

# The California current system during the last 136,000 years: response of the North Pacific High to precessional forcing

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## Abstract

Alkenone sea surface temperature (SST) records were generated from the Ocean Drilling Program's (ODP) Sites 1014 and 1016 to examine the response of the California Current System to global climate change during the last 136 ka. The temperature differences between these sites ( $\Delta SST_{NEP} = SST_{ODP1014} - SST_{ODP1016}$ ) reflected the intensity of the California Current and varied between 0.4 and 6.1 °C. A high  $\Delta SST_{NEP}$  (weaker California Current) was found for late marine isotope stage (MIS) 2 and early MIS 5e, while a low  $\Delta SST_{NEP}$  (stronger California Current) was detected for mid-MIS 5e and MIS 1. Spectral analysis indicated that this variation pattern dominated 23- (precession) and 30-ka periods. Comparison of the  $\Delta SST_{NEP}$  and SST based on data from core MD01-2421 at the Japan margin revealed anti-phase variation; the high  $\Delta SST_{NEP}$  (weakening of the California Current) corresponded to the low SST at the Japan margin (the southward displacement of the NW Pacific subarctic boundary), and vice versa. This variation was synchronous with a model prediction of the tropical El Niño–Southern Oscillation behavior. These findings suggest that the intensity of the North Pacific High varied in response to precessional forcing, and also that the response has been linked with the changes of tropical ocean–atmosphere interactions.

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## 1. Introduction

The California current system (CCS) is an eastern boundary current system in the North Pacific that consists of the southward California Current, northward Southern California Countercurrent, Davidson Current and California Undercurrent, as shown in Fig. 1 (Hickey, 1979). Previous studies finding decreased productivity (Lyle et al., 1992) and an increased meridional paleo-SST gradient (Doose et al., 1997; Mangelsdorf et al., 2000; Herbert et al., 2001) have suggested that the California Current declined during the last glacial maximum. This decline has been attributed to the depressed summer North Pacific High influenced by the semipermanent development of a high-pressure cell over the expanded Laurentide ice sheet on the North American continent (Kutzbach and Wright, 1985; Lyle et al., 1992). Recently, Yamamoto et al. (2004) noted that SST variation at the California margin did not fully

agree with the development of the Laurentide ice sheet but did correspond to SSTs at the Japan margin, suggesting basin-scale variation in North Pacific SSTs.

The relative intensities of the California Current and the Southern California Countercurrent are controlled by the development of the North Pacific High (Hickey, 1979). California Current water originates from the subarctic West Wind Drift and the subtropical North Pacific Current (Hickey, 1979). Subarctic water contributions and the intensity of the California Current increase during summer due to a strengthened North Pacific High (Hickey, 1979). The Southern California Countercurrent is a northward flow that occurs during most of the year except for spring; the countercurrent becomes seasonally intensified in late fall and winter (Hickey, 1979) and annually in El Niño years due to the weaker North Pacific High (Bograd and Lynn, 2001).

In this study, we compared alkenone-derived SST records at Ocean Drilling Program's (ODP) Sites 1014 and 1016 to clarify the response of the CCS to global climate change during the last 136 ka. Sites 1014 and 1016

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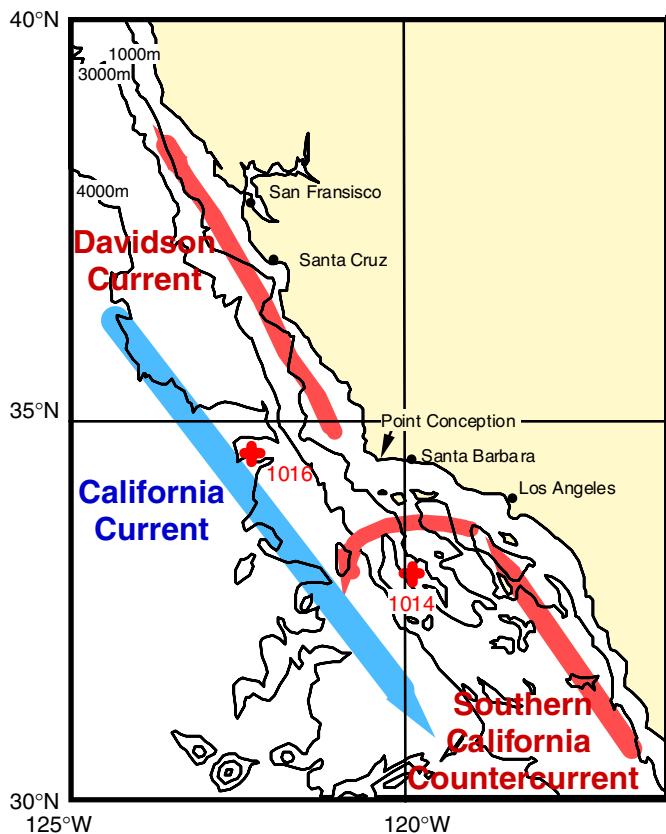


Fig. 1. Locations of ODP Sites 1014 and 1016 and the California Current System.

were drilled during ODP Leg 167 in 1996 (Lyle et al., 1997). Site 1014 is influenced by the northward invasion of the Southern California Countercurrent into the California Borderland region. Site 1016 is located along the main path of the California Current. The temperature difference between these sites must therefore reflect the relative intensities of the California Current and the Southern California Countercurrent.

Paleo-SSTs during the last glacial maximum (LGM) were estimated using alkenones, foraminifers, radiolarians and pollen at numerous locations along the California margin. Along the northern and central California margin, alkenone-derived temperatures (Prahl et al., 1995; Doose et al., 1997; Mangelsdorf et al., 2000; Seki et al., 2002) agreed with those estimated using radiolarian assemblages (Sabin and Pisias, 1996) and foraminiferal  $\delta^{18}\text{O}$  and their assemblages (Ortiz et al., 1997); these data indicate that temperatures were  $\sim 4^\circ\text{C}$  lower in the LGM than in the Holocene. In contrast, at the southern California margin (the California Borderland), alkenone-derived temperatures disagreed with foraminiferal and pollen-derived temperatures during the LGM. Alkenone temperature estimations have indicated that the LGM was 0–2  $^\circ\text{C}$  cooler than the Holocene (Herbert et al., 1995; Doose et al., 1997; Hinrichs et al., 1997), whereas temperature estimations using the sinistral/dextral ratios of

*Neogloboquadrina pachyderma*,  $\delta^{18}\text{O}$  of planktonic foraminifers, and foraminiferal assemblages have indicated that the LGM was more than 5  $^\circ\text{C}$  cooler than the Holocene (Kahn et al., 1981; Kennett and Ingram, 1995; Kennett and Venz, 1995; Thunell and Mortyn, 1995; Mortyn et al., 1996). Pollen assemblages at Site 893 agreed with foraminiferal SST estimations (Heusser, 1995). This “temperature disagreement” in the California Borderland remains an open question. We examined this issue by comparing variations in the unsaturation parameters of alkenones and alkenoates with planktonic foraminiferal  $\delta^{18}\text{O}$ .

## 2. Materials and methods

Cores 167-1014A-1H, 2H and 3H (ODP Site 1014; 0–16.9 m below the sea floor [mbsf]) were collected in the Tanner Basin ( $32^\circ 48' \text{N}$ ,  $118^\circ 54' \text{W}$ ) at a water depth of 1165 m (Lyle et al., 1997; Fig. 1). The sediment cores consisted of dark olive gray to dark gray hemipelagic clay interbedded nannofossil ooze containing foraminifers throughout (Lyle et al., 1997). Samples of approximately  $10 \text{ cm}^3$  in volume were taken every 20 cm and immediately frozen in  $\text{N}_2$ -filled Kapack bags. The age model (Table 1 and Fig. 2) was created by Oxygen Isotope Stratigraphy (Martinson et al., 1987) of the benthic foraminifera

Table 1  
Age-depth model of ODP Site 1014

Depth (mcd)	Method	Conventional age	( $\pm$ )	Age (years BP)	Reference
0.25	$^{14}\text{C}$ of BOM	2830	50	2297	1
1.05	$^{14}\text{C}$ of BOM	7800	60	8005	1
1.55	$^{14}\text{C}$ of BOM	11,310	60	12,643	1
1.95	$^{14}\text{C}$ of BOM	13,240	70	15,102	1
2.35	$^{14}\text{C}$ of BOM	16,450	80	18,766	1
2.75	$^{14}\text{C}$ of BOM	17,150	90	19,572	1
3.05	$^{14}\text{C}$ of BOM	18,000	90	20,550	1
Depth (mcd)	OIS			Age (years BP)	Reference
6.20	3.13			43,880	2
6.65	3.3			50,210	2
7.20	3.31			55,450	2
7.45	4			58,960	2
8.00	4.22			64,090	2
8.70	5			73,910	2
9.80	5.1			79,250	2
10.55	5.2			90,950	2
11.15	5.31			96,210	2
11.80	5.33			103,290	2
12.55	5.4			110,790	2
13.45	5.51			122,560	2
13.75	5.5			123,820	2
14.05	5.53			125,190	2
14.45	6			129,840	2
16.75	6.3			142,280	2
17.50	6.4			152,580	2
18.50	6.41			161,340	2

Reference 1: this study, 2: Hendy and Kennett (2000).

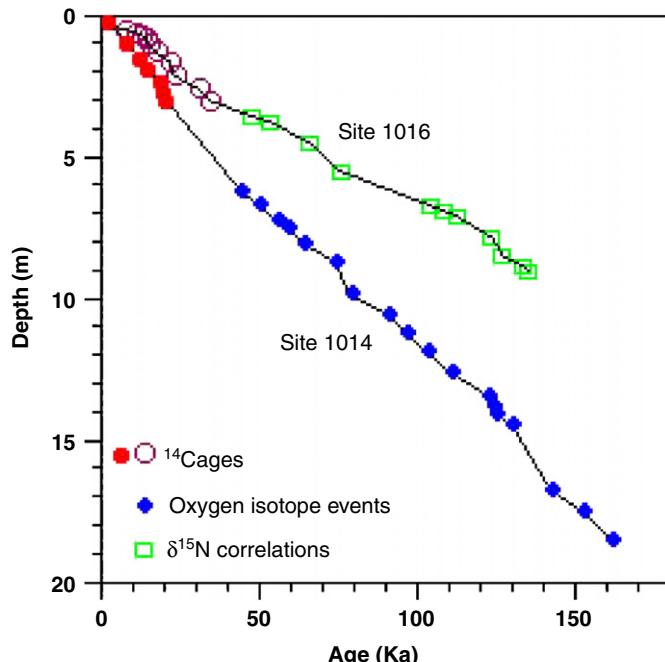


Fig. 2. Age-depth models of ODP Sites 1014 and 1016.

*Uvigerina* (Hendy and Kennett, 2000) and by calendar ages converted from the AMS  $^{14}\text{C}$  ages of seven bulk organic matter samples (2.30–20.55 ka). The  $^{14}\text{C}$  ages were converted to calendar ages using the CALIB4.3 marine98 program (Stuiver and Reimer, 1993) and an equation by Bard (1998) with a 633-year reservoir correction for the California margin (Ingram and Sounth, 1996).

Core 167-1016C-1H (ODP Site 1016; 0–9.2 mbsf) was collected off the coast of central California (34°32'N, 122°17'W) at 3834 m water depth (Lyle et al., 1997; Fig. 1). Core sediments consisted of gradationally interbedded dark gray to dark greenish gray diatom ooze and light greenish yellow to light olive gray diatom nannofossil ooze (Lyle et al., 1997). The core was sealed in a  $\text{N}_2$ -filled oxygen-impermeable plastic bag immediately after recovery and stored in a refrigerator for 3 months until shore-based sampling. Samples of approximately 10 cm<sup>3</sup> in volume were taken every 9 cm and immediately frozen. The age model was created with calendar ages converted from the AMS  $^{14}\text{C}$  ages of eleven bulk organic matter samples (5.66–37.72 ka) using the same program and reservoir correction method as for cores 167-1014A (Table 2 and Fig. 2). Correlating the  $\delta^{15}\text{N}$  values of this core with those of Site 1014 established that the age-depth model extended to 135.57 ka, as described in Section 3.1.

Alkenones and alkenoates were analyzed following the method of Yamamoto et al. (2000). The separation of all alkenones and alkenoates was enough for accurate quantification (Fig. 7 in Yamamoto et al., 2000). The alkenone unsaturation index  $U_{37}^{\text{K}}$  was calculated from the concentrations of di- and tri-unsaturated C<sub>37</sub> alken-2-ones

Table 2  
Age-depth model of ODP Site 1016

Depth (mcd)	Method	Conventional age	( $\pm$ )	Age (years BP)	Reference
0.18	$^{14}\text{C}$ of BOM	5550	40	5657	1
0.36	$^{14}\text{C}$ of BOM	7900	40	8128	1
0.54	$^{14}\text{C}$ of BOM	10,950	50	11,941	1
0.63	$^{14}\text{C}$ of BOM	13,130	70	14,339	1
0.72	$^{14}\text{C}$ of BOM	13,480	40	15,392	1
0.90	$^{14}\text{C}$ of BOM	15,090	50	17,201	1
1.17	$^{14}\text{C}$ of BOM	17,400	60	19,860	1
1.53	$^{14}\text{C}$ of BOM	21,380	90	24,513	1
1.98	$^{14}\text{C}$ of BOM	22,950	80	26,300	1
2.43	$^{14}\text{C}$ of BOM	30,440	200	34,427	1
2.88	$^{14}\text{C}$ of BOM	33,660	240	37,719	1

Depth (mcd)	Method	Age (years BP)	Reference
3.51	$\delta^{15}\text{N}$ correlation	48,100	1
3.78	$\delta^{15}\text{N}$ correlation	53,545	1
4.50	$\delta^{15}\text{N}$ correlation	65,914	1
5.49	$\delta^{15}\text{N}$ correlation	75,852	1
6.66	$\delta^{15}\text{N}$ correlation	104,290	1
6.93	$\delta^{15}\text{N}$ correlation	108,290	1
7.11	$\delta^{15}\text{N}$ correlation	112,752	1
7.83	$\delta^{15}\text{N}$ correlation	123,274	1
8.46	$\delta^{15}\text{N}$ correlation	127,050	1
8.82	$\delta^{15}\text{N}$ correlation	133,951	1
9.00	$\delta^{15}\text{N}$ correlation	135,573	1

Reference 1: this study.

[C<sub>37</sub>MK] using the following expression (Prahl et al., 1988):

$$U_{37}^{\text{K}} = [\text{C}_{37:2}\text{MK}] / [\text{C}_{37:2}\text{MK} + \text{C}_{37:3}\text{MK}].$$

Temperature was calculated according to the equation

$$U_{37}^{\text{K}} = 0.034T + 0.039,$$

where  $T$  is the temperature (°C) based on experimental results for cultured strain 55a of *Emiliania huxleyi* (Prahl et al., 1988); analytical accuracy was 0.24 °C in our laboratory. The culture-based calibration (Prahl et al., 1988) was nearly identical to the core-top calibration along the California margin (Herbert et al., 1998). Additional unsaturation indices for U<sub>36</sub><sup>me</sup> and K<sub>37:4</sub>/K<sub>37</sub> were calculated using the following equations (Prahl et al., 1995):

$$U_{36}^{\text{me}} = [\text{C}_{36:2}\text{ME}] / [\text{C}_{36:2}\text{ME} + \text{C}_{36:3}\text{ME}],$$

where [C<sub>36</sub>ME] is the concentration of C<sub>36</sub> fatty acid-methyl ester, and

$$K_{37:4}/K_{37} = [\text{C}_{37:4}\text{MK}] / [\text{C}_{37:2}\text{MK}] + [\text{C}_{37:3}\text{MK}] + [\text{C}_{37:4}\text{MK}].$$

To determine stable nitrogen isotope ratios, samples were combusted at 1020 °C in a Fisons Instruments EA1108 elemental analyzer (Beverly, MA, USA). The generated N<sub>2</sub> was introduced into a Finnigan Delta Plus isotope-ratio mass spectrometer (Thermo Electron Corp., Waltham, MA, USA) by using a He carrier. The ratio of

$^{15}\text{N}/^{14}\text{N}$  was expressed relative to the  $\text{N}_2$  in air. Machine drift during analyses was checked with 1- $\alpha$ -alanine ( $\delta^{15}\text{N} = 7.61\text{\textperthousand}$ ) every three samples. Accuracy was determined using interlaboratory-determined nitroarginine following the method of Minagawa et al. (1984).

We performed spectral analyses using the Blackman–Tuckey and Cross–Blackman–Tuckey methods provided in the Analyseries software package (Paillard et al., 1996).

### 3. Results and discussion

#### 3.1. Nitrogen isotopes and the age-depth model of Site 1016

Oxygen isotope stratigraphy could not be used to establish the age-depth model of Site 1016 because carbonates had dissolved at most intervals. Instead, we established that the age-depth model extended to 135.57 ka (Table 2 and Fig. 2) by correlating  $\delta^{15}\text{N}$  values of this core with those of Site 1014 (Fig. 3). Because the  $\delta^{15}\text{N}$  values of cores in the Northeastern Pacific rim from the Mexico to Oregon margins varied synchronously during the last glacial age (Kienast et al., 2002),  $\delta^{15}\text{N}$  could be used for stratigraphic correlation for the California margin. The  $\delta^{15}\text{N}$  values ranged from  $\sim 5\text{\textperthousand}$  to  $9\text{\textperthousand}$  (Fig. 3); this variation was synchronous with that in the Mexican margin (Ganeshram et al., 1995) except that the  $\delta^{15}\text{N}$  maximum for MIS 5a was more enhanced for Sites 1014 and 1016 than that at the Mexican margin. The  $\delta^{15}\text{N}$  values were generally 0–2‰ higher for the deeper site (Site 1016) than the shallower site (Site 1014). This difference

was consistent with the observation that the isotopic fractionation of  $\delta^{15}\text{N}$  between sinking particles and surface sediments tends to increase with increasing water bottom depth (Nakanishi and Minagawa, 2003).

#### 3.2. Paleo-SSTs at the California margin

At Site 1014, SSTs ranged from 12 to  $19^\circ\text{C}$  during the last 158 ka, whereas at Site 1016, SSTs ranged from 9 to  $18^\circ\text{C}$  during the last 136 ka (Fig. 3). The core-top temperatures of Sites 1014 and 1016 were  $\sim 16$  and  $\sim 14^\circ\text{C}$ , respectively, in agreement with present annual mean SSTs at these sites (Robinson, 1976). This result is concordant with the observation that core-top  $\text{U}_{37}^{\text{K}'}$ -derived temperatures represent the annual mean SSTs along the California margin (Herbert et al., 1998). At both sites, the SST increase preceded the MIS 1/2 and 5/6 boundaries by  $\sim 8$  and  $\sim 6$  ka, respectively (Fig. 3). This early warming was more evident at Site 1014 than at Site 1016 (Fig. 3), implying a stronger Southern California Countercurrent (Herbert et al., 2001) analogous to a modern El Niño condition (Bograd and Lynn, 2001) at the latest glacial stage. Strong cooling was observed around the MIS 2/3 and 3/4 boundaries (Fig. 3). *Coccolithus pelagicus*, a coccolithophore specific to subarctic waters in the North Pacific (Okada and McIntyre, 1977), was abundant during the same intervals at Site 1014 (Fig. 4; Tanaka, Y., unpublished data). These observations indicate an intensified California Current analogous to a modern La Niña condition during those periods.

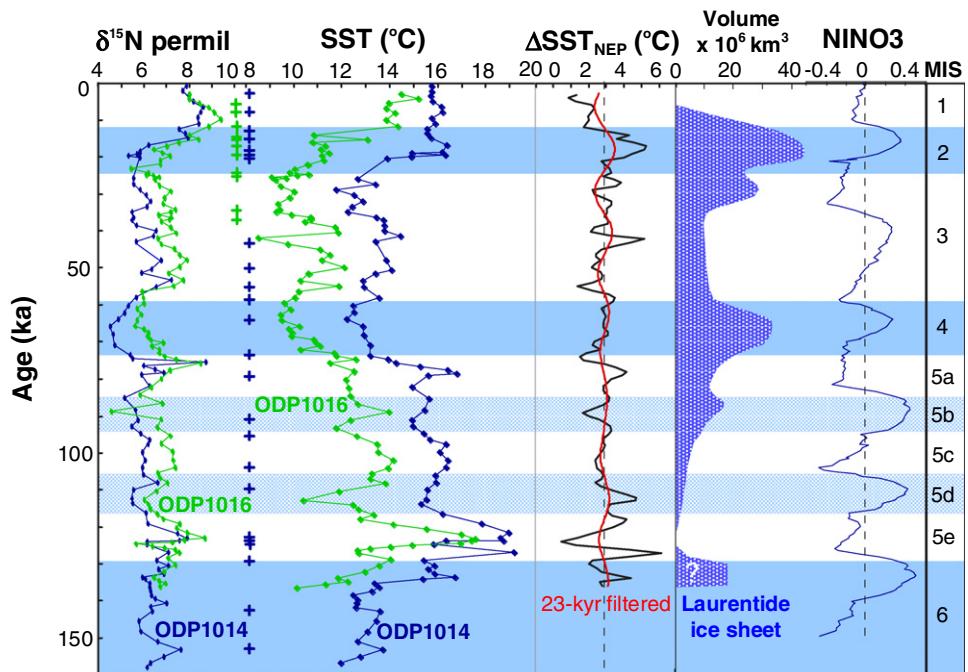


Fig. 3. Changes in sedimentary  $\delta^{15}\text{N}$  and alkenone  $\text{U}_{37}^{\text{K}'}$ -derived sea surface temperatures (SSTs) of ODP Sites 1014 and 1016 and the  $\Delta\text{SST}_{\text{NEP}}$  ( $\text{SST}_{\text{ODP}1014} - \text{SST}_{\text{ODP}1016}$ ) during the last 160 ka. Data on SSTs for Sites 1014 and 1016 are from Yamamoto et al. (2000, 2004). The calculated NINO3 index (Clement et al., 1999) and the estimated volume of the Laurentide ice sheet (Boulton et al., 1985) are shown for comparison. Blue and green crosses indicate the age controls for Sites 1014 and 1016, respectively, by  $^{14}\text{C}$  dating and oxygen isotope stratigraphy.

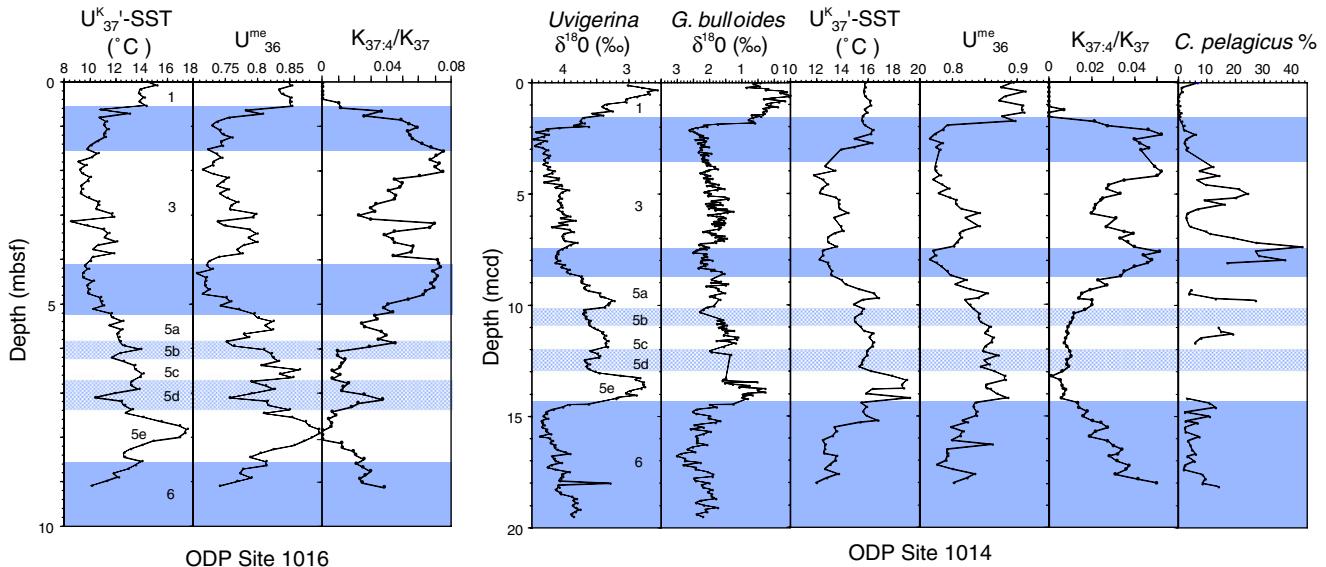


Fig. 4. The  $U_{37}'$ -derived SST,  $U_{36}^{me}$ , and  $K_{37:4}/K_{37}$  indices for Sites 1014 and 1016 and the abundance of *Coccolithus pelagicus* in relation to total coccoliths at Site 1014. The  $\delta^{18}\text{O}$  profiles of *Uvigerina* and *Globigerina bulloides* at Site 1014 (Hendy and Kennett, 2000) are shown for comparison. Data for SST,  $U_{36}^{me}$ , and  $K_{37:4}/K_{37}$  indices at Site 1016, and SST at Site 1014 are from Yamamoto et al. (2000, 2004).

In the Tanner Basin, the  $U_{37}'$ -derived paleo-temperatures during the latter half of MIS 2 at Site 1014 were more than 4 °C higher than those indicated by foraminiferal  $\delta^{18}\text{O}$  and assemblages in core AHF10164 (Kahn et al., 1981). Interestingly, the  $U_{36}^{me}$  and  $K_{37:4}/K_{37}$  profiles were similar to the  $\delta^{18}\text{O}$  profile of planktonic foraminifera *Globigerina bulloides* around MIS 2 (Hendy and Kennett, 2000) but differed from the  $U_{37}'$  profile at Site 1014 (Fig. 4). However, the  $U_{36}^{me}$  and  $K_{37:4}/K_{37}$  profiles were similar to the  $U_{37}'$  profile at Site 1016 (Yamamoto et al., 2000; Fig. 4). The  $U_{37}'$  profile showed good correlation with  $U_{36}^{me}$  and  $K_{37:4}/K_{37}$  at Site 1016; however,  $U_{37}'$  values from late MIS 2 samples were higher than those of other intervals at Site 1014 (Fig. 5). This finding suggests multiple sources of alkenones and alkenonates at Site 1014 in late MIS 2. At least two major sources are indicated. The first source may have been  $C_{36}\text{ME}$ - and  $C_{37:4}\text{MK}$ -rich species that bloomed in cooler water. The second source may have been  $C_{36}\text{ME}$ - and  $C_{37:4}\text{MK}$ -poor species that bloomed in warmer water. Sediment-trap studies at the Santa Barbara Basin demonstrated that when the spring upwelling was reduced under El Niño conditions, coccolithophore export production was enhanced particularly for the tropical species *Gephyrocapsa oceanica* (De Bernardi et al., 2005), while the export production of the planktonic foraminifers *Globigerina bulloides* and *N. pachyderma* (dextral) was reduced (Pak et al., 2004). Frequent invasions of subtropical water to the California Borderland during late MIS 2, which was analogous to modern El Niño conditions (Bograd and Lynn, 2001), may therefore have enhanced haptophyte blooms possibly by the second type of alkenone producer noted above and reduced the blooms of planktonic foraminifers and the first type of alkenone producer. This ecological setting possibly caused the  $U_{37}'$  to

record warmer temperatures than those by foraminifers and  $U_{36}^{me}$ . The high  $U_{37}'$  during late MIS 2 might more accurately reflect the frequent invasion of warmer water to the California Borderland.

### 3.3. Variations of the CCS on orbital timescales

Site 1014 experiences the northward invasion of the Southern California Countercurrent to the California Borderland, while Site 1016 is located along the main path of the California Current. Thus, the temperature difference between these sites reflects the relative intensities of the California Current and the Southern California Countercurrent. We obtained SST differences for the Sites 1014 and 1016 ( $\Delta\text{SST}_{\text{NEP}}$ ) for the last 136 ka by subtracting the SST of Site 1016 from that of Site 1014 (Fig. 3). The  $\Delta\text{SST}_{\text{NEP}}$  varied between 0.4 and 6.1 °C with an average of 2.9 °C. A high  $\Delta\text{SST}_{\text{NEP}}$  (weaker California Current) was observed in late MIS 2 and early MIS 5e, while a low  $\Delta\text{SST}_{\text{NEP}}$  (stronger California Current) was observed in mid-MIS 5e and MIS 1.

We evaluated the effects of time-scale uncertainty on the  $\Delta\text{SST}_{\text{NEP}}$  profile by comparing various  $\Delta\text{SST}_{\text{NEP}}$  profiles obtained under the assumption that the ages of Site 1016 shift in the direction of younger or older. Because the sample intervals of  $\delta^{15}\text{N}$  measurement were  $\sim 1.3$  ka at Site 1016, the maximum error in the direct correlation of  $\delta^{15}\text{N}$  must be  $\sim 1.3$  ka, although errors caused by other factors were not evaluated in detail. Shifting the ages of Site 1016 to younger and older directions within 3 ka did not change the  $\Delta\text{SST}_{\text{NEP}}$  profile, which shows high  $\Delta\text{SST}_{\text{NEP}}$  in late MIS 2, mid-MIS 3, MIS 5a and early MIS 5e and low  $\Delta\text{SST}_{\text{NEP}}$  in MIS 1, early MIS 3, the MIS 4/5a boundary and MIS 5b. One exception was mid-MIS 5e. Shifting to

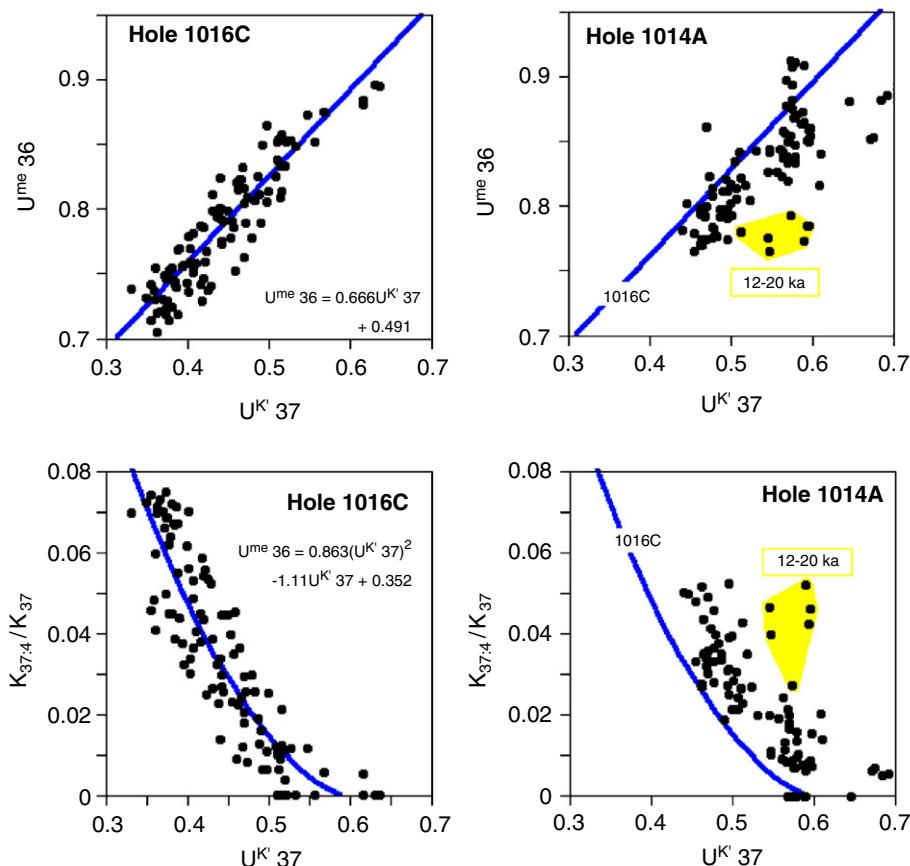


Fig. 5. Plots of  $U^{37} - U^{36e}$  and  $U^{37} - K_{37.4}/K_{37}$  for Sites 1014 and 1016 samples.

the younger direction by 1–3 ka weakened the negative peak at mid-MIS 5e.

Previous studies have suggested a decline in the California Current during MIS 2 based on decreased productivity (Lyle et al., 1992) and an increased meridional paleo-SST gradient (Doose et al., 1997; Mangelsdorf et al., 2000). Herbert et al. (2001) showed that an early warming at the California Borderland began  $\sim 24$  ka, and the SST reached a maximum value at  $\sim 16$  ka (late MIS 2), reflecting weakening of the California Current. This decline was attributed to a depressed summer North Pacific High influenced by the semipermanent development of a high-pressure cell over the expanded Laurentide ice sheet over continental North America (Kutzbach and Wright, 1985; Lyle et al., 1992). However, paleo-SST change at Site 1014 did not fully agree with the estimated volume change of the Laurentide ice sheet (Boulton et al., 1985). In particular, a low SST and abundant *C. pelagicus* indicate a strong California Current in late MIS 4; however, the second-largest expanse of the Laurentide ice sheet during the last glacial period also occurred during this time (Fig. 4). Thus, the volume of the Laurentide ice sheet may not have been the only factor affecting the CCS.

Spectral analyses indicated that the SST variation patterns of Sites 1014 and 1016 dominate 100-ka (eccentricity), 41-ka (obliquity), and 23-ka (precession) periods and

are highly coherent at these frequencies with the  $\delta^{18}\text{O}$  variation of the benthic foraminifera *Uvigerina* at Site 1014 (Fig. 6A and B). However, the variation pattern of  $\Delta\text{SST}_{\text{NEP}}$  dominated 23-ka and 30-ka periods (Fig. 6C).

Comparing the  $\Delta\text{SST}_{\text{NEP}}$  and the SST of MD01-2421 at the Japan margin revealed an anti-phase variation during the last 136 ka (Fig. 7). Both  $\Delta\text{SST}_{\text{NEP}}$  and the SST of MD01-2421 showed variations in 23-ka and 30-ka periods and lacked a 41-ka period (Fig. 6C and D). Cross-spectral analysis indicated that  $\Delta\text{SST}_{\text{NEP}}$  variation is highly coherent with the SST variation of MD01-2421 at 23-ka and 30-ka periods, and also indicated that the  $\Delta\text{SST}_{\text{NEP}}$  led the SST of MD01-2421 by  $169 \pm 12$  and  $127 \pm 11^\circ$  at 23- and 30-ka bands, respectively. The high  $\Delta\text{SST}_{\text{NEP}}$  at the California margin corresponded to the low SST at the Japan margin, and vice versa (Fig. 7).

At the Japan margin, the alkenone SST variation mainly reflects the latitudinal displacement of the subarctic boundary in summer (Yamamoto et al., 2004, 2005). The latitudinal position of the subarctic boundary associated with the westerly jet in summer is principally controlled by the relative intensities of the North Pacific High and the Okhotsk High (Nitta, 1987; Goes et al., 2001). A low SST reflects the southward displacement of the subarctic boundary, resulting from the weaker North Pacific High. The observed SST of alkenone maximum production

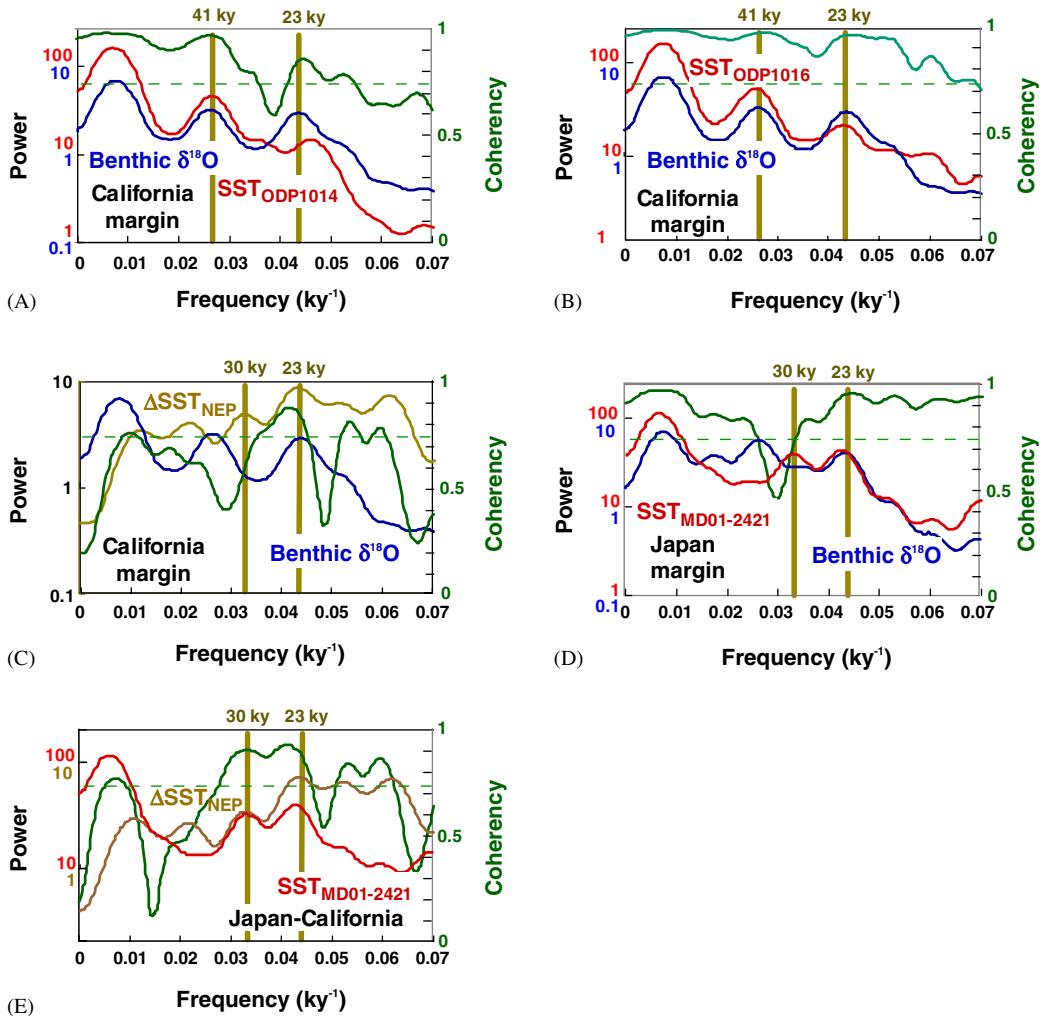


Fig. 6. Spectral densities and coherencies of (A) SST versus the  $\delta^{18}\text{O}$  of *Uvigerina* for Site 1014 at the California margin (3–136 ka, bandwidth = 0.011); (B) the SST of Site 1016 versus the  $\delta^{18}\text{O}$  of *Uvigerina* in Site 1014, California margin (3–136 ka, bandwidth = 0.011); (C)  $\Delta\text{SST}_{\text{NEP}}$  versus the  $\delta^{18}\text{O}$  of *Uvigerina* for Site 1014 at the California margin (3–136 ka, bandwidth = 0.011); (D) the SST versus the  $\delta^{18}\text{O}$  of benthic foraminifera *Uvigerina* for core MD01-2421, Japan margin (5–142 ka, bandwidth = 0.011) and (E) the  $\Delta\text{SST}_{\text{NEP}}$  versus the SST in core MD01-2421 (5–136 ka, bandwidth = 0.011). The horizontal dashed line indicates an 80% level of nonzero coherency.

season (July and August) at the site of MD01-2421 from 1950 to 2002 has positive correlation ( $r = 0.39$ ,  $n = 53$ ,  $p < 0.05$ ) with the intensity of the summer North Pacific High (Isono and Yamamoto, unpublished data). Thus, the high  $\Delta\text{SST}_{\text{NEP}}$  at the California margin and the low SST at the Japan margin both resulted from a weaker North Pacific High, and vice versa. The anti-phase relationship between  $\Delta\text{SST}_{\text{NEP}}$  and the SST of MD01-2421 at the precession band suggests that variation in the CCS at 23-ka periods is a part of the response of the North Pacific High to precessional forcing.

A 30-ka period was previously reported in the variations of atmospheric carbon dioxide concentration recorded in the Vostok ice core (Petit et al., 1999) and in the paleo-productivity records in the tropical Indo-Pacific (e.g., Beaufort et al., 2001, 2003). At the 30-ka band, the  $\Delta\text{SST}_{\text{NEP}}$  maxima (the weakest California Current) lagged the  $\text{CO}_2$  concentration minima by  $11 \pm 15^\circ$ , and the SST maxima of MD01-2421 lagged the  $\text{CO}_2$  concentration

maxima by  $39 \pm 23^\circ$  (Fig. 8). The phase lag between the  $\Delta\text{SST}_{\text{NEP}}$  maxima and the SST maxima of MD01-2421 was  $127 \pm 11^\circ$ ; the relationship was not anti-phase. This phase lag suggests that these variations were not a simple result of the response of the North Pacific High, but were interfered by an unknown factor at either the California or Japan margin.

#### 3.4. Linkage between the CCS and tropical Indo-Pacific climates

Both the  $\Delta\text{SST}_{\text{NEP}}$  and SST at the Japan margin are controlled by the intensity of the North Pacific High. The North Pacific High is influenced by modern interannual El Niño-Southern Oscillation (ENSO) through atmospheric teleconnections (Bograd and Lynn, 2001). In El Niño years, the tropical convection center moves to the central and eastern equatorial Pacific. This process weakens the North Pacific High in summer, resulting in a warming of

the California Borderland by the intensification of the Southern California Countercurrent (Bograd and Lynn, 2001). At the Japan margin, the depressed North Pacific High in the summer of El Niño years tend to delay the northward shifts of the westerly jet and oceanic subarctic boundary (Goes et al., 2001) and thus cool the Japan margin in summer (Nitta, 1987; Kawamura et al., 1998). In La Niña years, the tropical convection center moves to the

western equatorial Pacific, intensifying the summer North Pacific High and the California Current and cooling the California margin. This process excites the Pacific–Japan teleconnection pattern in summer (Nitta, 1987) and accelerates the northward shift of the subarctic boundary, warming the Japan margin (Nitta, 1987; Kawamura et al., 1998).

Zebiak–Cane ENSO model calculations for the past 150 ka demonstrated that seasonal anomalies in insolation driven by precession could have changed ENSO behaviors (see the calculated NINO3 in Fig. 7; Clement et al., 1999). In the modern Pacific, the Intertropical Convergence Zone (ITCZ) nears the equator in spring, generating a zonal symmetric response in atmospheric heating, while the eastern Pacific ITCZ moves north in autumn, generating zonal asymmetry in atmospheric heating. The model calculation assumes that when the perihelion occurs in spring, uniform atmospheric heating of the tropical Pacific generates a background state giving an El Niño-like response; for an autumn perihelion, asymmetric atmospheric heating generates a background state creating a La Niña-like response.

Beaufort et al. (2001) generated late Pleistocene sediment records of nannofossil-derived paleo-productivity along the equator in the Indian and Pacific oceans. Their study showed that variations in equatorial productivity have been caused primarily by both glacial–interglacial variability and precession-controlled changes in the east–west thermocline slope of the Indo-Pacific. Precession-controlled variations in productivity were linked to processes similar to the ENSO phenomenon and preceded changes in the  $\delta^{18}\text{O}$  of benthic foraminifera (Fig. 8), indicating that these changes did not result from ice sheet fluctuations.

At the precession band, the intensity minima of the North Pacific High was in phase with the April perihelion and the maxima of predicted NINO3 index (Clement et al.,

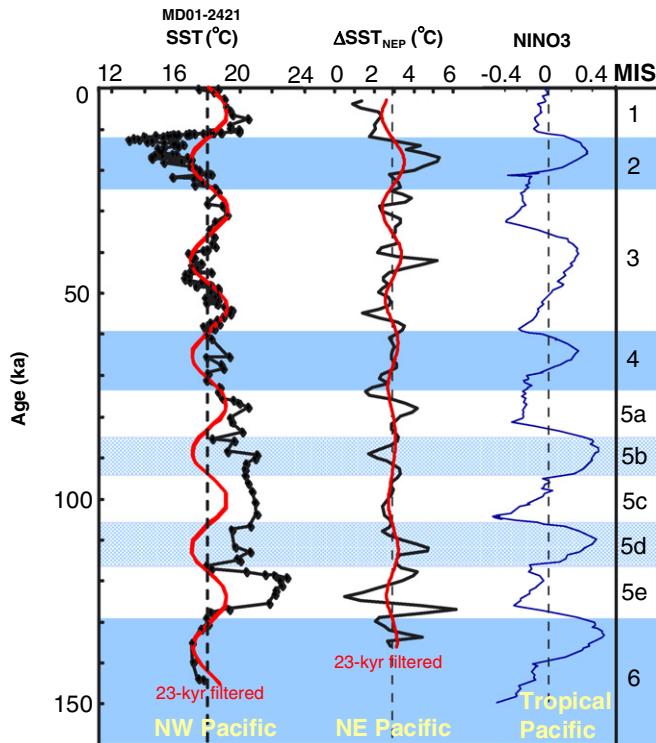


Fig. 7. The  $\text{U}_{37}^{\text{K}'}$ -derived SST at the Japan margin (core MD01-2421; Yamamoto et al., 2005), the  $\Delta\text{SST}_{\text{NEP}}$  at the California margin and the calculated NINO3 index (Clement et al., 1999) during the last 150 ka.

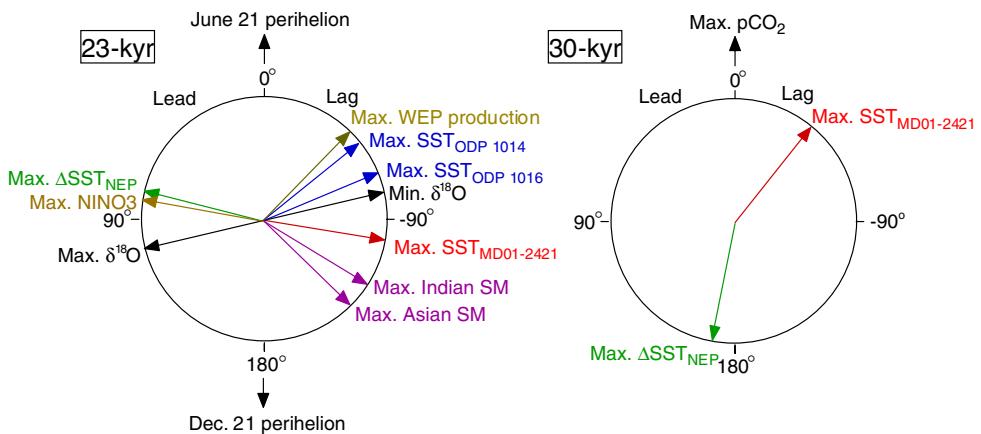


Fig. 8. Precession and 30-ka wheels showing the phases of the SST maxima in core MD01-2421 at the Japan margin, the SST maxima at Sites 1014 and 1016, the  $\Delta\text{SST}_{\text{NEP}}$  maxima at the California margin, SPECMAP  $\delta^{18}\text{O}$  maxima and minima (Imbrie et al., 1984), the predicted NINO3 (Clement et al., 1999), the maximum Indian summer monsoon (summer monsoon stack by Clemens and Prell, 2003), the maximum East Asian summer monsoon (abundance of *Cryptomeria japonica* in core MD01-2421, Igarashi and Oba, 2006), and the maximum primary production in the western equatorial Pacific (Beaufort et al., 2001). Positive and negative angles indicate phase leads and lags, respectively, relative to the precession minima (21 June perihelion) or the  $\text{pCO}_2$  maxima.

1999; Fig. 8). However, productivity variation in the tropical Indo-Pacific (Beaufort et al., 2001) was out of phase with the  $\Delta SST_{NEP}$  and the predicted NINO3 (Fig. 8). The El Niño-like phase of the thermocline slope lagged the  $\Delta SST_{NEP}$  maxima and the predicted NINO3 maxima by  $\sim 120^\circ$  (Fig. 8). Our results and those of the model prediction indicate that the weakest North Pacific High and the optimal phase of the El Niño-like condition lagged the ice volume maxima, while the productivity records (Beaufort et al., 1999) indicate that the optimal El Niño-like phase preceded the ice volume minima (Fig. 8). This phase discrepancy of long-term ENSO-like variability remains a key question because little is known about the behavior of tropical ocean–atmosphere interactions at orbital timescales, although shorter paleo-records in the eastern tropical Pacific and the tropical Andes regions were consistent with the model prediction (e.g., Koutavas et al., 2002; Moy et al., 2002).

The precession wheel in Fig. 8 illustrates that the strongest North Pacific High ( $\Delta SST_{NEP}$  minima) led by  $\sim 30^\circ$  ( $\sim 2$  ka) with the maxima of the Indian summer monsoon (based on the summer monsoon stack by Clemens and Prell, 2003) and East Asian summer monsoon (based on the abundance of a Japanese cedar [*Cryptomeria japonica*] pollen in the NW Pacific off Japan by Morley and Heusser, 1997; Igarashi and Oba, 2006). Meteorological studies demonstrated a close linkage between boreal summer monsoon (BSM) and ENSO; the stronger BSM corresponds to the La Niña state of the tropical Pacific, and the weaker BSM corresponds to the El Niño state (e.g., Meehl, 1987; Yasunari, 1991). Heusser and Sirocko (1997) reported a large centennial-scale fluctuation of pollen assemblages during the last deglaciation in ODP Site 893, Santa Barbara Basin, California margin. They suggested that correlative high-frequency variations for the northeast Pacific winter and southern Asian summer climates may reflect submillennial-scale ENSO-type forcing involving water vapor transport over the Indian and Pacific oceans. These modern and centennial-scale studies suggest the synchronous climate variations throughout the Indo-Pacific region. In contrast, our orbital-timescale records demonstrated the phase lead of the North Pacific High to boreal monsoon variations, suggesting either that the ENSO-like variability has been independent of BSM or that the ENSO-like variability has caused the change of BSM.

#### 4. Conclusion

The present study presents the new perspective that variation of the CCS is part of a basin-scale oceanic and atmospheric response in the North Pacific. The variation indicates that the North Pacific High varied in response to precessional forcing. Long-term ENSO-like variability and wave propagation by teleconnection (Heusser and Sirocko, 1997; Clement et al., 1999; Beaufort et al., 2001; Yamamoto et al., 2004) are potential driving forces of

the basin-scale climate response. Establishing the linkage between this basin-scale response and tropical ocean–atmospheric dynamics will be a key toward better understanding the role of the Indo-Pacific in global climate change.

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