloculata* (see Table 5 and 6). *Gt. rubescens* lives in the upper 50m and typifies the central subtropical regions of the Indian Ocean, where the waters are low in phosphate, very high in salinity, with intermediate levels of oxygen and temperature (Be, 1977). Another species with a relatively high species score is *Ga. bulloides* which also lives within 50-100mbsl and tolerates high salinities. However, *Ga. bulloides* is a subpolar-transitional species that likes waters high in phosphate. Component 3 could represent high salinity conditions in the upper surface layers; however, *Gs. sacculifer* responds negatively in component 3 which is a species also known to thrive in waters of high salinity.
Component 4 accounts for 7% variance and is dominated by the high positive species score of *Gs. sacculifer* and also associated with high negative species scores of *Gr. menardii*, *Gs. ruber*, *P. obliquiloculata*, *Gn. glutinata* (see Table 5 and 6). The significance of this component is difficult understand due to the monospecific dominance of *Gs. sacculifer* and the negative responses of fellow tropical-subtropical surface and sub-surface dwelling species. It is possible this component is also related to salinity as *Gs. sacculifer* is very tolerant to salinity extremes compared to the other planktonic foraminiferal species.

In the Q-mode factor analysis presented above, species *Ga. bulloides*, *N. dutertrei* and *Gr. menardii* are the dominant species of Factor 2 (Group 1) representing 11% of the variance in the species counts. However, the PCA analysis shows that the main reason for the variance in core BAR9403 is due to nutrient-rich water influencing the core region and the variance is not due to temperature. In order to understand this disparity between the two statistical techniques, component plots (bi-plots) will be used to define ecological groups within the principal component data as the species scores are complex (often an association occurs with temperature, nutrients and salinity). In addition, they may explain what physical-chemical parameter these groups are responding to.

In the component plots (see Fig. 35), the distance between the species relates to the similarity of the species and, therefore, the greater the angle between the vectors the greater the difference between the species. In addition, the vector or line length indicates the importance of the species in explaining the variance of the species counts. The species scores of the first two components are displayed in the plots above and explain 57% of the variance within core BAR9403 and noted as (1-2) behind the group number.
Figure 35
Bi-plot of Components 1 and 2 that explain 58% of the variance in core BAR9403 and reveal four ecological groups within the data. Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin.

The PCA analysis has split the Q-mode Group 1 (analyses above) into two groups (Group 1 (1-2) and Group 2 (1-2) revealing differing responses during the MIS’s. Group 1 (1-2) consists of species with positive scores in component 1 and 2. *N. dutertrei* is the principal species in Group 1 (1-2). The other primary species *N. pachyderma* (dextral) is separated by a significant angle from *N. dutertrei*, lying close to the Group 1(1-2) and Group 4 (1-2) boundary, thus reducing the correlation between these species (see Fig. 35). Both *N. dutertrei* and *N. pachyderma* (dextral) have a preference for a thermally, stratified water column, with a pronounced chlorophyll maximum (Thunell and Reynolds-Sautter, 1992). The samples from BAR9403 that are dominated by group one (see Table 7, indicated in lime green) are primarily at the base of the core during MIS 3, from 33 000-31 000 cal yrs B.P and, momentarily, at 29 000, 25 000 and 11 000 cal yrs B.P (see Table 7). The general patterns in Table 7 indicate...
cooler water with greater influx of nutrients during MIS 3 and around 11,000 yrs BP (possibly a Younger Dryas event) and the development of a DCM layer.

Group 2 (1-2) consists of species with positive species scores in the first component but negative scores in the second component. The primary species are *Ga. bulloides* and *Gr. menardii*, with secondary species *Gl. aequilateralis* and *Ga. falconensis*. The dominant species, *Ga. bulloides* and *Gr. menardii*, are both considered to be 'upwelling' indicators in tropical regions. Group 2 (1-2) does not show a response to the existence of colder water and possible nutrients within the photic zone during MIS 3 as displayed by the Group 1 (1-2) species (see Fig. 36). This may indicate that the mixed layer was stratified and nutrients did not reach the sea-surface and, therefore, 'upwelling' conditions were not established during MIS 3. However, this group does show an increase of abundance of 25% from ~17,000 yrs B.P to ~7,000 yrs B.P. as indicated by the sample scores and relative abundance graph (see Fig. 36 and Table 7), and may indicate that 'upwelling' conditions did occur during this time period.

Table 7

The positive and/or negative responses for each group compared to the sample (depth/age) in the first and second components of core BAR9403. Lime Green = Group 1(1-2) dominated by *N. dutertrei*, aqua blue = Group 2(1-2) dominated by *Ga. bulloides*, bright red = Group 3(1-2) dominated by *Gs. ruber* and *Gs. sacculifer*, mid dark blue = Group 4(1-2) dominated by *P. obliquiloculata*. Note how the blocks of colour relates to the dominant ecological group for that sample.
Group 3 (1-2) contains species with negative scores in the first and second components. Principal species *Gs. sacculifer* and *Gs. ruber* are highly correlated due to the small angle separating their scores. The secondary species include *Gl. calida* and *O. universa*. All of these species prefer tropical-subtropical water with oligotrophic conditions. In the present day, *Gs. sacculifer* is the most abundant species in tropical regions and *Gs. ruber* is the most abundant species in subtropical waters (Bé and Hutson, 1977).

Group 4 (1-2) contains species with negative scores in the first component and positive scores in the second component. The dominant species is *P. obliquiloculata* and associated with secondary species *Gn. glutinata*, *Gr. truncatulinoides* (dextral), *Gt. rubescens*, *Gt. tenellus*, *Gs. conglobatus*. Some of the species of this group preferentially live in deeper water (>100m depth) such as subtropical species *Gr.*
truncatulionoides (dextral) which lives between 100-2000m and is the deepest dwelling planktonic foraminifer. The other tropical-subtropical secondary species generally live between 50-100m with no known preference for nutrients, but they have relatively low scores in this group. The dominant species P. obliquiloculata has been associated with the South Equatorial Current (Bé and Hutson, 1977) and generally lives in tropical waters at ~100m water depth (Schiebel and Hemleben, 2005).

Part of Group 4 contains T. quinqueloba, Gr. ungulata, Gr. inflata, Gr. scitula, Gr. tumida and Gr. hexagona. These species have low scores lying on the mid-line between Group 4 and 1 (see Fig. 36). When questioning the association with Group 4, an additional group was investigated containing these low score mid-line species. However, a parallel response with the principal species of Group 4 was revealed. Species Gr. ungulata, Gr. scitula, Gr. tumida are deep-water dwellers with average occurrences between 100-300m. Therefore, it is assumed this group represents either the SEC or warm water at depth. The sample scores indicate this group dominates during MIS 3 between ~30 000 yrs B.P to 26 000yrs BP (see Table 7 and Fig. 36).
The combined relative percentage abundances of each ecological group previously defined by comparing Component 1 and Component 2 of the species counts in a bi-plot, projected over time and compared to the isotope record in core BAR9403. The highlighted section represents the timing of MIS 2 with the coldest stage at the LGM.
To compare the first four components of core BAR9403 species, scores for each component were plotted in various combinations within bi-plots (see Appendix 4). Most of the combinations confirmed that the main cause of variation in the planktonic foraminiferal counts is possibly due to nutrients. Component 2 versus component 3 defines the ecological groups within 39.1% of variance. The ecological groups of component 2 and 3 did not show strong patterns of abundance change through the MIS’s (see Appendix 4).

Figure 37
Ecological groups defined by comparing Component 1 versus Component 3. The bi-plots of Components 1 versus Component 3 explain 58% of the variance in core BAR9403, thus revealing four ecological groups within the data. Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin.

The ecological groups of Component 1 versus Component 3 are very similar to the ecological groups of Component 1 versus Component 2 (see Fig. 37 and 35). Group
1 (1-3) is dominated by ‘upwelling’ species *Ga. bulloides* and *Ga. falconensis*, while Group 2 (1-3) is dominated by *N. dutertrei* and *Gr. menardii*, with secondary species *Gl. aequilateralis* and *N. pachyderma* (dextral).

An inverse relationship between Group 1 (1-3) and Group 2 (1-3) is shown in the abundance curves (see Fig. 38). The ‘upwelling’ signal is reduced in Group 1 (1-3) compared to Group 2 (1-2) with *Gr. menardii* removed from the ‘upwelling’ group in 2(1-2) and placed in the DCM Group 2 (1-3). It is possible that the association of *Gr. menardii* and *N. dutertrei* is because of their preference for similar temperatures (both tropical-subtropical species) than a nutrient signal. This is because the addition of *Gr. menardii* does not change the patterns of the DCM group whereas it does change the signal and strengthens the ‘upwelling’ group of Group 2 (1-2; see Fig. 36 and 38). The species competition between *N. dutertrei* and *Ga. bulloides* was noted in previous studies and was linked to different food sources. This is evident in their individual species counts (see Fig. 28) and are the driving forces behind the patterns of Group 1 (1-3) and Group 2 (1-3); see Fig. 38.

Group 3 (1-3) show the similar patterns of relative abundance as seen in Groups 3 and 4 (1-2) with the increase of tropical-subtropical species during the Holocene and the MIS 2-MIS 3 transition. The principal species of Group 3 (1-3) include *Gs. sacculifer, P. obliquiloculata, Gs. ruber* and secondary species *Gs. conglobatus* and *Gl. calida*. These species dominate the modern-day tropical and subtropical provinces (Bé and Hutson, 1977).

Group 4 (1-3) is not as well defined as Group 4 (1-2) but it does also seem to show periodic pulses of deep water. This group is dominated by *Gn. glutinata, Gr. truncatulinoides* (dextral) and *Gt. tenellus.*
Figure 38
The combined relative percentage abundances of each ecological group from the comparison of Component 1 and Component 3 in core BAR9403. Note the inverse relationship of ‘upwelling’ Group 1 (1-3) dominated by *Ga. bulloides* and DCM Group 2 (1-3) dominated by *N. dutertrei*. 
Figure 39
Ecological groups defined by comparing Component 3 versus Component 4. The bi-plot of Components 3 versus Component 4 explain 22% of the variance in core BAR9403, thus revealing four ecological groups within the data. Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin.

Component 3 versus Component 4 explains ~22% of variance. By plotting Component 3 versus Component 4, it was hoped to gain insight into dominance of *Gt. rubescens* and *Gs. sacculifer* species scores and whether this is due to temperature or salinity effects. However, *Gt. rubescens* and *Gs. sacculifer* were not in the same group in the bi-plot. The variance displayed by *Gt. rubescens* was minor with a short vector length compared to the species with lower salinity tolerances and tropical and subtropical temperature preferences (see Fig. 39). A graph of the relative abundances of each group was also analysed (see Appendices 5). The graph displays more random patterns compared to previous component abundance graphs but shows increases of possible salinity groups during the MIS 2 (see Appendix 5). However, the larger abundance counts of *Ga. bulloides* and *N. dutertrei* appear to drown out the signal of *Gt. rubescens* and *Gs. sacculifer* respectively. The inter-relationship between *Ga. bulloides*
and *Gt. rubescens*, *Gs. sacculifer* and *N. dutertrei* may be due to similar timing of increased abundance of each of the species. MIS 2 is also believed to be a drier climate compared to the present and, therefore, could coincide with periods of increased salinity.

A more convincing pattern of variance due to salinity effects was found when comparing the species scores of component 1 versus component 4. In this bi-plot *Gs. sacculifer* and *Gt. rubescens* are placed in the same group (Group 4 (1-4)); see Fig. 40. Group 4 (1-4) also showed a different pattern of abundance compared to previous groups (see Fig. 41), with peak abundances over the LGM. Groups 1 to 3 (1-4) showed similar patterns of abundance change discussed above (see Fig. 41)

![Figure 40](image)

Ecological groups defined by comparing Component 1 versus Component 4 in core BAR9403. The bi-plot of Components 1 versus Component 4 explain 38% of the variance in core BAR9403, thus revealing four ecological groups within the data. Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin.
The abundance peaks of Group 4 (1-4) are not parallel with the abundance peaks of the warm group represented by Group 3 (1-4; see Fig. 41). *Gs. sacculifer* and *Gt. rubescens* are tropical and subtropical species respectively and their tolerance for high salinity conditions may explain why they are removed from the other warm water species to form Group 4 (1-4). This is supported by the abundance increase of Group 4 (1-4) around the LGM. Another abundance peak is seen during the Holocene which may be also temperature signal. However, SST remained high and stable around the time of peak abundances of Group 4 (1-4) (see Fig. 41) so it is possible a change in sea-surface salinity is causing the variation in Component 4.

The common situation with all of the bi-plots is that subtropical-tropical water conditions were maintained during the last 35 000 yrs B.P with temperature groups forming negative responses to the variance displayed in each component. Therefore, the major causes of variance in core BAR9403 are due to nutrients and the influx of deeper water and possibly a very minor cause of variance (7%) are due to salinity changes.
Figure 41

The combined relative abundances of each ecological group from the comparison of Component 1 and Component 4 in core BAR9403. Note the new ecological group 4 (1-4) which may be a signal for increased sea-surface salinity coloured in yellow. The previously noted DCM group 1 (1-4) coloured in lime green, ‘upwelling’ group 2 (1-4) in aqua green and warm water species Group 3 (1-4) in red.
8.2 Core MD61

8.2.1 Age Model

As stated in Chapter 7, radiocarbon dating and oxygen isotopes were used to produce an age model for MD61. Radiocarbon dates were obtained from twelve samples of 10mg of *Gs. ruber* in the upper core section (0-203.5cm; see Chapter 7). Three samples were beyond the radiocarbon calibration dataset of Hughen et al. (2004). Dates for the older part of the core were correlated with Bassinot et al. (1994) isotopic stages (see Fig. 42). The radiocarbon samples, and their corresponding ages are listed in Table 8. (see Appendix 6 for the complete age model), and the isotopic stages of MD61 shown in figure 42. The presence of pink *Gs. ruber* was noted at 120 800 yrs BP in accordance with the LAD in the Indian Ocean (Thompson et al. 1979).
Table 8
The radiocarbon ages obtained from ~10mg of species *Gs. ruber*. The two lab codes are because samples were initially prepared at LSCE, Gif-sur-Yvette, final analysis at ANU (see Methods Chapter 8). Ages calibrated within two standard deviations obtained with Calib 5.0 and calibration dataset of Hughen et al. (2004; see Methods Chapter 8) to date upper section of MD61.

<table>
<thead>
<tr>
<th>Lab code</th>
<th>Lab code</th>
<th>Depth (cm)</th>
<th>C^{14} dates</th>
<th>^{14}C cal yrs</th>
<th>95.4% (2(\sigma)) cal age range</th>
<th>Calibration Data</th>
</tr>
</thead>
<tbody>
<tr>
<td>MD61-40175</td>
<td>29604</td>
<td>0-1</td>
<td>1870±170</td>
<td>1364</td>
<td>982-1738</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40176</td>
<td>29605</td>
<td>17.5-18.5</td>
<td>2810±170</td>
<td>2466</td>
<td>2024-2854</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40177</td>
<td>29606</td>
<td>37.5-38.5</td>
<td>3070±180</td>
<td>2781</td>
<td>2327-3233</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40178</td>
<td>29607</td>
<td>45.5-46.5</td>
<td>3280±190</td>
<td>3037</td>
<td>2580-3513</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40180</td>
<td>29609</td>
<td>65.5-66.5</td>
<td>3960±190</td>
<td>3879</td>
<td>3405-4388</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40181</td>
<td>29610</td>
<td>81.5-82.5</td>
<td>4860±180</td>
<td>5072</td>
<td>4607-5530</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40182</td>
<td>29611</td>
<td>99.5-100.5</td>
<td>6030±190</td>
<td>6391</td>
<td>5944-6815</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40188</td>
<td>29617</td>
<td>135.5-136.5</td>
<td>7710±200</td>
<td>8109</td>
<td>7689-8508</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40186</td>
<td>29615</td>
<td>151.5-152.5</td>
<td>9670±210</td>
<td>10472</td>
<td>9956-10705</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40187</td>
<td>29616</td>
<td>175.5-176.5</td>
<td>12730±230</td>
<td>14283</td>
<td>13703-15025</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40185</td>
<td>29614</td>
<td>187.5-188.5</td>
<td>15180±250</td>
<td>17807</td>
<td>16953-18614</td>
<td>(Hughen et al., 2004)</td>
</tr>
<tr>
<td>MD61-40191</td>
<td>29619</td>
<td>203.5-204.5</td>
<td>20330±240</td>
<td>23749</td>
<td>22878-24376</td>
<td>(Hughen et al., 2004)</td>
</tr>
</tbody>
</table>
The δ¹⁸O curve from *Gs. ruber* in core MD61, with the radiocarbon samples indicated by the blue dots and the isotopic tie points of Bassinot et al. (1994) indicated by the bold numbers.

### 8.2.2 Isotope Analysis

The present-day δ¹⁸O value is -1.79‰ compared to a δ¹⁸O value of -0.40‰ at the LGM. The mean δ¹⁸O from the LGM to present is -1.50‰. As stated above, the response of the δ¹⁸O record to global sea-level during the LGM was 1.05‰ (Duplessy et al., 2002). This leaves a Δδ¹⁸O of 0.45‰ to be explained by temperature and/or salinity effects.

The lightest δ¹⁸O value occurred during the MIS 5 with a value of -2.50‰ with the penultimate glacial period (MIS 6.2) recording -0.49‰. The mean δ¹⁸O range from glacial stage MIS 6 to MIS 5 is -1.66‰.

The most extreme isotopic difference recorded in core MD61 is visible at the MIS 12 to MIS 11 transition. The heaviest δ¹⁸O value was recorded during MIS 12 with a value of -0.04‰ while the MIS 11 lightest δ¹⁸O value is -2.49‰ (see Fig. 43). The mean δ¹⁸O range from glacial stage MIS 12 to interglacial stage MIS 11 is -1.62‰.
which is slightly higher than the present day to LGM Δδ¹⁸O. However, the estimated ice-volume contribution to the δ¹⁸O signal at the MIS 11 to MIS 12 transition is ~1.5‰ (Chappell, 1998). Changes in SST and salinity are expected throughout this core due to the residuals in the isotope record after the ice-volume effect has been considered. The range in δ¹³C is from a high of 2.11‰ during MIS 10 and low of -0.05‰ during MIS 5.

Figure 43
The isotope record of core MD61 over the last 550 000 yrs BP obtained from *Gs. ruber* every 5cm. The oxygen isotope record is indicated in red versus the carbon isotope record in green. Foraminiferal counts were conducted up to 550 000 yrs BP and so only the isotopic results of this interval are shown.
8.2.3 Age versus Depth

The age versus depth profile of core MD61 shows a step-like pattern with increased sedimentation rates during interglacial periods and decrease rates during glacial periods (see Fig. 44). The overall sedimentation rate of core MD61 is 2.5cm/1000yrs. The highest sedimentation rate of 15cm/1000 yrs occurs during the Holocene but, during the other interglacial periods, the sedimentation rate is generally closer to 3-5cm/1000 yrs; the exception is MIS 11 which has a low sedimentation rate of 1.7cm/1000 yrs. The lowest sedimentation rate of 0.74cm/1000 yrs occurs during MIS 12. It appears that the sedimentation rate may be the result of varying river discharge. There is an increase of red brown sediment during the interglacial periods (see Figure 23, Chapter 8) and this is shown with an increase in sedimentation rate during MIS 5 and MIS 11.

8.2.4 Faunal Analysis

A total of 66,646 planktonic foraminifera were picked and identified within core MD61. This is equivalent to approximately 436 individuals per sample, at an average sample interval of 9cm. The most abundant species within core MD61 is the subtropical
species *Gs. ruber*. High abundances ≥ 30% are seen during the interglacial periods MIS 11, 9, 7, 5 with peak abundance (37.7%) during the Holocene. Other tropical and subtropical species follow this pattern (red colour in Fig. 45 and Appendix 7). Highest abundance of the tropical species *Gs. sacculifer* occurs during the last interglacial (18.2% peak; 12.7% average), with the next largest peak of 13.7% during MIS 9 but with an average abundance of 8.8%. Interestingly, MIS 11 only recorded a relative abundance peak of 11.4% but has a high average abundance of 9.6%. The tropical, deep-water dweller *P. obliquiloculata* registers a large increase in abundance during the last interglacial (MIS 5) and during MIS 7, with peak values of 17.9% (average 12.6%) and 19 % (average 9.4%) respectively compared to 6% (3.28% average) during the Holocene (see Fig. 45 and Appendix 7). MIS 11 records a lower than expected relative abundance with a peak value of 10.9% and an average abundance of 7.8%. Tropical ‘upwelling’ species *Gr. menardii* has two periods of abundance during the Holocene and MIS 11 with average abundance of 8.3% and 5.9% respectively. The highest relative abundance of 14.3% occurred during the Holocene from a general abundance of ~2% during other marine isotope stages (see Fig. 45 and Appendix 7).

The transitional, deep dwelling species *Gr. inflata* has the most distinct abundance change within core MD61 (see Fig. 45 and Appendix 7). This species is absent during the interglacial periods but achieves high relative abundances (~20-30%) during glacial periods. The highest relative abundance recorded for this species is during the MIS 2 and MIS 3 transition with a value of 32.6% and this may indicate a reduction in the temperature of the water column during glacial periods as this species prefers temperate water conditions.

The sub-polar to transitional species *Ga. bulloides* records its greatest abundance at MIS 11 and MIS 10 of 22% and 17.5% respectively and remains around 10% through all the isotope stages (see Fig. 45 and Appendix 7). As an indicator of cold nutrient-rich water, the increase in abundance of *Ga. bulloides* at MIS 11 is surprising. This could be a lag effect of increased nutrients during MIS 12, as shown by the δ13C record.

Subtropical DCM species *N. dutertrei* does not show a well defined pattern of abundance. *N. dutertrei* does seem to sustain its abundance of ~ 10% during MIS 14 to MIS 11, perhaps indicating nutrients around the thermocline but then reduces in abundance during glacial periods. *Gn. glutinata* has a more undefined pattern as it is achieves high abundance during interglacial and glacial periods. This species is considered ubiquitous in terms of temperature but has been associated to late bloom successions (Hilbrecht, 1996).
Figure 45
The relative abundance of planktonic foraminifera in core MD61 against age (yrs), δ¹⁸O and δ¹³C. *Gs. ruber, Gs. sacculifer, Gr. menardii* and *P. obliquiloculata* are species which favour tropical-subtropical water conditions and are indicated in red. *Gn. glutinata* and *N. dutertrei* are species which prefer cooler and nutrient rich water conditions and are indicated in green. *Gr. inflata* prefers transitional water masses, between subtropical and polar water temperatures and is indicated in purple. *Ga. bulloides*, a subpolar-transitional ‘upwelling’ species is indicated in dark blue. Glacial periods indicated by highlighted sections.
8.2.5 Sea Surface Temperature

For each sample there are 10 analogues which provide the best fit against the planktonic assemblages of core MD61. The SST estimate is calculated as the mean of the best 10 analogues per sample. The locations of the analogues are shown in Figure 46, and cover a wide latitudinal distance due to the glacial and interglacial cycles represented in core MD61. Many of the analogues are found from the studies of Martinez et al. (1999) and Prell et al. (1999).

![Figure 46](image_url)

The sites of cores-tops from the AUSMAT-F4 database that provide the analogues for the sea-surface temperature (SST) estimates in MD61. There are 10 core-top analogues per sample analysed in core MD61.

The quality of SST estimates is measured by the square chord distance (dissimilarity coefficient). The analogue estimate is reduced with values greater than 0.2, conversely squared-chord distances of <0.2, shows that the samples have good analogues in the AUSMAT-F4 database. The stages with reduced quality of SST estimates are during MIS 5 and MIS 7 (see Fig. 47). Higher relative abundances of tropical species *Gs. sacculifer* and *P. obliquiloculata*, compared to the Holocene, seem to result in larger square chord distances and higher estimated temperatures. Higher abundances of minor tropical to subtropical species, *Gs. conglobatus* and *O. universa*,
are also significantly higher during MIS 5 and 7 compared to the Holocene and could be reducing the quality of the analogue. The results from MIS 5 and MIS 7 will not be presented due to high square chord distances.

The sea-surface temperature record of MD61 is an estimate over five interglacial and five glacial periods (see Fig. 47). Most interglacial periods were equivalent or warmer than present. For example, the average temperature for the whole of the interglacial phase for MIS 11 = 28.3°C ± 0.84°C, MIS 9 = 27.9°C ± 0.84°C, and the Holocene = 27.8°C ± 0.84°C (see Fig. 47).

From the oxygen isotope record, it appears that the glacial period MIS 10 is colder than the LGM. The SST estimates confirm that MIS 10.2 is colder with a Tmean of 20.1°C ± 0.84°C, compared to the estimated Tmean of 21.3°C ± 0.84°C during the LGM (see Fig. 47). However, MIS 12.2 and MIS 6.2 are equivalent to the LGM with a Tmean of 20.8°C ± 0.84°C and 20.7°C ± 0.84°C respectively as these estimates are within the error range.

The most extreme glacial to interglacial transition was from the penultimate glacial period (MIS 6) to the last interglacial (MIS 5) with a temperature range of 8.7°C ± 0.84°C. A similar range of 8.6°C ± 0.84°C was also found from the transition of MIS 10 to MIS 9. The seemingly large transition in the isotope record from MIS 12 to MIS 11 only presented a temperature range of 7.8°C ± 0.84°C, while the smallest transition was 6.2°C ± 0.84°C from the LGM to the Holocene (see Fig. 47).

The presence of the Warm Pool is indicated by a summer SST (Tmax) above 28°C (Martinez et al., 1999). The Warm Pool reaches the site at the interglacial periods but is absent during the glacial and cooler periods with a general Tmax range of 22-25°C ± 0.87°C. Compared to the modern Tmax of 28.9°C ± 0.87°C, the warmest Tmax estimate of 29.9°C ± 0.87°C occurs during MIS 11 and there is good confidence of this estimate with a low square chord distance of 1.3. The coolest Tmax estimate of 22.2°C ± 0.87°C occurs during MIS 10. Likewise, MIS 10 records the coldest Tmin (coldest month) of 18°C ± 0.98°C. Seasonality (Tmax-Tmin) indicates that the glacial periods have a more extreme annual temperature range of ~ 4°C compared to ~ 2°C during the interglacial periods (see Fig. 47).

The thickness of the mixed layer was also estimated (in mbsl) which is believed to be the most relevant estimate in subtropical-tropical regions (Barrows and Juggins, 2005). A reference line is placed on the core-top mixed layer estimate of 36mbsl in figure 47 and shows that the mixed layer is thinner during interglacial periods compared to the glacial ones. The depth of the mixed layer is between 39-31mbsl with the
shallowest mixed layer depth recorded in MIS 11. During glacial periods the mixed layer depth lays between 63-34m below the sea surface (see Fig. 47).

The changing thickness of the mixed layer is also indicated in the estimated temperature at 50m, 100m and 150m water depths. Temperature at 50m is greatly reduced from \(~27-28°C \pm 0.84°C\) during the interglacial periods, to \(~20-21°C \pm 0.84°C\) during the glacial periods. Conversely, the temperature at 150m is more stable, ranging between \(~17-19°C \pm 0.84°C\) during interglacial periods to \(~15-16°C \pm 0.84°C\) during glacial periods. The exception is during MIS 5 with much higher temperatures being recorded at 150m (22-24°C \pm 0.84°C), but the high square chord distances are associated with the estimates from MIS 5 and should be viewed with caution (see Fig. 47).
Figure 47

The sea-surface MAT temperature record based on the faunal assemblages of core MD61 versus the $\delta^{18}O$ and $\delta^{13}C$ record through time. The glacial periods are shaded with the respective marine isotope stages indicated by the number.
8.2.6 Factor Analysis

Q-mode Factor analysis (Imbrie and Kipp, 1971) revealed two relevant factors in core MD61. The first two factors explain 93% of the variance in MD61 and the remaining factors are dismissed due to their low values and ability to explain the variance in the species counts (see Table 9). Factor 1 explains 84.8% of the variation within the planktonic foraminifera counts of core MD61.

Table 9
The variance of the species counts in core MD61 explained by each successive factor axis. The variance explained by each axis decreases from Factor 1 to Factor 5.

<table>
<thead>
<tr>
<th>Factor Axis</th>
<th>Factor Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.849</td>
</tr>
<tr>
<td>2</td>
<td>0.083</td>
</tr>
<tr>
<td>3</td>
<td>0.015</td>
</tr>
<tr>
<td>4</td>
<td>0.013</td>
</tr>
<tr>
<td>5</td>
<td>0.011</td>
</tr>
</tbody>
</table>

Factor 1 is dominated by *Gs. ruber* with a species score of 0.7 (see Table 10) and forms a weak association with *Ga. bulloides*, *Gs. sacculifer*, *P. obliquiloculata*, *Gr. menardii* (see Fig. 48). It appears that this group represents species with tropical and subtropical temperature preferences at the sea-surface and sub-surface levels. The inclusion of sub-polar-transitional and ‘upwelling’ species *Ga. bulloides* may be due to its high abundance during MIS 11 as seen in the relative abundances of the individual species.

Factor 2 is heavily dominated by *Gr. inflata* with a species score of 0.92 (see Table 10). *Gr. inflata* prefers transitional waters with little seasonal variation in salinity. *Gr. inflata* is a deep-water dweller with adults found up to 400m below the sea-surface (Hemleben et al., 1989).
Table 10

The species scores of Factor 1 and 2 explaining 93% of variance of the species counts in MD61. Species with large values are more important for explaining the variance in the species counts and species with similar species scores are correlated.

<table>
<thead>
<tr>
<th>Species</th>
<th>Species scores Factor 1</th>
<th>Species scores Factor 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ga. bulloides</td>
<td>0.25</td>
<td>-0.04</td>
</tr>
<tr>
<td>Ga. falconensis</td>
<td>0.14</td>
<td>0.06</td>
</tr>
<tr>
<td>T. quinqueloba</td>
<td>0.02</td>
<td>-0.01</td>
</tr>
<tr>
<td>Gl. rubescens</td>
<td>0.14</td>
<td>0.04</td>
</tr>
<tr>
<td>Gl. tenellus</td>
<td>0.05</td>
<td>-0.01</td>
</tr>
<tr>
<td>Gs. ruber</td>
<td><strong>0.70</strong></td>
<td>-0.30</td>
</tr>
<tr>
<td>Gs. sacculifer</td>
<td>0.24</td>
<td>-0.14</td>
</tr>
<tr>
<td>Gs. conglobatus</td>
<td>0.04</td>
<td>-0.02</td>
</tr>
<tr>
<td>O. universa</td>
<td>0.03</td>
<td>-0.02</td>
</tr>
<tr>
<td>Gl. aequilateralis</td>
<td>0.09</td>
<td>-0.05</td>
</tr>
<tr>
<td>Gl. calida</td>
<td>0.06</td>
<td>-0.02</td>
</tr>
<tr>
<td>Gn. glutinata</td>
<td>0.34</td>
<td>0.08</td>
</tr>
<tr>
<td>Gr. inflata</td>
<td>0.27</td>
<td><strong>0.92</strong></td>
</tr>
<tr>
<td>Gr. hirsuta</td>
<td>0.01</td>
<td>0.00</td>
</tr>
<tr>
<td>Gr. truncatulinoides</td>
<td>0.02</td>
<td>-0.02</td>
</tr>
<tr>
<td>Gr. truncatulinoides (d)</td>
<td>0.03</td>
<td>0.03</td>
</tr>
<tr>
<td>Gr. crassaformis</td>
<td>0.02</td>
<td>0.00</td>
</tr>
<tr>
<td>Gr. menardii</td>
<td>0.14</td>
<td>-0.09</td>
</tr>
<tr>
<td>Gr. tumida</td>
<td>0.01</td>
<td>-0.01</td>
</tr>
<tr>
<td>Gr. unguilata</td>
<td>0.03</td>
<td>-0.05</td>
</tr>
<tr>
<td>N. pachyderma (d)</td>
<td>0.03</td>
<td>0.02</td>
</tr>
<tr>
<td>N. dutertrei</td>
<td>0.30</td>
<td>0.01</td>
</tr>
<tr>
<td><strong>P. obliquiloculata</strong></td>
<td><strong>0.17</strong></td>
<td>-0.12</td>
</tr>
</tbody>
</table>
Figure 48
The two assemblage groups of planktonic foraminifera identified with Q-mode factor analysis (Imbrie and Kipp, 1971) in core MD61. Note the dominance of *Gr. inflata* and *Gs. ruber* over the other species in their corresponding group. Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin.

The dominance of *Gs. ruber* and *Gr. inflata* is obvious when the species scores of Factor 1 and 2 are plotted. Group 2 represents species with a positive score in Factor 1 and 2. The dominance of *Gr. inflata* in Group 2 is indicated by the vector length compared to the other species in the group. The other species in Group 2 have minor associations with *Gr. inflata* as indicated by the short vector lengths and larger angles between the species (see Fig. 48). The minor species of Group 2, in order of importance, are *Gn. glutinata*, *N. dutertrei*, *Ga. falconensis*, *Gt. rubescens*, *Gr. truncatulinoides* (dextral) *N. pachyderma* (dextral) and *Gr. hirsuta*. Some of these species are deep-water dwellers such as *Gr. truncatulinoides* and *Gr. hirsuta*, while others, like *N. dutertrei* and *Ga. falconensis*, have been associated with boundary currents in the Indian Ocean (Bé
and Tolderlund, 1971). However, the major dominance of *Gr. inflata* suggests this group represents the presence of cool oligotrophic deep water with a low salinity gradient.

Group 1 consists of species with a positive score in Factor 1 and a negative score in Factor 2, with *Gs. ruber* being the dominant species. There is a better relationship between *Gs. ruber* and the minor species as indicated by the smaller angles between the species compared to Group 2. This is seen in Figure 48 with the tropical, oligotrophic species *Gs. sacculifer, P. obliquiloculata* and *Gr. menardii* closely grouped together near the vector line of *Gs. ruber*.

When these groups are analysed against the isotope record and through time it shows the dominance of the two groups at different glacial and interglacial stages (see Fig. 49). Group 1 dominates the interglacial stages whereas Group 2 dominates in the glacial periods. Species *Gr. inflata* has been associated with the (South) Indian Central Waters of the Indian Ocean by Martinez et al. (1998) and Martinez et al. (1999) and may suggest that (South) Indian Central waters are influencing the core site during glacial periods.
Figure 49
The two ecological groups of planktonic foraminifera defined by Q-mode factor analysis in MD61. Group 1 containing the warm water species is most abundant during interglacial periods, is indicated in red. Group 2, which contains the transitional species *Gr. inflata*, dominates in the glacial periods; it is indicated in purple.
8.2.7 Principal Component Analysis

Analysis via the scree test of the Eigenvalues (Cattell, 1966) indicates that at least two to four components should be retained to describe the variation within the species of core MD61 (see Fig. 50). The main variance is explained by Component 1 (Eigenvalue 1) and the slope decreases significantly to Eigenvalue 5. The first four components were initially considered as the Eigenvalues of these components are above 5% and investigated for any additional ecological groupings (see Table 11). However, similar or only random groupings were found when comparing the lower value components (see Appendix 8-10), implying that most of the relevant ecological information is explained by the first three components.

![Scree Test](image)

**Figure 50**
The scree plot of the first 10 Eigenvalues from the species counts of MD61, showing how less variance is explained for each consecutive Eigenvalue.

**Table 11**
The variation described by each component (axis) in core MD61. As shown in the scree plot less variance is explained by the lower Eigenvalues

<table>
<thead>
<tr>
<th>Axis</th>
<th>Eigenvalue</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.547</td>
</tr>
<tr>
<td>2</td>
<td>0.108</td>
</tr>
<tr>
<td>3</td>
<td>0.089</td>
</tr>
<tr>
<td>4</td>
<td>0.070</td>
</tr>
<tr>
<td>5</td>
<td>0.043</td>
</tr>
</tbody>
</table>
For 148 observations, 23 variables were identified and analysed. In accordance with other studies (Martinez et al., 1998), species with abundances below 1% for all counts were excluded from the analysis. These species include S. dehiscens, H. pelagica, B. digitata, C. nitida, T. iota, Gr. scitula and N. pachyderma (sinistral).

Table 12
The species scores for the first four components described by principal component analysis in MD61. The highlighted numbers represent the dominant species of each component.

<table>
<thead>
<tr>
<th>Species</th>
<th>Component 1</th>
<th>Component 2</th>
<th>Component 3</th>
<th>Component 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ga. bulloides</td>
<td>-0.30</td>
<td>0.55</td>
<td>-1.97</td>
<td>-1.01</td>
</tr>
<tr>
<td>Ga. falconensis</td>
<td>0.20</td>
<td>-0.06</td>
<td>-0.81</td>
<td>0.82</td>
</tr>
<tr>
<td>T. quinqueloba</td>
<td>-0.03</td>
<td>0.27</td>
<td>-0.01</td>
<td>0.17</td>
</tr>
<tr>
<td>Gr. rubescens</td>
<td>0.13</td>
<td>1.46</td>
<td>-0.97</td>
<td>0.43</td>
</tr>
<tr>
<td>Gr. tenellus</td>
<td>-0.08</td>
<td>0.09</td>
<td>-0.22</td>
<td>0.15</td>
</tr>
<tr>
<td>Gs. ruber</td>
<td>-1.92</td>
<td>-3.61</td>
<td>0.02</td>
<td>1.80</td>
</tr>
<tr>
<td>Gs. sacculfer</td>
<td>-0.74</td>
<td>0.02</td>
<td>1.57</td>
<td>-0.47</td>
</tr>
<tr>
<td>Gs. conglobatus</td>
<td>-0.10</td>
<td>0.09</td>
<td>0.42</td>
<td>-0.01</td>
</tr>
<tr>
<td>O. universa</td>
<td>-0.11</td>
<td>-0.08</td>
<td>0.18</td>
<td>0.10</td>
</tr>
<tr>
<td>Gl. aequilateralis</td>
<td>-0.28</td>
<td>-0.69</td>
<td>0.25</td>
<td>-0.32</td>
</tr>
<tr>
<td>Gl. calida</td>
<td>-0.12</td>
<td>-0.18</td>
<td>-0.02</td>
<td>0.06</td>
</tr>
<tr>
<td>Gn. glutinata</td>
<td>0.20</td>
<td>-0.05</td>
<td>-2.63</td>
<td>1.17</td>
</tr>
<tr>
<td>Gr. inflata</td>
<td>4.21</td>
<td>-1.65</td>
<td>0.65</td>
<td>0.41</td>
</tr>
<tr>
<td>Gr. hirsuta</td>
<td>0.01</td>
<td>0.00</td>
<td>-0.04</td>
<td>-0.05</td>
</tr>
<tr>
<td>Gr. truncatulinoides (s)</td>
<td>-0.06</td>
<td>0.48</td>
<td>0.23</td>
<td>0.34</td>
</tr>
<tr>
<td>Gr. truncatulinoides (d)</td>
<td>0.14</td>
<td>0.27</td>
<td>-0.02</td>
<td>0.14</td>
</tr>
<tr>
<td>Gr. crassaformis</td>
<td>-0.01</td>
<td>0.46</td>
<td>0.11</td>
<td>0.41</td>
</tr>
<tr>
<td>Gr. menardii</td>
<td>-0.56</td>
<td>-0.74</td>
<td>-0.31</td>
<td>-1.95</td>
</tr>
<tr>
<td>Gr. tumida</td>
<td>-0.05</td>
<td>-0.03</td>
<td>-0.01</td>
<td>-0.05</td>
</tr>
<tr>
<td>Gr. ungulata</td>
<td>-0.27</td>
<td>-0.47</td>
<td>-0.32</td>
<td>-0.24</td>
</tr>
<tr>
<td>N. pachyderma (d)</td>
<td>0.08</td>
<td>-0.03</td>
<td>0.14</td>
<td>-0.41</td>
</tr>
<tr>
<td>N. dutertrei</td>
<td>-0.08</td>
<td>-0.97</td>
<td>0.58</td>
<td>-3.09</td>
</tr>
<tr>
<td>P. obliquiloculata</td>
<td>-0.54</td>
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Component 1, which explains 54.7% of variance within the species counts of core MD61, is heavily dominated by the positive score of Gr. inflata (see Table 12). There are no other significant positive species scores but there is a highly negative score from Gs. ruber. Component 1 appears to characterise the conditions favoured by Gr. inflata and suggests the occurrence of cool oligotrophic deep water over the core area during glacial periods. In the sample scores, this component dominates during the LGM and MIS 6 (see Table 13, purple).
Component 2 which accounts for 10.8% of variance are dominated by positive responses by and *P. obliquiloculata* and *Gt. rubescens* and is associated with negative responses of *Gs. ruber* and *Gr. inflata* (see Table 12). Species *P. obliquiloculata* is a tropical subsurface dweller and it has a similar species score as *Gt. rubescens*, suggesting an association with *Gt. rubescens* (a species that prefers warm water to exceed 10-12°C at 200m). Therefore this component may represent variance from warm, deep water. The negative response by *Gr. inflata* strengthens this assumption.

Component 3 accounts for 8.9% of the variance and is dominated by positive responses of *P. obliquiloculata* and *Gs. sacculifer* in association with strongly negative responses by *Gn. glutinata* and *Ga. bulloides* (see Table 12). *P. obliquiloculata* and *Gs. sacculifer*, are both tropical species, with *P. obliquiloculata*, reside lower in the water column compared to *Gs. sacculifer*. The negative responses from sub-surface to surface dwelling species (which also respond to nutrients) suggest this component represents tropical oligotrophic water at depth ~100m.

Component 4 accounts for 7% of variance and is dominated by positive scores from species *Gs. ruber*, *P. obliquiloculata* and *G. glutinata* and associated with strongly negative scores by *N. dutertrei*, *Gr. menardii* and *Ga. bulloides*. This component is very similar to component 3 but the dominance of *Gs. ruber* suggests this component represents subtropical oligotrophic conditions (see Table 12).
Table 13
The positive and/or negative responses for each group of foraminifera compared to the sample (depth/age) in the first and second components of core MD61. Purple = Group 1 (1-2) dominated by *Gr. inflata*. Group 2 (1-2) = red dominated by *Gs. ruber*. Group 3 (1-2) = deep red dominated by *P. obliquiloculata*. Group 4 (1-2) = yellow dominated by *Gs. rubescens*.

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A bi-plot of Components 1 and 2 that explain 65% of variance in core MD61 and reveal four ecological groups of foraminifera within the data. Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin of the plot.

The bi-plot of species scores from Component 1 versus Component 2 account for 65% of variance in core MD61 (see Fig. 51). Species with high positive species scores define Group 1. The principal species causing the variance in Group 1(1-2) is Gr. inflata, with a minor association of Ga. falconensis, Gn. glutinata and N. pachyderma (dextral). The species of this group appear to be temperate or transitional zone species (Gr. inflata and N. pachyderma) or have a widespread distribution in terms of temperature preference (Gn. glutinata). According to Martinez et al. (1999), Gr. inflata is a deep-water dweller living below 100m. The major dominance of Gr. inflata (as indicated by the vector length) suggests that this group represents transitional, oligotrophic water at depth (see Fig. 51).
Group 2 (1-2) is dominated by *Gs. ruber* which is most prolific in subtropical waters. The secondary species are also tropical to subtropical species but have an association to nutrient-rich water near coastal regions (*N. dutertrei, Gl. aequilateralis, Gl. calida, Gr. menardii* and *Gr. ungulata*). The dominance of *Gs. ruber* over the rest of the group as indicated by the vector lengths suggests that this group represents subtropical oligotrophic waters (see Fig. 51). Increased relative abundance of group 2 (1-2) is timed to the interglacial periods (see Fig. 52). In addition, there is an inverse relationship with the temperate, sub-surface species of Group 1(1-2).

Group 3 (1-2) is defined by species having a positive response in score 2 but a negative response to score 1. The principal species consist of *P. obliquiloculata* with secondary associations with *Ga. bulloides, Gr. truncatulinoides* (s) and *T. quinqueloba*. Less correlative associations as represented by *Gs. sacculifer* which lies towards Group 2 (1-2) with secondary associations with *Gr. tumida, Gs. conglobatus* and *Gr. tenellus*. *Gt. rubescens* also appears to lie towards Group 1 (1-2) with a secondary association with *Gr. truncatulinoides* (dextral). This group appears to contain species that have a preference for deeper water in a wide range of temperatures and the inclusion of *Gr. truncatulinoides* and *T. quinqueloba* suggest a homogenised mixed layer (see Fig. 52).

Group 4 (1-2) is dominated by *Gt. rubescens*. This small species occurs in low abundances at scattered localities in tropical and subtropical waters (Bé, 1977), and lives predominantly in the upper 50m of the water (Bé, 1977). *Gr. truncatulinoides* (dextral) is found in the tropical-subtropical areas (Kennett, 1979; Andrijanic, 1988), but lives in waters predominantly below 100-200m (Schiebel and Hemleben, 2005). The short vector lengths of this group (see Fig. 51) explain the least variance of the species counts. In the abundance graph, this group is quite random during glacial and interglacial periods and displays no convincing patterns (see Fig. 52).
Ecological groups revealed by analysing Component 1 versus Component 2 with PCA analysis in core MD61. The groups are compared to the δ¹⁸O and δ¹³C record of *Gs. ruber* through time. The glacial periods are highlighted and indicated by the even numbers.

**Figure 52**
Figure 53
A bi-plot of Component 1 versus Component 3 explaining 64% of variance in core MD61. Four ecological groups are revealed within the data. Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin of the plot.

The bi-plot of Component 1 versus Component 3, accounts for 64% of the variation in core MD61. Three more apparent groups are found in comparing score 3 with score 1 with 1% less variability explained. The separation of a nutrient group from temperature is more obvious compared to component 1 versus component 2 (see Fig. 53).

Again Group 1 (1-3) is dominated by *Gr. inflata* with strong positive values and a secondary association with *N. pachyderma* (dextral) (see Fig. 53). Species *Gr. inflata* and *N. pachyderma* (dextral) define Group 1 (1-3). Group (1-3) appears to be a stronger representation of the influence of transitional water at depths of ~100m than Group 1 (1-2) (see Fig. 52 and 54).
Group 2 (1-3) consists of species with a positive score in Component 1 and a negative score in Component 3. This group lays close to the boundary of Group 3 (1-3). The principal species is *Gn. glutinata* with secondary species *Gt. rubescens* and *Ga. falconensis* (see Fig. 53). This group does not show any obvious patterns of abundance in the graph with high abundance during glacial and interglacial periods (see Fig. 54) and this may be due to the ubiquitous nature of the dominant species *Gn. glutinata*.

The principal species of Group 3 (1-3) is *Ga. bulloides* with secondary species, *Gr. menardii, Gr. tumida, Gr. ungulata, T. quinqueloba, Gl. calida* and *Gt. tenellus*. *Ga. bulloides* is associated with 'upwelling' regardless of any geographical location. *Gr. menardii* also has been associated with 'upwelling' conditions in tropical water. It appears that this group represents 'upwelling' conditions and peaks of abundance are found during MIS 11 and the Holocene (see Fig. 54).

Group 4 (1-3) again has wide species associations but consists principally of *Gs. ruber, Gs. sacculifer* and *P. obliquiloculata*. Secondary species include *N. dutertrei, Gs. conglobatus, Gl. aequilateralis, Gr. truncatulinoides* (s), *O. universa, Gs. crassaformis*. This group appears to represent subtropical to tropical water conditions and increases in abundance during interglacial periods (see Fig. 54).

The increased variance of component 1 versus component 2 compared to Component 1 versus Component 3 may indicate that the temperature difference within the water column and the sea-surface is more important in explaining the variance in MD61 compared to the influence of nutrients. As seen in the Factor analysis, the influx of *Gr. inflata* during the glacial periods causes the most variance within MD61.
The ecological groups revealed by analysing Component 1 versus Component 3 with PCA in core MD61. The groups are compared to the δ¹⁸O and δ¹³C record of Gs. ruber through time. The glacial periods are highlighted and indicated by the even numbers. The purple Group 1 (1-3) indicates the relative abundance of transitional species. The grey group 2 (1-3) is a ubiquitous group. The aqua green Group 3 (1-3) indicates the relative abundance change of ‘upwelling’ species. The red Group 4 (1-3) represents the tropical-subtropical species.
8.3 Core MD2607

8.3.1 Age Model of MD2607

Figure 55
The $\delta^{18}O$ isotope record of core MD2607 from *Ga. bulloides* with the isotope tie points of Martinson et al. (1987) indicated on the record. Note the abrupt change in sedimentation rate after MIS 6.

The $\delta^{18}O$ record of planktonic foraminifera *Globigerina bulloides* was used to construct the age model of core MD03-2607. Individual isotope events 2.0, 2.2, 3.0, 4.0, 5.0, 5.1, 5.2, 5.3, 5.4, and 6.5 were identified in core MD2607 from the SPECMAP-stack of Martinson et al. (1987). Ages between the stratigraphic fix-points were obtained by linear interpolation using the Analyseries software of Paillard et al. (1996; see Fig. 55). For a full age model see Appendix 11.

8.3.2 Isotope analysis of core MD2607

The lightest $\delta^{18}O$ value of 0.43‰ was recorded during the Holocene while the heaviest $\delta^{18}O$ value of 2.71‰ was registered at the LGM (see Fig. 56). The mean $\delta^{18}O$ difference between these two stages is 2.05‰. The global difference in sea-level
accounts for 1.05% $\Delta \delta^{18}O$ between the LGM to present (Duplessy et al., 2002). Therefore, the residual difference in $\Delta \delta^{18}O$ of 1.00% must be due to a significant difference in SST and/or salinity between the LGM and present. The lightest $\delta^{18}O$ value recorded during MIS 5e was 0.44% and this indicates that the present day and last interglacial period registered similar conditions at this location.

Radiocarbon dating conducted on core MD2607 by Daniel Wilkins suggests the presence of a turbidite at 175-180cm (personal communication). Examination of this sample reveals ostracods with eye spots and suggests the sample contains some transported material from shallower depths. This is also indicated in the isotope record with steep peaks at 175cm/ ~15 110 yrs BP and was removed from the isotopic record.

Figure 56
The isotope record of core MD2607 over the last 175 000yrs BP obtained from Ga. bulloides every 5-30cm (see Chapter 8). The oxygen isotope record is indicated in red versus the carbon isotope record in green.
8.3.3 Age versus depth of MD2607

The age versus depth profile indicates a major increase in the sedimentation rate during the initial phase of MIS 6 (see Fig. 57). The average sedimentation rate of the isotopic stages are: Holocene = 11.6 cm/1000 yrs, MIS 2 = 20.8 cm/1000 yrs, MIS 3 = 11.8 cm/1000 yrs, MIS 4 = 5.9 cm/1000 yrs, MIS 5 = 9.81 cm/1000 yrs, MIS 6 = 41.1 cm/1000 yrs with an overall sedimentation rate of 18.7 cm/1000 yrs. Sediment studies of Feary et al. (2004), along the south coast of Australia, indicate the transgressive rate is an order of magnitude faster than rates measured during lowstand or highstand phases. This indicates that when the vast adjacent shelf is initially flooded, there is a pulse of sediment production and off-shelf export that decreases towards peak flooding. Gingele and De Deckker's (2005) clay analysis of MD2607 indicates the mouth of the Murray River was close to the core location during MIS 6 but not during the LGM and resulted in a high sedimentation rate of ~18.7 cm/1000 yrs.

Figure 57
The age versus depth profile of core MD2607 indicating a change in the sedimentation rates, especially at the base of the core from 1500-3270 cm. The sample points of the $\delta^{18}$O from Ga. bulloides are shown on the profile
8.3.4 Faunal Analysis of core MD2607

A total of 49,347 planktonic foraminifera were picked and identified from core MD2607. This is equivalent to an average of 429 individuals per sample, at a sample interval of 5-30cm. The interglacial periods recorded the presence of 25 species, while the glacial periods MIS 6 and MIS 2 recorded the presence of 19 and 22 species respectively. The samples of this core appear to be of higher diversity compared to other studies in the region such as Almond et al. (1993) and Li and McGowran, (1998), with the appearance of rare species *Gs. conglobatus* and *S. dehiscens* which were not found by Li and McGowran (1998) comprehensive study. This is probably due to the location of core MD2607 near the top of the continental slope and not on the continental shelf.

The highest species abundance of foraminifera in core MD2607 was recorded by *Ga. bulloides* (~45%) during the penultimate glacial period (MIS 6) and, generally, was the most abundant with an average abundance 19.5%. The other dominant species of core MD2607 include *Gr. inflata* (19.2%), *Gr. truncatulinoides* (sinistral; 12.4%), *Ga. falconensis* (11.5%) and *Gs. ruber* (8%; see Fig. 58 and Appendix 12)

This dominance of *Ga. bulloides* suggests the influence of cold, nutrient-rich water during certain isotopic stages in MD2607. *Ga. bulloides* is a sub-polar to transitional species which also responds to nutrients in the water column, especially blooms caused by ‘upwelling’. The other species that are in high abundance during MIS 6 are *T. quinqueloba*, reaching a peak of ~15%, which drops to < 5% after this glacial period. Two other species which respond to nutrients, *Gn. glutinata* and *N. dutertrei*, reach high abundances of 12% and with a peak abundance of 16% during MIS 6 (see Fig. 58 and Appendix 12). Conversely, the dominant warm and oligotrophic species *Gs. ruber* records its lowest abundance and is totally removed from the record (0%) at ~ 146 000, ~142 000 and 139 000 yrs B.P. Other warm water species *Gr. inflata*, *Ga. falconensis* and *O. universa* are reduced or removed from the record during MIS 6 (see Fig. 58 and Appendix 12).

The last interglacial (MIS 5) was dominated by subtropical and transitional species. Highest abundance was recorded by *Gr. inflata* with ~31% followed by *G. ruber* with ~15%; however, these species are more abundant during the Holocene. The most dramatic change in abundance during this stage is by dextrally coiled *Gr. truncatulinoides*, a tropical-subtropical species. This species has an abundance of <1% for all other stages but has a surge of abundance from 115 000 to 127 000 cal yrs B.P to ~12.5%. The tropical to subtropical species *Gt. rubescens* also records its highest
abundance during MIS 5 of 12% from a general abundance of ~5%. Minor tropical species also appear in the record during MIS 5 (see Fig. 58 and Appendix 12).

The LGM saw another increase in abundance of *Ga. bulloides* up to 39% compared to 10% during the Holocene; it is the most dominant species during the LGM. Rapid decreases in abundance was also seen in the sub-polar species *T. quinqueloba* around the LGM from 24% at ~23 000 yrs BP to <2% during the Holocene. Changes in abundance of both these species may be a result of cooling of the water column and possibly an influx of nutrients through ‘upwelling’ processes. Increases in abundance were also seen by the ubiquitous, but bloom species, *Gn. glutinata*, the sub-polar to temperate *Gr. truncatulinoides* (s) and the transitional *N. pachyderma* (d) (see Fig. 58 and Appendix 12).

Transitional to subtropical water conditions are indicated by the faunal assemblages during the Holocene. It appears from the faunal assemblage that the Holocene represent the warmest stage of MD2607, but this is discussed further below. The transitional species *Gr. inflata* is the most dominant species with a high abundance of 42.5% (see Fig. 58 and Appendix 12) at ~ 2500 cal yrs B.P. In addition, the subtropical-temperate species *Ga. falconensis* peaks at ~ 21.5% while, the subtropical-tropical *G. ruber* also peaks during the Holocene at ~19%. Notably, this is period is accompanied with a decrease of *Ga. bulloides* abundance by ~20% from the LGM. Species *T. quinqueloba* and *Gr. truncatulinoides* (s) reduce their abundance to <0.5% and <10% respectively during the Holocene while *Gn. glutinata* maintains an abundance during the LGM and Holocene of ~ 5% (see Fig. 58 and Appendix 12).

At low relative abundances of < 2%, minor species also confirm some of these patterns with the appearance of the tropical-subtropical species *Gl. aequilateralis, Gl. calida* as well as the subtropical species *O. universa* during the Holocene and MIS 5. Conversely, the polar species *N. pachyderma* (s) appears during the LGM, and has a peak in abundance (~2.5%) during MIS 6 (see Fig. 59).
Figure 58

The relative percentage abundance of the major planktonic foraminiferal species in core MD2607 plotted against SPECMAP Age, and the δ¹⁸O and δ¹³C record from *Ga. bulloides*. Species which favour tropical-subtropical water conditions are coloured in red. *Ga. bulloides*, *Gn. glutinata* are coloured in aqua blue for species, which prefer cooler and nutrient-rich water conditions. Transitional species are coloured in purple. Sub-polar species are coloured light blue and polar species dark blue.
The relative percentage abundance of minor planktonic foraminifera in core MD2607 plotted against SPECMAP Age, and the $\delta^{18}O$ and $\delta^{13}C$ record of a *Ga. bulloides*. Species which prefer tropical-subtropical water conditions are coloured red. Cold deep water species are coloured blue.
8.3.5 Sea-surface Temperature of MD2607

Figure 60
The sites of cores-tops from the AUSMAT-F4 database that provide the analogues for the sea-surface temperature (SST) estimates in core MD2607, with most analogues found between 45-30°S. There are many cores as there are 10 core top analogues per sample.

Most of the species counts in core MD2607 have an analogue within the Prell (1999) database (see Fig. 60). The highest mean temperature (Tmean) of 20°C ± 0.84°C was during the last interglacial (MIS 5e; ~ 125,000 yrs B.P.). This was 1.3°C warmer than the warmest mean estimate of 18.7°C ± 0.84°C for the Holocene. The modern-day annual estimate (Tmean) of 17.5°C ± 0.84°C is close to the general measured estimate of SST of 18°C (James et al., 1992) and confirms that the AUSMAT-F4 is providing good estimates of SST.

The increase in the Tmean temperature record during the last interglacial and present is due to the warmer month estimate (Tmax) and may indicate a stronger Leeuwin Current. The highest Tmax of 23.1°C ± 0.87°C was recorded at ~125 000 yrs BP compared to 21.9°C ± 0.87°C during the Holocene (see Fig. 61).

The LGM records a mean temperature of 11.4°C ± 0.84°C; however, the coldest mean temperature of 11°C ± 0.84°C was recorded during MIS 6. Tmin represents the coldest monthly temperature and the coldest recorded Tmin was during MIS 6 of 8.3°C.
± 0.98°C compared with the Tmin of 8.9°C ± 0.98°C during the LGM (see Fig. 61). However, these estimates are within the error range so the temperature difference is not significant between the two recorded glacial periods in core MD2607.

The annual mean SST (Tmean) indicates a temperature range of 9°C ± 0.84°C between MIS 6 to MIS 5, representing the largest SST transition from a glacial to interglacial. The LGM to the Holocene transition resulted in a 6.8°C ± 0.84°C difference from the LGM to a peak temperature at ~ 14 000 yrs BP (see Fig. 61).

During interglacial periods the Tmean estimates at the 50m-150m show a difference of ~3-3.2°C ± 0.84°C. During the glacial periods the difference between the 50m and the 150m estimate is ~ 1.4-1.7 °C ± 0.84°C. This indicates that temperature gradients are reduced during the LGM and MIS 6 suggesting a well mixed homogenised mixed layer (see Fig. 61). Conversely, there appears to be more stratification (possibly due to the Leeuwin Current) during interglacial periods.

The modern day estimate of the mixed layer is 71 mbsl and is indicated in figure 62. Compared to the Holocene the mixed layer is deeper during the glacial periods reaching a depth of ~125 mbsl. There is only one period where the mixed layer is shallower than present which occurs during MIS 5e (see Fig. 61).

Some samples record square chord distances higher than the 0.2 limit of analogue quality (see Fig. 61). This occurs most noticeably during glacial (see Fig. 61). The samples with high square chord distances have higher abundance of cold water species Ga. bulloides, Gr. truncatulinoides (s), with occasional higher abundance of T. quinqueloba. In addition, warm water species Gr. inflata and Gs. ruber have a low to very low abundance. The lower quality of the analogue for these periods may be due to the lower sampling of cores in the Southern Ocean to form an analogue with, and, therefore, these periods may be colder than what is presently shown in the SST record.
The sea-surface temperature record of core MD2607 versus the $\delta^{18}$O and $\delta^{13}$C record of *Ga. bulloides*. Tmean indicates the average annual temperature. Tmax records the temperature of the warmest month. Tmin records the coolest monthly temperature. The square chord distance indicates the quality of the analogue, <0.2 represents a good SST estimate. A reference line is placed at the modern day estimate of the mixed layer to indicate phases of deepening or shoaling of the mixed layer measured in meters below sea level (mbsl).
8.3.6 Factor Analysis of MD2607

The use of Q-mode factor analysis defines two important factors within MD2607. Factor one accounts for 81.6% of the variance in core MD2607 (see Table 14), and is dominated by *Gr. inflata* (0.54), with species *Gs. ruber*, and *Ga. falconensis* also recording high positive scores (see Table 15). The dominant species *Gr. inflata* is indicative of transitional water temperatures and prefers a thick mixed layer with oligotrophic to intermediate nutrient levels. Subtropical-transitional species *Gs. ruber* and *Ga. falconensis* reside higher in the water column compared to *Gr. inflata*. *Gs. ruber* represents the shallowest dwelling planktonic foraminifera and also prefers oligotrophic conditions in the mixed layer. Factor one suggests variance is due to the transitional-subtropical oligotrophic conditions in a deep mixed layer.

Table 14
The variance of the species counts in MD2607 explained by each successive factor axis. The variance explained by each axis decreases from Factor 1 to Factor 5.

<table>
<thead>
<tr>
<th>Factor axis</th>
<th>Factor value</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.816</td>
</tr>
<tr>
<td>2</td>
<td>0.112</td>
</tr>
<tr>
<td>3</td>
<td>0.023</td>
</tr>
<tr>
<td>4</td>
<td>0.014</td>
</tr>
<tr>
<td>5</td>
<td>0.008</td>
</tr>
</tbody>
</table>
Table 15

The species scores of the Q-mode Factor analysis based on the species counts of MD2607. Highlighted scores indicate the dominant species within each factor.

<table>
<thead>
<tr>
<th>Species Name</th>
<th>Factor 1 scores</th>
<th>Factor 2 scores</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ga. bulloides</td>
<td>0.54</td>
<td>0.64</td>
</tr>
<tr>
<td>Ga. falconensis</td>
<td>0.33</td>
<td>-0.09</td>
</tr>
<tr>
<td>T. quinqueloba</td>
<td>0.15</td>
<td>0.24</td>
</tr>
<tr>
<td>Gt. rubescens</td>
<td>0.09</td>
<td>-0.05</td>
</tr>
<tr>
<td>Gt. tenellus</td>
<td>0.00</td>
<td>-0.01</td>
</tr>
<tr>
<td>Ga. ruber</td>
<td>0.23</td>
<td>-0.33</td>
</tr>
<tr>
<td>O. universa</td>
<td>0.01</td>
<td>-0.03</td>
</tr>
<tr>
<td>Gl. aequilateralis</td>
<td>0.00</td>
<td>-0.02</td>
</tr>
<tr>
<td>Gl. calida</td>
<td>0.00</td>
<td>-0.00</td>
</tr>
<tr>
<td>Gn. glutinata</td>
<td>0.16</td>
<td>0.06</td>
</tr>
<tr>
<td>Gr. inflata</td>
<td>0.54</td>
<td>-0.61</td>
</tr>
<tr>
<td>Gr. hirsuta</td>
<td>0.03</td>
<td>-0.01</td>
</tr>
<tr>
<td>Gr scitula</td>
<td>0.03</td>
<td>0.01</td>
</tr>
<tr>
<td>Gr. truncatulinoides (s)</td>
<td>0.36</td>
<td>0.12</td>
</tr>
<tr>
<td>Gr. truncatulinoides (d)</td>
<td>0.01</td>
<td>-0.02</td>
</tr>
<tr>
<td>N. pachyderma (s)</td>
<td>0.00</td>
<td>0.01</td>
</tr>
<tr>
<td>N. pachyderma (d)</td>
<td>0.14</td>
<td>-0.03</td>
</tr>
<tr>
<td>N. dutertrei</td>
<td>0.15</td>
<td>0.04</td>
</tr>
<tr>
<td>P. obliquiloculata</td>
<td>0.00</td>
<td>-0.00</td>
</tr>
</tbody>
</table>

Factor 2 which accounts for 11.2% of variance in MD2607 is dominated by positive scores of *Ga. bulloides*, *T. quinqueloba* and *G. truncatulinoides* (sinistral) (see Table 15). All these species are dominant in polar to subpolar water temperatures. *Ga. bulloides* and *T. quinqueloba* reside in the upper mixed layer/photic zone, with *T. quinqueloba* bears symbionts in its test. However, *Gr. truncatulinoides* is a very deep water dweller found at depths down to 4000 mbsl and prefers a deep thermocline with a homogenised mixed layer. *T. quinqueloba* also prefers low temperature/salinity gradients in the mixed layer. Species *Ga. bulloides* is indicative of ‘upwelling’ environments and *T. quinqueloba* has been identified as being an ‘upwelling’ indicator in the Angola Basin (Van Leeuwen, 1989) and the optimal occurrence of *T. quinqueloba* was found in nutrient rich water (1.2 µg at/l phosphate) by Bé and Hutson, 1977). Conditions of ‘upwelling’ remove of any stratification which may be in the mixed layer, it appears that Factor 2 represents the influx of deep cool, nutrient-rich water into the core region with homogenised conditions in the mixed layer.
Figure 62
The ecological groups defined by comparing the species scores of Factor 1 compared to Factor 2 based on the species counts of core MD2607. Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin of the plot.

The first two factors explain ~93% of variation in core MD2607. The dominant species scores of Factor 1 and 2 are plotted in a bi-plot and reveal two ecological groups. Group 1 represents the species with high species scores in Factor 1 and the most dominant species is *Gr. inflata*, as indicated by the vector length. A close association between *Gs. ruber* and *Gr. inflata* is evident due to the low angle between their individual vectors (see Fig. 62).

Group 2 contains the species with positive species scores in Factor one and two. The dominance of *Ga. bulloides* is displayed by the length of the vector (see Fig. 62). The closer angle of association is displayed by *T. quinqueloba* compared to *Gr. truncatulinoides*, suggests that the variance in Factor 2 is due to the presence of sub-polar, nutrient-rich waters in the upper mixed layer and this is homogenised from a deep thermocline as indicated by the presence of *Gr. truncatulinoides* (sinistral).
When all of the species abundances of Group 1 and Group 2 are combined they reveal an inverse relationship during glacial and interglacial periods. It appears that there an increased influence of Group 1 during the interglacial periods and the increased influence of Group 2 during the glacial periods (see Fig. 63)
Figure 63
The species abundances of Factor 1 compared to Factor 2, versus the $\delta^{18}$O and $\delta^{13}$C record from 
*G. bulloides* in core MD2607. Factor groups defined by Factor analysis on the species counts 
of MD2607. Group 1 representing warm oligotrophic water and Group 2 representing cold 
water under the influence of 'upwelling' conditions and deep mixed layer.
8.3.7 Principal Component Analysis of core MD2607

The results from the principal component analysis of MD2607 consist of 115 observations from 19 variables. The first two components account for 72% of variability within the data (see Table 16). The main variance (60%) is explained in Component one (Eigenvalue 1) Using the scree test method two Eigenvalues could be retained to explain 72% of the variability in core MD2607 (see Fig. 64). Three Eigenvalues were considered as 8.5% of the variability is explained in the data of Component 3 but when the ecological groups of Component 3 were plotted similar ecological groupings of the first two components were found (see Appendix 13-14). Other Eigenvalues beyond 3 are dismissed due to repetitive mono-specific patterns of Gs. ruber and Ga. falconensis, which are accounted for in the preceding components (see Table 16).

Figure 64
The scree plot of the first ten Eigenvalues of MD2607.

Table 16
The Eigenvalues of the first five axes resulting from principal component analysis. The variance within the species counts of MD2607 is explained by each axis, which reduces for each successive axis.

<table>
<thead>
<tr>
<th>Axis</th>
<th>Eigenvalue</th>
</tr>
</thead>
<tbody>
<tr>
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<tr>
<td>2</td>
<td>0.122</td>
</tr>
<tr>
<td>3</td>
<td>0.085</td>
</tr>
<tr>
<td>4</td>
<td>0.060</td>
</tr>
<tr>
<td>5</td>
<td>0.040</td>
</tr>
</tbody>
</table>
Table 17
The species scores for the first five components described by principal component analysis in MD2607. The highlighted numbers represent the dominant species of each component.

<table>
<thead>
<tr>
<th>FullName</th>
<th>Score01</th>
<th>Score02</th>
<th>Score03</th>
<th>Score04</th>
<th>Score05</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ga. bulloides</td>
<td>-2.63</td>
<td>2.79</td>
<td>1.84</td>
<td>-0.06</td>
<td>-0.37</td>
</tr>
<tr>
<td>Ga. falconensis</td>
<td>0.36</td>
<td>-1.32</td>
<td>3.04</td>
<td>-0.79</td>
<td>2.52</td>
</tr>
<tr>
<td>T. quinqueloba</td>
<td>-0.90</td>
<td>-0.26</td>
<td>-1.49</td>
<td>-1.35</td>
<td>0.68</td>
</tr>
<tr>
<td>Gt. rubescens</td>
<td>0.21</td>
<td>-0.32</td>
<td>-0.24</td>
<td>0.60</td>
<td>0.23</td>
</tr>
<tr>
<td>Gt. tenellus</td>
<td>0.06</td>
<td>0.08</td>
<td>-0.05</td>
<td>0.12</td>
<td>-0.01</td>
</tr>
<tr>
<td>Gs. ruber</td>
<td>1.41</td>
<td>-0.48</td>
<td>1.82</td>
<td>1.79</td>
<td>-2.24</td>
</tr>
<tr>
<td>Q. universa</td>
<td>0.14</td>
<td>-0.01</td>
<td>0.01</td>
<td>0.17</td>
<td>-0.10</td>
</tr>
<tr>
<td>Gl. aequilateralis</td>
<td>0.09</td>
<td>0.06</td>
<td>-0.00</td>
<td>0.09</td>
<td>-0.04</td>
</tr>
<tr>
<td>Gl. calida</td>
<td>0.04</td>
<td>0.02</td>
<td>0.01</td>
<td>0.01</td>
<td>0.01</td>
</tr>
<tr>
<td>Gn. glutinata</td>
<td>-0.25</td>
<td>0.21</td>
<td>-0.15</td>
<td>0.26</td>
<td>0.26</td>
</tr>
<tr>
<td>Gr. inflata</td>
<td>3.0</td>
<td>2.64</td>
<td>0.10</td>
<td>-1.73</td>
<td>0.20</td>
</tr>
<tr>
<td>Gr. hirsuta</td>
<td>0.08</td>
<td>-0.01</td>
<td>-0.16</td>
<td>0.23</td>
<td>0.15</td>
</tr>
<tr>
<td>Gr. scitula</td>
<td>-0.07</td>
<td>0.18</td>
<td>0.05</td>
<td>0.18</td>
<td>-0.08</td>
</tr>
<tr>
<td>Gr. truncatulinoides</td>
<td>-0.49</td>
<td>-1.36</td>
<td>0.74</td>
<td>-3.05</td>
<td>-2.45</td>
</tr>
<tr>
<td>(s)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Gr. truncatulinoides</td>
<td>0.08</td>
<td>-0.05</td>
<td>-0.33</td>
<td>0.40</td>
<td>-0.40</td>
</tr>
<tr>
<td>(d)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>N. pachyderma (s)</td>
<td>-0.05</td>
<td>-0.03</td>
<td>-0.09</td>
<td>-0.01</td>
<td>0.03</td>
</tr>
<tr>
<td>N. pachyderma (d)</td>
<td>0.16</td>
<td>-0.44</td>
<td>-0.16</td>
<td>-0.24</td>
<td>0.77</td>
</tr>
<tr>
<td>N. dutertrei</td>
<td>-0.16</td>
<td>0.13</td>
<td>-0.04</td>
<td>0.51</td>
<td>-0.22</td>
</tr>
<tr>
<td>P. obliquiloculata</td>
<td>0.00</td>
<td>0.02</td>
<td>-0.05</td>
<td>-0.05</td>
<td>0.01</td>
</tr>
</tbody>
</table>

Strong positive scores of *Gr. inflata* and *Gs. ruber* dominate Component 1, which accounts for 60% of the variation in core MD2607 (see Table 17). These species are associated with warm subtropical to transitional water. *Gr. inflata* resides deeper in the water column compared to *Gs. ruber* and suggests a temperate, oligotrophic mixed layer to depths ~100m. This is confirmed by strong negative score from ‘upwelling’, subpolar to transitional species *Ga. bulloides*.

Component 2 is dominated by subpolar to transitional species *Ga. bulloides* and transitional species *Gr. inflata*. It seems that temperature is the main parameter associating these species together as *Gr. inflata* prefers oligotrophic water (Giraudeau, 1993; Giraudeau and Rogers, 1994), compared to *Ga. bulloides* in upwelling systems which prefers more eutrophic waters (Bé, 1977). It is that possible this component represents slightly cooler, transitional-subpolar water with a greater influence of cooler subsurface water compared to what Component 1 indicates.
Component 3 is dominated by high positive values of *Ga. falconensis* and positive scores of *Gs. ruber* and *Ga. bulloides*. The simplicity of the interpretation is reduced for this component. *Ga. falconensis* and *Gs. ruber* contain symbionts while *Ga. bulloides* is barren of symbionts. *Ga. falconensis* is found in temperate to subtropical water while *Gs. ruber* has a greater preference for subtropical water and *Ga. bulloides* is a subpolar to transitional species. Both *Ga. falconensis* and *Ga. bulloides* are associated with nutrients on the Lacepede Shelf with *Ga. falconensis* occurring in slightly higher temperatures than *Ga. bulloides* (Almond et al., 1993), while *Gs. ruber* is considered to be an oligotrophic species. With the dominance of species *Ga. falconensis* in component 3, there appears to be an association with both temperature and nutrients. Li et al. (1996b) recognised a mixed assemblage indicating mesotrophic conditions on the Lacepede Shelf and this may be the case in Component 3. The strong negative response of *T. quinqueloba* may be indicating a certain level of stratification leading to mesotrophic conditions and thermal layers in the photic zone. The interpretation of this group may be clearer in the bi-plots.

Four main ecological groups become obvious when comparing the species values of the first two components (see Fig. 65). Group 1 and 4 appear to be the most dominant groups due to the higher scores of there principal species. Group one is characterised by species with positive scores in component 1 and component 2. The principal species of this group, with a high positive score, is *Gr. inflata*. The secondary species of this group are *Gt. tenellus, Gl. aequilaterialis* and *Gl. calida*. This group dominates in interglacial periods, when the relative abundance of this group is combined and related to the δ¹⁸O record of MD2607 (see Fig. 66).

Group 2 is characterised by species with positive scores in Component 1 but negative scores in Component 2. The principal species is *Gs. ruber*, with a high positive score in Component 1 and low negative score in Component 2, and *Ga. falconensis* characterised by a low positive score in Component 1 and a higher negative score in Component 2. Secondary species in of this group are *N. pachyderma* (dextral), *Gt. rubescens, Gr. truncatulinoides* (dextral), *O. universa* and *G. hirsuta*. This group when graphed appears to represent warmer subtropical water compared to group 1, with higher abundance early to mid Holocene and during the last interglacial (MIS 5; see Fig. 66).
Figure 65
The bi-plot of Components 1 versus Component 2 explaining 60% of variance in core MD2607, revealing four ecological groups within the data. Minor species that do not influence the main variance of the two groups are removed for image clarity around the origin of the plot.

Group 3 consists of Gr. truncatulinoides (sinistral) and T. quinqueloba as principal species with secondary species N. pachyderma (sinistral; see Fig. 65). The species of this group have negative species scores in Component 1 and Component 2. Gr. truncatulinoides is the deepest-dwelling planktonic foraminifera (Schiebel and Hemleben, 2005), with the sinistral form associated with transitional to polar water (Kennett, 1979; Andrijanic, 1988). Species N. pachyderma (sinistral) and T. quinqueloba are associated with polar-subpolar water masses, prefer low temperature gradients in the water column and lives preferentially below 100m. Therefore, this group appears to represent cold subsurface dwellers. The relative abundance of this group increases during the glacial periods (see Fig. 66).
The principal species of Group 4 is *Ga. bulloides*, with a strong negative score in component 1 and component 2. The secondary species of group 4 are *G. glutinata*, *G. dutertrei* and *G. scitula*. Species *Ga. bulloides* is indicative of cold nutrient-rich water and resides in the upper water column. Species *Gn. glutinata* is a ubiquitous species but increases in abundance have been observed in late bloom successions. The strong score of *Ga. bulloides* compared to the minor species indicates that the group represents cold, nutrient-rich water being upwelled to the upper photic zone. The minor species of this group also have associations with nutrients within the water column. This group shows an increase in abundance during MIS 6 (see Fig. 66)
Figure 66

The ecological groups revealed by analysing Component 1 versus Component 2 with PCA analysis in MD2607. The groups are compared to the $\delta^{18}$O and $\delta^{13}$C record of *Ga. bulloides* through time. The glacial periods are highlighted and indicated by the even numbers. The purple Group 1 (1-2) indicates the relative abundance of transitional species. The red Group 2 (1-2) represents the tropical-subtropical species. The dark blue Group 3 (1-2) indicates sub-surface, cold water species. The aqua green Group 4 (1-2) indicates the relative abundance change of 'upwelling' species.
Table 18

The positive and/or negative responses for each group of foraminifera compared to the sample (depth/age) in the first and second components of core MD2607. Purple = Group 1(1-2) dominated by *Gr. inflata*. Group 2(1-2) = red dominated by *Gs. ruber*. Group 3(1-2) = dark blue dominated by *Gr. truncatulinoides*. Group 4 (1-2) = aqua green dominated by *Ga. bulloides*.

<table>
<thead>
<tr>
<th>Specmap Age (BP)</th>
<th>Depth (cm)</th>
<th>Sample Scores Component 1</th>
<th>Sample Scores Component 2</th>
<th>Sample Scores Component 3</th>
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</thead>
<tbody>
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</tr>
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The sample scores indicate the dominant ecological group of each sample (see Table 18). The colours match the group colours in Figure 67. This table illustrates that transitional group dominates during the late Holocene and occasionally during the last interglacial period (MIS 5). The subtropical group, containing the principal species *Gs. ruber*, is dominant during the last interglacial (5e) and initial stages of the Holocene. Deep-dwelling species of Group 3 (1-2) dominate at the LGM and occasionally during MIS 6. The upwelling species of Group 4 (1-2) dominate mostly during the MIS 6 glacial period.
Chapter 9

Discussion

9.1 Core BAR9403

9.1.1 Alteration within the mixed layer

During MIS 3, there was an increase in the relative abundance of *N. pachyderma* by ~10% and slightly higher abundances of *N. dutertrei* by ~5%. Comparatively, the tropical and subtropical oligotrophic species *Gs. ruber*, *Gs. sacculifer* and *Gr. menardii* were of lower abundance during MIS 3 (see Fig. 28 and Appendix 3). *N. dutertrei* and *N. pachyderma* are thermocline dwellers whose increased abundance is associated with the development of a Deep Chlorophyll Maximum layer (DCM; Fairbanks and Wiebe, 1980). A DCM is initiated when the upper water column is stratified and an increase in chlorophyll production occurs at the thermocline. A DCM was identified by Fairbanks and Wiebe (1980) and was observed as a seasonal feature along the Californian coast. Plankton trap data offshore California and Oregon (Sautter and Thunell, 1991; Ortiz et al., 1995) showed large populations of *N. dutertrei* develop in spring blooms and upwelling episodes were associated with high fertility about the base of the thermocline. In core-top samples Ding et al. (2006) revealed that in the Indian Monsoon Sumatra region high *N. dutertrei* abundances indicated conditions of low salinity and a shallow thermocline. It is also shown through the abundances of *Ga. bulloides* and *Gr. menardii* that upwelling conditions did not occur during MIS 3 (see Fig. 28 and Appendix 3) but, instead, the conditions favoured DCM. PCA analysis and the sample scores indicated that the DCM Group 1 (1-2), containing species *N. dutertrei* and *N. pachyderma*, dominated over the other ecological groups during MIS 3 and momentarily at 11 000 yrs BP (see Fig. 36).

The depth of the thermocline influences the rate of productivity of the DCM (Fairbanks and Wiebe, 1980), therefore the thinner the mixed layer, the greater the productivity of the DCM. In addition to SST, the MAT was used to estimate the mean annual depth of the mixed layer in core BAR9403 (see Fig. 31). Although the mean annual mixed layer is shown to be thicker during MIS 3, it is still relatively high in the photic zone at approximately 35-42mbsl. Thickening of the mixed layer due to
stratification has been observed within the Santa Barbara Basin by Pak and Kennett (2002) and this supports the abundance patterns described above. Therefore, MIS 3 is characterised by a highly stratified mixed layer leading to the development of a DCM in a slightly deeper and cooler thermocline location than present.

De Deckker et al.’s (2002) analyses of core BAR9442, located 1° south from BAR9403, indicated ‘blooms’ of the giant diatom Ethmodiscus rex between 28 000 yrs and 19 000 yrs BP. *E.* rex blooms occur when the water-column is permanently stratified, with a substantial increase in salinity and high levels of silica and nitrate near the sea-surface. This coincides with the period of stratification due to the development of a DCM layer within core BAR9403. This observation is supported by the benthic foraminifera analysis in core BAR9403 (Murgese, 2003), which also suggests a period of stratification from 35 000 to 15 000 yrs BP. In addition, a regional pattern of DCM development has been observed within the various seas of the Throughflow region by Linsley et al. (1985), Barmawidjaja et al. (1993); Ding et al. (2002) and Spooner et al, (2005) during MIS 3 and MIS 2.

The relative abundance of DCM species such as *N.* dutertrei reduced slightly during MIS 2 and the low abundance of upwelling species such as *Ga.* bulloides was maintained from MIS 3 to MIS 2. The LGM is characterised by a high abundance of tropical-subtropical species with a preference for oligotrophic conditions. This is supported by the Factor and PCA analyses that grouped *Gs.* ruber and *Gs.* sacculifer together and showed high abundance during the LGM (see Fig. 33 and 36). The samples scores also indicated that this group dominated over the other PCA groups during the LGM (see Table 7). The high abundance of these tropical-subtropical species suggest the water column was still stratified and nutrients may not be entrained into the upper surface layer through wind forced mixing which occurs presently (Sprintall et al., 2002).

Martinez et al. (1998) showed that the change in abundance of species, such as *N.* pachyderma, *N.* dutertrei and *Ga.* bulloides, vary around the region and depend on mixed layer thickness and the intensity of the Java Upwelling System. A more dynamic and productive Java Upwelling System encouraged by the SE Monsoon during the LGM is indicated in studies by Martinez et al. (1999), Takahashi and Okada (2000), Gingele et al. (2001) and Gingele et al. (2002). In addition, palaeoproductivity proxy investigations by Müller and Opdyke (2000) in the Timor passage indicate increased nutrients in the mixed layer during MIS 2. However, results from cores BAR9403 and BAR 9442 off the coast of Sumatra suggest that upwelling did not occur during the
LGM. Perhaps the upwelling signal found off the coast of Java by previous studies did not extend westward to the coast of Sumatra during the LGM.

The major change found in core BAR9403 is around the upwelling signal of *Ga. bulloides*, with increased abundance at ~14,000 yrs BP and which is maintained into the early Holocene (see Fig. 28 and Appendix 3). In addition, *Gr. menardii*, considered to be a tropical upwelling species, increases in abundance after the LGM to record its highest abundance at approximately 17 000 yrs BP and mirrors the high abundance of *Ga. bulloides*. *Gr. menardii* is abundant in upwelling regions where the thermocline is shallow and primary productivity high (Thunell and Reynolds, 1984; Martinez et al., 1998) and its increase in abundance is timed to the MAT estimate of a thinner mixed layer in MIS 2 (see Fig. 31). In addition, *Ga. bulloides* is also a nutrient opportunist and is known globally to bloom in areas of enhanced productivity and high nutrient levels at upwelling sites (Bé, 1977; Auras-Schudnagies et al., 1990; Kroon, 1990; Ganssen and Kroon, 2000).

The PCA analysis grouped *Gr. menardii* and *Ga. bulloides* together to represent upwelling Group 2 (1-2; see Fig 36) and the sample scores indicate that this group dominated from 17 000 yrs to 5800 yrs BP (see Table 7). Therefore, it appears that nutrients were reaching the sea-surface at 17 000 yrs BP. This is also indicated by the mean SST estimates with a reduction of SST at this time. This may indicate a strengthening of the Java Upwelling System after the LGM, perhaps due to the upwelling cell spreading westward from the coast of Java. Presently, upwelling along the coast of Java operates during the SE Monsoon as a result of southward Ekman transport (Wyrtki, 1962) but nutrients do not reach the surface due to an increased in the transport of the Throughflow (Bray et al., 1997). Analyses of Ding et al. (2006) provides evidence of a stronger Java upwelling system at ~ 8000 yrs BP from cores located in the Java/ Banda Sea area. It is possible the upwelling signal in core BAR9403 is evidence of a more intense monsoonal system relative to the LGM.

The abundance of oligotrophic species was also reduced from the 20 000-7000 yrs period and this indicates the removal of the stratified structure of the water column. Sea-surface temperature estimates also indicate that upwelling of nutrient rich, cooler water was reaching the sea-surface near the LGM-Holocene transition. At the LGM-Holocene transition De Deckker et al., (2002) recorded a high Ba_{excess}, and low *E rex* abundance.

There is evidence of increased nutrients early in MIS 2 and the initial stages of the Holocene through the interaction of *N. dutertrei* and *Ga. bulloides* (see Fig. 28 and
An increase in abundance of *Ga. bulloides* is associated with a decrease in *N. dutertrei* and this is due to the availability of different food resources. Offshore California, sediment trap studies have shown that in the later stages of upwelling are dominated by *Ga. bulloides*, a form associated with high zooplankton levels (Murray, 1995), while it has been observed that immediately after upwelling has ceased, *N. dutertrei*, a form associated with the thermocline, increases in abundance (Thunell and Reynolds-Sautter, 1992; Murray, 1995).

Murgese (2003) also analysed the benthic foraminifera assemblages in core BAR9403. The benthic record appears to contradict the planktonic foraminifera record by indicating that there was a reduction in productivity during 15 000-5000 yrs BP. Murgese (2003) suggested the wetter climate of this period restored the low salinity cap of the Indonesian Throughflow to the region and deepened the thermocline. The mixed layer estimate from core BAR9403 indicates that the mixed layer at the MIS 2-Holocene transition moved periodically between deeper to shallower depths compared to present. In addition the mixed layer estimate at the MIS 2-Holocene transition is shallower than during MIS 3 when the low salinity cap was removed. Perhaps the upwelling signal was lost from the benthic record due to the more intense, deep-water circulation also suggested by Murgese (2003).

The SST estimates from the 50-150m interval also suggest colder water was moving up from subsurface layers but this may be misrepresented by the MAT due to the increased abundance of *Ga. bulloides*. Perhaps the increased abundance of *Ga. bulloides* and *Gr. menardii* was due to nutrients being supplied by increased river discharge and the rapid rise in sea-level in the region at ~14 000 yrs BP (Hanebuth at al., 2000), but this will be discussed further below.

### 9.1.2 Sea-surface Temperature and Salinity

The MAT results of core BAR 9403 indicate that SST remained relatively stable through MIS 3 to the present (see Fig. 31). The lowest SST did not occur during the LGM in core BAR9403. The LGM recorded an average SST of 28.6°C, equivalent to present estimates. This supports the observations of the planktonic foraminiferal assemblages that upwelling did not occur during MIS 3 and 2. Importantly, if SST remained unchanged at the LGM, then the deviation in the δ¹⁸O record (1.03‰) must be due to a significant increase in sea-surface salinity.
This increase in salinity can be related to a regional decrease in precipitation and the removal of low salinity ITF from the core site during the LGM. A regional increase in salinity from these processes has already been observed by many studies (Ahmad et al., 1995; Linsley, 1996; Martinez et al., 1997 and De Deckker et al., 2003). In addition, studies on the vegetation history of the Indonesian region also indicate drier climate conditions culminating at the LGM (van der Kaars, 1989; Barmawidjaja et al., 1993; van der Kaars and Dam 1995, Wang et al., 1999b; van der Kaars et al., 2000). Clay analysis as a proxy for river discharge by Gingele et al. (2001) also indicates a dry climate during the LGM.

There is only one major deviation in the Tmean record after the LGM which results in a reduction of SST from 28.6°C to 26.1°C. The timing of this event is around 14 000 yrs BP (see Fig. 28 and 31) and is linked to the increase in Ga. bulloides. This date is significant in the region as it coincides with the ‘switching on’ of the NW Monsoon in Australia at 14 000 yrs BP estimated by Wyroll and Miller (2001), or the intensification of the general biannual Australasian Monsoon (van der Kaars and De Deckker 2002). An increase in monsoonal activity at 14 000 yrs BP have also been indicated by studies of the Fitzroy River and Lake Gregory (English et al., 2001; Bowler et al., 2001; Hesse et al., 2004).

**9.1.3 Timing of the Australasian Monsoon**

Core BAR9403 provides insight’s into the operation of the Australasian Monsoon due to its location. Usually, phytoplankton blooms associated with increased nutrients are limited due to the capping effect of the low salinity boundary layer of the Throughflow (sensu Lukas and Lindstrom (1991), which increases during the SE Monsoon (Tomczak and Godfrey, 1994). However, the area where the core is located is away from the influences of the Throughflow during the LGM as the Sunda Strait was closed at 32 000yrs BP (Hanebuth et al., 2000). This passageway was blocked due to lower sea levels culminating at the LGM so, regardless if the Throughflow was more saline or not, its absence from this region gives us insight of the dynamics of the water column without the barrier cap of the Throughflow.

The ocean currents in this region are influenced by the dynamics of the Australasian Monsoon. There appears to be periods of productivity change within the water column and this may relate to alteration within the monsoonal systems of the
region. Results from the MAT SST reconstruction and relative abundance of planktonic foraminifera indicate that vertical mixing is reduced during MIS 3 and earlier stages of MIS 2. The faunal abundances reveal there was stratification in the water column promoting the development of a DCM layer which continued into MIS 2. This is also supported by the E. rex evidence of De Deckker and Gingele (2002). Gingele et al. (2002) also suggested that the NW Monsoon was more dominant during the 35,000 to 20,000 yrs BP period in the eastern Indian Ocean, due to the direction and source of clays in the South Java Current. It is probable that the NW monsoon was dominant at this time. Dominance of the NW Monsoon would have promoted conditions of ‘downwelling’ and stopped nutrients reaching the sea-surface. Recent observations of phytoplankton bloom along the Lesser Sunda Islands over a three year period showed that during the NW Monsoon 1997 and 1999 the eastward SJC restrained the flow from the Straits along the Lesser Sunda Islands and ceased blooming (Asanuma et al., 2003). In addition, the estimates of the annual position of the mixed layer in core BAR9403 also reveals that the mixed layer was thicker during MIS 3 than present day estimate by approximately 10m (see Fig. 31).

The results from core BAR9403 agree with De Deckker and Gingele (2002) that no major upwelling was recorded during glacial times. However, they believed the monsoonal winds were absent thus permitting the ocean to be permanently stratified. This also coincided with the removal of the low salinity barrier layer which encouraged high levels of silica and nutrients to be maintained near the surface. Another theory of how high nutrients levels could be maintained near the sea-surface is provided by Broecker et al. (2000). They suggested that high silica-content Pacific thermocline water entering the Indian Ocean during glacial times and this would have been the source of nutrients for the proliferation of E. rex along a narrow belt centered at 9°S.

Recent data from Asanuma et al. (2003) during an anomalous monsoonal year (1998), indicated the surface layer were stratified (wind speeds were reduced during the NW and SE Monsoons) with a surface temperature more than 30°C. This high SST was maintained without wind driven mixing in the surface (Asanuma et al., 2003). This is believed to be a result of anomalies in the distribution of pressure systems between the Pacific and the Indian Ocean following an El Niño event, that (1) resulted in the eastward SJC which flowed away from the coast in the NW Monsoon. (2) No westward SEC was observed in the SE Monsoon. (3) The eastward SJC restrained the flow from the straits in the SE Monsoon and (4) Chlorophyll a concentrations ~ 1 mg m⁻³ were observed throughout the year. The persistence of conditions such as these may be what
was occurring during MIS 3 and the earlier stages of MIS 2 and supports van der Kaars and De Deckker (2002) who stated that the intensity of the Australasian Monsoon was reduced preceding the Holocene.

Due to the numerous observations of DCM development within the Indonesian Throughflow region, Spooner et al. (2005) suggested that the ITCZ was north of the Banda Sea during MIS 3 and MIS 2 most, if not all year round, promoting the trade winds to blow across the Banda Sea, without incursions of the ITCZ into northern Australia. This was also indicated by the SST maps of Barrows and Juggins (2005), who suggest that the ITCZ lied close to the equator during the LGM, reducing the variability of the system. This essentially has the effect of the operation of a ‘perpetual’ SE Monsoon and thus encouraging the reduction of the mixed layer. However, the development of a DCM layer also suggests that the intensity of the monsoonal system was reduced due to less wind forced mixing in the mixed layer. An (2000) provides a synthesis of evidence from loess deposits and palynology in China to suggest the summer monsoon of the East Asian Monsoon (SE Monsoon in the southern Hemisphere) was more dominant during MIS 3 and MIS 2. Stott et al. (2004) has also suggested that in the early Holocene the ITCZ was further north than present in the western tropical Pacific Ocean.

The abundance of oligotrophic species was reduced from the 20 000 to 7000yrs period and this indicates the removal of the stratified structure of the water column. Sea-surface temperature estimates also indicate that upwelling of nutrient rich, cooler water was reaching the sea-surface near the LGM-Holocene transition. At the LGM-Holocene transition De Deckker and Ginele (2002) recorded a high $\text{Ba}_{\text{excess}}$, and low $E. rex$ abundance. At approximately 14 000yrs BP, BAR9403 records its highest abundance of $Ga. bulloides$. This indicates that the Australasian Monsoon may have intensified at this time. Evidence of increased productivity in the Java Upwelling System in cores analysed by Ding et al. (2006), also suggested an enhanced East Asia Monsoon at ~ 8000 yrs BP.

Presently, during the SE Monsoon, the westward SEC and the southeasterly wind generated cyclonic eddies along the Sunda Islands. The blooming was observed over these cyclonic eddies, where nutrients were entrained to the surface (Asanuma et al., 2003). If the SE Monsoon was ‘stronger,’ then greater upwelling and eddy development would occur in the eastern Indian Ocean, which is not advantageous to the development of a DCM layer or for $E. rex$ blooms. De Deckker and Ginele (2002) also
record a substantial peak of carbonate deposition between 19 000 to 12 000 yrs suggesting an increase in primary productivity.

The Holocene appears to have more oligotrophic conditions in the mixed layer with increase abundances of *Gs. sacculifer* and *Gs. ruber* after 10-8 ka and a reduction of *Ga. bulloides* (see Fig. 28 and Appendix 3) and concurs with De Deckker et al. (2002) that the present oceanographic and climate system may have only been active over the last 10 ka.

### 9.2 Core MD61

In core MD61, the difference in $\delta^{18}O$ between the LGM and the Holocene is 1.50‰ (see Fig. 43), of which 1.05‰ is due to the global ice volume effect (Duplessy et al., 2002), and consequential change in sea-level. Using the oxygen-isotope fractionation relationship between CaCO$_3$ and water of 0.25‰/°C (Kim and O’Neil, 1997), the $\delta^{18}O$ residual of 0.45‰ only accounts for a SST difference of ~1.8 °C between the LGM and present. However, the MAT analogue technique suggests an average $T_{mean}$ difference of 6.8°C in SST between the LGM and the Holocene (see Fig. 47). Between MIS 12 and MIS 11 a $\delta^{18}O$ residual of 0.12‰ results after the sea-level estimate of 1.5‰ (Chappell, 1998) has been accounted for. Again, this only explains a 0.5°C SST difference, while the MAT SST range is 7.8°C between MIS 12 to MIS 11. It is possible the MAT SST estimates are not accurate but when investigated with species relative abundances and PCA analyses as discussed below, there is evidence that something is affecting the $\delta^{18}O$ record beyond the effects of sea-level and SST differences.

Waelbroeck et al. (2005) demonstrated that there is a difference between ‘equilibrium calcite’ and fossil foraminifera $\delta^{18}O$ due to a complex interrelationship of physical and chemical hydrological interactions and ecological effects. It is also possible that the variability in the $\delta^{18}O$ record has been affected by advective mixing of different water masses (Waelbroeck et al. 2005), and may indicate the increased influence of SICW (driven by the WAC) over the core site (MD61) during the glacial periods. This supports Holbourn et al. (2005) analyses of productivity fluxes in the Timor Sea (~13°S) who also suggested higher productivity during glacial periods due to the increased influence of the WAC.

Another component which may have affected the isotopic record is the influx of low salinity water from regional river systems such as the Fortescue River in Western
Australia. It is noted that ‘terrestrial’ red desert clays are a feature of the core during interglacial periods and may be a result of increased riverine flow during these periods. However, this would have made the difference between interglacial and glacial $\delta^{18}O$ larger. It is possible the salinity signal has been lost due to the influence of the SICW at the core site during the glacial periods.

The Factor and the PCA performed in MD61 indicates an alternation between tropical-subtropical, oligotrophic conditions during interglacial periods and subtropical-transitional, oligotrophic conditions at the sea and sub-surface during glacial periods (see Fig. 49 and 52). Importantly, this also suggests that there has been a change in the watermasses influencing the core site which can alter the $\delta^{18}O$ record. The details of these relationships will be explored further below.

### 9.2.1 Reduced Leeuwin Current

The persistence of tropical surface dwelling species such as *Gs. sacculifer* through all the Marine Isotopic Stages represented in core MD61 suggests that the Leeuwin Current was continually present during the last ~550,000 years BP (see Fig. 45 and Appendix 7). The relative abundance of *Gs. sacculifer* declined to its lowest abundance of 4\% during the LGM from an average abundance of 10-12.5\% during interglacial periods. The presence of *Gs. sacculifer* indicates that tropical ITF water was still present but considerably reduced possibly due to more arid glacial climates in Australia and sea-level (van der Kaars, 1991; van der Kaars and Dam, 1995, Gingele et al., 2001; De Deckker and Gingele, 2002; van der Kaars and De Deckker, 2002; 2003). This suggests that the Leeuwin Current still flowed southward at the location of core MD61 but in a reduced capacity due to the lower input of ITF water during glacial periods.

The relative abundance of subtropical to tropical surface dwelling species *Gs. ruber* was also reduced during glacial periods but it still remained a dominant species (~17.5\%). This suggests cooler subtropical to transitional conditions of the upper surface layer during glacial periods. Today, the upper section of warm, low salinity tropical water (ITF) becomes colder and the tropical characteristics are diminished as the current moves southward due to mixing with subtropical, salty waters (STW) from the south and heat loss via evaporation to the atmosphere (Rochford, 1969, Wyrtki, 1971, Fieux et al., 2005).
The greater reduction in relative abundance of *Gs. sacculifer* compared to *Gs. ruber* could represent a greater component of STW and the removal of ITF properties from the surface layer over the core site. This does not mean the southward flow of the Leeuwin Current was removed as presently the front of ITF and STF/SICW occurs at 24°S at the sea surface (Martinez et al., 1999) and characteristics of the Leeuwin Current are noted as far as the western coastline of Tasmania (Ridgeway and Condie, 2004). In addition, the glacial relative abundance of *Gs. ruber*, *Gs. sacculifer* and *Ga. bulloides* in core MD61 are of a similar magnitude to the relative abundance of these species measured between 24-30°S (southeastern Indian Ocean) by the core top study of Martinez et al. (1998); an area presently influenced by a greater component of STW compared to ITW.

In core MD61, the relative abundance of tropical and subtropical species was reduced during glacial periods (see Fig. 45 and Appendix 7). However, PCA revealed that the variability displayed by species living at the sea-surface was minor compared to the variability displayed by sub-surface species (see Fig. 52). It is shown that transitional species *Gs. inflata* was the most influential species during glacial periods and caused the most variability (54%) within the species counts (see Table 12 and 13). In the record of MD61, *Gr. inflata* appears during glacial periods with a peak abundance leading the $\delta^{18}$O isotopic minimum of the LGM. Conversely, during interglacial periods *Gr. inflata* records very low abundances of <1% (see Fig. 45 and Appendix 7).

The occurrence of *Gr. inflata* has been previously linked to South Indian Central Waters by Martinez et al. (1999) in the south-eastern Indian Ocean. Calcareous nannofossil studies such as Okada and Wells (1997) also suggested that SICW (driven by the WAC) increased in dominance during Marine Isotope Stage 2 and 6 (MIS 2 and 6) and this resulted in a weaker Leeuwin Current. Holbourn et al. (2005) also suggests the influence of the WAC during glacial periods from MIS 2 to MIS 12 due to increased palaeo-productivity signatures during these periods. However, as Holbourn et al. (2005) used benthic foraminifera they cannot comment on the location of the WAC but rather the influence of SICW/ITF frontal system at depth. Note from section 2.10.2 that presently the SICW/ITF frontal system at 500m depth is further north ~17°S than at the sea-surface.

In addition, Martinez et al. (1998) identified the ITF-(S)ICW front at the sea-surface by a significant decrease in the depth of the theromocline south of ~24°S. The AUSMAT-4 estimate of the mixed layer depth in MD61 indicates that the mixed layer became thicker during glacial periods and suggests that the ITF-SICW frontal system
may have moved north and be located above the core site during glacial periods. Using the AUSMAT-4 the best indication of the processes in the water column is with the mixed layer depth in tropical and subtropical locations (Dr Tim Barrows, personal communication). Therefore, a thicker mixed layer and the increase in the abundance of *Gr. inflata* indicates that there is an increased influence of SICW in the mixed layer during glacial periods.

Presently, one aspect of the seasonal variability of the Leeuwin Current is when the WAC strengthens and the Leeuwin Current is weakened during the austral summer. The WAC carries SICW and therefore, the results of core MD61 also suggest the WAC was more dominant during glacial periods and agrees with previous findings by Prell et al. (1980), Wells and Wells (1994), Barrows et al. (1996), Okada and Wells (1997), Barrows and Juggins (2005) and Holbourn et al. (2005).

Further evidence that the Leeuwin Current was not removed from the core site during glacial periods is shown by the relative abundance of species that respond to conditions of upwelling. Upwelling species such as *Ga. bulloides, Gr. menardii* and *Gn. glutinata* did not increase in abundance during glacial periods (see Fig. 45 and Appendix 7). This suggests the Leeuwin still flowed restricting the development of widespread upwelling along the western Australian coastline. It is a common paradigm that the along shore pressure gradient restricts upwelling along the western Australian coastline (Hanson et al., 2005). Due to the lack of evidence for sustained upwelling and the persistence of warm water species, it is possible that a thin veneer of warm water still existed to block the movement of cold nutrient rich waters from reaching the sea-surface. This would be analogous to what is seen presently further south of the core position from Perth to Cape Leeuwin where SST of ~20 °C exist and prevent prolonged steady state processes of upwelling (Gersbach et al., 1999). Benthic foraminifera were analysed by Murgese (2003) in core Fr10/95 GC17 (very close to the location of core MD61) showed high percentages of *C. wuellerstorfi* during the 31 000 to 18 000 yrs period, indicating oligotrophic conditions and increased ventilation by lateral advection of active bottom currents.

Estimates of water temperatures at the 50m water depth by the MAT analogue reveal a significant reduction of water temperature at 50m compared to the estimated temperature at 150m and suggests the movement of SICW towards the sea-surface (see Fig. 47). Another consideration is the SICW, being an oligotrophic watermass, may not carry enough nutrients to support a bloom. Hanscn et al. (2005) suggests that the modern day “upwelling” seen along the Gascoyne coastline is not of global standards of
other upwelling sites around the world. Nutrient levels were low as a result of being sourced from the base of the Leeuwin Current thermocline in the Capes and Ningaloo Currents (Gersbach et al., 1999; Hanson et al., 2005). In addition, the deepening of the mixed layer and the nutricline may reduce nutrient availability during glacial periods. Presently, the nutricline, defined by the depth of 1.0 mol/l of phosphate concentration (Molfino and McIntyre, 1990) is deeper further south of the core location at – 500m (Conkright, 2001).

The relative abundance of DCM species *N. dutertrei* also indicates that waters around the thermocline were disrupted during glacial periods, especially from MIS 10 until present. Presently, near the location of core MD61, a well-defined deep chlorophyll maxima is located near the nutricline at around 70m below sea level (Hanson et al., 2005). The thicker mixed layer suggested in glacial periods by the MAT would result in the thermocline/nutricline moving deeper in the water column reducing light availability and thereby reducing the productivity of the DCM layer. Each glacial period except for MIS 12 and MIS 4 resulted in a reduced abundance of *N. dutertrei*. This is supported by Martinez et al. (1999) who observed a conspicuous reduction of *N. dutertrei* and increase in *Gr. inflata* during the LGM. The thickness of the mixed layer may be a result of ‘caballing’ (downwelling) along the ITW-SICW front as suggested by Martinez et al. (1999), due to similar characteristics of the glacial ITW (colder and denser) with SICW.

It is indicated by the relative abundance and PCA analysis that the modern-day characteristics of the Leeuwin Current may have been different during glacial periods. This is supported by a reduction in SST (Tmean) of ~ 6-9°C in core MD61 (see Fig. 47). However, it can be presumed that the surrounding offshore waters would have also reduced in SST and increased in salinity during global glaciation. The relative difference in regional SST may indicate whether there is still a glacial Leeuwin Current flowing southward carrying relatively warm water into higher latitudes. Presently there is an obvious difference in SST and salinity between the flow of the Leeuwin Current and the surrounding waters (Ridgeway and Condie, 2004).

As stated in the methods chapter, SST maps were constructed from regional re-estimated SST by Barrows and Juggins (2005) from previous studies and combined with the results of MD61 (see Fig 67-73). However, the extent of Barrows and Juggins (2005) study is only to the LGM, as this is one of the coldest phases in the MD61 record, the dynamics of the Leeuwin Current during other glacial periods will be inferred from the LGM SST maps. Presently, the Leeuwin Current is seasonally variable
with a registered Tmax during March-April and Tmin during September (Conkright, 2001; see Fig 67 and 68), resulting in approximately a 4°C difference at each core site. The SST maps show this still occurs during the LGM as the Tmax and Tmin measures show a similar temp difference of approximately 4°C (see Fig 69-70).

Presently, the Tmax occurs when the Leeuwin Current is strongest in April whereas Tmin corresponds to the period where the Leeuwin Current is reduced, weakest in September – March (Pearce and Pattiaratchi, 1999). Seasonality within MD61 (Tmax – Tmin) is therefore a measure of the variability of the Leeuwin Current. A seasonality difference of 4.6°C is recorded during the LGM, which is equivalent to modern day estimates (Conkright, 2001), but 1.5°C more than seasonality measure of the MAT (3°C) during the Holocene. The increase in seasonality during the LGM is due to greater change in the Tmin estimates (September). In addition, seasonality is inversely proportional to the mixed layer depth. More variability displayed by the Tmin and the relationship with the mixed layer depth again suggests a stronger West Australian Current.
Map of the modern day SST of the southeast Indian Ocean during the warmest month (February). SST are taken from World Ocean Atlas 2001–February (Conkright, 2001) at the location of cores previously analysed by Prell et al. (1980), Wells and Wells (1994), Martinez et al 1999. Note the southward depression of the isotherms indicating the location of the Leeuwin Current.
Figure 68
Map of the modern day SST of the southeast Indian Ocean during the coldest month (September). SST are taken from World Ocean Atlas 2001–September (Conkright, 2001) at the location of cores previously analysed by Prell et al. (1980), Wells and Wells (1994), Martinez et al. 1999. Note the southward depression of the isotherms indicating the location of the Leeuwin Current.
Figure 69
A SST map of the warmest month during the LGM. Note the cooling of water at the location of MD61. Also note the isotherms that are still depressed southward (similar to today) but the temperature of water is equivalent to offshore SST. SST are taken from Barrows and Juggins (2005) from cores previously analysed by Prell et al. (1980); Wells and Wells (1994), Martinez et al 1999. The SST estimates of MD61 of this study are derived from the same MAT as Barrows and Juggins (2005).
Figure 70
A SST map of the coldest month during the LGM. Note the cooling of water at the location of MD61. Also note the isotherms that are still depressed southward indicating warmer water near the coast compared to offshore water (similar to today). SST are taken from Barrows and Juggins (2005) from cores previously analysed by Prell et al. (1980), Wells and Wells (1994), Martinez et al 1999. The SST estimates of MD61 of this study are derived from the same MAT as Barrows and Juggins (2005).
Figure 71
A SST map of the southeastern Indian Ocean indicating the difference between the modern SST and the LGM during the warmest month (Tmax). Note the large cell of cooling above MD61 possibly due to the location of the ITW/SICW front. Tmax SST are taken from Barrows and Juggins (2005) from cores previously analysed by Prell et al. (1980), Wells and Wells (1994), Martinez et al 1999. The SST estimates of MD61 of this study are derived from the same MAT as Barrows and Juggins (2005).
Figure 72
A SST map of the southeastern Indian Ocean indicating the difference between the modern coldest month SST (Tmin) and during the LGM the coldest month (Tmin). Note the large cell of cooling above MD61 possibly due to the location of the ITW/SICW front. Tmin SST are taken from Barrows and Juggins (2005) from cores previously analysed by Prell et al. (1980), Wells and Wells (1994), Martinez et al 1999. The SST estimates of MD61 of this study are derived from the same MAT as Barrows and Juggins (2005).
Figure 73
A zoomed in image of the SST around core MD61 indicating the cold cell over the core site during Tmax and Tmin months. Note the warm water flowing around the cold cell and the steep isotherm gradient above the core site indicating the location of the ITW/SICW front.
The SST estimates indicate that the boundary of the Warm Pool was not located above the core site during glacial periods, with the 28°C isotherm contracting to approximately 17.5°S (Tmax map; see Fig. 69). The removal of the Warm Pool to 18°S was initially suggested by Martinez et al. (1999) and supported by Barrows and Juggins (2005). This significant shrinking of the Warm Pool would reduce the amount of fresh warm source water (ITF) feeding the Leeuwin Current and this seems to be supported by the reduction of tropical surface species in core MD61 during glacial periods. The *Gs. sacculifer / N. dutertrei* ratio of Martinez et al. (1999) was used to indicate the location of the Warm Pool (see Appendix 15). This ratio in core MD61 was less than one during glacial periods again indicating that the Warm Pool was removed.

Comparing the modern SST maps with the LGM maps one can see that south of 25°S the temperatures along the coast become equivalent to the surrounding waters during Tmax but SST are still warmer along the coast during Tmin. This is also shown by the anomaly maps with greater cooling along the coast during Tmax compared to Tmin (see Fig. 71 and 72). Again, this may be due to the removal of the Warm Pool which is present in the core area during the modern Tmax period and will be discussed in more detail below. However, the isotherms are still depressed southward near the coast during Tmax and Tmin suggesting the Leeuwin Current still exists (see Fig 69 and 70). This is supported by the SST maps of Barrows and Juggins (2005) over the whole of the Indian Ocean. Again one can see the isotherms depressed southward near the coast suggesting the Leeuwin Current still flows (see Fig. 17b; Chapter 4). The regional SST maps during the LGM suggest that the strength of the Leeuwin Current was reduced as less warm water was being transported to higher latitudes and there was less of a difference of SST with surround waters. What also seems to be affecting the isotherm patterns is the shift northward of the ITW-SICW front with greater SST gradients north of MD61 and reduced SST gradients south of MD61.

Although the relative abundance counts do not indicate upwelling during glacial periods the SST maps indicate an area of colder water off the North West Cape in the location of MD61 and neighbouring core Fr10/95-17 during the LGM. It is also shown that warmer water surrounds this cold cell and suggests that the Leeuwin Current may flow around this anomalous cold cell (see Fig. 73). Presently, upwelling plumes are established off the North West Cape during the austral summer (Hanson et al., 2005) and, although the foraminifera abundances do not suggest upwelling, it appears that
colder water is reaching the surface for sustained periods as indicated in the Tmax and Tmin SST maps and the reduced temperature at 50m recorded by the MAT.

The cold band appears more widespread than the cold cell on the surface SST maps. This may be a signal of the ITF-SICW frontal system below the sea-surface which may breach the surface at the location of MD61 due to natural tendencies for upwelling in the area. If the dynamics of this system was maintained beyond the LGM, and the fossil record indicates it does, then MIS 10 may have a greater development of ITW-SICW front, due to significantly colder Tmean estimates during these glacial stages compared to the LGM. The glacial Tmean estimates of MIS 6 and MIS 12 were equivalent to the LGM as the Tmean estimates were within the Tmean error of 0.84°C.

It is possible that the reduction in SST is over estimated due to the influx of sub-surface species, which was noted by Martinez et al. (1999) and this cold band may be more of an indication of what is happening with the thermocline waters (sub-surface waters) rather than the surface flow. The estimates of SST and the thickness of the mixed layer is suggestive that the ITF-SICW frontal system that meets at the sea-surface presently at 24°S (Martinez et al., 1998) has moved 3-4° northward during the LGM. This is supported by the zoomed in image of the Gascoyne coast which shows that the SST reduces dramatically near 20°S which is also suggestive of a frontal system (see Fig. 74). There may not be as large difference between the location of the frontal system at depth and the surface location which is seen presently due to the reduced production of ITF waters during the LGM.

The establishment of large areas of cold surface water has been documented previously off the coast of Western Australia by (Prell and Hutson, 1972; Prell et al., 1979, 1980, CLIMAP, 1984, Prell, 1985; Wells and Wells, 1994, and Martinez et al., 1999). Wells and Wells (1994) found anomalous cooling of -6 to -7°C for cores along the Gascoyne Coast but only 3-4°C during MIS 6. It was argued that the Leeuwin Current was absent because of the large change in SST. However, reanalysis of SST of their cores by Martinez et al. (1999) and Barrows and Juggins (2005) combined with the results of MD61, indicate that this area is isolated, and only -3°C cooler, and it appears that warm water surrounds the cold patch at the sea-surface (see Fig 69 and 70). The existence of warm water was also indicated by the factor and PCA analysis on the species counts in MD61. This is also supported by the factor analysis of Martinez et al. (1999) who found a dominance of *Gs. inflata, Gn. glutinata, Ga. bulloides* and *Gs. ruber* in the principal Factor of the LGM suggesting seasonal influences of warm water due to the abundance of *Gs. ruber*.
If the Leeuwin Current was removed one would expect all along the coast to present a greater change in SST that cores away from the path of the Leeuwin Current. Wells and Well (1994) suggested this was the case but Barrows and Juggins (2005) showed the reduction of SST was overestimated, which is often found in transfer function estimates (see Chapter 6). In the SST maps, we see this only occurs at one point along the Gascoyne Coast (see Fig. 69-70).

Analysis of benthic foraminifera by Wells et al. (1994) did not find a strong signal for productivity during the LGM. However they suggested that during the penultimate glaciation (MIS 6) upwelling was established within the areas of low SST indicated by planktic foraminifera (Wells et al., 1994). In core MD61, there is no evidence of upwelling during MIS 6 and instead the faunal assemblages show a similar pattern as the LGM with the dominance of *Gr. inflata*. Therefore, a greater proportion of SICW was found within the mixed layer during MIS 6 compared to the interglacial periods. The absence of upwelling during MIS 6 has also been suggested by Okada and Wells (1997) as small placoliths (nannofossils) did not increase in abundance during MIS 6. The increase in abundance of *Gr. inflata* during glacial periods in core MD61 appears to support the findings of Okada and Wells (1997) who found the reduced abundance of nanofossil *Florisphaera profunda* was due to lower temperatures in the lower photic zone rather than increased productivity in the upper photic zone during MIS 2 and 6.

A study on organic carbon from cores Cores 53GC04, 53GC07, 57GC19 and 57GC15 by Veeh et al. (2000) on the Exmouth Plateau and the Perth Basin showed little support for enhanced productivity off Western Australia during the LGM and dismissed that the Leeuwin Current was removed and replaced by the West Australian Current during the LGM. This is significant because a previous study by McCorkle et al. (1994) on the same core material used by Veeh et al. (2000) suggested that there was a significant increase in productivity during the LGM. McCorkle et al. (1994) used mass accumulation rates of benthic foraminifera to indicate increased productivity during the LGM but Veeh et al. (2000) suggests the increase in MAR found by McCorkle et al. (1994) may be elevated due to recycled organic matter during sea-level lowstands of the LGM.

Takahashi and Okada (2000) suggested the Leeuwin current was weakened at the site of core Fr 10/95-17 but did not reach core Fr10-95-20 during the LGM. However, these assumptions were based on angle types of species *Gephyrocapsa* (nannoplankton) to infer SST which cannot be converted to a numerically estimate.
Further south of MD61, in the SST maps (see Fig. 69-70), one can see that SST is equivalent to further offshore but the isolines bend southward near the coast suggesting warmer water is still present along the shelf break. In addition, there does not appear to be a cold cell along the Capes Current coastline (34°S) during the LGM which would be expected to expand if the Leeuwin Current was removed.

It is suggested from core MD61 that the cold cell along the Gascoyne Coast is an isolated phenomenon due to the location of the ITW-SICW frontal system, not due to removal of the Leeuwin Current. This cold cell was not due to processes of ‘upwelling’ in the classical sense as suggested by Wells and Wells (1994) and Wells et al. (1994) as the mixed layer was significantly deeper during glacial periods and upwelling species such as *Ga. bulloides* did not respond. This cold cell is more likely to be due to the breaching of the ITW-SICW front at the sea-surface, as presently the Gascoyne Coast appears to represent a weak spot in the alongshore pressure gradient (Hanson et al., 2005).

**9.2.2 Variability of the Leeuwin Current**

Ridgeway and Condie (2004) recently confirmed that the along shore pressure gradient is the dominant force driving the Leeuwin Current along the western coastline of Australia but also acknowledged that the seasonality in the strength of the Leeuwin Current is due to variations in both along shore pressure gradient and the coastal wind stress (south-westerly winds). For the flow of the Leeuwin Current to be interrupted or weakened during glacial periods, then the strength of the along shore pressure gradient must be reduced or the along shore pressure gradient is maintained and the opposing wind stress is stronger.

The evidence from MD61 suggests that less ITW was entering the Leeuwin Current during the LGM. Less input of ITW suggests a reduced geostrophic gradient as seen during El Niño events (Meyers 1996; Feng et al., 2003). In addition, the greater influence of SICW suggests a greater presence of the West Australian Current. The greater dominance of the West Australian Current suggests either that wind conditions are favourable for its flow, which is at its maxima during November-January presently, and/or the alongshore pressure gradient is less. What would aid in the dominance of SICW is the apparent movement northward of the subtropical front as reported by (Prell et al., 1979; Howard and Prell, 1992; Martinez et al., 1994; Passlow et al., 1997), which drives the production of SICW.
The difficulty in the interpretation in core MD61 is that it appears that the ITF/SICW front is over the core site during glacial periods thickening the mixed layer. One would expect the mixed layer to thin if the Leeuwin Current was reduced away from the core site (further south) as there would be less onshore transport resulting in downwelling. However, as the front has moved northward it can be assumed the pressure gradient has been reduced.

As shown in Chapter 2, the interaction with the Throughflow and the Leeuwin Current is complex and the Leeuwin Current is not just an offshoot of the ITF. This is due to a complex interrelationship of circulation in the Indo-Australian Bight with the SJC and SEC and wind directions. Maximum strength of the ITF is observed during the austral winter in August/September (Wyrski, 1987; Molcard et al., 1996), timed with the SE Monsoon, but this does not result in a stronger Leeuwin Current. This actually corresponds with a minimum of transport of the Leeuwin Current. The Leeuwin Current, by this time of the year, is weak and the West Australian Current dominates circulation along Western Australia (Tomczak and Godfrey, 1994).

However, due to observations by Godfrey (1996), Martinez et al. (1999) suggested that the along shore gradient should be stronger during the LGM because of upwelling off the coast of Java encouraging the flow of the Throughflow from the Pacific and into the SEC. Godfrey (1996) suggested that as the ITW flows into the SEC some water is carried southward out of the SEC by action of Ekman transports associated with strong trade winds. As this water losses heat there is a southward decrease in the steric height in the eastern Indian Ocean. This longshore steric height cannot be balanced geostrophically so flow accelerates southward (Cresswell and Golding, 1980; Godfrey, 1996).

As shown in figures 69-70 compared to figures 67-68 and the fossil assemblages, it appears that less ITW is entering the Leeuwin Current at approximately 20°S. In addition, the SST maps of Tmin and Tmax show the SST gradient is greater during modern-day by 1-1.5°C, respectively taken at from reference points 20°S, 115°E to 33°S 115°E. Tmax modern has the largest SST gradient of 6°C and the smallest gradient of 4.5°C occurs during both the Tmax and Tmin in the LGM. This could suggest the alongshore pressure gradient is less during the LGM and probably for the other glacial periods recorded in core MD1.

As Morrow and Birol (1998) state, “the variability in the systems of the eastern Indian Ocean are directly forced by the strong seasonal monsoons, which provide local
and remote wind forcing; but is also influenced by the remote ocean forcing between the Pacific and Indian ocean via the Indonesian Throughflow”.

It is possibly the conditions presented for the LGM are due to a persistent SE Monsoon. The Leeuwin Current first reduces in strength in September and the change in wind direction, initiates the equatorward wind stress during the onset of the SE Monsoon (Godfrey and Ridgeway, 1985). This suggestion is based on the faunal assemblages, as it is evident less ITW is entering the Leeuwin Current and there appears to be an increased dominance of the West Australian Current. The West Australian Current is strongest in September–November (Wijffels et al., 1996). The persistence of a SE Monsoon would weaken the Leeuwin Current but not completely remove it (as seen presently), and result in ITW water being fed into the SEC. It has been suggested by De Deckker (1998), Martinez et al. (1999), Takahashi et al. (2000) and Gingele et al. (2001) that a more saline ITF during the LGM (glacial periods) would encourage the flow of ITW into the SEC, which is also at its maximum strength during the SE Monsoon.

The PCA analysis in MD61 isolated *Gs. rubescens* into a Group 4 (1-2; see Fig. 52). This species lives in the upper 50m and typifies the central subtropical regions of the Indian Ocean, where the waters are low in phosphate, very high in salinity, with intermediate levels of oxygen and temperature (Bé, 1977). It is possible this species represents a salinity component in the faunal assemblages, as it appears to be in core BAR9403. The abundance of this group is erratic but the samples scores indicates this group dominates during glacial periods MIS 10, MIS 8, MIS 4 and sporadically during MIS 6 (see Table 13), while *Gr. inflata* completely dominates during the LGM.

The persistence of the SE Monsoon during the LGM has been suggested by Barrows and Juggins (2005) and Spooner et al. (2005). Barrows and Juggins (2005) suggest this is due to the ITCZ sitting close to the equator throughout the year and removing the incursion phase of the NW Monsoon into northwestern Australia. The persistence of the SE Monsoon and, therefore, the reduction of the NW Monsoon may be the reason why there is more of a difference in SST along the coast compared to the surrounding waters in the Tmin map compared to the Tmax map (see Fig 67 and 68). Perhaps the removal of the Warm Pool and the reduction of warmer/wetter monsoon period is why less warm, low salinity water is building up in the Throughflow region to be transported by the Leeuwin Current. A reduced NW Monsoon would also reduce the strength of the SJC which delivers water into the Leeuwin Current during the later stages of the NW Monsoon and thereby aid in the reduction of the Leeuwin Current.
While during $T_{min}$ a geostrophic flow is being encouraged by the strengthening of the SEC and warmer water is transported southward. This hypothesis will be explored in the southern core site MD2607 as the Leeuwin Current is wind driven along the southern coast and is strongest when the westerly winds dominate (timed to when the SE Monsoon strengthens in northern Australia).

The fossil record in core MD61 is a signal for the overall seasonal variation of the Leeuwin Current. The evidence from the fossil record of core MD61 suggests that the Leeuwin Current was weaker during glacial periods due to a strengthened West Australian Current during glacial periods either because the alongshore pressure gradient was reduced and/or the northward along shore winds were increased. Without direct measurements, it is difficult to say if it is due to increased wind or reduced alongshore pressure gradient. Regardless, the steric height appears to be strong enough to prevent upwelling as conditions of downwelling were encouraged by the steric gradient south of the core site. In addition, the SST maps indicate that the Leeuwin Current still flows.

9.2.3 Enhanced Leeuwin Current

The expansion of tropical water (ITW) over the MD61 core site would provide a favourable habitat for tropical, oligotrophic planktonic foraminifera. An increase in the abundance of these species could also imply that the Leeuwin Current was stronger at certain periods in the past. Comparison between the interglacial periods recorded in MD61 such as the Holocene and Isotope Stages 7, 9 and 11 shows that the relative abundance of *Gs. ruber* and *Gs. sacculifer* was equivalent to modern-day abundances (see Fig. 45 and Appendix 7). However, an increase in relative abundance of tropical species *Gs. sacculifer* during MIS 5e suggests that the Leeuwin Current may have been stronger compared to the present day. In addition, the relative abundance of another tropical species, *P. obliquiloculata*, indicates warmer sub-surface conditions and possible enhancement of the Leeuwin Current during MIS 5 and 7 (see Fig. 45 and Appendix 7).

PCA analysis of component 1 versus component 2 combined tropical surface and sub-surface species into Group 3 (1-2). The peaks in relative abundance of Group 3 (1-2) indicates that the Leeuwin Current was stronger and possibly thicker during MIS 5, MIS 7 and MIS 11 compared to present (see Fig. 52). The influence of the Leeuwin Current at greater depths is indicated by the species scores in Component 2 and 3 which
accounts for ~ 11% of variability in the relative abundance counts. The sample scores also indicate the dominance of Group 3 (1-2) over other groups during MIS 5, 7 and 11 (see Table 13). In comparison, the Holocene period is dominated by the subtropical Group 2 (1-2) and MIS 9 was equally dominated by subtropical and tropical groups (see Fig. 52). The increased influence of ITW at the sub-surface and the reduction of SICW indicate that the WA Current has weakened and it can be inferred that the Leeuwin Current is stronger as a result.

The increased influence of ITW during interglacial periods is also shown by the SST which indicates most interglacial periods were equivalent or warmer than the present Tmean of 27.5°C. During MIS 5e, a peak Tmean of 29.1°C is recorded, while MIS 11 records a peak Tmean of 28.6°C (see Fig. 47). Both of these estimates are significant as it is beyond the error range of 0.83°C. Seasonality is most reduced during MIS 5 with only 1°C difference between Tmax and Tmin (see Fig 47), indicating warmer surface and sub-surface waters especially during the Tmin estimate (~September).

The results from core MD61 show that the tropical component of the Leeuwin Current was thicker with a greater influx of tropical ITW into the flow. A greater influx of tropical waters suggests a stronger along shore pressure gradient and therefore supports previous evidence of an enhanced Leeuwin Current during MIS 5. Wells and Wells (1994) found an SST increase during MIS 5 due to warmer winter temperatures up to 3°C and the increase development of coral fringing reef systems along the coast of Western Australia (Collins et al., 2003) who all suggest an enhanced Leeuwin Current during the last interglacial period (MIS 5e). In addition, the results from core MD61 indicate a stronger Leeuwin Current also occurred during MIS 7, part of MIS 9 and 11.

9.2.4 Upwelling events

There is evidence for upwelling in core MD61; however, this occurs during interglacial periods. A major upwelling event was centred around the later stages of MIS 11 with large increase (12%) in the relative abundance of Ga. bulloides. Conditions of upwelling are also indicated by tropical upwelling species Gr. menardii, towards the transition with glacial period MIS 10. These two species are grouped in principal component analysis Group 3 (1-3) and show a peak in abundance at the MIS 11. The other periods of higher abundance of Group 3 (1-3) is during the Holocene (see Fig. 54).
The evidence of upwelling during interglacial periods initially appears counterintuitive, i.e. if the Leeuwin Current was removed, then an upwelling system equivalent to offshore Peru or Namibia would exist. This is because the modern-day dynamics of the Leeuwin Current, positioned on the shelf break, promotes downwelling. Upwelling pulses are seen today along the shallow shelfal regions along the western coastline driven by the cool equatorward coastal Ningaloo and Capes Currents during the austral summer (Pearce and Pattiaratchi, 1998; Pearce and Pattiaratchi, 1999; Gersbach et al., 1999; Hanson et al., 2005). This process occurs when the Leeuwin Current is weaker with the current being pushed further offshore. Hanson et al. (2005) describe the processes of upwelling to be a combination of a) the depth of the Leeuwin Currents nutrient-depleted mixed layer, b) the strength and duration of upwelling favourable winds, and c) the geographic location. Presently, core MD61 is located under the pathway of the oligotrophic Leeuwin Current but according to Hanson et al. (2005), near the boundary with the more nutrient rich shelfal waters influenced by the Ningaloo Current. The Ningaloo and Capes counter-Currents are strongest during the austral summer driven by northward winds.

One parameter we can measure on MD61 core site is the thickness of the mixed layer. It was shown that the mixed layer was deep during glacial periods, believed to be due the presence of the ITW/SICW front. The periods of increased abundance of upwelling PCA group 3 (1-3) are when the mixed layer reduces in thickness and seemingly permit the upwelling of more nutrient rich waters to the sea-surface (see Fig. 54). Glacial evidence presented above, the mixed layer would thin when the ITW/SICW front is removed; possibly back to 24°S, where it is found today. This is significant as upwelling waters of the Ningaloo Current are sourced from the base of the thermocline where DCM layers develop (Hanson et al., 2005).

The reason upwelling may not be occurring during other interglacial periods may be due to wind direction. It is shown above that there is an influx of ITW water during interglacial periods suggesting a greater steric height and possibly reduced along shore winds, therefore, conditions similar to March-June period of the present day when the Leeuwin current is strongest. The seasonality of the Leeuwin Current is reduced during warmer phases such as MIS 5 as indicated by a warmer Tmin signal and therefore, it is possible the alongshore wind system was different. The slightly weaker Leeuwin Current during the Holocene, due to the greater influence of the along shore winds, may be why the transport of the Leeuwin Current seems higher during the other interglacial periods recorded in MD61. In addition, the greater transport of warm water
by the Leeuwin Current may bath the shelfal regions suppressing and overcoming the counter currents.

It has been shown that the NW Monsoon is believed to have been intensified at approximately 14 000 yrs BP (Wyroll and Miller, 2001) and this may provide more ideal wind conditions for upwelling near the core location. The peak of the upwelling group in Holocene occurs between the 15000 to 5000 yrs period and may represent the increased intensity of the general Australasian Monsoonal system as suggested by van der Kaars and De Deckker (2002). Due to similar orbital parameters during MIS 11 and present (Berger and Loutre, 2002; Loutre and Berger, 2003), a greater presence of the NW Monsoon may also have occurred during MIS 11. Another factor that could be contributing to the 'upwelling signal' is the increase of fluvial discharge as a consequence of increased precipitation. From 14 000 to 5000 yrs BP the region was characterised by substantial summer rains (Veeh et al., 2000; Gingele et al. 2001; van der Kaars and De Deckker, 2002), which is timed to the upwelling signal in core MD61. The timing of the upwelling events during MIS 11 and the Holocene is associated with the deposition of red clays from river discharges. However, the upwelling species did not increase during other interglacial periods with similar sedimentation such as in MIS 5. This discredits the possibility that foraminiferal species may be responding to nutrients delivered to the core site via the river sediment load.

The benthic foraminifera analysed by Murgese (2003) in the eastern Indian Ocean did not respond to higher primary productivity from MIS 3 to present but this may be due to strong bottom currents and lateral transport of nutrients. However, nannofossil evidence along the western coastline indicates mild upwelling for MIS 7 and 5 in core RS96G21 (Okada and Wells, 1997).

9.2.5 Climate Change

Many authors have acknowledged the effect of the Leeuwin Current on the regional climate (Maxwell and Cresswell, 1981; Pearce and Walker, 1991; Allan and Haycock, 1993; Hutchins and Pearce, 1994; Pearce and Pattiaratchi, 1999; Ansell et al., 2000; Feng et al., 2003). This seems just as significant during glacial periods. Mean SST estimates indicate that the glacial periods were generally 6-9 °C lower than present around 20-22 °C (see Fig 47). Conversely, SST of interglacial periods especially MIS 5, 7 and 11 were around 28-29 °C. The estimates of SST during interglacial periods are significant as they are above the 27.5°C transition point where deep cloud convection increases (Waliser and Graham, 1993). During glacial periods, the reduced SST would
remove the necessary processes at the ocean atmosphere interface and result in a significant reduction of cloud cover and hence precipitation.

The SST record of core MD61 seems to link with regional indications of an arid climate during glacial periods (Gingele et al. 2001, van der Kaars and De Deckker, 2002, van der Kaars and De Deckker, 2003). The Leeuwin Current is the only eastern boundary current with a large heat transfer from the ocean to the atmosphere, with values equivalent to a western boundary current (Josey et al., 1999) and the ocean–atmosphere heat flux is largest along the southwest Australian coastline where the annual heat loss exceeds 50Wm$^{-2}$ (Morrow et al., 2003). Therefore, it is highly likely that the Leeuwin Current contributes to the regional aridity signal with less warm water being carried to higher latitudes resulting in reduced hydrological convection processes during the glacial periods.

The age versus depth profile of core MD61 revealed a step-like pattern with increased rates of sedimentation corresponding to interglacial periods. The interglacial periods are also characterised by the increase of reddish brown sediment (see Fig. 23) in the core. The colour of this sediment is typical of the arid soils of inland areas of Western Australia. This represents oxide and sesqui-oxide coatings acquired by quartz and other minerals under sub-tropical aerobic weathering environments (Shulmeister et al., 2004). The increase of red clays suggests an increase in precipitation and river discharge during interglacial periods compared to glacial periods. It also appears there is no effect of dust accumulation during glacial periods as the sedimentation rate is very low during glacial periods and the sediment consisted of foraminiferal sands and calcareous ooze.

Presently, the Pilbara and Gascoyne regions have a hot dry climate with at times intense rains associated with summer cyclones (Department of Environment and Heritage, 2001; see Fig 74). As tropical cyclones originate in oceanic areas where SST exceeds 28°C (Webster and Stretch, 1978, Martinez et al., 2002), this process is significantly reduced during glacial periods and contribute to the aridity of the region. Veeh et al. (2000) found significantly lower input of terrigenous sediments during the LGM on the Exmouth Plateau from the flux patterns of aluminium and thorium compared to the Holocene.
Figure 74

The late Holocene appears to be drier than the early Holocene in the regional terrestrial records (Magee and Miller, 1998; Luly, 2001; Wyrwoll and Miller 2001; van der Kaars and De Deckker, 2002 and van der Kaars and De Deckker 2003). Interestingly, the PCA analysis suggests a progressively greater dominance of the subtropical group 2 (1-2) compared to the tropical group 3 (1-2) of the late Holocene but this does not seem to be represented in the SST estimates. In addition, the benthic record of Murgese (2003) indicates a highly diverse assemblage suggesting prevalent oligotrophic conditions from 5ka to present.

Core MD61 captures MIS 11 which is believed to be analogous for a future greenhouse climate. In core MD61, the mean SST at the warmest peak (MIS 11.3) was 28.6 °C compared to the present day estimate of 27.5°C, which is beyond the error range 0.83°C. The Tmax estimates 28.9°C and 29.9°C and Tmin of 25.7°C and 27.4°C were
recorded for the Holocene and MIS 11 respectively and these temperature differences are also significant.

The SST estimates from MD61 agree with terrestrial evidence from Europe (Rousseau et al., 1992; Loutre and Berger, 2003) and marine evidence from New Zealand (Howard, 1997; King and Howard, 2000) that suggests that MIS 11 was warmer than the Holocene and MIS 5. The Chinese loess–soil sequence shows that the summer monsoon was particularly strengthened during MIS 11 and 5 (Guo et al., 2000; Loutre and Berger, 2003). Moreover, sea-level highstands have been reported from Alaska, England, Bermuda and the Bahamas (Hearty et al., 1999; Kindler and Hearty, 2000; Loutre and Berger, 2003) and seem to agree with model estimates that both MIS 11 and the future are characterized by small amount (if any) of continental ice, with almost no variation during the whole interval (Loutre and Berger, 2003).

The SST estimates together with the PCA analysis of MD61 indicate a thicker flow of tropical water in the upper sea-surface. If the predictions of MIS 11 of being an analogous future climate are correct then it can be inferred there would be greater cyclone activity along the West Australian Coast bringing increased precipitation compared to present. The Leeuwin Current would also positively reinforce the characteristics of this interglacial period by delivering a greater amount of warm, low salinity water to higher latitudes.

9.3 Core MD2607

9.3.1 Enhanced Leeuwin Current

Along the southern coastline of Australia, the present-day distribution of planktonic foraminifera is highly influenced by the Leeuwin Current. The Leeuwin Current establishes a strong longitudinal gradation from warmer assemblages in the west to more temperate assemblages in the east near the Lacepede Shelf (Li et al., 1996a; Li et al., 1996b; Li and McGowran, 1998 and Li et al., 1999). Presently, the ratio of Gs. ruber to Gr. inflata is used to determine the strength of the Leeuwin Current along the southern coastline (Li et al., 1996a; Li et al., 1996b and Li and McGowran, 1998). While, at the core location near the Lacepede Shelf, the dominance of Gr. inflata is attributed to the presence of the Leeuwin Current (Li et al., 1996b and Li and McGowran, 1998). In addition, stratification occurs in the water column due to the
warm water of the Leeuwin Current. Almond et al. (1993) suggest that the greater diversity in interglacial periods is due to stratification of the water column as more niches in the water column can be exploited by the planktonic foraminifera.

The examination of the planktonic foraminifera in core MD2607 reveals the dominance of subtropical and transitional species during interglacial periods alternating with sub-polar species dominating in the glacial periods see Fig. 58 and Appendix 12). In addition, the interglacial periods have a higher diversity of planktonic foraminifera (25 species) compared to the glacial periods due to greater stratification of the water column.

Most of the variance within core MD2607 (~60%) is due to the changing relative abundance of subtropical species *Gs. ruber* and transitional species *Gr. inflata* during glacial and interglacial periods (see Fig. 58 and Appendix 12). Although both species dominate during interglacial periods, the PCA analysis separated the species into a transitional and subtropical group. The relative abundance change of the transitional group 1 (1-2) containing *Gr. inflata* and the sample scores indicate that this group is more dominant from approximately 8600 yrs BP to present (see Fig. 66). Conversely, the subtropical group containing *Gs. ruber* is more dominant during the 14 000 to 8600 years period and during the last interglacial period (see Fig. 58 and Appendix 12). This fulfils Li et al’s. (1996a) and 1999) findings where a greater ratio of *Gs. ruber* versus *Gr. inflata* indicates a ‘warmer’ (stronger) Leeuwin Current and a greater ratio of *Gr. inflata* would indicate slightly cooler (weaker) conditions. This may indicate that the Leeuwin Current was stronger during the last interglacial and the early Holocene compared to present.

The SST record at 50-150m indicates the development of a thermocline during interglacial periods and early MIS 3 possibly due to the presence of the Leeuwin Current. There is greater warming at the 50m interval, around 17.5°C during the Holocene and MIS 3-4, and 19.5°C during the last interglacial, compared to approximately 16-15°C at 100m and 150m (see Fig. 61). During glacial periods the 50-150m temperatures appear to be more uniform at around 11°C. The last interglacial has the largest SST gradient and this suggests the Leeuwin Current was stronger compared to present.

Thermal skins of warm high salinity water develop in the Great Australian Bight due to surface heating during summer (Herzfeld and Tomczak, 1997; Herzfeld, 1998), and joins the Leeuwin Current in winter near the eastern edge of the Great Australian Bight (Ridgeway and Condie, 2004). The maximum eastward extent of this Great
Australian Bight water is beyond 136°E in autumn, and the maximum SST is 2-3°C higher than surrounding water (Herzfield, 1998). It is assumed that GAB water and the Leeuwin Current would influence core MD2607 during interglacial periods. This was observed by Li et al. (1996a) on recent shelf sediments, and suggested a high *Gs. ruber*/*Gr. inflata* ratio that may signal the existence of a warm shelf current as it shifts to inner and outer shelf locations. It is also possible that the production of Great Australian Bight water may by enhanced during apparently warmer interglacials such as during MIS 5.

On the south western shelf, relict miliolid and discorbid benthic foraminifera are interpreted to have formed in shallow shelf to strandline embayments and lagoonal environments during numerous sea-level excursions particularly during MIS 3 and MIS 4 (Li et al., 1999). As Cann and Clarke (1993) show the Leeuwin Current delivers tropical-subtropical species to the southern coastline. However, warm, high-salinity lagoon water may spill onto the slope during eustatic sea-level change during interstadal periods and may be able to maintain the presence of these species. In core MD2607, the abundance of *Gs. ruber* during MIS 3 and MIS 4 may indicate the build up of warm water on the shelf, as species *Gs. ruber* has a high salinity tolerance (Bé and Hutson, 1977). However, *Gr. inflata* prefers a deep thermocline with little seasonal variation in salinity (Hilbrecht 1996). Therefore, the parallel increase in abundance of *Gr. inflata* and *Gs. ruber* suggests the presence of the Leeuwin Current during the interglacials.

A good indicator for a stronger Leeuwin Current is the large increase in dextral coiling *Gr. truncatulinoides* around MIS 5e (see Fig. 58 and Appendix 12). This very-deep dwelling species is representative of waters north of 25°S (Bé and Tolderlund, 1971; Bé and Hutson, 1977). A significant pulse of warm water must have occurred to increase this species abundance from near 0% at all other marine isotope stages to 12.5% during MIS 5. In addition, minor tropical-subtropical species such as *Gt. rubescens*, *Gl. aequilateralis* and *O. universa* also appear in the record with higher abundance during MIS 5 compared to the Holocene (see Fig. 58 and Appendix 12).

The SST reconstructions of core MD2607 also indicates that MIS 5e was warmer than present, which may suggest a strengthened Leeuwin Current along the south coast. A 2.5°C increase in *T* mean is recorded during MIS 5 and this is associated with a greater change in *T* max compared to *T* min (see Fig. 61). The warmest month (*T* max) is probably associated with both the warm saline water body formed in the eastern Australian Bight and the Leeuwin Current. Just looking at the SST estimates, it
is difficult to determine if the warming is due to shelfal waters or the influence of the Leeuwin Current. However, the timing of increased SST are synchronous with the northern core (MD61) indicating the Leeuwin Current was influential in increasing subtropical species and associated SST increases. In core MD61, a peak in abundance of tropical species compared to sub-tropical species occurred between 15000-4600yrs. This is similar to the period of dominance of *Gs. ruber* over transitional species *Gr. inflata* in core MD2607 (14 000 to 8600 yrs BP). Likewise MIS 5e is the warmest period for core MD61 and MD2607.

The SST estimates of core MD2607 are close to the 2-3°C increase in SST estimate by Murray-Wallace and Belperio (1991) on bivalve *Anadara trapezia* during high sea levels of the Glanville Formation during MIS 5. But less than the estimated 5°C increase in SST of Cann and Clarke (1993) for the occurrence of tropical species *Marginopora* in the Spencer Gulf. Generally, the results from core MD2607 agree with other studies that more warm water is entering the southern coastline during the last interglacial. The other evidence along the southern coastline of a stronger Leeuwin Current is the succession of estuarine warm water bivalve *Anadara trapezia* along the south-west coastal margin during MIS 5 and MIS 7 (Kendrick et al., 1991), and the high sea-level shorelines during MIS 5 of *Marginopora* and *Anadara* in the Glanville Formation, South Australia (Hails and Gostin, 1984; Murray-Wallace and Belperio, 1991; Cann, 1992; Cann and Clarke, 1993; Belperio, 1995; Murray-Wallace, 1995; Murray-Wallace et al., 2000). Coastal bitumen and amber found along the southern Australian coast were probably also transported through the same route from Indonesia (McGowran et al. 1997).

Regionally, the climate was more humid during MIS 5. Lakes in the region were filled, vegetation cover increased and dune activity and dust propagation ceased (Gingele et al., 2005). The increased flow of warm water along the south coast would promote evaporation and probably result in greater precipitation to this region.

**9.3.2 Glacial dynamics**

The two glacial periods represented in core MD2607 appear to have different water characteristics due to changing levels of nutrients in the water column. There is evidence of cooler eutrophic water within the water column with peak upwelling during MIS 6 (see Fig. 58 and Appendix 12). In addition, PCA sample scores (see Fig. 66), indicate Mode Water is more influential during the LCM compared to present.
The increase in abundance of *Gr. truncatulinoides* (s) during glacial periods causes this species to become one of the dominant species of core MD2607 (see Fig. 58 and Appendix 12). There was a 15% increase in abundance in *Gr. truncatulinoides* (s) during the LGM and 10% increase during MIS 6 compared to the Holocene. In the PCA analysis *Gr. truncatulinoides* (s) dominates Group 3 (1-2) which is associated with *T. quinqueloba* and *N. pachyderma* (s) (see Fig. 66). Species *N. pachyderma* (s) and *T. quinqueloba* are associated with polar-subpolar water masses, prefer low temperature gradients in the water column and live preferentially below 100m (Manighetti and Northcote, 2000). Group 3 (1-2) dominates the sample scores during the LGM with a 30% increase in abundance, and dominates sporadically during MIS 6.

The significant increase in relative abundance of very deep dwelling species *G. truncatulinoides* (s) during the glacial periods compared to deep-dwelling species such as *N. pachyderma*, *Gr. crassaformis*, *Gr. hirsuta* and *Gr. scitula*, suggests a deepening of the thermocline (see Fig. 58 and 59). In comparison, a higher proportion of deep-dwelling species may indicate a shallower thermocline (e.g. Ravelo et al., 1990). This is also represented in the MAT analogue estimate of the mixed layer depth with the mixed layer reaching a maximum depth of 125m during MIS 2 and MIS 6 compared to 71m below the sea-surface during the Holocene (see Fig. 61). There is also a uniform reduction of the temperature estimates at 50 to 150m of 5°C during the LGM and MIS 6 suggesting a well mixed homogenised mixed layer.

Lohmann (1992), Martinez (1994) and Martinez (1997) suggested that *Gr. truncatulinoides* (s) could be used to indicate Mode Water. Mode Water is produced by winter convection and overturning south of the STC (McCartney 1997). Mode water is identified by the presence of a thick mixed layer of homogeneous temperature and salinity (a thermostad). Further evidence is shown by the increase in relative abundance of *T. quinqueloba* during glacial periods, as *T. quinqueloba* prefers a low temperature gradient within the mixed layer. In addition, the higher the abundance of *Gr. truncatulinoides* (s) suggests the thicker thermostad (Martinez 1997). This appears to be supported in MD2607 with the maximum thickness of the mixed layer occurring at the LGM. This relationship of the depth of the thermocline with deep and very deep dwelling species was observed in the Tasman Sea by Martinez (1994) and Martinez (1997).

Presently, in the Australian sector, SAMW is circulated by an anticyclonic gyre which brings it to the continental margin at depths of 450-850m (McCartney, 1977; Edwards and Emery, 1982) and isolated from the sea surface by warmer surface waters
(Passlow et al., 1997). However, Li et al. (1996b) indicate that a mesostrophic water conditions occurs on the Lacepede Shelf due to the interaction of subtropical-transitional oligotrophic waters sources from the Leeuwin Current and more nutrient rich waters sourced from the Southern Ocean. It appears from the relative abundance of *Gr. truncatulinoides* (s) and *T. quinqueloba* and the response of Group 3 (1-2) that stratification is reduced and conditions are more eutrophic during the LGM with a pronounced movement and thickening of Mode Water over the core site.

The increased influence of Mode Water during LGM coincides with a decrease in Tmean of approximately 6°C ± 0.84°C compared to present (see Fig. 61). After global sea-level affects the residual in δ¹⁸O for the LGM to present is 1.00‰. This accounts for a SST difference of approximately 4°C using the oxygen isotope fractionation relationship between CaCO₃ and water of 0.25‰/°C (Kim and O’Neil, 1997). The discrepancy of the δ¹⁸O and SST records may also indicate that water from difference sources is influencing the core site during the LGM (Waelbroeck et al. 2005).

9.3.2.1 Upwelling

*Ga. bulloides* characterises upwelling situations regardless of their geographical position (Hemleben et al., 1989). This species dominates the fossil assemblage of MD2607 particularly during glacial periods and a peak in abundance during MIS 6 (~45%). Other species that respond to nutrients within the water column such as *Gn. glutinata* and *N. dutertrei* also increase in relative abundance during glacial periods especially MIS 6 (see Fig. 58 and Appendix 12). This suggests that the processes of upwelling did occur during glacial periods and was stronger during MIS 6.

The PCA analysis and the relative abundance counts suggest that stratification from the Leeuwin Current was reduced especially during MIS 6 and this resulted in upwelling of sub-polar nutrient rich waters to the sea-surface (see Fig. 66). Subtropical species *Gs. ruber* is removed from the record during MIS 6 and only present in minor amounts during the LGM (see Fig. 58 and Appendix 12). Transitional species *Gr. inflata* is also significantly reduced during the glacial periods with a greater reduction during MIS 6. The reduction of a warm surface layer is also indicated by sub-polar species *T. quinqueloba* during the LGM. Its dramatic increase in abundance may indicate a change in the vertical temperature gradient of the water as *T. quinqueloba* prefers water with very little seasonal change in salinity and low vertical temperature gradients (Manighetti and Northcote, 2000). There is a general an inverse relationship
between the relative abundance of *Ga. bulloides* and *Gr. inflata* in core MD2607 (see Fig. 58 and Appendix 12). Olson and Smart (2004) observed this pattern in the NW Atlantic Ocean, with increased abundances of *Gr. inflata* related to periods with lowered nutrient availability and warmer surface waters.

Presently, upwelling of the colder fertile bottom water occurs in various parts of the southern coastal margin (Bunt, 1987; James et al., 1997). These biofacies can be attributed to the influence of cold water from the Southern Ocean (Li et al., 1999), driven by the West Wind Drift. However, the planktonic foraminiferal assemblages of the Bonney Upwelling system have not been analysed and could be contributing to the upwelling signal.

Passlow et al. (1997) study of planktonic and benthic foraminifera, further to the east of MD2607 on the Victorian coastline (38°51'S, 141°03'E) also had a marked increase in *Ga. bulloides* during LGM and MIS 6. Benthic assemblages were also characteristic of increased productivity levels during glacial periods. Almond et al. (1993) examined benthic foraminifera from the slope sediments in the western Great Australian Bight area. Higher infaunal numbers indicated the delivery of more organic carbon to the sediment and provides evidence of intensified productivity, implying intensified upwelling during glaciations. Studies by James et al. (2000), Holbourn et al. (2002) and Feary et al. (2004) on the presence of bryozoan mounds during glacial periods suggest, the lowered sea level, the weakened Leeuwin Current, and increased upwelling, provided enhanced carbon flux and nutrients, contributing to prolific bryozoan growth and mound development on the southern coastline. The findings of core MD2607 agrees with Almond et al. (1993), Passlow et al. (1997), James et al. (2000), Holbourn et al. (2002) and Feary et al. (2004) that there was a marked strengthening of cool upwelling water over the continental slope during glacial periods, especially during MIS 6. The areas of upwelling appear more widespread that wind driven Bonney Upwelling seen presently to the east of core MD2607. Upwelling is more likely due to West Wind Drift along the whole southern coastline driven closer by frontal migration as suggested by these previous studies.

Feary et al. (2000) speculated that there may be a causal relationship between the unusual occurrences of bryozoan-dominated biogenic mounds formed at low sea-level stands and the presence of high salinity brines and high H$_2$S and CH$_4$ dissolved gas. Feary et al. (2004), postulated a decreased hydrostatic head would have accompanied falling sea levels, and seepage of the gases (from gas hydrates) may have encouraged the enhanced biogenic activity that produced the carbonate mounds.
However, sedimentological analysis by Gingele et al. (2005) revealed there was no indication of anoxic conditions in core MD2607 and neighbouring core MD2611.

Gingele et al. (2005) used proxies such as concentrations of aragonite, low and high magnesium calcite, total carbonate, total organic carbon, sulphur and δ¹³C of planktonic foraminifera *Globigerina bulloides* on core MD2607 and revealed cyclic increases of productivity during glacial periods. Additionally, the covariance of organic carbon, sulphur and δ¹³C of *Ga. bulloides* suggested that the strong variations in the organic carbon record have a marine origin (Gingele et al., 2005). It was proposed by Gingele et al. (2005) that as MD2607 site is under the “Eastern Australian Dust Plume” dust could have fertilized surface waters as there was no indication of the palaeo-Murray River being a major source of nutrients. Planktonic foraminiferal analyses in this study suggest that the likely cause of increased nutrients in the core location is due to upwelling of AAIW and/or Circumpolar Deep Water which is presently situated between 850-1100 mbsl (Passlow et al., 1997) and 1200 to 4000 mbsl (Emery and Meincke, 1986), respectively. As core MD2607 is located above the Murray Canyon System, it is possible the canyon system provided a conduit for nutrient rich bottom water to shelf edge during glacial periods.

As pointed out by Gingele et al. (2005) another factor that could also be affecting the δ¹³C signal is the world wide phenomena attributed to a change in the preformed isotopic composition of southern source waters (Oppo and Fairbanks, 1989; Schneider et al., 1992) indicated by the minima at 13 000 yrs BP and during MIS 6.

### 9.3.3 Migration of the Subtropical Convergence

In core MD2607, the sinistral form of *N. pachyderma* remains a minor species with a maximum relative abundance of only 2.5% during the penultimate glaciation. Studies such as Hemleben et al. (1989), Howard and Prell (1992), Martinez (1994) and Passlow et al. (1997) have suggested the position of the STC can be traced using shifts in the ratio of sinistral *N. pachyderma*. Martinez (1994) and Passlow et al. (1997) palaeoceanographic studies of the Tasman Sea and the southern waters of Australia used the 40% abundance of *N. pachyderma* (s) as an indicator of the location of the front. Based on 40% *N. pachyderma*, Martinez (1994) estimated the STC has moved to 45°S during MIS 6, 4 and 2 and to 43°S during MIS 6, 10 and 12 (Martinez 1994). It appears from Martinez et al. (1994 a,b) that the greater upwelling signal during MIS 6 in core
MD2607 may be due to a greater movement of the STC northward during MIS 6 bringing cold, more nutrient rich water closer to the core location.

It appears that the front did move north of its current position during the glacial periods but not as much as the proposed 6 degrees to above the core location of MD2607, as the faunal abundances would have a much greater response over the core location. The evidence from core MD2607 supports Passlow et al. (1997) planktonic foraminiferal study that the location of the STC was registered by core E27-30 south of Tasmania (45°04′S, 147°13′E) but the STC was not registered by core E55-6 along the coast of Victoria (38°51′S, 141°03′E). The only other study on the south-eastern continental margin region was by Well and Okada (1997) (38°34′S and 140°37′E). Their study was based on planktonic and benthic foraminifera and nannofossils abundances and they suggested that the STC was over the core site during the LGM due to increases in productivity. However, in Well and Okada, (1997) study *N. pachyderma* (s) was only ≤ 8% during glacial periods and the LGM was not recovered due to a hiatus in the core.

### 9.3.4 Glacial Leeuwin Current

It appears that the Leeuwin Current was still occasionally present on the southern coastline during the LGM but the characteristics of the current were altered and its impact on the dynamics of the region was significantly less than at present. The existence of an occasional warmer flow, comparative to the Mode Water, over the core site is due to the presence of *Gr. inflata* with a relative abundance of 10% during the LGM; still a prominent species in the fossil record. *Gr. inflata* is a transitional species and according to Li et al. (1996a), Li et al. (1996b), Li and McGowran (1998) is a present-day indicator of the Leeuwin Current on the Lacepede Shelf. It is also shown in MD2607 that increased abundance of *Gr. inflata* indicates warmer water during the Late Holocene. In addition, subtropical species *Gs. ruber* was still present in the record during the LGM. However, *Gs. ruber* was removed from the record and the abundance of *Gr. inflata* decreased further during MIS 6 (see Fig. 58 and Appendix 12). A further decrease in the abundance of these species during the penultimate glacial period (MIS 6) suggests a source of warmer water was present during the LGM. In addition, the minor tropical-subtropical species seem to indicate sporadic pulses of warm water between the interglacial periods (see Fig. 58 and Appendix 12).

As stated in earlier the presence of *Gs. ruber* could be due to shelfal water maintaining its population in cold periods. However, warm saline conditions would be
unsuitable for *Gr. inflata* as this species requires a deep thermocline and has preferences for water masses with little seasonal variation in salinity (Hilbretch, 1996). The presence of these species suggesting the Leeuwin Current was still operating during the LGM.

The abundance of subtropical to transitional species *Ga. falconensis* is maintained at 15-20 % during the LGM but is almost absent during MIS 6. *Ga. falconensis* also responds to eutrophic conditions and has been observed in nutrient plumes along the southern coastline presently (Li et al., 1996a). The increased dominance of Mode Water and upwelling conditions suggests eutrophic conditions during the LGM and MIS 6. The relatively high abundance of *Ga. falconensis* during the LGM suggests warm water was still being carried into the core area, while, its reduction in MIS 6 suggests colder water was present.

The most extensive planktonic foraminifera study, previous to this study, was conducted by Almond et al. (1993) and they inferred that the Leeuwin Current was absent during the LGM and MIS 6. Almond et al.’s. (1993) study was conducted on a short core (439cm, 102GC08) on the eastern slope of the Great Australian Bight. Accurate comparisons of marine isotope stages are difficult as there is no chronological record for core 102GC08, but very similar abundances patterns for *Gs. ruber* and *Gs. inflata* occur between MD2607 and 102GC08. At the phase interpreted to be the LGM and MIS 6, *Gr. inflata* still represents 20% of the fossil assemblage and *Gs. ruber* represent <5%. These abundances are higher than MD2607 during the LGM. However, together the two cores suggest a west to east gradient of warmer to cooler species during the LGM. A progressive longitudinal decrease of warm-water species and their abundance from west to east is observed today due to the processes of the Leeuwin Current (Li et al. 1996a; Li et al. 1996b; Li and McGowran, 1998 and Li et al., 1999).

Holbourn et al. (2002) presents benthic foraminiferal evidence of comparatively warmer water during the LGM compared to MIS 6 along the shelf break of the Great Australian Bight. The $\delta^{13}$C and faunal abundances of benthic foraminifera suggest higher temperatures and conditions of downwelling. This is supported further by the slower growth of bryozoan mounds at the shelf edge, whereas, sites further offshore showed extensive mound development during MIS 2 and 4 (James et al., 2000; Holbourn et al., 2002).

Almond et al. (1993) suggested that the greater diversity in the interglacial periods is due to stratification of the water column so, therefore, more niches in the water column can be exploited by the planktonic foraminifera. The diversity of species is reduced to 19 species during the penultimate glacial periods (MIS 6) compared to the
22 species present at the LGM and 25 species present during the Holocene. The findings of stratification of the water column during the LGM supports the idea of comparatively warmer waters near the shelf break which also links to the findings of Holbourn et al. (2002). In addition, the findings of MD2607 suggest less comparatively ‘warm’ water is reaching the core site during MIS 6, further reducing stratification and resulting in the dominant upwelling processes during this period.

Other studies along the southern shelf by Li et al. (1996a, 1996b), McGowran and Li (1998) and Li et al. (1999) could not resolve whether the Leeuwin Current shut down during glacial times due to reworking on the shelf sediment and possibly the resolution of the dredge studies.

Ridgeway and Condie (2004) recently established that the main driving force of the Leeuwin Current along the southern coastline was the westerly wind regime which switches on at May to November. This was also indicated by model studies of Middleton and Cirano (1999), and Cirano and Middleton (2004) where the westerly wind stress at the coast drives an onshore Ekman mass transport which leads to coastal downwelling and eastward shelf edge flow (Ridgeway and Condie, 2004).

One factor that might assist in carrying water from the western coastline is the apparent increase in the mid-latitude westerlies during the LGM (Colhoun, 1991; Markgraf et al., 1992; Harrison, 1993). However, Hesse and McTainsh, (1999), Wyrrwoll et al. (2000), and Hesse et al. (2004) suggested increased glacial atmospheric dust loadings and dust deposition are due more to source area aridity, greater dust production than to stronger atmospheric circulation over the mid-latitude source areas. Although Hesse and McTainsh (1999) suggest there is evidence of a persistent westerly flow during the LGM. The persistence of westerly conditions during the LGM is also supported by Gingele and De Deckker (2005). A striking feature of the SST record is a peak in SST at ~ 14 000 yrs BP and could signify the strengthening of the monsoonal and westerly wind systems.

McGowran and Li (1998) point out that the flow of shelf water is consistently eastward, like the West Wind Drift, due to Coriolis forcing and the prevalent wind from the south-west. It is shown from the SST maps that comparatively warmer water is still being transported by a weak Leeuwin Current along the western coastline during the LGM. Therefore, it is not unlikely that transitional water may still occasionally flow along the southern coastline during the LGM as the driving mechanisms along the southern coastline are still in place during the LGM.
It has been estimated that the STF moved northward by approximately 2° during the LGM (Prell et al., 1979). If the STF maintained its modern day form as described by Belkin and Gordon (1996; see Fig. 7), then it appears the front may not have blocked the flow of the Leeuwin Current along the southern coastline during the LGM. Passlow et al., 1997 did not find any evidence that the STC moved to 38°51'S during the LGM. Recent evidence from Armand (1997), Sturm (2003) and Gersonde et al. (2005) suggests the movement of the STC northward was minimal during the LGM with most changes occurring at the Polar Front. However, an additional 2° shift northward of the STF during MIS 6 (Martinez et al., 1994b) appears to be sufficient to block the flow of the Leeuwin Current.

Even if the Leeuwin Current was still occasionally present, the amount of 'warmer' water it would carry would be fairly insignificant in terms of affecting the climate compared to the impact of the subpolar water that dominates the fossil assemblages during glacial periods. Aridity was widespread in the region during glacial periods (Wilford, 1984; Gardner et al., 1987; Wasson and Clark, 1988; Wasson, 1989; Nanson et al., 1988, 1990, 1992, 1995; Harrison, 1993; Rea, 1994; Croke et al., 1996 Hesse, 1994; Allan and Lindesay, 1998; Hesse and Mettanish 1999; Hope et al., 2004; Gingele and De Deckker 2005) and the significant reduction in the flow of the Leeuwin Current would have more of an impact on aridity than the occasional presence of transitional water along the shelf break. It is shown by the regional terrestrial studies that a reduction in precipitation was more significant than the reduction in temperature in developing widespread aridity in the LGM. The occasional presence of a minor flow of transitional water would not be sufficient to induce precipitation in the area.

Presently, the water of the region is a mix of transitional water with subtropical elements brought to the region by the Leeuwin Current. It has been shown in the northern core (MD61) and the regional SST maps, that the characteristics of the Leeuwin Current appears to have changed on the west coast with a greater element of subtropical-transitional water in the upper layer (possibly with STW characteristics) during the LGM. The glacial Leeuwin Current was significantly weakened but occasionally transitional water may be carried along the southern coast causing a relative difference with the main mass of sub-polar water off the coastline. Past studies argue that the Leeuwin Current was absent because tropical-subtropical species are removed (Almond et al., 1993’ Wells and Okada, 1997). A much stronger Leeuwin Current would be required to maintain subtropical conditions in a glacial ocean and as Wells and Okada (1997) point out that the general cooling during the glacial periods
may be sufficient to prevent the survival of tropical-subtropical species. Perhaps, we need to look at transitional species to understand if the Leeuwin Current still operates on the southern shelf as the main water type of offshore water is subpolar during glacial periods.
Chapter 10

Conclusion

The findings of this study combine the information of faunal abundances, an estimate of SST and an estimate of the mixed layer depth to provide a more complete picture of the upper water column, offshore West and South Australia and Sumatra for the middle to late Quaternary. The findings of this study have captured two main patterns of flow displayed by the Leeuwin Current that relate to the glacial and interglacial cycles.

The results provide evidence that the Leeuwin Current was still active in the core area of MD61 during glacial periods. However, the characteristics of the Leeuwin Current were different compared to the present. The faunal abundances and SST indicate a reduction in temperature and an increase in salinity at the MD61 core site, suggesting less ITW was being delivered to the Leeuwin Current from the Warm Pool area, which had also decreased in size. The SST maps also showed the Leeuwin Current was weaker during glacial periods and there was less of a temperature difference between the Leeuwin Current and the offshore water along the western coastline. However, the Leeuwin Current was still strong enough to prevent upwelling along the western coastline during the glacial periods. It is assumed that the characteristics of the water above the MD61 core site are similar to what is found presently at approximately 25-32°S due to the relative thickness of the SICW and ITW and relative abundances of the planktonic foraminifera compared to the core top study (modern abundances) of Martinez et al. (1998). In general, the faunal assemblages changed from tropical-subtropical to transitional assemblages at MD61 site during glacial periods.

The weakening of the Leeuwin Current along the western coastline resulted in a further reduction of flow as the Current moved southward, incorporating the characteristics of the offshore cooler saltier water as it does presently. It is believed the Leeuwin Current reached the core location of MD2607, at the upper continental slope of the Lacepede Shelf, during the LGM but was absent during MIS 6 due to the northward movement of the STC blocking its flow. Evidence for the presence of the Leeuwin Current on the southern coastline during the LGM was the gradational change in comparatively warmer foraminifer species towards the east of the Great Australian Bight compared to the abundances seen by Almond et al. (1993) on the western side of
the Great Australian Bight. In addition, there was greater stratification at the sea-surface during the LGM compared to MIS 6, thus allowing the persistence (although at low abundance) of subtropical-transitional species compared to the dominant sub-polar assemblages during glacial periods.

However, the influence of the Leeuwin Current on the dynamics of the region was greatly reduced during the LGM compared to present. The reduction of the Leeuwin Current on the southern coastline resulted in upwelling which was further enhanced during MIS 6 when the Leeuwin Current was absent. Increased productivity appears to be widespread along the southern coastline during glacial period in studies away from the boundary of the STC. The reduction of the Leeuwin Current may also have aided in the development of arid glacial climates in the region with less warm water being carried along the southern coastline disrupting the evaporation/precipitation cycle of the region.

Most of the variation at the northern and southern core sites occurred from present sub-surface water masses moving into the lower-upper mixed layers during glacial periods. There was a greater influence of SICW at the site of MD61 while at the southern core MD2607 there was an increase in SAMW. Both of these water masses are produced at or north of the STF and the dominance of these waters masses suggest greater production of these water masses and the movement of the STF northward during glacial periods. The greater influence of SICW in the eastern Indian Ocean was also indicated by the northward migration of the ITW/SICW frontal system by 3-4° to over the MD61 core site which resulted in a 6-9°C decrease in SST, a thickening of the mixed layer and the dominance of transitional species during glacial periods. The dominance of SICW also suggests that the Western Australian Current was relatively stronger during the glacial periods which aided in reducing the strength of the Leeuwin Current.

Wind direction and alteration to the along shore pressure gradient also played a part in altering the current dynamics of the study region. Core BAR9403 offshore Sumatra contributed to a better understanding of the Australasian Monsoon which acts as an external forcing mechanism of the Leeuwin Current. Compared to the cores of the Leeuwin Current, the SST in BAR9403 did not change during the LGM due to stratification at the sea surface. The reduction of vertical mixing and or upwelling at BAR9403 also suggests that the general monsoonal system was weaker during the LGM. This is observed during El Niño events when the Throughflow and the Leeuwin
Current are weakened and the West Australian Current is strengthened due to the reduction of the alongshore pressure gradient.

Previous studies in the region indicate that the monsoon system was weakened due to the ITCZ being situated at the equator. This results in the NW Monsoon not being able to move into northern Australia during the austral summer and essentially the conditions of the SE Monsoon persist throughout the year. The findings from this study confirm that this condition may have persisted during glacial periods. Evidence for a persistent SE Monsoon is shown by a weak but not completely removed Leeuwin Current, the movement of ITW into the SEC, the strengthening of the West Australian Current, and the occurrence of warmer water on the southern coast of Australia during the LGM which is driven by westerly winds which were dominant in southern Australia when the SE Monsoon operated further north.

Past studies have argued that the Leeuwin Current was removed because the tropical-subtropical component of the current was reduced and due to an increase in salinity at the sea-surface. This study reveals that the temperature and low salinity characteristics of the Leeuwin Current changed but it was still operating during glacial periods as the forcing mechanism of the current were still in place, permitting its flow to higher latitudes.

The findings from this study add to the regional understanding of the dynamics of the Leeuwin Current during interglacial periods, with the Leeuwin Current appearing to be stronger during MIS 5, MIS 7, MIS 11 and occasionally during MIS 9, compared to the Holocene. The Leeuwin Current appears to be stronger due to increases in SST and the response of subsurface species to warmer temperatures suggesting a thicker Leeuwin Current. The warming of the subsurface layers was aided by the southward movement of the ITW/SICW front and STC which reduced the mixed layer at both the northern and southern core sites. Stratification due to the Leeuwin Current is a feature in the southern core during interglacial periods.

There is an upwelling signal during MIS 11-10 and during the Holocene in MD61 along the western coastline. The upwelling conditions of these periods could be due to the southward movement of the ITW/SICW, reducing the thickness of the mixed layer and allowing nutrient-rich water from the base of the thermocline to be upwelled, as seen presently along the Gascoyne Coast. Upwelling could also be due to the re-intensified Australasian Monsoon system driving the Ningaloo and Capes Currents, as these Currents presently operate during the NW Monsoon.
There is a synchronicity with warmer interglacial periods such as MIS 5, between the northern and southern cores of the Leeuwin Current. A stronger and thicker flow of the Leeuwin Current has significant climatic implications. This is particularly significant for the analogous future climate of MIS 11 which had a SST approximately 2°C warmer during the summer and winter compared to present. It is anticipated that in the future the Leeuwin Current will provide a conduit from the warmest oceanographic feature on Earth, the Warm Pool, to the Southern Ocean aiding in the reduction of the Antarctic polar ice caps.
11.1 Future Work

This study suggests that the Leeuwin Current reached the southern core site during the LGM. This finding questions the current literature which states that the Leeuwin Current was not present on the southern coastline during glacial periods. Holbourn et al. (2002) presents data that also indicates warmer water on the slope of the western Great Australian Bight during the LGM, but as this was a benthic study, it was difficult to determine the source of this warm water. It is apparent that more planktonic foraminifera studies need to be conducted on the southern coastline to get an understanding of the processes at the sea-surface, where the Leeuwin Current is situated. Assemblage counts from pelagic organisms, from cores along the coast would determine if a longitudinal temperature gradient existed during the LGM and thereby, indicate the presence/absence of the Leeuwin Current. As state previously, it must be understood that the characteristics of the Leeuwin Current would have changed in terms of salinity and temperature as with other water bodies during glacial periods. Focusing on the maintenance of temperate fauna may also suggest the Leeuwin Current existed along the shelf break, when the offshore water masses had more subpolar characteristics.
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APPENDICIES
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