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Palaeoceanography of the Western Pacific Warm Pool During the Last Glacial Maximum:
Long-term Climatic Monitoring of the Maritime Continent

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Introduction

Until recently, palaeoceanographic changes in the North Atlantic, through the Atlantic Conveyor Belt, were put forward as a plausible mechanism to explain Pleistocene global oceanic circulation and climate changes (e.g. Broecker and Denton 1989). This view has been challenged by the realisation that oceanographic and climatic phenomena in the tropics may also significantly contribute to global change (e.g. Hirst and Godfrey 1993). The Western Pacific Warm Pool (WPWP), the warmest region of the global ocean (>28°C, Yan *et al.* 1992), is part of the El Niño-Southern Oscillation phenomenon (ENSO) and, as such, variations in its extent and dynamics through time play an important role in global climate (e.g. Godfrey 1996). Furthermore, the Indonesian passageway, as the connection at the surface between the Pacific and Indian Oceans, constitutes a major component of the renewal of the North Atlantic Deep Water (NADW; e.g. Gordon 1986). However, this view has been questioned by a number of researchers (e.g. Pariwono *et al.* 1986; Rintoul 1991; Schmitz and McCartney 1993; Schmitz 1995; Godfrey 1996; Wijffels *et al.* 1996; Macdonald and Wunsch 1996; Macdonald 1998; Ganachaud and Wunsch 2000) who consider the Indonesian throughflow as a minor constituent for the advection of NADW, but probably a major contributor to the exchange of upper water in the Southern Hemisphere. Furthermore, it has been suggested that the wind field around Australia exerts a major influence on the magnitude of the throughflow (e.g. Pariwono *et al.* 1986; Godfrey 1996; McBride 1998), and in particular, the strength of the wind field in the Southern Ocean as well as the topographic barrier imposed by New Zealand (Godfrey 1996).

If this last scenario is valid, then we have to look for palaeoceanographic and palaeoclimatic evidences all around Australia when trying to reconstruct the WPWP and the dynamics of the Indonesian throughflow during the last glacial maximum (LGM), a time when wind strength and, in concordance, oceanic currents in the Southern Ocean may have been stronger (Petit *et al.* 1981; de Angelis *et al.* 1987); this would have caused a more intense throughflow between the Pacific and Indian Oceans.

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Recent interest in reconstructing the WPWP during the LGM include the works of Thunell *et al.* (1994) and Martinez *et al.* (1997, 1999), who compiled previous palaeoceanographic studies in the western Pacific (e.g. CLIMAP 1981; Thompson and Shackleton 1980; Wells and Wells 1994; Martinez 1994b; Miao *et al.* 1994).

The present paper is a review of published data up to 1997 when this paper was accepted for publication in this special issue [some updated references have been added herein] and an examination of the palaeoceanographic and palaeoclimatic conditions of the Indonesian throughflow under a LGM scenario when 1) sea level was lowered by 125 ± 4 m (e.g. Yokoyama *et al.* 2000, 2001), 2) the Indonesian passageway was reduced in extent, 3) precipitation in the region was reduced by ~30% (e.g. Flenley 1979; van der Kaars and Dam 1995), 4) sea-surface temperatures were lower by ~2°C (e.g. CLIMAP 1981; Thunell *et al.* 1994), and 5) the monsoons were less intense in August and more intense in February (e.g. Duplessy 1982; Sirocko *et al.* 1996). This chapter largely relies on published information, in addition to new palaeoceanographic data from the eastern Indian Ocean. As with oceanographic studies for the present, attention will be given to the sources of the throughflow water, its physical parameters (salinity and temperature), its paths, and a discussion of its possible effects on global climate.

Oceanography of the Western Pacific Warm Pool

The Indonesian Throughflow

The Indonesian region is characterised by at least eight deep basins (and seas with water depths generally reaching more than 4500 m) connected through relatively shallow sills (Tomczak and Godfrey 1994).

The high atmospheric and ocean temperatures in the region induce evaporation of oceanic water, formation of low pressure cells and precipitation; the overall balance is a gain of freshwater from precipitation. Consequently, SSTs are high and SSSs relatively low (Wyrtki 1961). As indicated above, waters with SSTs exceeding 28°C are referred to as the WPWP and occur in response to the west-flowing equatorial currents (e.g. Tomczak and Godfrey 1994; Gordon and Fine 1996) (Figure 9.1).

During the Southern Hemisphere winter (August), trade winds blow over Australia's northwest region from the southeast; during the Southern Hemisphere summer (February), winds take the opposite direction. Maximum throughflow occurs in August as a result of a lowering of sea level south of Indonesia in response to the strong SE trade winds. Consequently, water moves from the Pacific Ocean to the Indian Ocean as a response to a higher steric height in the former region (e.g. Godfrey and Golding 1981; Godfrey and Ridgway 1985). Estimates of the volume of water transported from the Pacific to the Indian Ocean varies between 2 Sv and 24 Sv (1 Sverdrup, Sv = 10^6 m^3/s) (Tomczak and Godfrey 1994) or more conservatively, 7 Sv to 18.6 Sv (Murray and Arief 1988; Godfrey 1996). Recent estimates of ~19 Sv in August and ~3 Sv in February evidence the wide range of monsoonal variation of the

throughflow and the difficulty of the problem of measuring it (Godfrey 1996; but see also Wiffels *et al.* 1996 and McBride 1998).

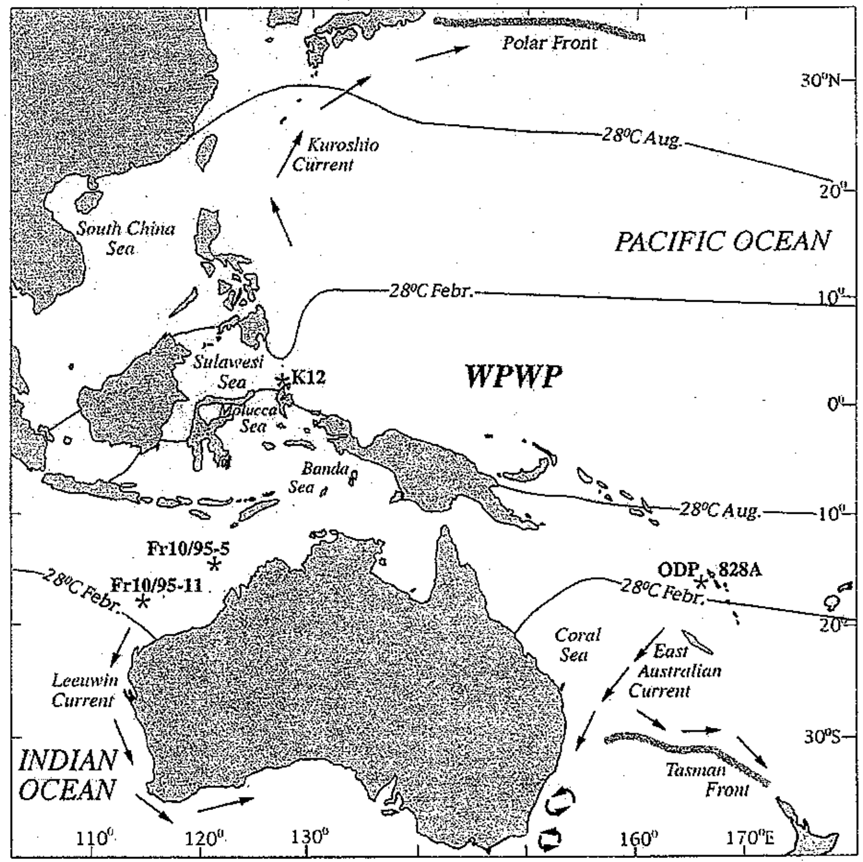

Fig. 9.1 Map of the western Pacific showing the location of cores (stars) and the limits of the Western Pacific Warm Pool (WPWP; the 28°C isotherm) in August and February (from Levitus et al. 1994). The Kuroshio, Leeuwin and East Australian Currents are indicated as outflows of the WPWP.

The WPWP and the atmospheric pressure system are coupled phenomena that operate as a west-east oscillator causing El Niño events to occur every 2 to 5 years (e.g. Enfield 1989; Clarke 1991; Yan *et al.* 1992). Water that passes the Indonesian throughflow eventually spreads into the Indian Ocean as the west-flowing South Java and the South Equatorial Currents, and as the south-flowing Leeuwin Current (e.g. Godfrey and Ridgway 1985; Wijffels *et al.* 1996).

As early as 1961, Wyrtki noticed the sharp contrast between the WPWP surface mixed-layer characterised by SST >28°C, SSS <34‰ and density st <22.0, and by intermediate water masses being colder, saltier and with a st =27.7. Its boundary is a strong discontinuity layer located between 120 and 160 m water depth, as indicated by a sharp density variation; temperature (i.e. the 25°C isothermal) is a good proxy of the density gradient because salinity changes little in the sub-surface (Wyrtki 1961). Localised seasonal upwelling areas in the WPWP are the only places where the surface layer is disrupted (Wyrtki 1961).

The uppermost 1000 m of the Indonesian throughflow are occupied by a modified Pacific Central Water mass whose vertical salinity profile progressively becomes uniform prior to reaching the Indian Ocean. Subsequently, the latter mixes with Indian Central Water to restore this gradient (Tomczak and Godfrey 1994). One possibility is that, in the Indonesian throughflow, turbulent mixing occurs at sill depths, thus causing a vertical homogenisation of salinity, but not of water temperature which retains its gradient (Tomczak and Godfrey 1994). Another explanation would be that most of the flow passing through the upper thermocline of the Lombok Strait consists of low-salinity, well oxygenated water of North Pacific origin (Gordon and Fine 1996). In contrast, only a small amount of high salinity water may pass through the lower thermocline of the east Indonesian throughflow as export flux from the South Pacific (Gordon and Fine 1996). Most of the Indonesian throughflow is assumed to contain water of low salinity, North Pacific origin (the North Equatorial Current) and moving from Mindanao through Makassar Strait and then towards the Lombok, Savu and Timor Straits into the Indian Ocean (Gordon and Fine 1996). Conversely, the more saline South Pacific water (the South Equatorial Current) is mostly recycled in the Halmahera Eddy and returns to the Pacific (e.g. Gordon 1986; Gordon and Fine 1996). No claim for a turbulent mixing mechanism is made in the latter interpretation to explain uniform salinity in sub-thermocline waters. It should be noted that the WPWP displays strong seasonal SSS variations that can reach 2‰ in the Banda and Arafura Seas (Wyrtki 1961).

Upwelling occurs in the Banda and Arafura Seas during August, when the SE monsoon forces the Monsoon Current through the Java and Flores Seas. As the volume of water entering the Indonesian region from the Pacific Ocean through the Halmahera Sea is insufficient to compensate the volume of water transported to the west (and to the south through the Timor Sea), upwelling occurs thus bringing cold and salty sub-surface water to the sea-surface (the Subtropical Lower Water located at 125-150 m; Wyrtki 1958). In February, the opposite mechanism occurs; downwelling is induced by the NW monsoon that brings an excess of water to the Banda and Arafura Seas. At ~1000 m depth, the excess of water in the Banda Sea finds its way into the Indian Ocean (Wyrtki 1958).

Upwelling south of Java occurs during August when the throughflow is at its maximum. It is expected that SSTs would be low (Wyrtki 1962). However, the reduced steric heights in the region then cause the throughflow to increase and warm water to flood into the eastern Indian Ocean (Godfrey 1996).

Outflows of the Western Pacific Warm Pool (WPWP)

As indicated above, difficulties still exist to precisely calculate the Indonesian throughflow partly due to the fact that part of the WPWP is recycled into the North Pacific through the north-flowing Kuroshio Current and into the South Pacific through the south-flowing East Australian Current. Both the Kuroshio and the East Australian Currents are fed by the North and South Equatorial Currents and, presumably, would be stronger if the Indonesian throughflow were closed (refer to modelling results by Godfrey and Golding 1981; Hirst and Godfrey 1993). A third important outflow of the WPWP comprises the South Equatorial and Leeuwin Currents in the Indian Ocean. Consequently, under a LGM scenario when the Indonesian throughflow could have been highly restricted due to a sea-level drop of 125 ± 4 m (e.g. Yokoyama *et al.* 2000, 2001), all possible outflows of the WPWP (and their associated seas) should be considered; namely, the NW Pacific and the South China Sea, the SW Pacific and the Coral Sea, and the eastern Indian Ocean.

The South China Sea has a maximum depth of ~4300 m and connects the northwest extreme of the Indonesian throughflow with the North Pacific. Ocean currents in this region respond to the monsoon forcing with a general circulation that is clockwise during the SW monsoon in August, and anti-clockwise during the NE monsoon in February when some water is supplied to the Kuroshio Current. Salinities in the South China Sea vary seasonally in response to monsoonal rains, i.e. lower SSS in the eastern South China Sea occur during the SW monsoon (Tomczak and Godfrey 1994).

The Coral Sea is an open sea receiving the influx of the South Pacific Equatorial Current that is affected by seasonal variations and gets stronger during the Southern Hemisphere winter (when Trade Winds are stronger). At ~18°S, the South Equatorial Current diverges: part of it travels to the north into the Solomon Sea to ultimately feed the North Equatorial Counter Current or the Kuroshio Current – and part of it travels southward to feed the East Australian Current that eventually deviates to the east at the Tasman Front (at ~30°S), or ultimately disappears as warm-water eddies in the southern portion of the Tasman Sea (Tomczak and Godfrey 1994).

In the eastern Indian Ocean, the Leeuwin Current is a narrow eastern boundary current that runs along the Western Australian margin to eventually reach the Great Australian Bight in the Southern Ocean transporting up to ~5 Sv of water (e.g. Tomczak and Godfrey 1994). The action of the Trade Winds on the westward flowing South Equatorial Current (in the Indian Ocean) triggers a southward Ekman transport whose cooling at high latitudes results in a lower steric height in the eastern Indian Ocean and causes the Leeuwin Current to flow southward (e.g. Cresswell and Golding 1980; Godfrey 1996). In a LGM scenario, a reduced throughflow, and presumably, a less intense Leeuwin Current would imply that the Ekman transport was less efficient in lowering the steric height of the eastern Indian Ocean. Recent work on modern-day clay distribution on the sea floor in the eastern Indian Ocean by Gingele *et al.* (2001a), in contrast with LGM clay distribution in the same area (Gingele *et al.* 2001b), confirm that the Leeuwin Current, if at all existent at the LGM, would have taken the direction of the central Indian Ocean.

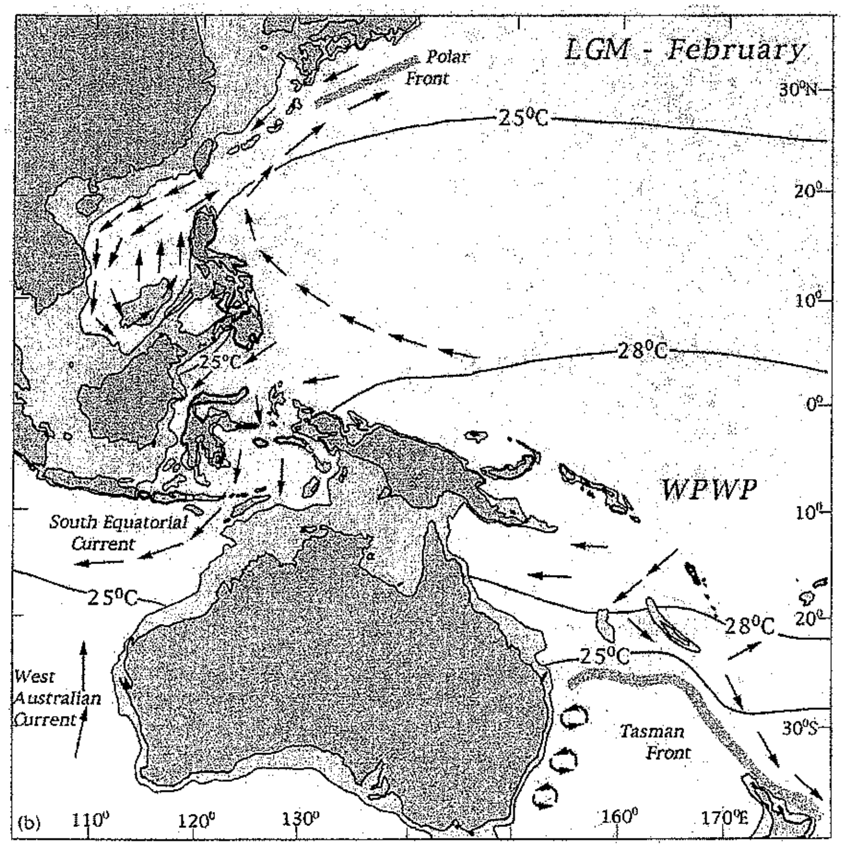

Fig. 9.2 Palaeoceanographical representation of the WPWP during the LGM for the austral a) summer (February), and b) winter (August). The 28°C and 25°C SST isotherms are from CLIMAP (1981), Anderson et al. (1989), Miao et al. (1994), Wang et al. (1995) and Barrows et al. (1996). Reconstruction of the South China Sea relies on information obtained in Wang et al., (1995), the Polar Front from Thompson and Shackleton (1980) and Oda and Takemoto (1992), the Tasman Front from Martinez (1994b), and the Leeuwin Current from Wells and Wells (1994).

Table 9.1. Core locations. Selected sites in the WPWP.

Core	Latitude	Longitude	Water depth (m)
K12	02°41'20"N	127°44'10"E	3510
ODP 828A	15°47.34'S	166°17.04'E	3086.7
Fr10/95-5	14°00.55'S	121°01.58'E	2472
Fr10/95-11	17°38.57'S	114°59.93'E	2458

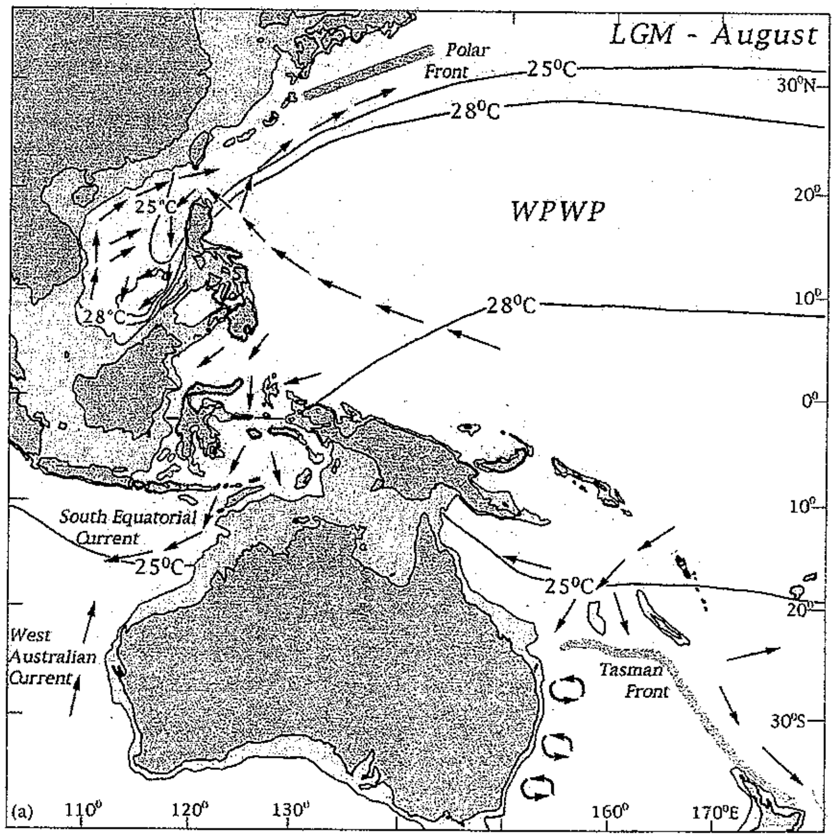

Fig. 9.2b

Materials and methods

As indicated above, much of this presentation relies on published information as well as on four sediment cores carefully selected as to represent the pattern of fluctuations in the WPWP. Planktonic foraminifera analyses from cores K12 from the Molucca Sea (Barmawidjaja *et al.* 1993), ODP 828A from the eastern Coral Sea (Martinez 1994a, 1994b), Fr10/95-11 and Fr10/95-5 from the eastern Indian Ocean (see Table 9.1 and Figure 9.1 for core locations) are used to derive sea-surface temperatures (SST) for the last ~30,000 years. The modern analogue technique (MAT) is applied herein to reconstruct SSTs and is regarded as more reliable than the transfer function method used in previous studies (e.g. Prell 1985; Thunell *et al.* 1994; Ortiz and Mix 1997). The MAT world data base (Prell

1985) was modified by, 1) adding a new foraminiferal data set of core-tops recently collected in the eastern Indian Ocean (Martinez *et al.* 1998) and from around Australia (Barrows and Hunt 1996) and, 2) removing data points from the North Pacific and North Atlantic Oceans as they are not regarded as analogous to conditions in the Southern Hemisphere. This modified MAT world database provides better modern analogues for the Indonesian region. We also selected *N. dutertrei*, a characteristic intermediate to deep-dwelling species (Bé 1977; Hemleben *et al.* 1989), in order to determine the influence of intermediate waters (and the depth of the deep chlorophyll maximum, DCM) as well as deciphering the possible bias in estimating SSTs when using the MAT.

Results

Oceanographic parameters in the WPWP during the LGM: A brief review

Sea-surface temperature patterns

The reduction in extent of the WPWP during the LGM was already implied by the CLIMAP (1981) results that show a 28°C SST isotherm restricted to the western Pacific in August (Southern Hemisphere winter) and split in two areas in February (Southern Hemisphere summer) (Figure 9.2). The CLIMAP (Climate/Long-Range Investigations, Mapping and Prediction 1981) results also suggested: 1) a reduced change in SST in the WPWP area, 2) a stronger seasonality in the WPWP area, 3) the absence of the WPWP in the South China Sea, the eastern Indian Ocean, and most of the Coral Sea, 4) a warm-water tongue in the Coral Sea apparently representing a more dynamic East Australian Current, 5) the equatorward deflection of isotherms in the NW Pacific Ocean, and 6) the equatorward deflection of isotherms in the eastern Indian Ocean. Anderson *et al.* (1989) confirmed some of these results and showed no evidence of a more dynamic East Australian Current, but the northern displacement of the 18° to 22°C isotherms (see Figure 9.2). The northward displacement of cool subtropical waters, i.e. at the Tasman Front, was confirmed by factor analyses on assemblages of planktonic foraminifera (Martinez 1994b), benthic foraminifera (Kawagata 2001) and recent SST reconstructions in the Tasman Sea (Barrows and Hunt 1996).

Similarly, Miao *et al.* (1994) and Wang *et al.* (1995) showed that the 28°C isotherm was restricted to the South China Sea in August (see Figure 9.2). The equatorward displacement of isotherms in the NW Pacific Ocean, implying an eastward deflection of the Kuroshio Current by the southern deflection of the Oyashiro Current (the southward migration of the Polar Front), was also suggested (e.g. Thompson and Shackleton 1980; Ujiié *et al.* 1991; Oda and Takemoto 1992).

Finally, the results of Wells and Wells (1994) and Barrows *et al.* (1996) seem to agree with the CLIMAP SST reconstructions for the eastern Indian Ocean, thus supporting the idea of van Andel *et al.* (1967) that the West Australian Current was stronger during the LGM (see Figure 9.2). Van Andel *et al.* (1967) also suggested that SSTs at the Timor Sea region were considerably cooler due to the blockage of warm currents through Torres Strait, although recent work by

Spooner (2001) clearly demonstrates that there was little SST change in the Banda Sea east of Timor during the LGM. The reduced SSTs in the eastern Indian Ocean and the lowering of sea-level in northern Australia caused the disappearance of the Gulf of Carpenteria as a major cyclogenic area (Webster and Streten 1978). Because tropical cyclones today originate in oceanic areas where SST exceeds 28°C (i.e. within the WPWP), then important climatic consequences are expected from the exclusion of cyclones in the eastern Indian Ocean that would therefore face a reduced rainfall (Webster and Streten 1978).

Beside microfossil analyses for the reconstruction of sea-surface conditions in the western Pacific, Sr/Ca in corals (e.g. Beck *et al.* 1997), the U_{37}^{k} technique (Lyle *et al.* 1992), and oxygen-isotope analyses of planktonic foraminifera (Broecker 1986) have been used. Unfortunately, corals studied so far have been collected at the margins or outside the WPWP and Sr/Ca ratios have not been measured for the LGM (Beck *et al.* 1997). The oxygen-isotope records of planktonic foraminifera for the western Pacific appeared to support CLIMAP results (Broecker 1986). However, those records came from low-sedimentation rate areas where the glacial-interglacial $\delta^{18}O$ difference is probably blurred by bioturbation (Broecker 1986; Martinez *et al.* 1997). U_{37}^{k} analyses for the central Pacific and the Indian Ocean tend to confirm the CLIMAP results (Lyle *et al.* 1992; Rostek *et al.* 1993; Sonzogni *et al.* 1998).

Overall, the WPWP (between 20°S and 20°N) did register a reduced drop in SST (<1.5°C) as confirmed by Thunell *et al.* (1994) who used the modern analogue technique (MAT) on planktonic foraminifera assemblages. The reliability of the MAT and the transfer function method to infer past SSTs will be discussed below with new data available from the WPWP region.

Sea-surface salinity changes

Sea-surface salinities in the Bonaparte Depression (NW Australia) during the LGM were interpreted to be similar to today's based on palaeontological and sedimentological inferences (van Andel *et al.* 1967). However, van Andel *et al.* (1967) also noted that in order to keep SSS close to normal in the Bonaparte Depression, river runoff had to be substantially lower; similarly, the occurrence of caliche nodules in the Sahul Rise was interpreted as an indicator of more arid conditions in Australia.

On the basis of a higher relative abundance of *N. dutertrei* during the LGM, it was suggested that SSSs were reduced in the Sulu Sea (Linsley *et al.* 1985), and increased in the Molucca Sea (Barmawidjaja *et al.* 1993). The latter researchers argued that the presence of *N. dutertrei* in both region should be regarded as indicative of a 'deep' chlorophyll maximum closer to the sea-surface, the presence of upwelling, and possibly saltier conditions (by as much as 1.9‰) all over the western Pacific. Barmawidjaja *et al.* (1993) also found an increase in the content of *Neogloboquadrina pachyderma* right-coiling form (=labelled *N. pachyderma* R herein) in the Molucca Sea that they postulate to indicate lower SSTs.

From a recent compilation done on deep-sea cores from the WPWP, large planktonic $\delta^{18}O$ differences were interpreted to represent a general increase of

SSS during the LGM (Martinez *et al.* 1997). This interpretation relies on the assumption that SST reconstructions, by applying the MAT, are reliable. Sea-surface salinity was found to increase by ~1‰ at the centre of the WPWP and probably to a higher value south of ~12°S and north of ~8°N (Figure 9.3), thus implying that evaporation minus precipitation was higher in the region for the LGM. To interpret this phenomenon, two approaches were then followed: (1) the first one consists of using *a residual approach* where the glacial-interglacial change in SSS resulted from the difference of the LGM and Holocene $\delta^{18}O$ signals in planktonic foraminifera, after a correction for the ice-volume effect (~1.2‰; Fairbanks 1989) and SST variations (Thunell *et al.* 1994), and (2) the second one being *a mathematical approach*, where the SSS difference is calculated as:

$$\Delta SSS = SSS_{LGM} - SSS_{Recent} = (\Delta\delta^{18}O - a - b\Delta SST)/c$$

where SSS_{LGM} is the local salinity during the LGM, and SSS_{Recent} represent today's local salinity. $\Delta\delta^{18}O$ is the difference between the LGM and the modern values, a is the ice-volume effect, ΔT is the SST difference between the LGM and the Recent, b is the slope of the $\delta^{18}O$ versus temperature relationship, and c is the slope of the $\delta^{18}O$ versus salinity relationship (Broecker 1989; Rostek *et al.* 1993). Sea-surface salinity during the LGM (SSS_{LGM}) was derived by assuming an ice-volume effect of 1.2‰, the slope of the $\delta^{18}O$ versus temperature relationship being -0.23‰/°C for *Globigerinoides sacculifer* and -0.2‰/°C for *Globigerinoides ruber* (Duplessy *et al.* 1981), and the slope of the $\delta^{18}O$ versus salinity relationship being 0.5$\delta^{18}O$‰/S‰ (Broecker 1989) (Figure 9.3 and Martinez *et al.* 1997). The SST difference between the LGM and the Recent along a latitudinal transect in the region was taken from MAT results on planktonic foraminiferal assemblages (Thunell *et al.* 1994).

As mentioned before, both approaches seem to indicate that SSS during the LGM was higher by ~1‰ at the centre of the WPWP, implying that evaporation minus precipitation was higher with important consequences for the coupled ocean-atmosphere system, the fresh-water balance and global circulation.

Sea-surface currents and the wind field

As indicated above, physical oceanographers have found that the Indonesian throughflow responds to the dynamics of currents and wind fields around Australia and most particularly in the Southern Ocean. CLIMAP's (1981) results indicate that, on a global scale, the wind field and oceanic currents were more dynamic during the LGM than today, whereas the overall albedo was larger due to drier conditions on land and increased polar ice cap. In fact, the wind strength in the Southern Ocean during the LGM has been suggested to have been stronger, based on the presence of aeolian dust (Petit *et al.* 1981) and chloride concentrations in ice cores (de Angelis *et al.* 1987). This view has been supported by others (e.g. Sarnthein *et al.* 1981; Rea 1994) who argued that the increase upwelling in equatorial regions in the Pacific and Atlantic Oceans was due to a stronger Southern Ocean circulation induced by a stronger wind field.

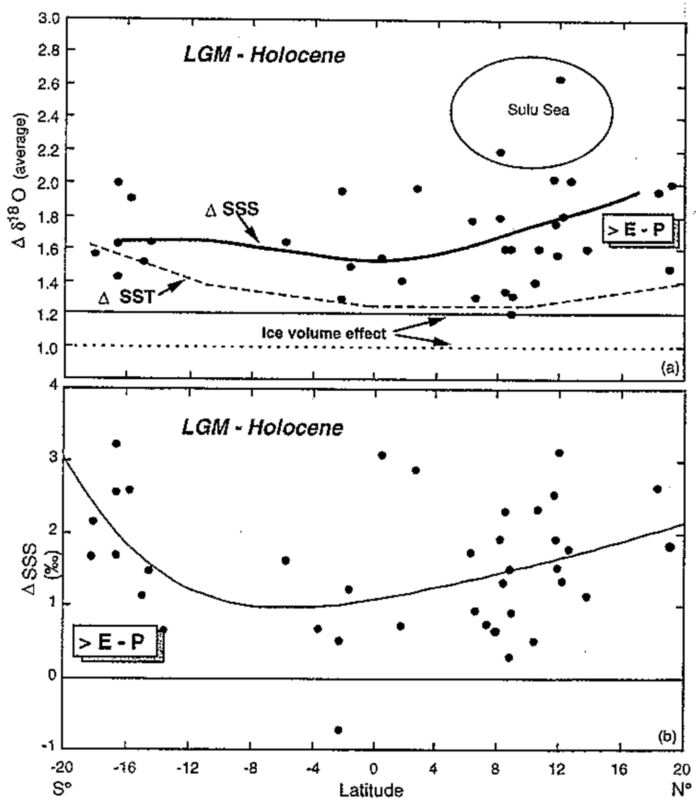

Fig. 9.3 SSS changes in the WPWP. Residual approach a), and mathematical approach b). In a), glacial-interglacial D$\delta^{18}O$ values are plotted against latitude; residuals attributable to salinity anomalies (DSSS) are derived after applying a $\delta^{18}O$ 1.2‰ ice-volume effect (Fairbanks 1989), and correcting for SST changes (Thunell et al. 1994). In b), the LGM and Holocene SSS difference (DSSS) is plotted against latitude. Note that SSS and evaporation minus precipitation (>E-P) was higher in all the WPWP during the LGM. Correlation lines were obtained as polynomial fits of order 5. A $\delta^{18}O$ 1.0‰ ice-volume effect as suggested by Schrag et al. (1996) is also indicated by a horizontal dashed line in a). For further explanations see text and Martinez et al. (1997).

It is estimated that the wind increased by a factor of 30% to 70% (Crowley and Parkinson 1988). In contrast to the idea of a stronger wind field in the Southern Ocean, and linked to the northern shift of the Antarctic Polar Front, Bareille *et al.* (1994) showed that aridity and the larger exposure of land would be sufficient to explain the dust peaks in ice-cores and deep-sea cores. Furthermore, there is no definitive explanation to the northern shift of the Antarctic Polar

Front, as modelling results show that increasing the strength of the wind by a factor of 0.5 to 2.0 would increase the speed of the water flow without changing the position of the Front. In contrast, by shifting meridionally the present wind field, a change in the strength of the current and the position of the Front (Klinck and Smith 1993) would occur. The dust argument can easily be challenged in contrast with the one favouring chloride input. As pointed out by Crowley and North (1991), the concentration of sea salt in the atmosphere is proportional to the strength of the wind and, consequently, higher values in Antarctic ice-cores can only be explained by a wind increase of 20% to 50% or perhaps even more.

Apart from the study of Wang and Li (1995) in the South China Sea, there is no other attempt to quantitatively model oceanic currents in the WPWP for the LGM. Sea-surface currents in the South China Sea were inferred to follow a similar pattern to today's; i.e. a clockwise circulation in August, and an anti-clockwise circulation in February (see Figure 9.2). However, the southward displacement of the Polar Front in the North Pacific (Thompson and Shackleton 1980; Oda and Takemoto 1992), and the northward displacement of the Tasman Front (Martinez 1994b) were shown to imply a weakening of both the Kuroshio and East Australian Currents respectively during the LGM. As mentioned above, the same reasoning has been applied to the eastern Indian Ocean where the Leeuwin Current would have been weaker (van Andel *et al.* 1967; Wells and Wells 1994; McCorkle *et al.* 1994).

The ocean general circulation model (OGCM) prepared by Hirst and Godfrey (1993) suggests that because the Kuroshio and East Australian Currents are fed by the North and South Equatorial Currents, these would be presumably stronger if the Indonesian throughflow were closed. However, this model was created in an attempt to understand circulation patterns for the present Indonesian throughflow. Consequently, LGM boundary conditions, such as the reduction in SST in high latitudes, were not included in their model. In contrast, an ocean general circulation model (OGCM) for the LGM by Lautenschlager *et al.* (1992) suggests that the strength of the Kuroshio Current and East Australian Current would have decreased during the LGM, thus supporting the Thompson and Shackleton (1980), and Martinez (1994b) results. However, even though the Tasman Front was displaced northward from ~30 to ~26°S during the LGM, tropical planktonic foraminifera species (such as *G. ruber* and *G. sacculifer*) were found east of Tasmania during the LGM (refer to Factor 1 of Martinez 1994b). The presence of the foraminiferal assemblage represented by Factor 1 almost 20° south of their normal habitat, suggests a transport by warm-water eddies such as occurs today. Another possibility is that this fauna may indicate that the East Australian Current was still operating during the LGM and that not all of its volume was diverted towards the central Pacific along the Tasman Front.

SST estimates from sediment core material

Core K12 from the Molucca Sea

Our SST estimates for Core K12, a high-sedimentation rate core from the Molucca Sea (Barmawidjaja *et al.* 1993), are presented in Figure 9.4. The latter

authors did not carry out SST reconstructions in this core. Temperatures in August (Northern Hemisphere summer) are higher than 28°C during most of the Holocene and during two short intervals prior to the LGM. The lowest SSTs were reached during the LGM (~25.6°C) and during a short interval prior to the LGM. The SST pattern for February (Northern Hemisphere winter) is similar, but the 28°C value is only exceeded during brief intervals in the Holocene (Figure 9.4).

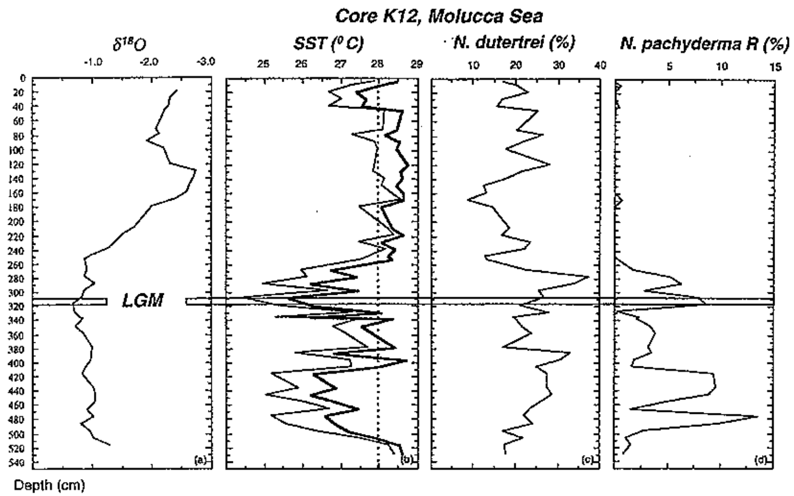

Fig. 9.4 Core K-12, Molucca Sea. a) $\delta^{18}O$ record smoothed to 3-point moving average, from Barmawidjaja et al. (1993); b) February SST (bold line) and August SST (plain line) MAT estimates; c) N. dutertrei percentage curve; d) N. pachyderma R percentage curve. Note the correspondence of the LGM with low SSTs and high percentages of N. dutertrei and N. pachyderma.

The pattern of SSTs in core K12 shows a similarity with the percentage abundance of *N. dutertrei* and *N. pachyderma* R, i.e. higher abundances of these species correspond to low SSTs, and conversely, lower abundances correspond to high SSTs. Noticeably, *N. pachyderma* R almost disappears at the end of the glacial stage (at ~250 cm) whereas *N. dutertrei* displays a declining trend from the LGM towards the Holocene reaching a minimum (~9%) at 170 cm, i.e. at the beginning of the Holocene.

In the MAT, as in the Transfer Function method, it is assumed that the distribution of foraminifera species are mainly controlled by SST. As pointed out by Barmawidjaja *et al.* (1993), the increased content of *N. dutertrei* and *N. pachyderma* R in core K12 may indicate the presence of a deep chlorophyll maximum within the photic zone. The effects of these species on SST estimates will be discussed below.

Core ODP 828A from the eastern Coral Sea

Core ODP 828A is a very-high sedimentation rate core from west of Vanuatu in the eastern Coral Sea (Martinez 1994a, b; Martinez *et al.* 1997; Figure 9.5). As with many Pacific deep-sea cores, carbonate dissolution is severe during interglacial intervals, whereas carbonate preservation is good during glacial intervals (Martinez 1994a). Consequently, SST estimates between 4 and 16 m (a preservation interval) in that core are considered herein as highly reliable; at this interval, SSTs in February are invariably above 28°C and, in many cases, above 29°C (Figure 9.5). In August, in contrast, SSTs show an average of ~27°C dropping occasionally to 25°C and rising to >28°C. Noticeably, SSTs in August show higher amplitude fluctuations than SSTs in February (Figure 9.5).

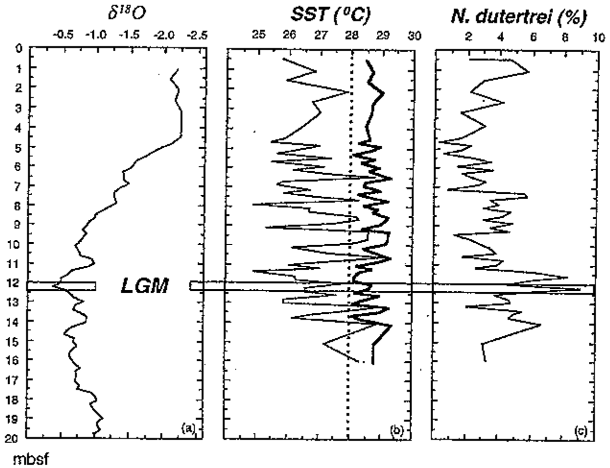

Fig. 9.5 Core ODP 828A, Vanuatu (eastern Coral Sea). a) $\delta^{18}O$ record smoothed to 3-point moving average, from Martinez et al. (1997); b) February SST (bold line) and August SST (plain line) MAT estimates; c) N. dutertrei percentage curve. Note the correspondence of the LGM with high percentages of N. dutertrei, but the lack of any conspicuous drop in SST.

The percentage abundance of *N. dutertrei* shows maximum values during the LGM, a declining trend towards the Holocene and a minimum value at the beginning of the Holocene (Figure 9.4). In contrast to core K12 from the Molucca Sea, the percentage abundance of *N. dutertrei* is low in core ODP 828A; in the latter, it is the percentage of other deep-dwelling species (Martinez 1994b) which engender a minimum bias in the estimation of SSTs using the MAT.

Cores Fr10/95-5 and Fr10/95-11 from the eastern Indian Ocean

February SST reconstructions on core Fr10/95-5 display a steadily declining trend, from temperatures >28°C in the late Holocene to ~26°C during the glaciation phase (Figure 9.6). There is no significant drop in SST during the LGM as it is observed between 30 and 45 cm depth interval that corresponds to a trough in the $\delta^{18}O$ record (heavy values). The 28°C value is exceeded only during the Holocene, particularly above the 20 cm depth interval; this interval also corresponds to maximum values (>1) for the *G. sacculifer/N. dutertrei* ratio which Martinez *et al.* (in press) suggested to be a good proxy of the WPWP boundary in the eastern Indian Ocean. August SSTs display a declining trend from ~25°C in the late Holocene to ~23°C during the glaciation phase, thus paralleling February SSTs (Figure 9.6).

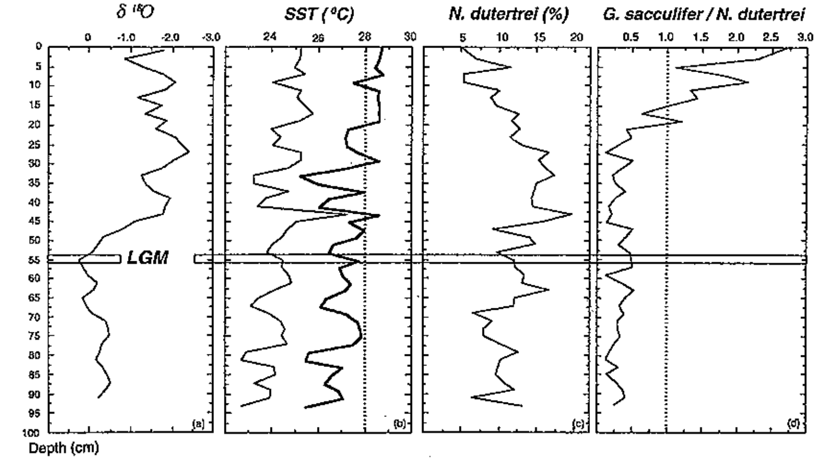

Fig. 9.6 Core Fr10/95-5, eastern Indian Ocean. a) $\delta^{18}O$ record smoothed to 3-point moving average, from a work in preparation; b) February SST (bold line) and August SST (plain line) MAT estimates; c) N. dutertrei percentage curve; d) G. sacculifer/N. dutertrei ratio. Note the correspondence of the LGM with intermediate percentages of N. dutertrei, but the lack of a conspicuous drop in SST. Note also that the 28°C value is exceeded during the late Holocene in correspondence to a >1 G. sacculifer/N. dutertrei ratio.

In contrast to the previous two cores (K12 and ODP 828A), maximum percentages of *N. dutertrei* do not coincide with the LGM, but to the beginning of the Holocene instead. A declining trend from a maximum of ~20% during the early Holocene to a minimum of ~5% during the late Holocene is also observed (Figure 9.6). During the LGM, values vary between ~10 and ~15%.

August and February SST reconstructions on core Fr10/95-11 parallel each other and display a maximum reduction of ~3°C during the LGM (Figure 9.7).

The 28°C value is exceeded all throughout the core, except during the glacial phase. As with core Fr10/95-5, the *G. sacculifer/N. dutertrei* ratio shows a good correspondence with warm intervals, i.e. values exceeding 1.0 which correspond to SSTs in excess of 28°C, whereas values <1.0 correspond to the glacial interval (see Figure 9.7).

As with cores K12 and ODP 828A, maximum percentages of *N. dutertrei* show a maximum value close to the LGM, and a declining trend towards the beginning of the Holocene (Figure 9.7). Percentages of *N. dutertrei* in this core are comparable to those in core ODP 828A.

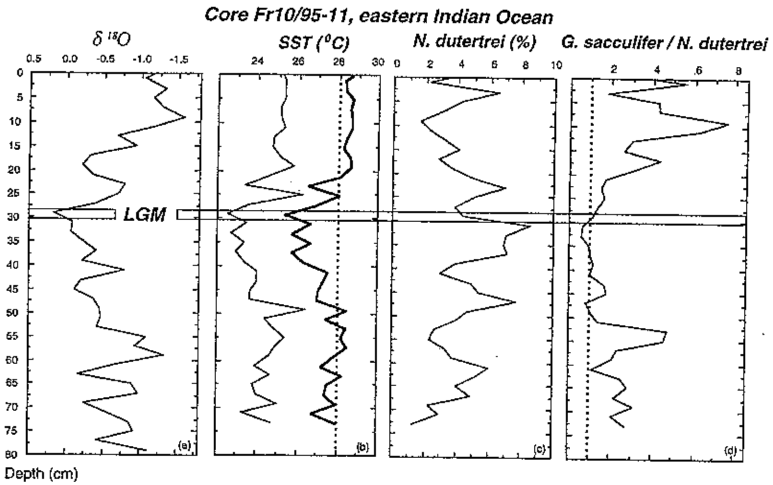

Fig. 9.7 Core Fr10/95-11, eastern Indian Ocean. a) $\delta^{18}O$ record smoothed to 3-point moving average, from a work in preparation; b) February SST (bold line) and August SST (plain line) MAT estimates; c) N. dutertrei percentage curve; d) G. sacculifer/N. dutertrei ratio. Note the correspondence of the LGM with high percentages of N. dutertrei,

Discussion

SST reconstructions and the percentage abundance of N. dutertrei

By comparing SST reconstructions for the LGM in cores K12, ODP 828A, Fr10/95-5 and Fr10/95-11 with maps for the same time-slice (Figures 9.2(a) and 9.2(b); reconstructed herein using a composite of previously published data for the LGM), a number of discrepancies appear. These are: 1) the 28°C isotherm in August was located north of the Molucca Sea, 2) the 25°C isotherm was located south of the Molucca Sea, and 3) the 25°C isotherm was located slightly northward in the eastern Indian Ocean in August. Our new results may imply that the WPWP was further reduced in extent in comparison to the reconstructions

presented in Figure 9.2. Similarly, the almost constant SSTs in the eastern Coral Sea (as represented by the ODP 828A record) imply that the region was enclosed within the WPWP for the last ~30,000 years in February and outside of it in August. This finding strongly contrasts with Beck *et al.*'s (1997) Sr/Ca SST reconstructions for a Vanuatu coral which suggest a 6.5°C reduction in temperature 10,000 years ago. Seasonality in the region, however, gives similar results using the two palaeothermometers, i.e. ~3°C.

Before discussing the implications of these results, we should examine the statistical methods, based on plankton assemblages, used to reconstruct SSTs. Reconstruction of sea-surface environmental parameters using the transfer function method and the MAT are based on similar assumptions, namely, that the ecological requirements of foraminiferal species have remained unchanged during the Pleistocene period and that the yearly average SST and SSS are directly related to ecological parameters of surface waters (Imbrie and Kipp 1971). The reliability of both methods has been tested by several authors (e.g. Molfino *et al.* 1982; Prell 1985; Le 1992; Ortiz and Mix 1997). Concordance in estimating SST for the past using different microfossil groups, i.e. foraminifera, coccolithophorids and radiolarians, was pointed out as a strength of the transfer function method (Molfino *et al.* 1982). However, Molfino *et al.* (1982) also showed that, despite the ability to discern water masses using groups of microfossils with diverse ecological requirements, in low latitudes major dissimilarities are found that may be due to other ecological factors more important than SST, such as nutrients and thermocline depth. The implied assumption in the transfer function and the MAT is that foraminiferal species (as well as radiolarians and coccolithophorids) record sea-surface conditions despite their transit through different water depths during their life-cycle or the almost exclusive intermediate to deep-water habitat of some species. It may be that, following our recent investigations on 54 core-tops from the eastern Indian Ocean (Martinez *et al.* in press), the distribution of planktonic foraminifera in that area is directly linked with SST and nutrients. Then, a change in the thermocline in the past may require a dropping in optimum SST requirements for some species. If that was the case, past SST reconstructions would have to be modified accordingly and this would explain the discrepancies obtained between SSTs and land temperature reconstructions using pollen assemblages. As indicated above, we wish to explore the influence of intermediate waters and its possible bias when estimating past SSTs, by selecting the percentage abundance of *N. dutertrei*, a characteristic intermediate to deep-dwelling species (Bé 1977; Hemleben *et al.* 1989).

Neogloboquadrina dutertrei is mainly a tropical-subtropical, herbivorous species considered to preferentially occur in upwelling regions (Bé 1977; Thiede 1975, 1983; Thunell *et al.* 1983; Hemleben *et al.* 1989). Based on glacial-interglacial $\delta^{18}O$ difference in *N. dutertrei*, Shackleton (1977) suggested that this species may have preferred warmer waters and 'followed' a particular density level in a saltier glacial ocean. However, this view has been changed by the observation that the depth of the chlorophyll maximum, rather than density, is the major controlling factor on the distribution of planktonic foraminifera (Fairbanks and Wiebe 1980; Ravelo *et al.* 1990). *N. dutertrei* has been found in abundance during the upwelling season (e.g. in the Panama Basin) when the thermocline is shallow and primary production is high (Thunell and Reynolds 1984), or during a post-upwelling stage (e.g. in the San Pedro basin, NE Pacific Ocean) when the upper water column becomes warmer (between 12° to 20°C) and stabilises, and with the chlorophyll maximum being strong, coinciding with abundant numbers of diatom species (Sautter and Thunell 1991). In his global distribution study, Hilbrecht (1996, 1997) shows that *N. dutertrei* exhibits narrower tolerances during summer, i.e. SST above 21°-22°C with an optimum at ~27°C, a temperature of ~15°C at 200 m, SSS of ~35‰ (but it can be found in waters with SSS between ~34‰ and ~37‰) and a surface density of ~22.5. Hilbrecht (1996) also points out that *N. dutertrei* increases when *Globigerina bulloides* decreases. The latter species is also regarded as typical of upwelling systems and shows preference for productive cold waters (e.g. Bé 1977; Hemleben *et al.* 1989). Thus, the exclusion of one of the two species in eutrophic regions may reflect a different availability of phytoplankton preys and the advantageous competence of one species over the other.

Our results corroborate the suggestion by Barmawidjaja *et al.* (1993) that the relative increase of *N. dutertrei* (and *N. pachyderma* R when present) during the LGM, accounts for most of the drop in SSTs in the WPWP (see Figures 4 and 7). In fact, high percentage abundances (>30%) of *N. dutertrei* are only recorded in core K12 from the Molucca Sea where SST dropped by ~3°C during the LGM; in other cores, the abundance of the species is low coinciding with a reduction in SST during the LGM. We also agree with the interpretation that the deep chlorophyll maximum was probably within the photic zone and nutrients were more abundant and close to the surface; seasonal upwelling was also more frequent and intense. However, we do emphasise that it may well be that the drop in SST is apparent and most probably assemblages of planktonic foraminifera reflect a shallower thermocline where SST remained almost the same as today. Similarly, we also consider the alternative possibility that even though the MAT accurately compares down-core assemblages with modern analogues, and those similar assemblages presumably can be translated in SST estimates, it may be that those tropical species whose optimum is close to the maximum SST of the present ocean, did live in slightly colder waters (perhaps by ~1° to 3°C) where nutrients were abundant and close to the sea-surface.

A shallower nutricline/thermocline (and stronger oceanic upwelling) during the LGM implies an increase in the strength of oceanic divergence along the equatorial Pacific. This can be easily explained by evidence of stronger wind fields around Australia during the LGM. However, the intensification of trade winds also implies a more efficient accumulation of warm water in the western Pacific and a deepening of the thermocline as happens today during non-El Niño years (e.g. Enfield 1989). Consequently, to solve this apparent contradiction, the El Niño events may have been presumably more frequent, thus favouring upwelling by the equatorial divergence of currents during the LGM.

SSS reconstructions and the pattern of sea-surface currents

In today's Indonesian throughflow, most of the surface water is of low-salinity North Pacific origin rather than of a high-salinity South Pacific origin (e.g.

Wyrtki 1961; Gordon and Fine 1996; Godfrey 1996). Under a saltier LGM scenario (SSS higher by ~1‰ at the centre of the WPWP), we suggest that the source of Indonesian throughflow water would mainly have been of South Pacific origin. Today, a salinity maximum (35.2‰ to 35.5‰) is found at ~100 m to 150 m water depth and is referred to as the Subtropical Lower Water originating at mid latitudes (Wyrtki 1961). Also this water mass becomes diluted by the abundant fresh water precipitating in the Indonesian region (Gordon and Fine 1996). Under an increased evaporation minus precipitation balance in the region (Martinez *et al.* 1997), salty South Pacific water would not be significantly diluted and could occur closer to the sea-surface, therefore explaining not only the regional increase in SSS, but also the shallow deep chlorophyll maximum as suggested by Barmawidjaja *et al.* (1993). It may also be that the mixed layer of the ocean during the LGM was further enriched in nutrients supplied by river runoff from the extensively exposed continental shelves of the region. Iron – a biolimiting element in many regions of the ocean (e.g. Martin 1990) – may have been more efficiently supplied to the sea-surface by the intense wind field promoting primary productivity and increasing export production of organic matter out of the euphotic zone (e.g. Herguera 1994; Kawahata and Suzuki 1994, 1995). Already, recent work by Hesse (1997) implies a greater supply of aeolian dust during the LGM at least for the Exmouth Plateau. However, more recently Hesse and McTainsh (1999) postulated that mid-latitude transport of dust during the LGM was no more prominent than during the Holocene • at least in the Tasman Sea • but that wind strength was more predominant in high latitudes. Further, Gingele *et al.* (2001b) confirm that winds at low latitudes north of Australia were quite different from today's arrangement. Nevertheless, the relative decrease in the percentage abundance of *N. dutertrei* in core Fr10/95-5 (Figure 9.6) may indicate that upwelling conditions close to the Australian continent were significantly reduced as compared to other regions of the WPWP. Another interpretation would be that the nutricline was closer to the surface and the thermocline much steeper compared to today. In a saltier (and cooler?) glacial ocean, we would expect denser surface waters in the Indonesian throughflow moving, and perhaps, sinking at intermediate depths in the eastern Indian Ocean; this would be the glacial Leeuwin Current. It could be that the Leeuwin Current did not circumnavigate the West Australian coast but instead migrated westward to the centre of the Indian Ocean between latitude 10° and 20°S (as the South Equatorial Current?). Gingele *et al.* (2001b) agree to this proposed scenario.

Conclusions

A review of the palaeoceanography of the western Pacific Warm Pool (WPWP), focusing on sea-surface temperature (SST) and salinity (SSS) reconstructions and the pattern of currents, was presented. New SST reconstructions, by using the modern analogue technique (MAT), are shown for cores K12 (Molucca Sea), ODP 828A (Vanuatu, eastern Coral Sea), and Fr10/95-5 and Fr10/95-11 (eastern Indian Ocean). The WPWP, a seasonally and inter-annual changing feature of the ocean, was saltier (by ~1‰) and highly reduced in extent during the Last

Glacial Maximum (LGM). However, the Indonesian throughflow may have been stronger in response to stronger wind fields in the Southern Ocean and around, as well as north, of Australia. The reduction in extent of the WPWP was accompanied by the equatorward migration of the Tasman Front in the South Pacific, and the Polar Front in the North Pacific. The Kuroshio and East Australian Currents may still have been strong at that time. The Leeuwin Current may have been saltier and colder and still in operation during the LGM, but its overall path different. Reduced SSTs during the LGM may be a bias of the increase abundance of the intermediate to deep-dwelling planktonic foraminiferal species *Neogloboquadrina dutertrei* and *N. pachyderma* right-coiling. We support a previously proposed idea that during the LGM, the deep chlorophyll maximum was within the photic zone, the thermocline gradient steeper and primary productivity was higher in the WPWP. Under a LGM scenario, when wind fields were stronger, oceanic divergence may have been more efficient.

Tropical planktonic foraminifera may have lived in a slightly colder WPWP thus biasing SST reconstructions (by ~1°to 3°C ?).

Epilogue: Significance of palaeoceanographic work to modern conditions in the Maritime Continent north of Australia

We have demonstrated through the palaeoceanographic investigations of some selected cores in the WPWP that this area is subject to some long-term, very dynamic changes. The evaporation-precipitation balance did change, and so did the nutricline/thermocline depth. The position of the latter two may have resulted from a different wind strength as postulated for the LGM. The glacial ocean, even in the WPWP, would have been different even though high SSTs were maintained close to 28°C.

The WPWP is not only a dynamic region, but it is also climatically a very significant region of the globe. It is where high values of heat and moisture transfer occur between the ocean and atmosphere. This area is also directly linked with the eastern Pacific through ENSO signals.

Our preliminary study of deep-sea cores provides a basis for future work on other aspects linked to climate change in the WPWP. These include: 1) vegetational changes and ecological successions (to be detected through pollen examination in the cores; see van der Kaars and De Deckker submitted); 2) fire intensity and frequency (through the analysis of charcoal particles in the cores); 3) mineralogical changes (through compositional changes of sediment; see Gingele *et al.* 2001a, 2001b) - some of which may be related to erosional changes on land (some caused by human activity?); and 4) aeolian transport and volcanic dust. The evidence of all these can be obtained using the same cores, and therefore would provide a link between oceanic, continental as well as atmospheric phenomena.

Future study of the physicochemical phenomena linked with the palaeoceanography of the WPWP promises to be very exciting indeed, and ought to help better understanding the interplay of many different components, including human-related ones, in what is best called the Maritime Continent north of Australia.

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