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Last deglacial sea-surface temperature evolution in the Southeast Pacific compared to climate changes on the South American continent

Jung-Hyun Kim*, Ralph R. Schneider, Dierk Hebbeln, Peter J. Müller, Gerold Wefer

Fachbereich Geowissenschaften, Universität Bremen, Klagenfurter Straße, 28359 Bremen, Germany

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Abstract

Applying the alkenone method, we estimated sea-surface temperatures (SSTs) for the past 33 kyr in two marine sediment cores recovered from the continental slope off mid-latitude Chile. The SST record shows an increase of 6.7°C from the last ice age (LIA) to the Holocene climatic optimum, while the temperature contrast between LIA and modern temperatures is only about 3.4°C. The timing and magnitude of the last deglacial warming in the ocean correspond to those observed in South American continental records. According to our SST record, the existence of a Younger Dryas equivalent cooling in the Southeast Pacific is much more uncertain than for the continental climate changes. A warming step of about 2.5°C observed between 8 and 7.5 cal kyr BP may have been linked to the early to mid-Holocene climatic transition (8.2–7.8 cal kyr BP), also described from equatorial Africa and Antarctica. In principal, variations in the latitudinal position of the Southern Pacific Westerlies are considered to be responsible for SST changes in the Peru–Chile current off mid-latitude Chile. © 2002 Elsevier Science Ltd. All rights reserved.

1. Introduction

Climate changes during the transition from the Last Ice Age (LIA, 21 ± 2 cal kyr BP) (Mix et al., 2001) to the Holocene, the so-called Termination I, have been intensively studied in various regions of the world ocean (e.g., Broecker et al., 1988; Labracherie et al., 1989; Duplessy et al., 1992). Among such studies, the alkenone unsaturation ratio ($U_{37}^{K'}$ index) has been successfully used for reconstructing sea-surface temperature (SST) changes in the Atlantic (Zhao et al., 1995; Rühlemann et al., 1999; Bard et al., 2000), the Indian (e.g., Bard et al., 1997), and the Pacific Oceans (Prahl et al., 1995; Doose et al., 1997; Pelejero et al., 1999), and in the Mediterranean Sea (Cacho et al., 1999, 2000, 2001).

However, very little information exists on temperature changes in the Southeast Pacific during Termination I, in particular, the southern part of the Peru–Chile Current (PCC). Previous paleoenvironmental reconstructions in this region were based primarily on land studies using pollen (e.g., Heusser, 1990; Markgraf, 1993), glacier (e.g., Lowell et al., 1995), and snowline reconstructions (e.g., Clapperton, 1993). More recently, marine sediment cores GIK 17748-2 and GeoB 3302-1 recovered from the continental slope off Chile at ≈33°S provided first high-resolution records of both terrestrial and marine climate for this area based on elemental and clay mineral composition and foraminifera assemblages (Lamy et al., 1999; Marchant et al., 1999). The climatic information derived from these terrestrial records was mainly about precipitation changes induced by latitudinal shifts of the Southern Hemisphere westerly wind system. However, the influence of strong upwelling in this Pacific eastern boundary current system did not allow us to apply modern analog or transfer function techniques on foraminiferal assemblages down-core in order to estimate past SSTs. From other coastal upwelling areas, it is well known that past changes in the abundances of certain foraminiferal species are predominantly related to variations in thermocline structure and nutrient conditions and not directly to past temperature variability (e.g., Prell and Van Campo, 1986; Mix and Morey, 1996; Little et al., 1997). Under these conditions, foraminiferal transfer function and modern analog techniques can lead to erroneous temperature estimates (e.g., Cayre and Bard, 1999; Hendy and Kennett, 2000). We therefore consider

*Corresponding author. Tel.: +49-421-218-8929; fax: +49-421-218-3116.

E-mail address: jungkim@allgeo.uni-bremen.de (J.-H. Kim).

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alkenone thermometry as a more appropriate method for paleotemperature estimation in this region. For upwelling areas and their surrounding filamentous zones, it has been shown that alkenone unsaturation ratios from sediment cores yield reliable estimates for past annual mean SST values (Prahl et al., 1995; Müller et al., 1997; Cayre and Bard, 1999; Kirst et al., 1999; Müller and Fischer, 2001). Accordingly, here we present SSTs for the PCC system off Chile, based on temporal variations in the $U_{37}^{K'}$ index over the past 33 kyr. The data provide new information about the regional characteristics for the LIA to Holocene temperature contrast and the pattern of temperature changes during Termination 1 for this eastern boundary current system. Moreover, this new SST record allows detection of similarities and discrepancies in the transient behaviour between oceanic and terrestrial climates in the Southeast Pacific, on the South American continent and in other climate-sensitive regions, and thus improves our understanding of interhemispheric and interoceanic teleconnections associated with paleoclimatic changes at millennial time scales.

2. Modern conditions

Detailed overviews about the prevailing oceanic and atmospheric systems in the study area are given in previous studies (e.g., Shaffer et al., 1995; Strub et al., 1998). Here we briefly summarize the major features in the oceanic and atmospheric circulation system that may have induced SST changes in the past.

2.1. Atmospheric system

As illustrated in Fig. 1a, the study area lies at the northern limit of the Southern Hemisphere westerly wind belt. This boundary maintains its position permanently south of 38°S, providing high precipitation to the continent throughout the year (Miller, 1976). On the contrary, north of 34°S, the subtropical high-pressure system has a major influence on the climatic system, causing high aridity in this area (Miller, 1976). The equatorial subsurface water of the Gunther undercurrent (GUC) reaches the surface within coastal upwelling driven by southerly to southeasterly winds. These upwelling favourable winds occur in northern and central Chile north of 35°S throughout the year but are restricted to the austral summer between 35° and 38°S (Strub et al., 1998). South of 38°S, onshore-blowing westerly winds generally prevent coastal upwelling. Previous paleoceanographic studies imply that in the past, latitudinal displacements of the Southern Pacific Westerlies influenced the whole oceanographic system along the Chilean continental slope and continental paleoclimate (Heusser, 1989; Villagrán, 1993; Veit, 1996; Lamy et al., 1999).

2.2. Oceanic currents

As shown in Fig. 1b, the northward flowing PCC (Subantarctic surface water), branching from the Antarctic circumpolar current (ACC) at approximately 40–45°S (Boltovskoy, 1976) primarily controls the oceanographic setting of the study area. Off the mid-latitude Chilean coast, the PCC is separated into the oceanic Peru–Chile current (PCC_{ocean}) and the coastal Peru–Chile current (PCC_{coast}) by the poleward flowing Peru–Chile countercurrent (PCCC; subtropical surface water) (Strub et al., 1998). Beneath these surface currents close to the coast, the poleward directed GUC (equatorial subsurface water) is between 100 and 400 m over the continental slope and outer shelf (Strub et al., 1998).

3. Material and methods

3.1. Sample collections

We analysed five core top sediment samples taken during the CHIPAL cruise (SO-102) with R/V Sonne (Hebbeln et al., 1995) (Fig. 1, Table 1) to verify previous results from other upwelling areas that annual mean temperatures are documented in the sedimentary alkenone signal within the error limits of this method (Herbert et al., 1998; Müller et al., 1998). In addition, we analysed two sediment cores GIK 17748-2 and GeoB 3302-1 to reconstruct temporal variations in SSTs (Fig. 1, Table 1). Core GIK 17748-2 was recovered from the southern part of the Valparaiso basin on the continental slope at a water depth of 2545 m off Valparaiso, mid-latitude Chile (32°45′S; 72°02′W, core length 383 cm) during R/V Sonne cruise SO 80 (Stoffers et al., 1992). This core consists primarily of light brown–olive clayey silts and silty clays with three distinct turbiditic sandy layers (9–16, 47–54, and 160–177 cm) (Marchant et al., 1999). These turbiditic layers are assumed to be geologically "instantaneous" and were subtracted from the sedimentary sequence (Hebbeln et al., submitted). The core was sampled for alkenone analysis at depth intervals of 5 cm between 0 and 178 cm core depth and 10 cm between 178 and 383 cm core depth, where sedimentation rates are significantly higher than above. Core GeoB 3302-1 was taken ≈60 km further to the south on the continental slope at a water depth of 1498 m (33°13′S; 72°06′W, core length 412 cm) during R/V Sonne cruise SO 101 (Von Huene et al., 1995) and principally consists of olive–olive-brown clayey silts. This core was also sampled at intervals of 5 cm.

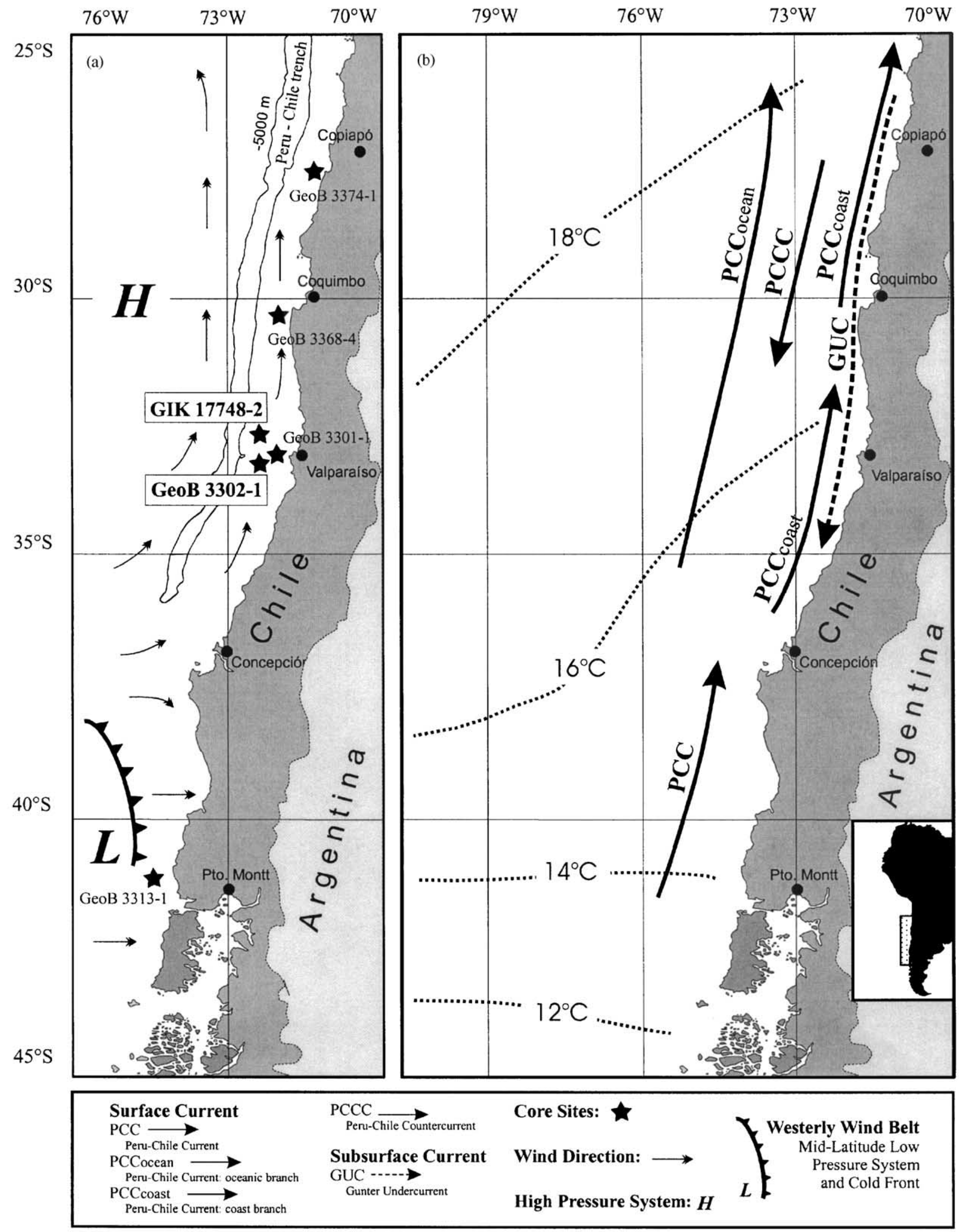

Fig. 1. Maps showing modern atmospheric and oceanographic conditions in the study area: (a) present position of the southern westerly wind system including a mid-latitude cyclone with its associated cold-front, and the subtropical high-pressure system, compiled after Strub et al. (1998), and the position of the investigated core-tops and cores, and (b) ocean surface and subsurface currents in the Southeast Pacific off mid-latitude Chile, compiled after Shaffer et al. (1995) and Strub et al. (1998), and annual mean SSTs at 0 m water depth (Levitus and Boyer, 1994).

Table 1

Core name	Gear type	Latitude	Longitude	Water depth (m)		
(a) Positions and water depths for sediment cores investigated in this study						
GIK 17748-2	Gravity corer	32°45′S	72°02′W	2545		
GeoB 3301-1	Multicorer	33°09′S	71°59′W	969		
GeoB 3302-1	Gravity corer	33°13′S	72°06′W	1498		
GeoB 3313-1	Gravity corer	41°00′S	74°27′W	852		
GeoB 3368-4	Multicorer	30°22′S	71°58′W	3240		
GeoB 3374-1	Multicorer	27°28′S	71°10′W	1352		
Core name	**Depth in core (cm)**	**Modern Annual mean SST (°C)**	**$C_{37:3}$ (ng/g)**	**$C_{37:2}$ (ng/g)**	**$U^{K'}_{37}$**	**SST[a] (°C)**
(b) Modern annual mean sea-surface temperatures (SSTs) at 0 m water depth from Levitus and Boyer (1994) *and SSTs derived from alkenone measurements on surface sediments*						
GIK 17748-2	3	15.9	557	749	0.574	16.1
GeoB 3301-1	0–1	15.7	590	850	0.590	16.5
GeoB 3313-1	0–1	14.0	1684	1735	0.507	14.0
GeoB 3368-4	0–1	16.8	438	738	0.628	17.7
GeoB 3374-1	0–1	17.1	873	1664	0.656	18.5

[a] $U^{K'}_{37} = 0.044 + 0.033T$ (Müller et al., 1998).

3.2. Alkenone analysis and SST estimates

The analytical procedure for the determination of $U^{K'}_{37}$ is described in detail elsewhere (e.g., Müller et al., 1998; Müller and Fischer, 2001). Briefly, long-chain, unsaturated ketones (alkenones) were extracted from 1 g freeze-dried and homogenized sediment samples using an UP200H ultrasonication disruptor probe (Hielscher GmbH, S3 Micropoint, 200 W, amplitude 60%, pulse 0.5 s) and successively less polar solvent mixtures of methanol and methylene chloride (CH_3OH, CH_3OH/CH_2Cl_2 1:1, CH_2Cl_2), each for 3 min. Prior to extraction, the samples were spiked with octacosane acid methyl ester as internal standard. A 3 μl aliquot of the final extracts was analysed by capillary gas chromatography using a Hewlett-Packard (HP) 5890 Series II plus gas chromatograph equipped with a DB5MS fused silica capillary column (60 m, 0.32 mm i.d., and 0.1 μm film thickness). The oven temperature was programmed from 50°C to 250°C at 25°C/min, from 250°C to 290°C at 1°C/min, and an isothermal period maintained for 26 min at 290°C, followed by a further heating step from 290°C to 310°C at 30°C/min and a 10 min isothermal hold at 310°C.

We calculated the alkenone unsaturation index from $U^{K'}_{37} = (C_{37:2})/((C_{37:2}) + (C_{37:3}))$, where $C_{37:2}$ and $C_{37:3}$ represent the di- and tri-unsaturated C_{37} alkenones, respectively (Brassell et al., 1986). The $U^{K'}_{37}$ values were converted into temperature values applying the global core-top calibration of Müller et al. (1998) ($U^{K'}_{37} = 0.033T + 0.044$). The comparison of SST estimates from surface sediments with modern atlas values (Levitus and Boyer, 1994) suggests that off Chile, alkenone-based SST estimates closely resemble annual mean temperatures of the surface-mixed layer (Table 1). The precision of the measurements ($\pm 1\sigma$) was better than 0.009 $U^{K'}_{37}$ units (or 0.26°C), based on 27 replicate extractions on different days of a laboratory internal reference sediment (CC 1076-2) from the South Atlantic. The data of core GIK 17748-2 and GeoB 3302-1 presented in this paper will be archived in the German climate database Pangaea (http://www.pangaea.de).

3.3. Stratigraphy

Following the accelerator mass spectrometry (AMS), ^{14}C-based stratigraphic concept for these cores developed by Marchant et al. (1999) and Lamy et al. (1999), and the refined analyses by Hebbeln et al. (submitted), core GIK 17748-2 covers the past 16 kyr, while core GeoB 3302-1 covers the time span between 32.7 and 13 cal kyr BP. We did not take into account the upper 18 cm of core GeoB 3302-1 for this study because of the inaccurate age model and very low sedimentation rate.

Linear sedimentation rates (LSR) in core GIK 17748-2 varied between 13 and 66 cm/kyr during Termination 1 and between 9 and 16 cm/kyr for the Holocene (Fig. 2a). LSR in core GeoB 3302-1 is in the range of 16–43 cm/kyr during the last glacial period and 5–8 cm/kyr during Termination 1 (Fig. 2b). We combined the two SST curves from cores GIK 17748-2 and GeoB 3302-1 to obtain one record, including the LIA, Termination 1, and the Holocene.

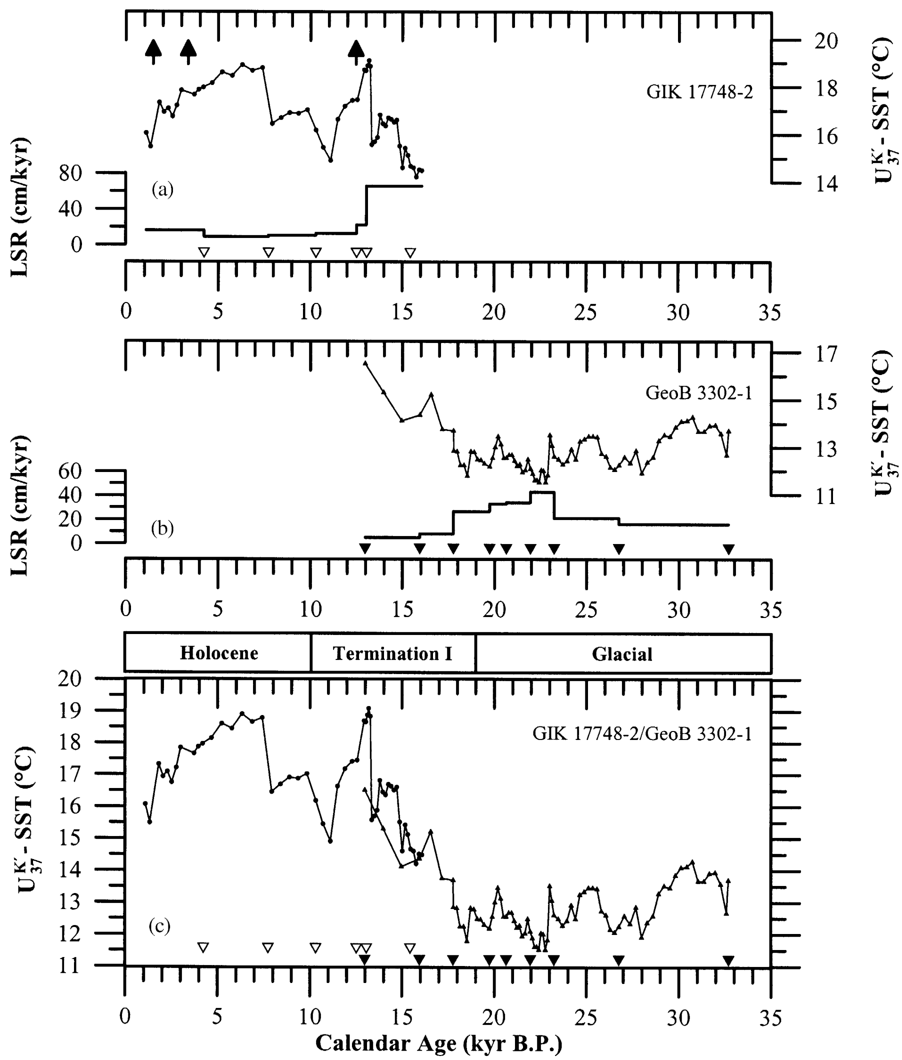

Fig. 2. Time series of alkenone-derived SSTs, calculated after Müller et al. (1998) and linear sedimentation rates (LSR) of cores GIK 17748-2 (a) and GeoB 3302-1 (b), and the combined alkenone-derived SST time series (c), plotted against calendar-year time scale. The age models for cores GIK 17748-2 and GeoB 3302-1 are based on AMS-^{14}C dates obtained from previous studies (Lamy et al., 1999; Marchant et al., 1999) and three additional AMS-^{14}C dates on core GeoB 3302-1 (Hebbeln et al., submitted). Arrows indicate the places where three turbiditic layers were subtracted from the sedimentary sequence. Open and solid triangles indicate the age control points of core GIK 17748-2 and GeoB 3302-1, respectively. The ages for Termination 1 are fixed after Fairbanks (1989) and Mix et al. (2001).

4. Results and discussion

4.1. SST record over the past 33 kyr

Fig. 2c shows the variations in alkenone-derived SST over the past 33 kyr. This combined time series is here divided into three intervals: the Last Glacial Stage, Termination 1, and the Holocene. Throughout the Last Glacial Stage including the late marine Oxygen Isotope Stage (OIS) 3 and OIS 2, SSTs varied within the range of 11.5–14.5°C. The period of the LIA (21 ± 2 cal kyr BP) (Mix et al., 2001) showed an average SST of 12.3°C with the lowest value of 11.5°C between 23 and 22 cal kyr BP The initial warming during Termination 1 started around 18.5 cal kyr BP. During Termination 1, SSTs increased by about 7°C, attaining the maximum value of about 19°C at 13 cal kyr BP. This initial warming trend was interrupted by a short cooling event between 13.8 and 13.3 cal kyr BP. After the temperature maximum at 13 cal kyr BP was reached, SSTs decreased continuously to 15°C at about 11 cal kyr BP, and recovered afterwards to 17°C. During the Holocene, a warming pulse of about 2.5°C occurred between 8.0 and 7.5 cal kyr BP. The highest Holocene values around 19°C were reached between 7.5 and 6 cal kyr BP. Subsequently, after the Holocene climatic optimum, the Southeast Pacific SSTs decreased by about 3°C towards a value similar to modern annual mean SST (15.7°C) in this area (Levitus and Boyer, 1994).

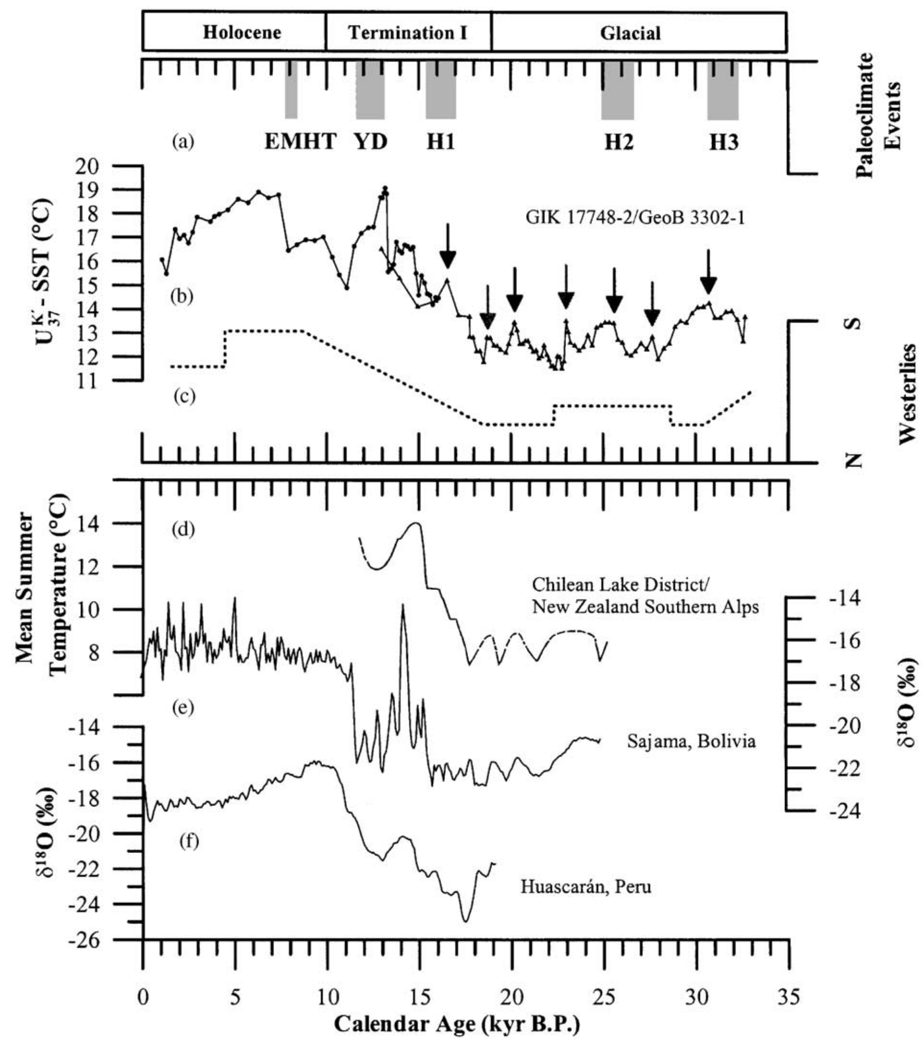

Fig. 3. Comparison of the Southeast Pacific SST record (b) to (a) the short-lived paleoclimate events of H3 (32.4–30.8 cal BP; Vidal et al., 1997), H2 (26.8–25 cal BP; Vidal et al., 1997), H1 (17–15.5 cal BP; Vidal et al., 1997), YD (13–11.5 cal BP; Rutter et al., 2000), and EMHT (8.2–7.8 cal BP; Stager and Mayewski, 1997). The ^{14}C ages in Vidal et al. (1997) were converted into calendar ages using the CALIB program (Stuiver et al., 1998). H, YD, and EMHT denote Heinrich event, Younger Dryas, and early to mid-Holocene climatic transition, respectively. Arrows indicate the timing of SST maxima; (c) relative latitudinal position of the Southern Pacific Westerlies taken from Lamy et al. (1999); (d) mean summer temperature record for the Chilean Lake District and the New Zealand Southern Alps, which was compiled by Denton et al. (1999a) based on pollen studies of Heusser et al. (1999) and glacier studies of Denton et al. (1999b); (e and f) oxygen stable isotope records from the Sajama and Huascarán continental ice cores (Thompson et al., 1998).

4.2. Implications for climate variations in the Southeast Pacific

4.2.1. The Last Glacial Stage

Similar to Chilean pollen records from Taiquemó (42°10.25′S, 73°35.50′W, 170 m above sea level (a.s.l.)) (Heusser et al., 1999), our alkenone-derived SST record from the Southeast Pacific is characterized by 5-kyr rhythms during the Last Glacial Stage, with SST maxima centred at about 30.5, 25.5, 20.5, and 16.5 cal kyr BP (Fig. 3b). If we take into account other less-prominent SST maxima at about 27.5, 23, and 18.5 cal kyr BP, SST fluctuations were more frequent, showing roughly 2.5-kyr cycles. It is yet unknown as to which mechanism generated climatic oscillations in southern Chilean vegetation and SSTs off mid-latitude Chile at sub-Milankovitch time scales. However, it is noteworthy that the timing of the three SST maxima observed in our alkenone-derived SST record from the Southeast Pacific resembles, within uncertainties of the age model, the time intervals of the Heinrich events in the North Atlantic (Vidal et al., 1997; Elliot et al., 1998; Bard et al., 2000) (Fig. 3a). Therefore, the 2°C fluctuations in the 5 kyr cycles in the Southeast Pacific may be considered to some extent as a response to modifications in thermohaline circulation originating in the North

Atlantic. This implication from the Chilean alkenone-derived SST record that the eastern boundary current system off South America experienced warming in response to cooling and cessation of NADW formation in the North Atlantic can be supported by experiments with a global ocean circulation model that propose such a warming for the PCC in reaction to meltwater-induced destabilization of the thermohaline overturn in the North Atlantic (e.g., Schiller et al., 1997). However, high-resolution glacial SST records with much better absolute age control have to be produced in order to better constrain the anti-phase behaviour between the Southeast Pacific and the North Atlantic suggested from our SST record and ocean circulation model experiments.

The Chilean alkenone-derived SST record reveals a SST contrast of 6.7°C between the LIA and the Holocene climatic optimum, but a temperature contrast of only 3.4°C between present and LIA temperature values. The magnitude of SST increase between the LIA and the present is comparable to that of the CLIMAP (1981) and Broccoli and Marciniak (1996) reconstructions based on faunal transfer functions and an atmosphere-mixed layer ocean model applied to the study area. However, pollen evidence from the Chilean Lake District and the Isla de Chiloé (41–43°S) (Denton et al., 1999a; Heusser et al., 1999; Moreno et al., 1999), as well as glacial sediments and a landform study in the Strait of Magellan (52–53°S) (Benn and Clapperton, 2000), and stable isotopes from Peruvian and Bolivian ice cores (Thompson et al., 2000) (Fig. 3e and f) suggested that surface air temperatures at the LIA were 6–8°C lower than the present. This can be supported by the hydrogen isotope record of moss fragments preserved in a sediment core from Estancia Harberton in southern Tierra del Fuego (Markgraf and Kenny, 1997). Thus, the cooling at the sea surface at temperate latitudes in the Southeast Pacific had probably the same magnitude, if Termination 1 from the LIA into the Holocene climate optimum is considered. However, the temperature contrast between the LIA and modern conditions is two times less than that reported for the adjacent landmasses. This may result from the 3°C cooling of the ocean surface during the late Holocene, which is not clearly documented in the continental records. Nonetheless, the marine temperature contrast of 3°C between modern and LIA conditions is of sufficient magnitude for generating the Last Glacial Patagonian ice field (Hulton et al., 1994; Hulton and Sugden, 1997) with the continental temperature decrease even stronger than required. A similar twofold difference between marine and continental magnitudes of LIA cooling is apparent for the Southeast Atlantic (Kirst et al., 1999; Schneider et al., 1999: $\Delta T = 3-4$°C) and South Africa (Stute and Talma, 1998: $\Delta T =$ 5–6°C).

The above-cited South American paleoclimatic studies based on pollen (e.g., Heusser, 1989), glacier and snowline retreat (Caviedes, 1990; Clapperton et al., 1995), glacier modelling (Hulton et al., 1994; Hulton and Sugden, 1997), and clay mineral composition in marine sediments (Lamy et al., 1999) suggested that the position of the southern westerly cyclonic storm tracks at the LIA was displaced northward by about 5–7° in latitude (Fig. 3c), concomitant with a northward expansion of Antarctic sea ice (e.g., Heusser et al., 1999). Such a northward shift of the westerlies from their present position between ≈40° and 50°S probably had also moved the entire oceanographic system (Fig. 1) along the Chilean continental slope towards the north. Marine evidence for such a northward movement of the Southern Pacific Westerlies during the last glacial period is present in bulk sediment proxies (e.g., organic carbon, biogenic opal, and carbonate) and planktic foraminifera accumulation rates (Hebbeln et al., submitted). Basically, colder temperatures and higher productivity of mid-latitude Chile were related to a more northern position of the Southern Pacific Westerlies, while warmer temperatures and lower productivity resulted from a more southern position. Therefore, we assume that this fundamental relationship between the position of the Southern Pacific Westerlies and regional hydrography off mid-latitude Chile caused most of the SST difference between the LIA and the Holocene in the study area.

4.2.2. Termination I

According to the alkenone-derived SST record, deglacial warming of the PCC system occurred in three steps of about 3°C warming each (Fig. 3b). The initial warming during Termination I started at about 18.5 cal kyr BP and lasted until 16.5 cal kyr BP. The second and third warming steps occurred at about 15 and 13 cal kyr BP, respectively. This observation of the stepwise warming trend in the PCC system is in contrast with the view of a continuous warming in South America during Termination I (e.g., Ashworth and Hoganson, 1993). For the ocean, such a stepwise deglacial-warming pattern has also been described for the Indian Ocean, related to the deglacial changes in monsoonal circulation (Sirocko et al., 1996; Bard et al., 1997).

Inferred from the changes in the terrigenous mineral composition of the two cores, post-glacial southward migration of the Southern Pacific Westerlies started at about 18.5 cal kyr BP (Lamy et al., 1999) (Fig. 3c). Therefore, a substantial portion of the first step in post-glacial SST rise from 12°C to 15°C was probably related to this southward migration of the average position of westerly cyclonic storm tracks in the southern mid-latitudes. The timing of the initial deglacial warming in our SST record corresponds to those from a pollen-based temperature record from the

Chilean Lake District (40–43°S) (Denton et al., 1999a) and a stable isotope record from the Huascarán ice core (Thompson et al., 1998), showing the onset of Termination I at about 17.5 cal kyr BP (Fig. 3d and f). This shows that the onset of the deglacial warming in the ocean paralleled that on the South American continent. Possible causes for small temporal differences between these marine and continental records could be an inaccuracy of age models and/or different sensitivity of different proxies to temperature changes.

However, in the Sajama ice core record from Bolivia, the onset of the deglaciation started at about 16 cal kyr BP, 1500 yr later than in the other records. Whether this lag is a true climatic feature or a stratigraphic problem remains unsolved. Remarkably, the most prominent deglacial warming event in the Sajama ice core between 15 and 14 cal kyr BP correlates well with the second marine warming step in our SST record, in the Chilean Lake District record, and in the Huascarán ice core record. The timing of the second warming step, which correlates to the onset of the Bølling/Allerød warm period in the Northern Hemisphere, and its occurrence in all the records in South America and the Southeast Pacific suggest the global nature of this warming event. The first warming step then indicates the lead of the onset of the last deglaciation in the Southern Hemisphere that is obvious from many other climate records (e.g., Charles et al., 1996; Blunier et al., 1998; Kim et al., 2002).

The SST record shows the third warming event at about 13 cal kyr BP, which has no corresponding event in the adjacent terrestrial records. Therefore, we attribute this feature to a regional marine event in the Southeast Pacific. Its origin will be discussed later in context with the strong mid-Holocene warming observed between 8 and 7.5 cal kyr BP (Fig. 3b).

4.2.3. Younger Dryas

The existence of an equivalent cooling event in South America during the Younger Dryas (YD) (13–11.5 cal kyr BP, e.g., Rutter et al., 2000) has been a controversial issue. Previously, based on palynological investigations, Markgraf (1991, 1993) and Villagrán (1993) postulated that there is no evidence of such a cooling event, in contrast to Heusser and Streeter (1980), and Heusser and Rabassa (1987). Recently, based on pollen records from sediments of small lakes in southern Chile (44°–47°S), Bennett et al. (2000) reinforced that there was no cooling in southern Chile and even elsewhere in the Southern Hemisphere during the YD, showing evidences for a continuous forest development and increasing plant diversity, and thus a period of stable, or possibly slightly increasing temperatures.

On the other hand, based on a multi-proxy record from proglacial Lake Mascardi on the east slope of the southern Andes (41°10′S, 71°53′W; 3554 m a.s.l.), Ariztegui et al. (1997) showed a significant advance of the Tronador ice cap during the YD. This interpretation was opposite to previous views based on pollen and beetles data that warming was uniform and rapid to the east of the Andes with no significant temperature and/or precipitation changes equivalent to the YD cooling event (e.g., Markgraf, 1993; Villagrán and Armesto, 1993). In addition, Thompson et al. (1995, 1998, 2000) reported reliable evidences for a YD-type cooling event from Sajama (Bolivia, 18°06.0′S, 68°53.0′W, 6542 m a.s.l.) and Huascarán (Peru, 9°06.4′S, 77°36.5′W, 6048 m a.s.l.) continental ice cores (Fig. 3e and f), with strong support from other studies carried out in the low-land and high-mountain areas of tropical South America (e.g., Clapperton, 1993; Ledru, 1993; Servant et al., 1993; Clapperton et al., 1997) and in southern South America (Denton et al., 1999a) (Fig. 3d) as well.

With respect to the YD, our alkenone-derived SST record from the Southeast Pacific is puzzling and does not help to sort out whether Southern Hemisphere climate changes were similar or different from those in the Northern Hemisphere. A strong cooling trend in the Southeast Pacific is centred at 11 cal kyr BP and afterwards, at about 10 cal kyr BP, temperatures increased to values similar to modern annual mean SST (Fig. 3b). Moreover, if the warming at about 8 kyr BP is considered, this cooling event lasted for 5000 yr, much longer than the duration of the YD (1500 yr) (Fig. 3b).

Many observations support the view that the YD cooling event is global or nearly global in its distribution (e.g., Ivy-Ochs et al., 1999; Rutter et al., 2000), suggesting an atmospheric propagation of the YD climatic impact. Therefore, it is likely that the existence of a cooling event in the Southeast Pacific regions associated with the YD cooling event should be a consistent feature during Termination I. However, based on our alkenone-derived SST record, this global harmony of a YD cooling event is not apparent, and the accurate timing and duration for the Southeast Pacific still remains under discussion.

One possible explanation might be that changes in the Chilean alkenone-derived SST during Termination I were associated with variations in the advection of high-nutrient, cold ACC waters by the PCC caused by the shift of the Southern Pacific Westerlies, analogous to the situation during the Last Glacial Stage. That is, variations in the Chilean alkenone-derived SST in the Southeast Pacific during Termination I might have been governed by a mechanism that was responsible for global climate changes, e.g. YD, but may additionally reflect Southern Ocean climate changes that are not well documented for the subtropical regions.

4.2.4. The Holocene

SSTs recovered from the cooling event at 10 cal kyr BP and for 2000 yr remained at a level similar to the modern condition. Then, SSTs increased abruptly within 500 yr by 2.5°C during the early Holocene (Fig. 3b). The striking warming at 8.0–7.5 cal kyr BP in the Southeast Pacific might be related to paleoclimatic changes at the early to mid-Holocene climatic transition (EMHT, 8.2–7.8 cal kyr BP), observed in diatom-derived climate records from the Lake Victoria (equatorial East Africa) and sodium concentration changes in the Taylor Dome (Antarctic) and GISP2 (Greenland) ice cores (Stager and Mayewski, 1997). At the EMHT, Lake Victoria records showed a high degree of wind-driven water mixing and high lake levels, indicating intensified monsoon winds in the tropics. On the other hand, GISP2 and Taylor Dome records implied weakened meridional circulation and atmospheric mixing, revealing rapid decreases of sodium concentration at that time (Stager and Mayewski, 1997). Such a short-lived paleoclimatic change between 8.0 and 7.5 cal kyr BP was also observed in marine and lacustrine records in the central Mediterranean region (Ariztegui et al., 2000).

The warming event in the Southeast Pacific at the EMHT probably resulted from post-Pleistocene warming and weakening of polar atmospheric circulation that was caused by a decline in the extent of sea ice, and thus the poleward shift of circumpolar air masses (Stager and Mayewski, 1997). This hypothesis is supported by the southernmost position of the Southern Pacific Westerlies that was inferred from grain-size and clay mineral studies for the early Holocene (Lamy et al., 1999) (Fig. 3c). However, the magnitude of this warming is striking and thus it seems unlikely that the warming solely resulted from the movement of the Southern Pacific Westerlies.

Recent climatological data suggest that the latitudinal position of the Southern Pacific Westerlies is strongly related to the strength of the Southeast Pacific subtropical high (e.g., Markgraf, 1998), which is in turn closely associated with the El Niño-Southern Oscillation (ENSO) (Rutllant and Fuenzalida, 1991; Cerveny, 1998). For example, during the warm phase of ENSO (El Niño events), pressure is anomalously low over the Southeast Pacific, leading to a weakening of the Southeast Pacific subtropical high and a consequent northward shift of the Southern Westerlies (Villalba et al., 1996).

This implies that changes in ENSO variability/intensity also in the past were closely related to the strength of the Southeast Pacific subtropical high and thus to the shifts of the Southern Pacific Westerlies. Extremely warm SSTs off mid-latitude Chile in the Southeast Pacific may have been caused by two superimposed processes. Warming due to the southward shift of cold Southern Ocean waters associated with the movement of the Southern Pacific Westerlies and extra warming due to the advection of warmer waters within the southern limb of the South Pacific subtropical gyre at the time of southernmost position of the Southern Pacific Westerlies. This would further imply that subtropical gyre circulation was intensified during the mid-Holocene. Rodbell et al. (1999) presented a record of deposition events from an alpine lake in Ecuador, showing a weaker ENSO variability/intensity during the mid-Holocene. Such a reduced ENSO variability/intensity during the mid-Holocene has been supported by model experiments (Clement et al., 2000; Liu et al., 2000). In addition, a consistent picture from various model studies is that the annual mean SSTs over most of the central-eastern equatorial Pacific were colder than today at that time (Bush, 1999; Otto-Bliesner, 1999; Clement et al., 2000; Liu et al., 2000) due to the strengthened Pacific trades (Liu et al., 2000). This in turn implies that the Southeast Pacific subtropical high was stronger than today and that the Southern Pacific Westerlies moved pole ward. This assumption is in accordance with the southernmost position of the Southern Westerlies at that time as shown by Lamy et al. (1999) and now also implied by extraordinary high SSTs off mid-latitude Chile.

On the other hand, the SST decrease during the late Holocene was then related to the onset of the modern state of ENSO at about 5 cal kyr BP (Rodbell et al., 1999) associated with on average weaker subtropical gyre circulation and atmospheric high-pressure conditions, and causing slight northward re-movement of the Southern Pacific Westerlies (Fig. 3c).

This Holocene shift in poleward movement of the Southern Pacific Westerlies and weakening in ENSO variability/intensity, causing the extraordinary warm SSTs between 7.5 and 5 cal kyr BP, may have taken place also at about 13 cal kyr BP but with much shorter duration, and congruent with Bølling/Allerød warming in the Northern Hemisphere. More high-resolution climate records for ENSO variability/intensity and strength at mid-Holocene and Bølling/Allerød time intervals are still required for better understanding to what extent ENSO variability/intensity in the equatorial Pacific was connected to Northern Hemisphere and high latitude climate change in the past.

5. Conclusions

SSTs in the Southeast Pacific were reconstructed for four surface sediment samples and two sediment cores recovered from the continental slope off mid-latitude Chile, using the alkenone ($U_{37}^{K'}$) method. The comparison of SST estimates from surface and core top sediments with modern atlas SST values (Levitus and Boyer, 1994) suggests that the alkenone-based SST estimates represent annual mean temperatures of the surface mixed layer.

Down-core SST values for the Southeast Pacific show an increase of 6.7°C from the LIA into the Holocene climatic optimum (about 6 cal kyr BP) but a temperature contrast of only 3.4°C between the LIA and the present, comparable to previous estimates, e.g. the CLIMAP.

During the Late Glacial Stage, SST record from the Southeast Pacific was characterised by 5 and 2.5-kyr cycles. The initial warming of the last deglaciation started at about 18.5 cal kyr BP, which corresponds to that observed in South American continental records. Small temporal differences between these marine and continental records could be due to slight inaccuracy of age models and/or different sensitivity of different proxies to temperature changes. During the YD, a rapid cooling trend was revealed in the Southeast Pacific SST record although the cooling event lasted longer than the YD observed in the North Atlantic regions. Therefore, according to the SST record, the existence of a YD equivalent cooling in the Southeast Pacific is much more uncertain than for the South American continental climate.

During the Holocene, an extremely rapid warming pulse of about 2.5°C at 8.0–7.5 cal kyr BP occurred in the Southeast Pacific. This warming trend is probably related to the early to mid-Holocene climatic Transition (EMHT) between 8.2 and 7.8 cal kyr BP observed in equatorial east Africa and the Antarctic. After the Holocene climatic optimum, the Southeast Pacific cooled by about 3°C towards modern temperatures of 15.7°C in this area.

In principal, SST changes off mid-latitude Chile may have been closely associated with variations of the position of the Southern Pacific Westerlies as they influenced the northward advection of cold ACC waters by the PCC. However, SST changes at millennial time scales during the Last Glacial Stage may rather have been linked to oscillations in thermohaline circulation initiated in the North Atlantic because the timing of the SST maxima in the glacial 5-kyr rhythms corresponds fairly well to that of Heinrich events in the North Atlantic. The extraordinary warm SSTs at about 13 and 7.5–5 cal kyr BP can be explained by a weakening in ENSO variability/intensity that caused the advection of warmer waters within the southern limb of the South Pacific subtropical gyre at the time of southernmost position of the Southern Pacific Westerlies.

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