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The palaeoceanography of the Leeuwin Current: implications for a future world

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Abstract

Long-term progressive changes of the Leeuwin Current are linked to plate and ocean basin 'geography' and Cenozoic global climates and palaeoceanography. Suggestions of the presence of a proto-Leeuwin Current as early as late Middle to Late Eocene times (*c.* 35–42 Ma) cannot be verified by the fossil record of the western margin of Australia. "Leeuwin Current style" circulation around Australia was certainly established by the early Oligocene, in response to palaeogeographic changes in the Tasman Strait. This, followed by tectonic reorganisation of the Indonesian Archipelago throughout the Miocene, provided a palaeogeographic setting, which by the Pliocene was essentially that of today. The subsequent history of the Leeuwin Current comprises climatically-induced changes operating over orbital and sub-orbital temporal scales. Specifically, the advent of Pleistocene-style climates, especially over the last 800 000 years, and their associated interglacial – glacial states provide the two end-member climate-ocean states that have characterised Leeuwin Current activity during that time. Indications of the nature of these contrasting states is provided by: (i) the Last Interglacial (*c.* 125 Ka) during which sea level was higher by some +4 m, and with higher sea surface temperatures (SSTs) clearly indicating a more 'active' Leeuwin Current; and (ii) the Last Glacial Maximum (21 Ka), during which sea level was some 130 m lower than today, resulting in massive shelf extensions along the coast of Western Australia, accompanied by reduced Indonesian Throughflow, lower low latitude SSTs and changes in the Western Pacific Warm Water Pool, and with these changes, possibly reduced Leeuwin Current activity. Sub-orbital scale fluctuations in current strength are driven by global climate change associated with El Niño – La Niña events as well as regional climatic changes driven by volcanism. These forcing mechanisms operate at time scales well within the reach of human experience, and provide important comparative data for predicting the response of the Leeuwin Current to climate change predicted for this century. Studies of the impact of changes in the vigour of the Leeuwin Current on shallow marine communities are in their infancy. Coupling climate models with geological analogues provide important research agenda for predicting the trajectory of future changes to the Leeuwin Current and their impacts on the marine biota of coastal Western Australia.

Keywords: palaeoceanography, Leeuwin Current, Last Interglacial, Last Glacial Maximum

Introduction

The geological history of the Leeuwin Current involves a combination of re-arrangements of global geography in terms of both the continent positions and associated plate tectonic changes, and a global climate regime which brought the Earth out of an essentially 'ice free' world into the present Ice Age. Presently, the Leeuwin Current is a relatively strong southward flow of warm, low salinity tropical water along the Western Australian coast – from near North West Cape to Cape Leeuwin and on to the Great Australian Bight (Cresswell 1991) – driven by a meridional steric high gradient. Given the Leeuwin Current's general controls, of a southward pressure gradient force balanced by northward wind stresses, we see the Leeuwin Current being driven by hemisphere-scale climate constraints, as

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well as itself having a possible impact on the climate regime of the western margin of Australia. With the predicted progression of the Earth into a warmer future climate setting, the Leeuwin Current will be placed in a world which is being steered away from the distinctive glacial-interglacial climate signature of the last *c.* 800 000 years (Loutre & Berger 2000; Archer & Ganopolski 2005). Given the importance of the Leeuwin Current in controlling marine biogeography in the region, understanding its likely response to these changes remains to be established. In this contribution we use the geological record to elucidate the range of states that the Leeuwin Current has been able to adopt over long-time scales – information that may prove helpful in constraining model predictions of future events.

The history of the Leeuwin Current over plate tectonic time scales

The long-term geological history of the Leeuwin Current is linked to the relative position and organisation of the Australian plate (*e.g.*, Li & Powell 2001) juxtaposed against changes in global ocean and atmosphere climates that characterise specific time periods within the last *c.* 30 myrs (see Zachos *et al.* 2001). In their overall function, two 'ocean-gateways' – the Indonesian Throughflow and the Tasmanian Gateway – are clearly of importance and their roles figure prominently in discussions of the early history of the Leeuwin Current.

Establishment of the Tasmanian Gateway

Shafik (1990) and McGowran *et al.* (1997) suggest an early presence of a "proto-Leeuwin Current" stemming from late Middle to Late Eocene time (35–42 Ma). They specifically relate this to a wider plate reorganisation and a new phase of seafloor spreading marked by Chron C19 (41–42 Ma). Along the Western Australian coast the Middle Eocene Merlinleigh Sandstone of the central Carnarvon Basin preserves a colonial coralline assemblage, which in combination with the associated mollusc assemblage, suggests warm ocean conditions off this part of the coast during that time (McNamara & Scott 1983). The distinctive Merlinleigh assemblage may represent an outlier of the Eocene Austral Indo-Pacific Province. However, it is not at all clear that its presence can be unequivocally attributed to the presence of a proto-Leeuwin Current, as it could simply be an expression of regional SSTs. The same doubt can be expressed with respect to the evidence presented by McGowran *et al.* (1997), whose palaeontological data support warm water along the margins of southern Australia during this time, but whether this necessarily implies the presence of a proto-Leeuwin Current remains an open question.

Recent studies of Middle-Late Eocene (Bartonian-Priabonian, 40.4–33.9 Ma) foraminifers (Haig & Mory 2003) and molluscs (Darragh & Kendrick 2008) from the southern Carnarvon Basin near Kalbarri, have found no evidence of significantly warm SSTs over that time. Affinities of the mollusc assemblage lie substantially with those of the later Eocene of temperate southern Australia. This inference is in accord with the absence from the assemblage of tropical and subtropical foraminifers (Haig & Mory 2003) and corals (Darragh & Kendrick 2008). Hence there is at present no compelling evidence for the presence of any Austral Indo-Pacific (*i.e.*, tropical) provincial fauna in the Kalbarri Eocene assemblage, nor any clear manifestation of what could be interpreted as an Eocene Leeuwin Current phenomenon. It should be noted that Haig & Mory (2003) state that the absence of tropical and subtropical species of foraminifera suggests a cooling trend over this interval, comparable with that noted for southern Australian waters by McGowran *et al.* (2000).

McGowran *et al.* (1997) make the claim of evidence for proto-Leeuwin Current activity at *c.* 33 Ma with additional evidence of activity later in the Oligocene. These claims and those of an earlier (Eocene) existence of a proto-Leeuwin Current have to be seen within the context of more recent work on the timing of the opening of the Tasmanian Gateway, which facilitated flow along the southern margins and into the Pacific. Stickley *et al.* (2004) demonstrate that it was only after 30.2 Ma that the region saw the establishment of stable, open-ocean, warm-temperate, oligotrophic settings characterised by siliceous-carbonate ooze deposition. The evidence presented by the authors indicates that deepening of this early Oligocene Tasmanian Gateway initially produced an eastward flow of relatively warm surface waters into the southwestern Pacific Ocean, for which they also adopted the term ''proto-Leeuwin'' Current. It now seems clear that the tectonic changes that produced a deep Tasmanian 'gateway' between Australia and Antarctica were essentially complete by *c.* 30 Ma (Stickley *op. cit.*), and hence somewhat later than originally envisaged by McGowran *et al.* (1997).

With the opening of the Tasmanian Gateway, the oceanic circuit around the Australian continent was complete. At that time the Australian continent lay well south of its present position, with the region north of Australia lacking the flow constrictions presently imposed by the Indonesian Arc and associated terrains. With the opening of the Southern Ocean, Australia essentially represented an island, setting up a circulation forced by Godfrey's (1989) "Island Rule", which describes a circulation pattern forced by wind stress (Sverdrup circulation) setting up variations in steric height. With an open ocean north of Australia and the existence of the Tasmanian Gateway, tropical water, in part sourced from the Pacific, could now flow along the western margins of the continent, continue into the Bight, and return back into the Pacific. Hence, we see the first palaeogeographic setting for a Leeuwin Current type 'circulation' that bears resemblance to what is seen today, but one which is neither related to the temperature and salinity differences that are of fundamental importance to the dynamics of the present Leeuwin Currrent (Godfrey & Weaver 1997), nor does justice to the complexity of the flow-paths, that are now recognised in contributing to current dynamics (Domingues *et al*. 2007).

The Western Australian fossil record gives some suggestion that a SST pattern somewhat resembling that of today was in place by the Paleogene-Neogene transition. Inner shelf limestones of the Middle Miocene Poivre Formation of Barrow Island (McNamara & Kendrick 1994) feature diverse echinoid and molluscan assemblages, each with a strong tropical component. No comparable faunas are known from either the Perth or Bremer Basins but the Colville Sandstone of the marginal Eucla Basin (equivalent to the Nullarbor Limestone (Lowry 1968, 1970)) provides comparative evidence from the south coast for that time.

The mollusc assemblage from the Poivre Formation (71 spp.) is strongly tropical in character with notable representations of tellinoid and venerid bivalves and cerithioidean, strombid, cypraeid and conid gastropods. Elements in common with the Eucla Basin Middle Miocene (Colville Sandstone, Nullarbor Limestone) are few and apparently limited to the Tethyan relictual genus *Campanile*, the Australian endemic *Zoila*, and the species *Terebellum terebellum* (Linneaeus). The last mentioned species, which is extant, occurs in the Middle Miocene Nullabor Limestone, and the Miocene of Kerala, Java and Saipan (McNamara & Kendrick 1994).

This comparison of Middle Miocene echinoids and molluscs from Barrow Island and the Eucla Basin recognises a weak tropical component in the southern Australian assemblages, consistent with a remote, but only diffuse and sporadic proto-Leeuwin Current, capable of exerting no more than a slight influence on the faunal composition of inshore waters at that time. McGowran *et al.* (1997) however, argue for strong Leeuwin Current activity coincident with the Early–Middle Miocene boundary (*c.* 18–16 Ma). They place this event into a global sea level high context, in which a rising trend from the Late Oligocene to the latest Early Miocene is replaced by lowering sea levels in the Late Miocene. Coincident with the 'high-stand' of these events are globally warm SSTs, and the authors envisage a strong Leeuwin Current embedded within the global events.

The Indonesian Throughflow

Given the contribution that the Indonesian Throughflow makes to the source region of the Leeuwin Current, understanding the terrain restrictions and bathymetric controls north of Australia that presently direct the throughflow from the Pacific into the Indian Ocean, is of fundamental importance. An overview of the geological evidence points to the Indonesian Gateway being a feature with a long history (*e.g.*, Nishimura & Suparka 1997; Nishimura 2002; Gaina & Müller 2007). However, it now seems likely that it began to close by the Early Miocene, as indicated by deep water exchange stable isotope signatures between the two oceans (summary in Kuhnt *et al.* 2004). Subsequent restriction events on surface and intermediate water exchange will have occurred during the ensuing intervals of the Miocene to Pleistocene, but the evidence for such events is ambiguous and poorly constrained (Kuhnt *et al.* 2004).

Cane & Molnar (2001) claim that the Indonesian Gateway took on an expression resembling its present form during the last 3–4 myrs. In their view, the development of the geological-relief-bathymetric constraints offered by the tectonic re-organisation did not impact only on the details of the geography of the throughflow, possibly leading to a general cooling of the Indian Ocean (*cf.* Gordon *et al.* 2003, who advocate cooling due to monsoonal controls), but also may have led to changes in throughflow sources. Cane & Molnar (2001) envisage that at present much of the water constituting the Indonesian Throughflow is sourced from the North Pacific. Warm southern Pacific sourced water presently flows westward along the Equator in the Southern Equatorial Current. On reaching the island of Halmahera, it becomes part of the North Equatorial Counter Current and flows eastward. Cane & Molnar (*op. cit.*) propose that with New Guinea further south and with Halmahera being a smaller island, the warmer water from the South Pacific would have passed into the Indian Ocean rather than be involved in its present eastward transport path. However, recent particle track modelling reveals that the equatorial Pacific is also a source of Leeuwin Current water (Domingues *et al.* 2007). Nevertheless, the presently available views suggest that it has only been for the last few million years (essentially the Pliocene, 5–1.8 Ma) that a terrain arrangement north of Australia existed which imposed an exchange pattern between the Pacific and Indian Ocean resembling that of today. For a 'full' Leeuwin Current to have existed at that time, in addition to a restricted gateway, the temperature and salinity structure requirements over the Western Pacific Warm Pool and adjacent Indian Ocean would also have had to be met to provide the Indian Ocean and Pacific Ocean sources required (Godfrey & Weaver 1991; Domingues *et al*. 2007). In addition, a trade wind regime in the Pacific Ocean was necessary to contribute to the set-up of sea level in the western equatorial Pacific. Sato *et al.* (2008) present evidence that point to oceanographic conditions associated with the Indonesian Throughflow, similar to those of today, were established around 3–4 Ma, and that the Western Pacific Warm Water Pool had essentially taken on its modern form by that time.

The fauna of the early Late Pliocene Roe Calcarenite (*c.* 3–4 Ma (Beu & Darragh 2001; Kendrick unpublished data; Ludbrook 1978)) point to the existence of a strong Leeuwin Current at that time. The Roe Calcarenite is exceptional in the Australian Tertiary for its high species diversity and faunal composition. Its substantial warm-water component combines taxa descended from locally established Miocene lineages with others likely to be Late Pliocene immigrants from extraneous tropical or subtropical waters, consistent with a significant Leeuwin Current presence along the southern coast of Western Australia at the time of deposition.

The appearance in the early Late Pliocene of the Eucla Basin of a substantial, immigrant, warm-water faunal surge along the southern coast of Western Australia, presumably derived from an assertive Leeuwin Current, may bear some relationship to the timing of geological events in the Indonesian region (Cane & Molnar 2001), leading to the genesis of a distinct Leeuwin Current with controlling processes and dynamics akin to those at present. However, these inferences need to be placed within a more general context that recognises that the Middle Pliocene was a time during which mean global temperatures were substantially warmer than present (*e.g.*, Haywood *et al.* 2000) and with CO_2 levels higher than pre-industrial values (Raymo *et al*. 1996). Hence climatic factors also may have played a substantial role in the amplification of the current. The climatic effect suggested for current activity during the mid-Pliocene is a harbinger of the first-order control on the Leeuwin

Current, to be firmly established during Pleistocene times and continuing today.

Leeuwin Current activity over orbital time scales – the last *c.* 2 million Years

Once the terrain arrangement north of Australia resembled that of the present, the details of global climate dynamics became the sole drivers of Leeuwin Current events. Over the time scales of the last *c.* three million years, the Earth attained a full expression of the present ice age. Of special relevance is that during the last 1.75 myrs, the SST regime of the Western Pacific Warm Pool has been relatively stable (Garidel-Thoron *et al.* 2005), providing Leeuwin Current controls that may have been not dissimilar from those operating over Late Quaternary time-scales. It has now been firmly established that the Milankovitch mechanism provides the commonly termed "tuning fork" of global climates for this time. The dominant orbital periodicities operating in the Milankovitch mechanism centre on ~ 100 kyrs (eccentricity); 40 kyrs (obliquity-tilt); and 23kyrs and 19 kyrs (precession). Evidence for the operation of these mechanisms is given by the overwhelming weight of palaeoenvironmental studies and climate theory and modelling that now exists (*e.g.*, Lea 2004; Ruddiman 2007; Saltzman 2002; IPCC 2007). The Earth system responds to these changes by adjustments in ocean-ice volume distributions and internal arrangements of the global climate system with concomitant impacts on land-atmosphere-ocean characteristics and dynamics. Milankovitch insolation variations can lead to a direct response in ocean temperature (*e.g.*, Liu *et al.* 2003), which is an important driving control of the Leeuwin Current (Godfrey & Weaver 1997).

An insight into the likely impact of these variations can be obtained from idealised experiments using coupled ocean-atmosphere general circulation models. An example of this approach is the application of the Fast Ocean Atmosphere Model (FOAM), with variation in tilt and precession changes (see Wyrwoll *et al*. 2007 for details of the experiments). FOAM is a fully coupled ocean-atmosphere model without flux adjustment. The atmosphere component of FOAM is a fully parallel version of the NCAR CCM2, in which the atmospheric physics are replaced by those of CCM3 (Jacob 1997). The atmosphere component has R15 resolution (equivalent to grid spacing of about 7.5° longitude and 4° latitude). The ocean component was developed following the Geophysical Fluid Dynamics Laboratory Modular Ocean Model (GFDL MOM) with a resolution of 1.4° latitude and 2.8° longitude, 32 levels.

The results of the experiments show that both tilt and precession exert a significant control on SSTs, the surface wind field and precipitation patterns on the wider region that provides the setting for Leeuwin Current activity (Figure 1). Clearly the resolution of FOAM prevents a realistic exploration of what these changes may imply for Leeuwin Current activity. But the aim of the experiments was to draw attention to the likely controls that orbital changes may provide. A consideration of the possibility of a more direct Milankovitch forcing of the Leeuwin Current is more relevant to interglacial stages when the accompanying global boundary conditions are less complex. Liu *et al.* (2003) provide an example of the insight model studies offer when considering the ocean response to orbital forcing during an interglacial period.

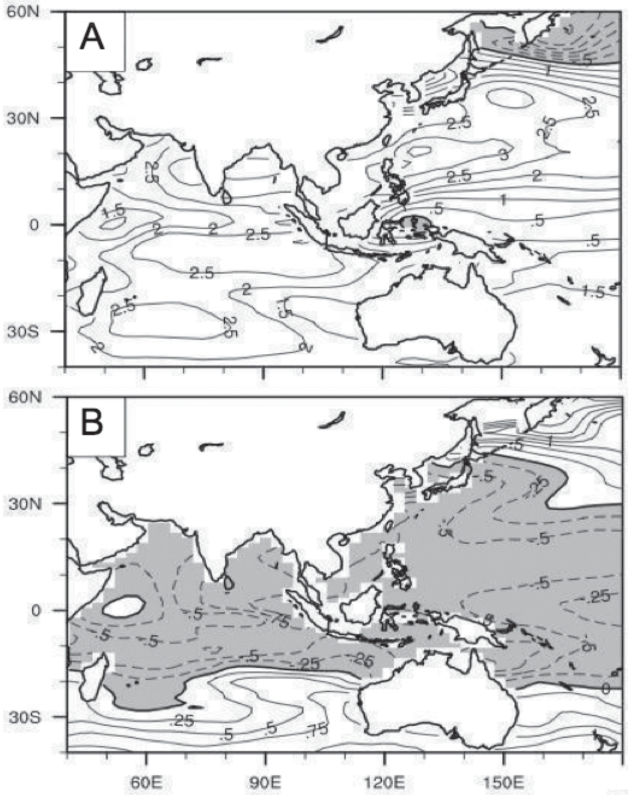

Figure 1. JFM Sea Surface Temperature (°C) of: (a) averaged Perihelion effects showing the difference between Precession in the Southern Hemisphere and Precession in the Northern Hemisphere(PS – PN); (b) averaged Tilt effects showing the differences between Tilt High and Tilt Low (TH – TL) for FOAM experiment (see Wyrwoll *et al*. 2007 for details of these experiments).

For the last 1–2 million years, the onshore marine sequences of the Perth Basin provide good indications that Leeuwin Current activity has varied significantly in that time. There, the Ascot Formation represents the Pliocene – Pleistocene transition. The predominant lithology is that of a grey, siliciclastic, mostly unlithified, richly fossiliferous arenite: subordinate carbonate is a feature of the older, more easterly member, for which diagnostic fossils include the bivalve *Cucullaea* and the pelagic janthinid gastropod *Hartungia*, both also present in the Roe Calcarenite. The Early Pleistocene younger member of the Ascot Formation records the initial appearance of the genus *Pecten* Müller in southern Australia, probably from a Mediterranean – Atlantic source via South Africa and the Antarctic Circumpolar Current (Beu & Darragh 2001; Beu 2006). Being entirely subsurface, no contact between the older and younger members of the Ascot Formation has as yet been observed, nor has superposition. In view of their consistent faunal discrepancies, any contact, if and where present, is likely to be unconformable, as envisaged by Beu & Darragh (2001, Figure 6).

Mollusc assemblages of the Ascot Formation are extensive but largely unrecorded beyond the limited listing of taxa in Kendrick *et al.* (1991) but, compared with that of the Roe Calcarenite, are clearly younger, include a distinct endemic element and are deficient, even devoid of, unambiguously warm-water taxa. This is true in particular for the younger Early Pleistocene Member. The formation as a whole signifies a regressive-cooling trend for the terminal Pliocene-Early Pleistocene, which was probably operative since the ending of Roe Calcarenite time.

In overview, the later and terminal Pliocene and entire early Pleistocene in southern Western Australia presents a picture of progressive oceanic cooling and regression, with the associated fossil record devoid of any recognisable warm-water component comparable with that so characteristic of the Roe Calcarenite. Throughout this interval no evidence can be recognised for the presence of the Leeuwin Current on regional scales in southwestern Australia.

A notable event that occurred in the coastal-marine sequences of the Perth Basin is the change from siliciclastic deposition (represented by the Ascot Formation) to the carbonate style of deposition of the Tamala Limestone (Kendrick *et al.* 1991). The change to a carbonate style of sedimentation may be interpreted as being the direct result of higher rates of carbonate productivity, but other explanations may also apply (see Kendrick *et al. op. cit.*). However, the evidence for higher sea surface temperatures during the Middle Pleistocene, as compared to those during the time of deposition of the Ascot Formation, is indicated by the molluscan faunas, which signify greater tropical and subtropical affinities than evident in the Ascot Formation. The weak representation of tropical and sub-tropical elements in the mollusc fauna of the older Plio-Pleistocene units of the Perth Basin suggests that the Leeuwin Current was then either of lesser importance than during the Middle Pleistocene, or not active. Unfortunately, at this stage we have no absolute age control of these events.

There is now evidence that there was an increase in ΔSST gradient of ~ 3.5 to ~ 5°C between the eastern and western equatorial Pacific during the interval, 1.2–0.8 myrs ago (Garidel-Thoron *et al*. 2005). The suggestion is that this would have strengthened the Walker Circulation and led to a shift from a more El Niño state to La Niña conditions. Variations in Leeuwin Current strength to El Niño-Southern Oscillation (ENSO) variations are well known, with strong and weak Leeuwin Current flow characterising La Niña and El Niño states, respectively (Pearce & Philips 1988; Li & Clarke 2004). The shift from a more dominant El Niño state to La Niña conditions during the period 1.2–0.8 myrs ago may be a factor worth considering in accounting for the Ascot Formation/Tamala Limestone contrasting evidence of Leeuwin Current activity.

Of the interglacial sea level events that are evident in the Middle and Late Pleistocene marine record of the western margin of Australia (Kendrick *et al*. 1991; Bastian 1996), it is only for the Last Interglacial centred on 125 Ka (Marine Isotope Substage (MIS) 5e) that reliable numerical dates are available. Previous to the present sea level high stand, the Last Interglacial was the last time that a comparable sea level high to that of the last *c.* 6 000 years has occurred. The Last Interglacial was a time during which global climates were generally similar to today but with global temperatures higher than at present (Kukla *et al*. 2002; North Greenland Ice Core Project Members 2004).

A stronger than present Leeuwin Current – the significance of the Last Interglacial

Western Australia presents a relatively stable continental margin in a far-field location, with a comprehensive Last Interglacial stratigraphic record readily lending itself to the reconstruction of sea level events. The recognised events attest to a Last Interglacial sea level high stand which is pervasive in the coastal geomorphology of much of Western Australia. The high stand produced a sea level of about +4 m and persisted from between 128±1 Ka and 116±1 Ka (Zhu *et al.*1993; Eisenhauer *et al.*1993; Stirling *et al*. 1995, 1998). The likelihood that during this time two sea level high events occurred has been an issue of discussion for many years (Marshall & Thom 1976; Kendrick *et al.* 1991) and is now attracting renewed interest (Andrews *et al*. 2007; Hearty *et al*. 2007). Greenstein *et al.* (2005) have recently proposed the likelihood of evidence for two Last Interglacial sea level high stands in the Exmouth Gulf, but the dating in support of this claim remains to be completed.

The Last Interglacial is of fundamental importance in demonstrating that the Leeuwin Current influences the geographic reach of extensive coral reef development along the Western Australian coast. Because of their high initial uranium content, corals are well suited for ^{234}U – ^{230}Th dating, and an impressive suite of dates is now available for many sites along the Western Australian coast (Szabo 1979; Veeh *et al*. 1979; Kendrick *et al.* 1991; Zhu *et al.* 1993; Stirling *et al.* 1995, 1998; Eisenhauer *et al.* 1996; McCulloch & Mortimer 2008). In this respect the Last Interglacial differs fundamentally from the earlier interglacials present in the Quaternary sea level record of the southern part of Western Australia that essentially lack coral reef-building events.

Hermatypic corals of the reef-building genus *Acropora* are prominent in suitable shallow water habitats of Australian tropical seas, and in Western Australia forming elaborate coral “gardens”, which today extend south to the Houtman Abrolhos (Veron & Marsh 1988). In addition, three species of *Acropora* have been found living near Jurien Bay and one has become established recently at Rottnest Island, the latter situated at the extreme latitudinal limit of *Acropora* survival (Marsh 1993). This distribution differs fundamentally from that of the Last Interglacial, during which fringing coral reef development extended well south of the Houtman Abrolhos. The prominent reef exposed at Fairbridge Bluff on Rottnest Island (Szabo 1979; Playford 1988), rich in *Acropora* (tabular and stag-horn forms) and other hermatypic corals, was among the earliest recognised evidence widely seen as attesting to a southward extension of tropical reef assemblages during the Last Interglacial.

A tropical coral fauna, similarly attributed to the Last Interglacial, occurs at Fremantle with evidence of a coral-rich limestone bar, once located between Rous Head and Arthur's Head and removed during construction of the Fremantle inner harbour in 1892–1897, and elsewhere as far south as Augusta. There, a richly fossiliferous, emergent deposit is exposed along the eastern side of the entrance channel to Hardy Inlet and continues into an adjacent quarry. Associated with *Acropora* specimens from here are the tropical pearl oyster *Pinctada fucata* and the gastropod *Strombus (Euprotomus) vomer*. Further east at Cheyne Bay a shelly, coralline aggregate, well above present sea level is cemented to a crystalline basement and forming a small promontory about 0.5 km north of the mouth of the Eyre River, at the western end of Cheyne Bay. Coralline material from this deposit includes both *Acropora* and a specimen of the tropical genus *Cyphastrea,* the latter not known living south of Cockburn Sound (L.M. Marsh *pers. com.).* At the Pallinup River mouth a specimen of *Acropora* has been obtained from a lithified shore deposit lying at about 3 m above low water mark. This is currently the most marginal record for the genus known from southern Western Australia.

Last Interglacial emergent reefs along the coast of Western Australia expose coral assemblages that are remarkably well preserved, allowing for a great number of taxa to be identified with a degree of certainty that compares closely with modern material. The Last Interglacial reefs exposed from Rottnest Island to North West Cape span some 12 degrees of latitude. Today the region comprises the Western Overlap Zone between the tropical Dampierian and temperate Flindersian provinces (Wilson & Gillett 1980). Greenstein & Pandolfi (2008) censussed Late Pleistocene reef coral communities preserved at five separate localities between Rottnest Island and North West Cape and compared their data with those published for adjacent modern reefs (Figure 2). Between-community (beta) diversity was markedly different between times owing to the extended geographic ranges of tropical-adapted taxa during the Last Interglacial compared to today (Figure 3).

While the palaeoecological evidence is clearly consistent with an enhanced Leeuwin Current during the Last Interglacial, geochemical data seem to question this claim. Corals through their trace element composition (Sr/Ca, Mg/Ca and U/Ca) provide quantitative estimates of sea surface temperatures (see Bradley 1999; Cole 2005), including seasonal differences. Using Sr/Ca ratios from the Last Interglacial at Ningaloo, McCulloch and Esat (2000) obtained summer maximum temperatures ranging from 26°C to 28°C and winter minima ranging from 23°C to 20°C with an annual mean temperature of 24°C. These SSTs estimates are slightly cooler than present and the authors recognise that their results sit uncomfortably with the palaeoecological evidence indicative of a stronger Leeuwin Current during the Last Interglacial. They advocate the view that this apparent paradox may be resolved by advocating increased summer maximum temperatures rather than winter minimum SSTs,

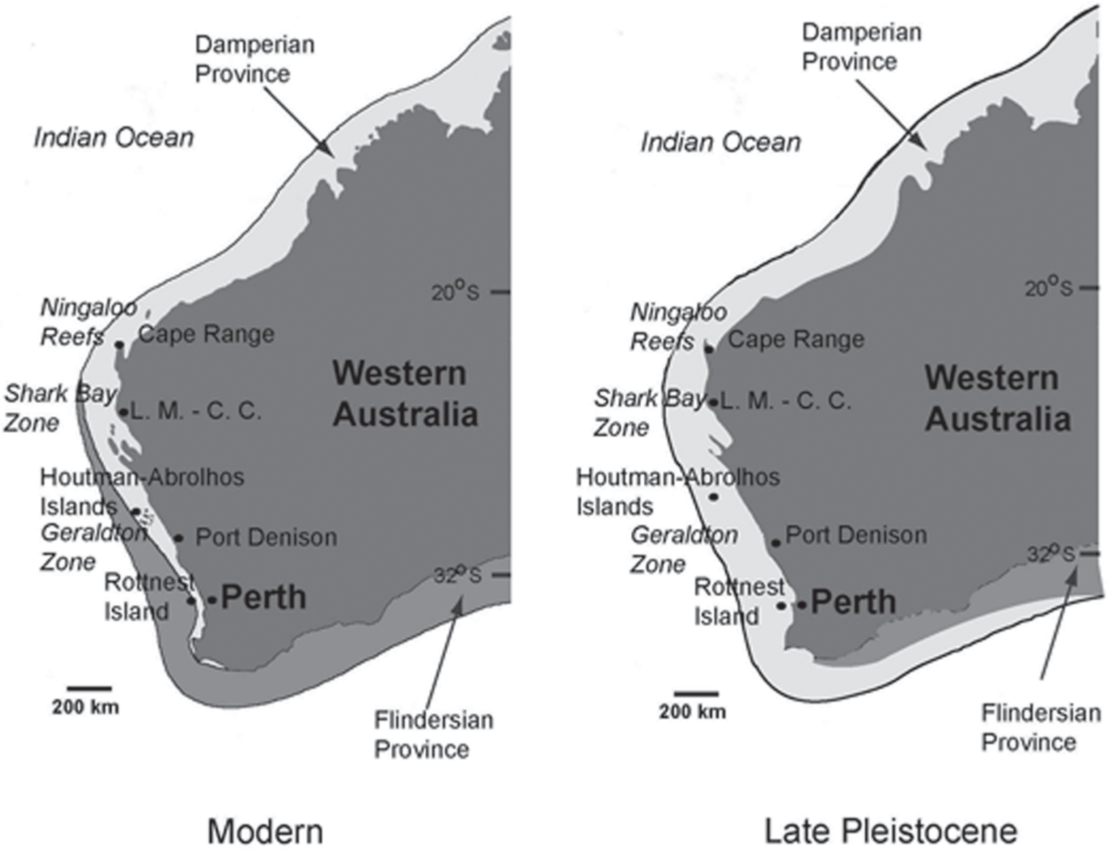

Figure 2. Modern and Pleistocene localities used for comparative study by Greenstein & Pandolfi (2008) and their disposition with respect to the Dampierian and Flindersian provinces today and during Late Pleistocene time. Disposition of modern provinces after Wilson & Gillett (1980); exact location of the "overlap zone" during Late Pleistocene time is unknown. Modern reef zones after Veron & Marsh (1988); L. M. – C. C. = Lake Macleod – Cape Cuvier.

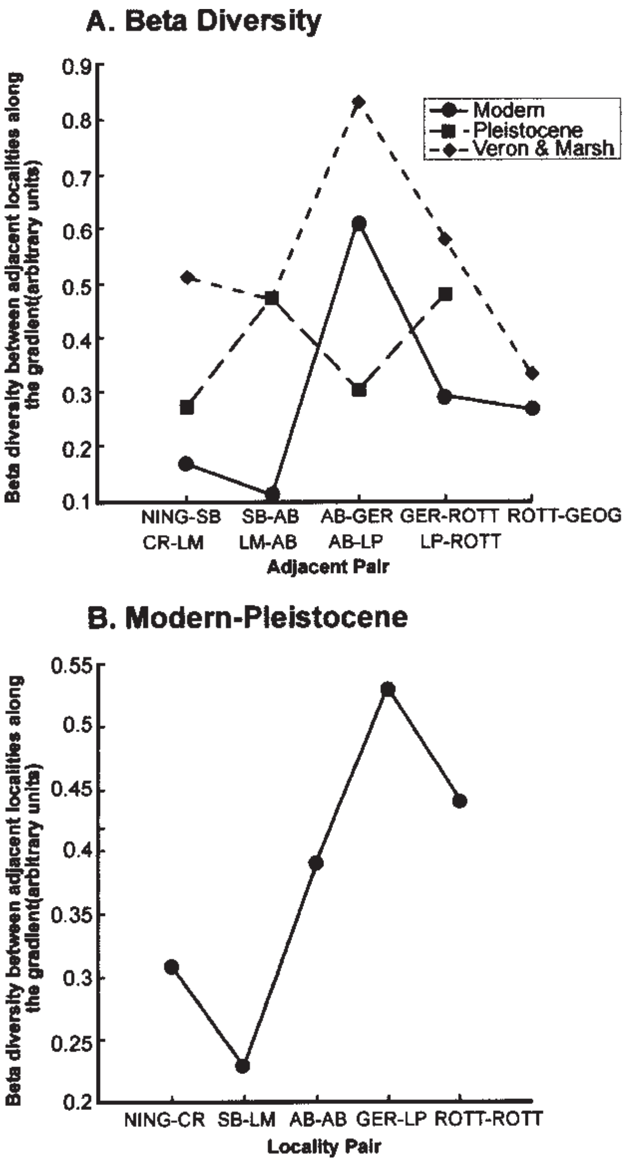

Figure 3. Analysis of beta diversity presented by Greenstein & Pandolfi (2008). A) Beta diversity computed between adjacent localities for modern (full data set of Veron & Marsh 1988; culled data set using only modern taxa that are also represented in Pleistocene reefs) and Pleistocene reef coral communities. A comparison to an additional modern locality south of the study area (Geographe Bay) indicates that lower beta turnover resumes within the Flindersian Province. For modern reefs NING = Ningaloo, SB = Shark Bay, AB = Abrolhos, GER = Geraldton, ROTT = Rottnest Island, GEOG = Geographe Bay. For fossil reefs CR = Cape Range, LM = Lake Macleod-Cape Cuvier, AB = Abrolhos, LP = Leander Point, ROTT = Rottnest Island. Gradient beta diversity and additivity are indicated in upper left. B) Beta diversity between modern and Pleistocene localities present at each latitude, abbreviations as in A, above.

suggesting that during Last Interglacial summers the Leeuwin Current was more intense and delivered a larger volume of warm water to the southern coast of Western Australia. They draw attention to a modern analogue, the high latitude reefs of southern Japan, where coral calcification occurs mainly in the summer months in response to the warm Kuroshio Current. Additionally, McCulloch & Esat (2000) were unable to provide a reliable age for the coral analysed but associated it with an assemblage dated at 128–122 Ka.

There are other claims of varying SSTs during the Last Interglacial. Wells & Wells (1994) in their reconstruction of the ocean circulation off Western Australia indicate a significant reduction in SSTs at 130 Ka. They suggest that at 120 kyrs, winter SSTs off the North West Cape were 3°C higher than at present, with a 'pool' of colder water to the west. In adopting this inference, it needs to be recognised that the SSTs of Wells and Wells tend to underestimate SST (Barrows *et al.* 2000). On the basis of ^{234}U – ^{230}Th ages, Stirling *et al.* (1998) concluded that coral growth associated with a Last Interglacial sea level high stand along the Western Australia margin spans the period *c.* 128–121 Ka, seeing this as a period of reef growth after which reef growth was arrested due to lower SSTs. They considered two alternate, though not mutually exclusive possibilities: (i) reef termination may have been related to a sudden switching off of the Leeuwin Current at *c.*121 Ka, a consequence of global climate events; (ii) the absence of coral reefs may simply be due to a more oscillating sea level history after *c.* 121 Ka, preventing reef growth. However, in the Houtman Abrolhos a coral reef fringe is preserved in a sheltered location in the Wallabi Group and has been dated to 117–116 Ka (Zhu *et al.* 1993). Furthermore, in cores from the Ningaloo reef barrier there is clear evidence of coral growth until *c.* 115±2 Ka (Collins *et al.* 2003). At least for these areas, the indications are that reef growth may not have been constrained by lower SSTs during the later part of the Last Interglacial.

In summary, it is clear that our understanding of Last Interglacial events, both in terms of sea level changes and likely differences in SSTs, is incomplete. Nevertheless it can be confidently concluded that the palaeoecological data from southwestern Australia indicate a stronger than present Leeuwin Current during at least part of the Last Interglacial. If we follow the argument of Stirling *et al.* (1998) that coral growth (at least for the region south of the Houtman Abrolhos) spans the period *c.* 128–121 Ka, and that subsequently reef growth was arrested due to lower sea surface temperatures, recent general circulation experiments allow an exploration of the mechanisms by which this occurred.

Wyrwoll &Valdes (2003) undertook a sensitivity experiment to establish the response of the northwest Australian monsoon to a precession driven, high insolation event that characterised the low latitudes of the Southern Hemisphere at 115 Ka. The sensitivity experiment used the high resolution Hadley Center climate model – HadAM3 version of the U.K. Meteorological Office's Unified Model (Pope *et al.* 2000). One result of the experiment showed a significant increase in lower tropospheric southerly winds along the coast of Western Australia during March (Figure 4). In light of the fact that the Leeuwin Current is controlled by the balance between a southward pressure gradient force and a northward wind stress, this could relate to reduced Leeuwin Current activity during the later part of the Last Interglacial. However, why a stronger Leeuwin Current existed at other times during the Last Interglacial, remains an open question.

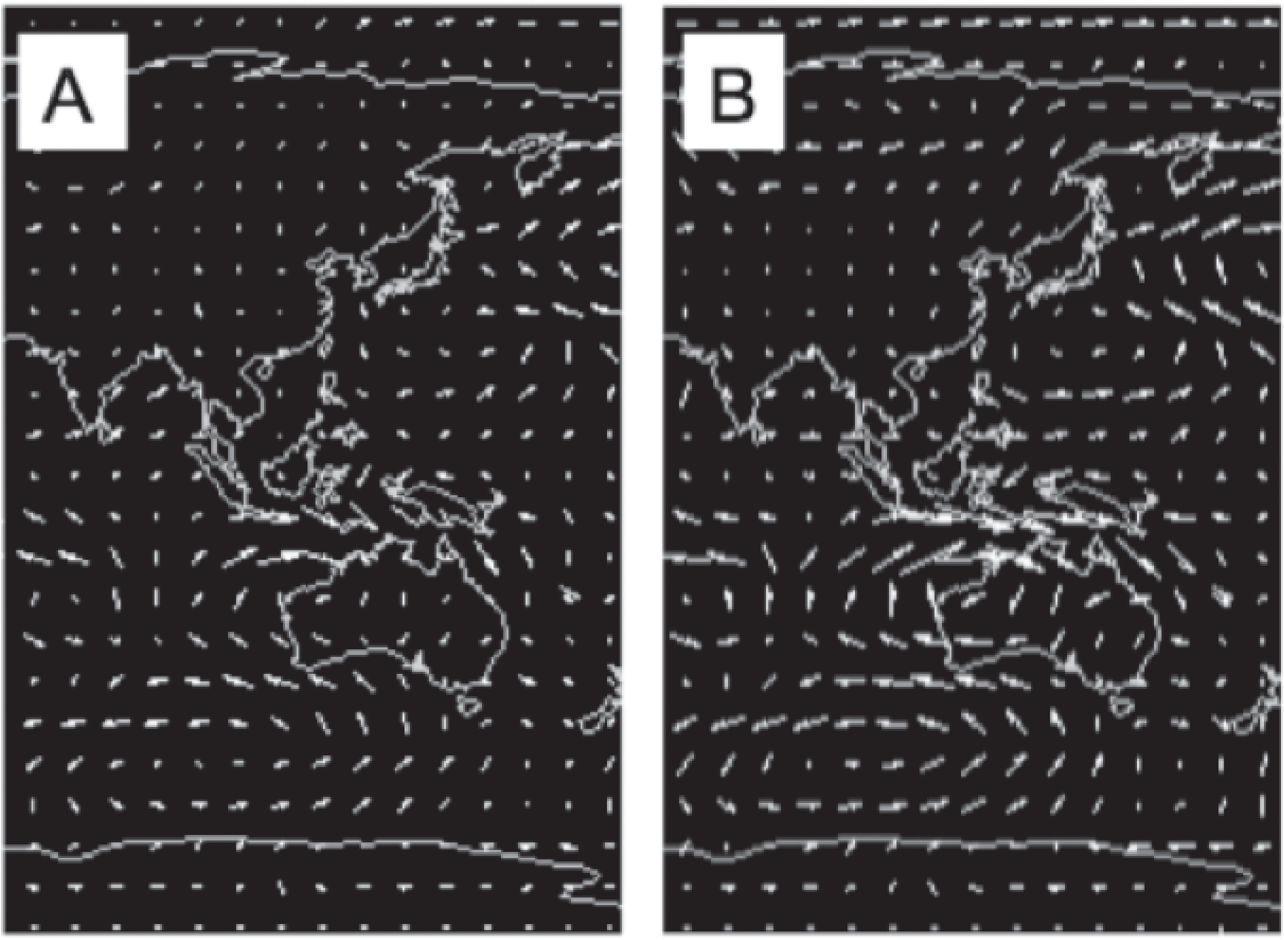

Figure 4. HadAM3 model results for 115 Ka (March) indicating increased strength (with respect to present) of southerly winds along the western Australian margin: (a) Surface winds; (b) Winds at 850 hPa.

Sea level and Leeuwin Current events during the Late Pleistocene

Figure 5 provides a detailed sea level history from the Last Interglacial to the present. It is clear that for much of this time sea level was lower than present. Consequently, only evidence recovered from ocean cores exists, and even that is very limited. Rivers *et al.* (2007) suggested that there is evidence during Marine Isotope Stage 3 (MIS 3) (*c.* 30–60 Ka) for the presence of a warm-water current in the Great Australian Bight, and relate this to the Leeuwin Current. Takahashi (2000) recognises a weakening of the Leeuwin Current during the end of MIS 3.

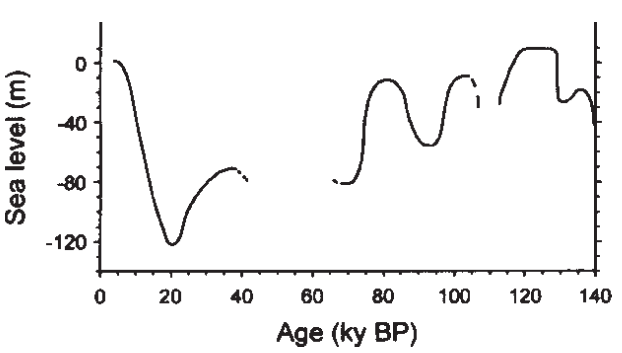

Figure 5. Sea level history for the last *c.* 140 000 years (after Cutler *et al.* 2003).

The Last Glacial Maximum

Because it marks the nadir of the last glacial, the Last Glacial Maximum (LGM), *c.* 21 Ka, provides a profound contrast to the Last Interglacial, prompting interest in the status of the Leeuwin Current at that time. Massive global-scale reorganisation of both atmospheric and oceanic circulations characterised the LGM, with sea levels some 130 m lower than at present. With the lowering of sea level, a massive extension of the land area of Western Australia occurred and with it, major changes in the Indonesian Archipelago (Figure 6). In Western Australia, land extensions were most dramatic over the northwestern region, while further south they were less defined, notably in the North West Cape region – where the shelf is at its narrowest. How these changes in geography may have affected Leeuwin Current activity remains to be explored but given the partial dependence of the Leeuwin Current on the Indonesian Throughflow (Domingues *et al.* 2007) more constricted passages with the low sea levels of the LGM should have reduced its strength. At present there are differing views

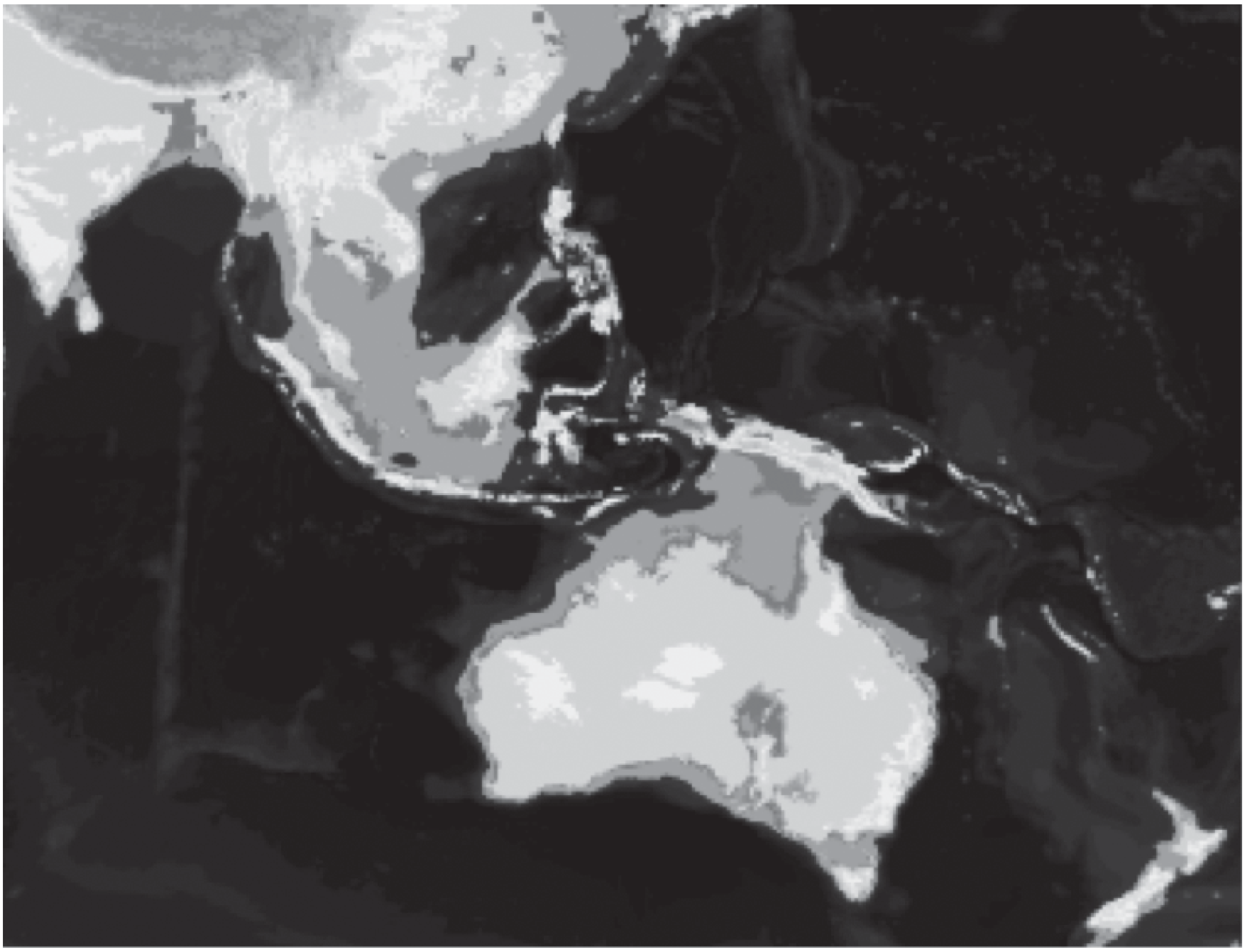

Figure 6. Land extension over the wider Australian-Asian region at the Last Glacial Maximum. Sea level at that time was ~130 m lower than present.

on the state of the Leeuwin Current during the LGM. An overview of the eastern Indian Ocean circulation and more specifically Leeuwin Current activity during this time is provided by Wells & Wells (1994). They conclude that a significant change in SSTs occurred at the time, amounting to a maximum winter (summer) 6°C (7°C) decrease in SSTs in the region of North West Cape and attribute these changes to a more active West Australian Current at that time, in the absence of the Leeuwin Current. Results of work in the Great Australian Bight (Rivers *et al.* 2007) concur with the view that, during much of Marine Isotope Stage 2 (25–11.4 kyrs), the Leeuwin Current was weak or absent – a view also supported by Takahashi (2000).

A wider evaluation of SSTs at the LGM by Barrows & Juggins (2005) suggests that the Leeuwin Current was still active at that time, even though tropical cooling of up to 4.1°C was proposed. These authors recognise an equatorward displacement of isotherms in higher latitudes, indicating colder water transport along the west coast of Australia into low latitudes. According to Barrows & Juggins (*op. cit.*), at the LGM the greatest cooling of 3–4°C occurred between 17–22°S, indicating a stronger West Australian Current at that time. The lower SST estimates of Barrows & Juggins are significantly lower than those of Wells & Wells, who advocated reductions of up to 7°C for this region. Of great significance is their accompanying claim of an active Leeuwin Current at the LGM, albeit transporting cooler water than today (Barrows & Juggins 2005). They see an active Leeuwin Current as being the reason for an apparent absence of an eastern boundary upwelling regime. In support of their view the authors cite Veeh *et al.* (2000), who found little evidence for increased upwelling along the coast of Western Australia during the LGM and had also drawn attention to the likelihood that the Leeuwin Current was active at that time.

Latest Pleistocene

Investigations of chemically- and biologically-mediated precipitation of carbonate materials have been claimed to provide indications of the onset of the Leeuwin Current after a supposed absence during the LGM. James *et al.* (2004) suggest that ooid formation on the northwest shelf required higher salinity than at present and gave ^{14}C ages of 15.4–12.7 Ka for ooid formation. In their interpretations, the salinity levels required for ooid formation are thought to indicate the absence of the Leeuwin Current at that time. In support they cite the work of Gingele *et al.* (2001 a & b) who, on the basis of clay mineralogy, suggest that the Leeuwin Current onset occurred at *c.* 12 Ka, claims that need to be reconciled with the views of Barrows & Juggins (2005) and Veeh *et al.* (2000) of an active Leeuwin Current at the LGM.

While the issue of a possible active/inactive Leeuwin Current at the LGM (or the degree to which this may have been the case) cannot be resolved at present, there can be no doubt that the Leeuwin Current was fully established during the final few millennia of the Pleistocene. Drilling of the Houtman Abrolhos coral reef build-up has shown that by *c.* 10 Ka, a coral reef complex was firmly established (Collins *et al.* 1993; Wyrwoll *et al.* 2006). An intriguing aspect of the seismic stratigraphy obtained from the Houtman Abrolhos is the existence of a reflector (recognised in the Easter Group) at a depth of ~ 40 m. It seems likely that this 'reflector' indicates the 'surface' on which late Pleistocene reef colonisation took place. Reef accumulation rates established for the interval *c.* 10 Ka to present (~ 6 m/kyrs) would require the onset of reef growth to date to *c.* 13 Ka, necessitating a fully established Leeuwin Current prior to that time in order to allow coral colonisation.

The Leeuwin Current over Holocene and historical time scales and associated ENSO variations

Records of Leeuwin Current activity during the last few thousand years are obtained by analyses of the δ^{18}O composition of coral colony skeletons. Massive corals provide the best-known avenue for developing annually resolved climate proxy records throughout the tropical oceans (*e.g.*, Cole 1996; Dunbar 1996; Gagan *et al.* 2000). As they grow, these corals record, in the chemical composition of their aragonitic skeletons, information about the physical and chemical environment of the surrounding seawater. This attribute, combined with rapid growth rates (10–20 mm/year), their longevity (centuries to millennia) and the presence of annual bands (analogous to tree rings), means that they can yield high resolution (*e.g.*, monthly) records of past environmental change over hundreds of years, with a temporal resolution of +/- a few weeks. In order to absolutely resolve temperature vs. seawater composition effects on skeletal δ^{18}O, it is necessary to use combined oxygen isotopic and Sr/Ca analysis. The recently developed technique for quantifying both SST and water composition (salinity) changes using combined tracers has proved extremely effective in both living and fossil corals (McCulloch *et al.* 1994, 1996, 1999; Gagan *et al.* 1998; Tudhope *et al.* 1998; Crowley *et al.* 1999; Ren *et al.* 2003).

An attempt to obtain an indication of Leeuwin Current activity over centennial timescales is provided by Kuhnert *et al.* (1999, 2000). In determining the oxygen isotope (δ^{18}O) composition of cores from coral colonies of *Porites lutea* from both the Houtman Abrolhos and Ningaloo Reef, SST records were developed for the last 200 years and 116 years, respectively. A general trend identified from both sites was a long-term increase in SST amounting to 0.7°C/century for the Houtman Abrolhos and 1.5°C for Ningaloo Reef. The apparent increase in SSTs of 1.5°C is considered problematic and likely to be too high due to uncertainties in the δ^{18}O-SST calibration. Regardless of this uncertainty, the coral record does suggest that the overall warming observed is not coupled to a change in the seasonality of the Leeuwin Current. Allan & Haylock (1993) had already recognised increases in SSTs off the northern and central-west region of Australia (also see Lough 2008), corroborating the trend identified in the coral record. More recently, Feng *et al.* (2005) point to an additional overall Indian Ocean warming of 0.6°C in the period, 1991–2000. Similarly, the continental shelf off Western Australia has seen an increase in SSTs of ~ 0.6°C over the past 50 years (Pearce & Feng 2007).

In the Ningaloo Reef coral record there are indications of short-lived decreases in SST related to volcanic eruptions (Kuhnert *et al.* 2000). The suggestion is of a 1°C SST decrease associated with the Pinatubo eruption (1991) and a possible signal associated with the Krakatau event (1883). In previous work Gagan & Chivas (1995) also identified the Pinatubo event in the stable isotope record of corals from the Ningaloo Reef. They were able to relate this to SST changes induced by the Pinatubo eruption in the Western Pacific Warm Pool (WPWP), some 2.5 years previously and transmitted to the Ningaloo region via the Leeuwin Current. These authors suggest that the eruption-induced cooling of the WPWP could have impacted on the duration (1991–1994) of a negative phase of the Southern Oscillation. It is clear from these data and wider considerations (Cane 2005) that some of the variability evident in Leeuwin Current activity can be attributed to volcanic events and underscores the sensitivity of the system to regional events operating over human time-scales.

As already noted, the importance of El Niño-Southern Oscillation (ENSO) in determining the strength of the Leeuwin Current is well established. However, the attempts to link the 200-year coral δ^{18}O proxy record of SSTs from the Houtman Abrolhos to possible variations in Leeuwin Current activity have met limited success, finding only a weak covariation between Leeuwin Current strength and coral δ^{18}O (Kuhnert *et al.* 1999). A periodicity of 3–5 years is evident in the coral record and some consideration has been given to the possibility that this may be indicative of ENSO events. But it is noteworthy that a similar periodicity is not evident in the Ningaloo Reef record, in which a 9 year periodicity in SSTs is dominant.

It is well known that ENSO events have a complex history when considered over long time scales (*e.g.*, Tudhope *et al.* 2001; Turney *et al.* 2004; Rein *et al.* 2005). Model results suggest that seasonal changes in insolation alone can result in changes in ENSO characteristics (Clement *et al.* 2000; Cane 2005). Rein *et al.* (*op. cit.*) provide a comprehensive appraisal of the record of ENSO events over the last 20 kyrs. They show that a phase of stronger El Niño activity started in Peru around 17 Ka. Very strong El Niño activity occurred during the early and late Holocene, especially during the second and third millennia B.P. El Niño events were weak before and during the beginning of the Younger Dryas, during the middle of the Holocene and during Medieval times. An earlier compilation by McGregor & Gagan (2004) drew on coral δ^{18}O records throughout the west Pacific to provide details of the variation of ENSO occurrence during the Holocene. They recognise 12 events/century for the period 7.6–7.1 Ka; 8 events/century for the period, 6.1–5.4 Ka; and 6 events/century at 6.5 Ka. During 2.5–1.7 Ka, the coral records indicate large and protracted $\delta^{18}O$ anomalies indicative of particularly severe El Niño events. They note specifically, that the 2.5 Ka Madang Papua New Guinea corals record a protracted 4-year El Niño, like the 1991–1994 event, but almost twice the amplitude of the 1997–1998 event (Tudhope *et al*.2001). In addition, they recognise that the 2 Ka Muschu Island (Papua New Guinea) coral $\delta^{18}O$ record shows a severe 7 year El Niño, longer than any recorded Holocene or modern event. This is supported by the findings of Woodroffe *et al.* (2003) who analysed a late Holocene coral record from Christmas Island (central Pacific) and found evidence for an extreme El Niño that was twice the amplitude of the 1997–1998 event. Significant variations are also evident in the more recent historical record (*e.g.*, Allan *et al.* 1996). Given the strong and generally accepted relationship between the strength of the Leeuwin Current and ENSO events, the now-established record of ENSO makes it clear that the Leeuwin Current must have shown marked variations over a range of time-scales. Such variability must have imprinted itself on the biota leading it to acquire a 'plasticity' that equipped it to accommodate ENSO variability or, alternatively, forcing marked changes in population size resulting in 'bottle-neck' effects (B. Knott, *pers. com.*). Observed 'plasticity' in the biota should inform ecological notions as to whether marine communities represent loose associations of component taxa that respond individually to environmental change, or more integrated associations that are predictable in space and time.

Some lessons for the future?

Forcing mechanisms

A challenge facing us is predicting the response of the Leeuwin Current to anticipated future greenhouse climate states and likely accompanying biotic changes. Since the trajectory of human induced climate change is forecast to continue into the foreseeable future (IPCC 2007), predicting the impact of climate change on the abundance and distribution of organisms has become increasingly urgent. The western margin of Australia provides no exception to this, with possible future changes in Leeuwin Current activity having potentially far-reaching implications for the marine biota. An early attempt at considering the likely response of the Leeuwin Current to global warming and the implications this carries for marine organisms was provided by Pattiaratchi & Buchan (1991). Their attempt at identifying the possibilities of changes in forcing mechanisms was greatly hampered by the limitations of global climate models at the time of study. These limitations were highlighted by their inability to establish the likelihood of possible ENSO changes under enhanced greenhouse conditions – a foremost consideration in any assessment of future Leeuwin Current activity. The restrictions imposed by model limitations were still evident by the time of Cane's (2005) review, and even today only guarded conclusions can be drawn from available simulations. The Fourth IPCC report (IPCC 2007) concludes that present model simulations still give no consistent indication of discernible changes in projected ENSO amplitude or frequency in the 21st century.

Biological response

Whatever models may ultimately predict about future climate states and their impact on Leeuwin Current activity it is clear that at the most general level, the Earth has experienced significant anthropogenic warming over the last *c.* 100 years. We know that 'warm Earth' periods evident in the global climate record have imprinted themselves on Leeuwin Current activity – as evidenced in Western Australia by the fossil assemblages preserved in the Roe Calcarenite and the Last Interglacial (MIS 5e) deposits. Given this evidence, it would seem sensible to turn to past 'warm Earth' events and use these as analogues to explore the nature of Leeuwin Current activity and the biological response under a future warm climate state. The use of 'geological analogues' to complement model climate predictions (*e.g.*, IPCC 2007) is not without potential pitfalls. One of the most challenging is that past 'warm' states were insolation driven, while future climates are a response to anthropogenic greenhouse forcing. Nevertheless, using past 'warm Earth' Leeuwin Current states may help to shed some light on the course of future events.

At the global-scale the Last Interglacial was warmer than the present interglacial, clearly evidenced by proxy palaeoclimate data-sets (Kukla *et al.* 2002) and reflected in higher than present sea levels (Stirling *et al.* 1998), indicating lower ice volumes. It was clearly a time of stronger than present Leeuwin Current activity, which had a profound impact on shallow marine ecosystems adjacent to coastal Western Australia. Many of the important components of these systems have been preserved as fossils and palaeoecological studies of these deposits allow us to begin to understand ecosystem responses to climate change over spatial and temporal scales unavailable to modern ecological studies. The results may then be used as a step towards understanding the likely responses of shallow marine communities to future global warming. The assumption underlying this approach is, of course, that in a future warmer Earth, Leeuwin Current activity will mirror past 'warm Earth' responses.

The results of recent work (Greenstein & Pandolfi 2008) predicting the response of coral taxa to future greenhouse induced Leeuwin Current changes, provides insight into the potential that this approach holds. Already, present climate change is resulting in noticeable range expansions of coral taxa in the Caribbean (Vargas-Ángel *et al.* 2003; Precht & Aronson 2004) and the Indo-Pacific (Marsh 1993) provinces. The genus *Acropora* was reported living for the first time from Rottnest Island by Marsh (1993), who attributed the presence of *A. youngei* to both increased temperature as well as lack of competition from macroalgae. Range expansions of individual coral taxa will likely continue as temperature increases over the next century. Greenstein & Pandolfi (2008) predict that certain acroporid and faviid corals would appear in regions south of the Houtman Abrolhos Islands in the future. Although such predicted geographic shifts of coral taxa will not mitigate ecological and economic losses resulting from localised mortality, range expansions of important contributors to tropical reef coral communities may allow them to persist in "temperature refugia" in the wake of future climate change.

Greenstein & Pandolfi (2008) also found that less faunal change occurred between fossil coral communities between North West Cape and Rottnest Island compared to the adjacent modern coral communities offshore. A similar decrease in beta diversity between communities may occur as tropical-adapted coral taxa expand their geographic ranges south along the coast of Western Australia in the future. This effect may be amplified by preferential survival of particular coral taxa (*e.g.*, Done 1999). Although the reasons for the correlation are controversial, the role of biodiversity in enhancing ecological stability has been demonstrated on small spatial and temporal scales (*e.g.*, McCann 2000; Naeem & Li 1997; McGrady-Steed *et al.* 1997). However, it remains to be demonstrated whether regional ecological stability and beta diversity are correlated. In summary, this recent comparative study of Pleistocene and modern corals in coastal Western Australia elucidates the likely impact of increasing Leewuin Current activity on coral faunas in the region. Additional work predicting the impact on coral community membership remains to be attempted.

Concluding discussion

A unique aspect of geological study of evidence for the Leeuwin Current is that it allows a distinction to be made between factors that affected the current over significantly different time scales. For example, the early history of the Leeuwin Current is characterised by changes that took place over time intervals quite beyond human experience. The interplay of tectonics (deepening of the Tasmanian Gateway followed by emplacement of today's Indonesian Archipelago) and ocean circulation took place over tens of millions of years, and established the initial foundation for "Leeuwin Current style" along the western margin of Australia. Whether prior to this, the evidence of warm sea surface temperatures should be ascribed to the existence of a "proto-Leeuwin Current" can be questioned. It may be more appropriate to relate the existence of the Leeuwin Current to its forcing mechanisms, and not simply claim its existence on the basis of elevated SSTs.

Once the Leeuwin Current *sensu stricto* was emplaced by these processes 2–3 million years ago, its geologic history comprised alternate waxing and waning in response to forcing mechanisms related to climate rather than tectonics. Although shorter term climatic changes certainly affected the Leeuwin Current earlier in its history, the resolution afforded by younger stratigraphic sections as well as geochemical data preserved in modern and fossil coral skeletons has allowed researchers to investigate the climate-current relationship at two increasingly finer temporal scales as the recent is approached.

The first of these is Milankovitch-style orbital forcing of climate, which has been shown to affect Leeuwin Current activity on scales of 10–100 Ka at least since the onset of Pleistocene time, two million years ago. These cycles were likely responsible for the pronounced variability in the strength of the Leeuwin Current during the Pleistocene as well as the changeover to a carbonate-dominated, subtropical-tropical sedimentary regime in coastal Western Australia during the Middle Pleistocene. The best-studied of the orbital-scale events are the Last Interglacial and Last Glacial Maximum. Sedimentary deposits and fossil assemblages from each interval have yielded information about SST variability and estimates of the strength of the Leeuwin Current. Range expansions and contractions of important marine benthic taxa in response to current variability have been documented from coastal and offshore deposits produced during each interval.

The finest temporal resolution of Leeuwin Current history comes from climate proxies obtained by geochemical analyses of fossil and modern coral skeletons. Perhaps it is this history that is most relevant to issues of contemporary marine management. The coral proxy record shows clearly that the Leeuwin Current is sensitive to changes in global climate (*e.g.*, El Niño-La Niña events) as well as subtle changes in regional climate (*e.g.*, impact of volcanic eruptions), both of which operate over human timescales. Given the sensitivity apparent from the recent geological past, the Leeuwin Current is certain to respond to anthropogenically induced warming over this century. Hence continued application of geological analogues (using the recent geological past to inform predictions of the biologic response to future change) and climate modelling (to predict the timing and magnitude of that change) remain an essential, albeit challenging research arena.

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References

Allan R J & Haylock M R 1993 Circulation features associated with the winter rainfall decrease in Southwestern Australia. Journal of Climate 6: 1356–1367.

Allan R, Lindesay J & Parker D 1996 El Niño Southern Oscillation climate variability. CSIRO Publishing.

Andrews J E, Portman C, Rowe P J, Leeder M R & Kramers J D 2007 Sub-orbital sea-level change in early MIS 5e: New evidence from the Gulf of Corinth, Greece. Earth and Planetary Science Letters 259: 457–468.

Archer D & Ganopolski A 2005 A moveable trigger: fossil fuel CO2 and the onset of the next glaciation. Geochemistry, Geophysics, Geosystems 6: Q05003, doi10.1029/2004GC000891.

Barrows T T & Juggins S 2005 Sea-surface temperatures around the Australian margin and Indian Ocean during the Last Glacial Maximum. Quaternary Science Reviews 24: 1017–1047.

Barrows T T, Juggins S, De Deckker P, Thiede J & Martinez J I 2000 Sea-surface temperatures of the southwest Pacific Ocean during the last glacial maximum. Paleoceanography 15: 95–109.

Bastian L V 1996 Residual soil mineralogy and dune subdivision, Swan Coastal Plain, Western Australia. Australian Journal of Earth Sciences 43: 31–44.

Bernasconi S M, Andres M S & McKenzie J A 2002 850,000-year record of sea -level fluctuations and margin evolution from Great Australian Bight, Site 1127, ODP Leg 182. American Geophysical Union, Fall Meeting, Abstract #PP12A-0327.

Beu A G 2006 Marine molluscs of oxygen isotope stages of the last 2 million years in New Zealand. Part 2. Biostratigraphically useful and new Pliocene to Recent bivalves. Royal Society of New Zealand 36: 151–338.

Beu A G & Darragh T A 2001 Revision of southern Australia Cenozoic fossil Pectinidae (Mollusca : Bivalvia). Transactions of the Royal Society of Victoria 113: 1–205.

Bradley R S 1999 Paleoclimatology: reconstructing climates of the Quaternary. Academic Press, London.

Cane M A 2005 The evolution of El Niño, past and future. Earth and Planetary Science Letters 230: 227–240.

Cane M A & Molnar P 2001 Closing of the Indonesian seaway as a precursor to east African aridification around 3–4 million years ago. Nature 411: 157–162.

Clement A C, Seager R & Cane M A 2000 Suppression of El Niño during the mid-Holocene by changes in the Earth's orbit. Paleoceanography 15: 731–737.

Cole J 1996 Coral records of climate change: understanding past variability in the tropical ocean-atmosphere. *In*: Climatic variations and forcing mechanisms of the last 2000 years (eds P D Jones, R S Bradley & J Jouzel). NATO ASI Series, 41: 331–353.

Cole J E 2005 Holocene coral record: windows on tropical variability. *In*: Global change in the Holocene (eds A Mackay, R Battarbee, J Birks & F Oldfield). Hodder Arnold, London, 168–184.

Collins L B, Zhu Z R, Wyrwoll K-H, Hatcher B G, Playford P E, Chen J H, Eisenhauer A & Wasserburg G J 1993 Late Quaternary facies characteristics and growth history of a high latitude reef complex: the Abrolhos carbonate platforms, eastern Indian Ocean. Marine Geology 111: 203–221.

Collins L B, Zhu Z R, Wyrwoll K-H & Eisenhauer A 2003 Late Quaternary structure and development of the northern Ningaloo Reef, Australia. Sedimentary Geology, Special Volume: Late Quaternary Reef Development 159: 81–94.

Cresswell G R 1991 The Leeuwin Current – observations and recent models. Journal of the Royal Society of Western Australia 74: 1–14.

Crowley T J, Quinn T M & Hyde W T 1999 Validation of coral temperature calibrations. Paleoceanography 14: 605–615.

Cutler K B, Edwards R L, Taylor F W, Cheng H, Adkins J, Gallup C D, Cutler P M, Burr G S & Bloom A L 2003 Rapid sea-level fall and deep-ocean temperature change since the last interglacial period. Earth & Planetary Science Letters 206: 253–271.

Darragh T A & Kendrick G W 2008 Silicified Eocene molluscs from the Lower Murchison district, Southern Carnarvon Basin, Western Australia. Records of the Western Australian Museum 24: 217–246.

Domingues C A, Maltrud M E, Wijffels S E, Church J A & Tmczak M 2007 Simulated Langrangian pathways between the Leeuwin Current and the upper-ocean circulation of the southeast Indian Ocean. Deep-Sea Research 54: 797–817.

Done T J 1999 Coral community adaptability to environmental change at the scales of regions, reefs and reef zones. American Zoologist 3: 66–79.

Dunbar R B 1996 Eastern Pacific corals monitor El Niño/Southern Oscillation, precipitation, and sea surface temperature variability over the past 3 centuries. *In*: Climatic variations and forcing mechanisms of the last 2000 years (eds P D Jones, R S Bradley & J Jouzel). NATO ASI Series, 41: 373–405.

Eisenhauer A, Chen J, Wasserburg G J, Collins L B, Zhu Z R & Wyrwoll K-H 1993 Holocene sea level determinations relative to Australia. Earth & Planetary Science Letters 114: 529–547.

Eisenhauer A, Zhu Z R, Collins L B, Wyrwoll K-H & Eichstätter R 1996 The Last Interglacial sea level change: new evidence from the Abrolhos Islands, West Australia. Geologische Rundschau 85: 606–614.

Feng M, Meyers G & Church J 2005 How our sea temperature has changed. Indian Ocean Climate Initiative, Climate Note 3/05.

Gagan M K, Ayliffe L K, Hopley D, Cali J A, Mortimer G E, Chappell J, McCulloch M T & Head M J 1998 Temperature and surface-ocean water balance of the mid-Holocene tropical western Pacific. Science 279: 1014–1018.

Gagan M K, Ayliffe L K, Beck J W, Cole J E, Druffel R B, Dunbar R B & Schrag D P 2000 New views of tropical paleoclimate from corals. Quaternary Science Reviews 19: 167–182.

Gagan M K & Chivas A R 1995 Oxygen isotopes in western Australian coral reveal Pinatubo aerosol-induced cooling in the Western Pacific Warm Pool. Geophysical Research Letters 22: 1069–1072.

Gaina C & Müller C 2007 Cenozoic tectonic and depth/age evolution of the Indonesian gateway and associated back-arc basins. Earth-Science Reviews 83: 177–203.

Garidel-Thoron de T, Rosenthal Y, Bassinot F & Beaufort L 2005 Stable sea surface temperatures in the western Pacific warm pool over the past 1.75 million years. Nature 433: 294–298.

Gingele F X, De Deckker P & Hillenbrand C D 2001a Clay mineral distribution in surface sediments between Indonesia and NW Australia – source and transport by ocean currents. Marine Geology 179: 135–146.

Gingele F X, De Deckker P & Hillenbrand C D 2001b Late Quaternary fluctuations of the Leeuwin Current and paleoclimates on the adjacent land masses: clay mineral evidence. Australian Journal of Earth Sciences 48: 867–874.

Godfrey J S 1989 A Sverdrup model of the depth-integrated flow for the World Ocean allowing for island circulations. Geophysical and Astrophysical Fluid Dynamics 45: 89–112.

Godfrey J S & Weaver A J 1991 Is the Leeuwin Current driven by Pacific heating and winds? Progress in Oceanography 27:225–272.

Gordon A L, Susanto R D & Vranes K 2003: Cool Indonesian throughflow as a consequence of restricted surface layer flow. Nature 425: 824–828.

Greenstein B J, Pandolfi J M & Blakeway D R 2005 A fossil reef from the Last Interglacial, Western Australia. Coral Reefs 24: 593.

Greenstein B J & Pandolfi J M 2008 Escaping the heat: range shifts of reef coral taxa in coastal Western Australia. Global Change Biology 14: 1–16.

Haig D W & Mory A J 2003 New records of siliceous, marine, later Eocene from Kalbarri, Western Australia. Journal of the Royal Society of Western Australia 86: 107–113.

Haywood A M, Valdes P J & Sellwood B W 2000 Global scale paleoclimate reconstructions of the middle Pliocene climate using UKMO GCM: initial results. Global Planetary Change 25: 239–256.

Hearty P J, Hollin J T, Neumann A C, O'Leary M J & McCulloch M 2007 Global sea-level fluctuations during the Last Interglaciation (MIS5e). Quaternary Science Reviews 26: 2090–2112.

Hirst A C & Godfrey J S 1993 The role of the Indonesian through-flow in a global ocean GCM. Journal of Physical Oceanography 23: 1057–1086.

IPCC 2007 Climate Change 2007: The physical science basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (eds S Solomon, D Qin, M Manning, Z Chen, M Marquis, K B Averyt, M Tignor & H L Miller). Cambridge University Press, Cambridge.

Jacob R 1997 Low frequency variability in a simulated atmosphere ocean system. Ph.D thesis, University of Wisconsin-Madison, 177 pp.

James N B, Bone Y, Kyser T K, Dix G R & Collins L B 2004 The importance of changing oceanography in controlling lat Quaternary carbonate sedimentation on a high-energy, tropical oceanic ramp: north-western Australia. Sedimentology 51: 1179–1205.

Kendrick G W, Wyrwoll K-H & Szabo B J 1991 Pliocene-Pleistocene coastal events and history along the western margin of Australia. Quaternary Science Reviews 10: 4 19–439.

Kuhnert H, Pätzold J, Hatcher B, Wyrwoll K-H, Eisenhauer A, Collins L B, Zhu ZR & Wefer G 1999 A 200-year coral stable isotope record from a high-latitude reef off Western Australia. Coral Reefs 18: 1–12.

Kuhnert H, Pätzold J, Wyrwoll K-H & Wefer G 2000 Monitoring climate variability over the past 116 years in coral oxygen isotopes from Ningaloo Reef, Western Australia. International Journal of Earth Science 88: 725–732.

Kuhnt W, Holbourn A, Hall R, Zuvela M & Käse R 2004 Neogene history of the Indonesian Throughflow. *In*: Continent-ocean interactions in the East Asian marginal seas (eds P Clift, D Hayes, W Kuhnt & P Wang). American Geophysical Union Monographs 149: 299–320.

Kukla G J, Bender M L, de Beaulieu J-L, Bond G, Broecker W S, Cleveringa P, Gavin J E, Herbert T D, Imbrie J, Jouzel J, Keigwin L D, Knudsen K-L, McManus J F, Merkt J, Muhs D R, Muller H, Poore R Z, Porter S C, Seret G, Shackleton N J, Turner C, Tzedakis P C & Winograd I J 2002 Last Interglacial climates. Quaternary Research 58: 2–13.

Lea D W 2004 The 100 000.yr cycle in tropical SST, greenhouse forcing, and climate sensitivity. Journal of Climate 17: 2170–2179.

Li K J & Clarke A J 2004 Coastline direction, interannual flow, and the strong El Niño currents along Australia's nearly zonal southern coast. Journal of Physical Oceanography 34: 2373–2381.

Li Z X & Powell C McA 2001 An outline of the palaeogeographic evolution of the Australasian region since the beginning of the Neoproterozoic. Earth-Science Reviews 53: 237–277.

Liu Z, Brady E & Lynch-Steiglitz J 2003 Global ocean response to orbital forcing in the Holocene. Paleoceanography 18: 19:1–19:20. doi:10.1029/2002PA000819.

Lough J M 2008 Shifting climate zones of Australia's tropical marine system. Geophysical Research Letters 35 L14708, doi 10.1029/20008GL034634.

Loutre M F & Berger A L 2000 Future climate changes: are we entering an exceptionally long interglacial? Climate Change 46: 16–19.

Lowry D C 1968 Tertiary stratigraphic units in the Eucla basin in Western Australia. Geological Survey of Western Australia Annual Report for 1967: 36–40.

Lowry D C 1970 The geology of the Western Australian part of the Eucla basin. Geological Survey of Western Australia Bulletin 122: 201pp.

Ludbrook N H 1978 Quaternary molluscs of the western part of the Eucla Basin. Geological Survey of Western Australia Bulletin 125: 1–286.

Marsh L M 1993 The occurrence and growth of Acropora in extra-tropical waters off Perth, Western Australia. *In*: Proceedings of the 7th International Coral Reef Symposium (ed R H Richmond), University of Guam Press, UOG Station, Guam. Vol 2: 1233–1238.

Marshall JF & Thom B G 1976 The sea-level in the last interglacial. Nature 263: 120–121.

McCann K S 2000 The diversity-stability debate. Nature 405: 228–233.

McCulloch M T &Mortimes G E 2008 Application of the U-Th decay series to dating of fossil and modern corals using MC-ICPMS. Australian Journal of Earth Sciences 55: 955–965.

McCulloch M T & Esat T 2000 The coral record of last interglacial sea levels and sea surface temperatures. Chemical Geology 169: 107–129.

McCulloch M T, Gagan M K, Mortimer G E, Chivas A R & Isdale P J 1994 A high resolution Sr/Ca and δ18O coral record from the Great Barrier Reef, Australia, and the 1982–1983 El Niño. Geochimica et Cosmochimica Acta 58: 2747–2754.

McCulloch M T, Mortimer G E, Esat T M, Xianhua L, Pillans B & Chappell J 1996 High resolution windows into Early-Holocene climate: Sr/Ca coral records from the Huon Peninsula. Earth and Planetary Science Letters 138: 169–178.

McCulloch M T, Tudhope A W, Esat T M, Mortimer G E, Chappell J, Pillans B, Chivas A R & Omura A 1999 Coral record of equatorial sea-surface temperatures during the penultimate deglaciation at Huon Peninsula. Science 283: 202–204.

McGowran B, Li Q, Cann J, Padley D, McKirdy D M & Shafik S 1997 Biogeographic impact of the Leeuwin Current in southern Australia since the late middle Eocene. Palaeogeography, Palaeoclimatology, Palaeoecology 136: 19–40.

McGowran B, Archer M, Bock P, Darragh T A, Godthelp H, Hageman S, Han S J, Hill R, Li Q, Maxwell P A, McNamara K J, Macphail M, Mildenhall D, Partridge A D, Richardson J, Shafik S, Truswell E M & Warne M 2000 Memoir of the Association of Australasian Palaeontologists 233: 405–470.

McGrady-Steed J, Harris P M & Morin P J 1997 Biodiversity regulates ecosystem predictability. Nature 390: 162–165.

McGregor H V & Gagan M K 2004 Western Pacific coral $\delta^{18}O$ records of anomalous Holocene variability in the El Niño-Southern Oscillation. Geophysical Research Letters 31: 10.1029/2004GL019972.

McNamara K J & Kendrick G W 1994 Cenozoic molluscs and echinoids of Barrow Island, Western Australia. Records of the Western Australian Museum, Supplement 51.

McNamara K J & Scott J K 1983 A new species of Banksia (Proteaceae) for the Eocene Merlinleigh Sandstone of the Kennedy Range, Western Australia. Alcheringa 7: 185–193.

Naeem S & Li S 1997 Biodiversity enhances ecosystem reliability. Nature 390: 507–509.

Nishimura S 2002 Topics on Neogene evolution of Pacific Ocean gateways, a summary of IGCP-355. Revista Mexicana de Ciencias Geológicas 19: 2002–273.

Nishimura S & Suparka S 1997 Tectonic approach to the neogene evolution of Pacific-Indian seaways. Tectonophysics 281: 1–16.

North Greenland Ice Core Project Members 2004 High-resolution record of Northern Hemisphere climate extending into the last interglacial period. Nature 432: 147–151.

Pattiaratchi C B & Buchan S J 1991 Implications of long-term climate change for the Leeuwin Current. Journal of the Royal Society of Western Australia 74: 133–140.

Pearce A F & Feng M 2007 Observations of warming on the Western Australian continental shelf. Marine and Freshwater Research 58: 914–920.

Pearce A F & Phillips B F 1988 ENSO events, the Leeuwin Current and larval recruitment of the western rock lobster. Journal de Conseil International pour l'Exploration de le Mer 45: 13–21.

Playford P E 1988 Guidebook to the geology of Rottnest Island. Geological Society of Australia, Perth and the Geological Survey of Western Australia.

Pope V D, Gallani M L, Rowntree P R & Stratton R A 2000 The impact of new parameterisations in the Hadley Centre climate model-HadAM3. Climate Dynamics 16: 123–146.

Precht W F & Aronson R B 2004 Climate flickers and range shifts of reef corals. Frontiers in Ecology and the Environment 2: 307–314.

Raymo M E, Grant B, Horowitz M & Rau G H 1996 Mid-Pliocene warmth: stronger greenhouse and stronger conveyor. Marine Micropaleontology 27: 313–326.

Rein B, Lückge A, Reinhardt L, Sirocko F, Wolf A & Dullo W-C 2005 El Niño variability off Peru during the last 20,000 years. Paleoceanography 20: PA4003, doi:10.1029/2004PA001099.

Ren L, Linsley B K, Wellington G M, Schrag D P, Hoegh-guldberg, 2003 Deconvolving the $\delta^{18}O$ seawater component from subseasonal coral $\delta^{18}O$ and Sr/Ca at Rarotonga in the southwestern subtropical Pacific for the period 1726 to 1997. Geochimica et Cosmochimica Acta 67: 1609–1621.

Rivers J M, James N P, Kyser T K & Bone Y 2007 Genesis of palimpsest cool-water carbonate sediment on the continental margin of Southern Australia. Journal of Sedimentary Research 77: 480–494. Doi:10.2110/jst.2007.046.

Ruddiman W F 2007 Earth's Climate: past and future. W H Freeman & Company. New York. 388pp (Second edition).

Saltzman B 2002 Dynamical paleoclimatology: generalized theory of global climate change. Academic Press. San Diego. 354pp.

Sato K, Oda M, Chiyonobu S, Kimoto K, Domitsu H & Ingle C I 2008 Establishment of western Pacific warm pool during the Pliocene: evidence from planktic foraminifera, oxygen isotopes, and Mg/Ca ratios. Palaeogeography, Palaeoclimatology, Palaeoecology 262: 140–147.

Shafik S 1990 The Maastrichtian and early Tertiary record of the Great Australian Bight Basin and its onshore equivalents on the Australian southern margin: a nannofossil study. BMR Journal of Australian Geology & Geophysics 11: 473–497.

Stickley C E, Brinhuis H, Schellenberg S A, Sluijs A, Röhl U, Fuller M, Grauert M, Huber M, Warnaar J & Williams G L 2004 Timing and deepening of the Tasmanian Gateway. Paleoceanography 19: PA4027, doi:10.1029./2004 PA001022.

Stirling C H, Esat T M, Lambeck K & McCulloch M T 1998 Timing and duration of the Last Interglacial: evidence for a restricted interval of widespread coral reef growth. Earth & Planetary Science Letters 160: 745–762.

Stirling C H, Esat T M, McCulloch M T & Lambeck K 1995 High-precision U-series dating of corals from Western Australia and implications for the timing and duration of the Last Interglacial. Earth & Planetary Science Letters 135: 115–130.

Szabo B J 1979 Uranium-series age of coral reef growth at Rottnest Island, Western Australia. Marine Geology 29: M11–M15.

Takahashi K 2000 Calcareous nanoplankton in the southeastern Indian Ocean: modern biogeography and paleoceanography for the last 30,000 years. Ph.D. Thesis, Department of Earth & Planetary Sciences, Graduate School of Science, Hokkaido University, Japan.

Tudhope A W, Chilcott C P, McCulloch M T, Chappell J, Shimmield G B & Ellam R 1998 ENSO-style climate variability during glacial and interglacial climate regimes: Evidence from geochemical analysis of annually-banded massive corals in Papua New Guinea. EOS Transactions, American Geophysical Union 79(45): F493, AGU Fall Meeting 1998.

Tudhope A W, Chilcott C P, McCulloch M T, Cook E R, Chappell J, Ellam R M, Lea D W, Lough J M & Shimmield G B 2001 Variability in the El Niño – Southern Oscillation through a glacial-interglacial cycle. Science 291: 1511–1517.

Turney C S M, Kershaw A P, Clemens S C, Branch N, Moss P T & Fifield L K 2004 Millennial and orbital variations of El Niño/Southern Oscillation and high-latitude climate in the last glacial period. Nature 428: 306–309.

Vargas-Ángel B, Thomas J D & Hoke1 S M 2003 High-latitude *Acropora cervicornis* thickets off Fort Lauderdale, Florida, USA. Coral Reefs 22: 465–473.

Veeh H H, Schwebel D, van de Graaff W J E & Denman P E 1979 Uranium-series age of coralline terrace deposits in Western Australia. Geological Society of Western Australia Journal 26: 285–292.

Veeh H H, McCorkle D C & Heggie D T 2000 Glacial/interglacial variations in sedimentation on the West Australian continental margin: constraints from excess ^{230}Th. Marine Geology 166: 11–30.

Veron J E N & Marsh L M 1988 Hermatypic corals of Western Australia: records of annotated species lists. Records of the Western Australian Museum, Supplement 2. Western Australian Museum, Perth.

Wells P E & Wells G M 1994 Large-scale reorganization of ocean currents offshore Western Australia during the late Quaternary. Marine Micropaleontology 24: 157–186.

Wilson B R & Gillett K 1980 Australian Shells. AH & AW Reed Pty. Ltd, Sydney.

Woodroffe C D, Beech M R & Gagan M K 2003 Mid-late Holocene El Niño variability in the equatorial Pacific from coral microatolls. Geophysical Research Letters 30: 10.1029/2002GL 015868.

Wyrwoll K-H & Valdes P 2003 Insolation forcing of the Australian monsoon as controls of Pleistocene mega-lake events. Geophysical Research Letters 30: 2279–2283.

Wyrwoll K-H, Zhu Z R, Collins L B & Hatcher B G 2006 Origin of blue hole structures in coral reefs: Houtman Abrolhos, Western Australia. Journal of Coastal Research 22: 202–208.

Wyrwoll K-H, Liu Z, Chen G, Kutzbach, Liu X 2007 Sensitivity of the Australian summer monsoon to tilt and precession forcing. Quaternary Science Reviews 26: 3043–3057.

Zachos J, Pagani M, Sloan L, Thomas E & Billups K 2001 Trends, rhythms, and aberrations in global climate 65 Ma to present. Science 292: 686–693.

Zhu Z R, Wyrwoll K-H, Collins L B, Chen J H, Wasserburg G J & Eisenhauer A 1993 High-precision U-series dating of Last Interglacial events by mass spectroscopy: Houtman Abrolhos Islands, Western Australia. Earth & Planetary Science Letters 118: 281–293.