FIG. 4 The coral reef off the H. Steinitz Marine Biology Laboratory in Eilat at ~6 m depth. *a*, Under normal conditions (15 December 1994); *b*, at the peak of the algal bloom in spring 1992 (6 April 1992). The filamentous green alga is *Enteromorpha* sp. Corals growing on elevated substrates (seen in *b*) were not covered by algae and were therefore not damaged.

(A. Miroz, personal communication), but its effect on corals was unfortunately not examined. We note, however, that events of El Niño/Southern Oscillation (ENSO) occurred in both 1983 and 1992 (ref. 19). Although several studies[1,15] singled out volcanic eruptions as the dominant cause for these two cold winter anomalies in the Middle East, and indeed, in Eilat eight of the 15 ENSO events[20] during the past 45 years coincided with warmer-than-average winters, it is possible that the co-occurring ENSO events affected the magnitude of the anomalies[20,21].

The large algal bloom in spring 1992 was an ecological amplification of a rather short-term air-temperature anomaly. To the best of our knowledge, the weak vertical stratification that makes such amplification possible is unique to deep marginal seas with shallow sills.

Large volumes of water exchange heat with the atmosphere during the relatively short period of deep mixing. Therefore, sea surface temperatures in Eilat do not decrease much below 21 °C (average lowest daily sea surface temperature from 1988 to 1995 was 20.7 °C with a range of 0.7 °C, compared with 11.2 °C with a range of 4 °C for the minimum air temperatures). Palaeoclimate techniques using stable isotopes in marine sediments and corals[2,22] would therefore be insensitive to such small interannual variations in local water temperatures. On the other hand, our findings indicate that records of massive eutrophication and spring blooms in the Gulf of Eilat, which are likely to be found in fossil corals[23], may become useful indicators of interannual variations in air temperature. □

Received 21 July; accepted 6 September 1995.

1. Robock, A. & Mao, J. *Geophys. Res. Lett.* **19,** 2405–2408 (1992).
2. Reiss, Z. Hottinger, L. *The Gulf of Aqaba, Ecological Micropaleontology* (Springer, Berlin, 1984).
3. Murray, S. P., Hecht, A. & Babcock, A. *J. mar. Res.* **42,** 265–287 (1984).
4. Paldor, N. & Anati, D. A. *Deep-Sea Res.* **26,** 661–672 (1979).
5. Wolf-Vecht, A., Paldor, N. & Brenner, S. *Deep-Sea Res.* **39,** 1393–1401 (1992).
6. Knauss, J. A. *Introduction to Physical Oceanography* (Prentice-Hall, New Jersey, 1978).
7. Benayahu, Y. & Loya, Y. *Proc. 3rd Int. Coral Reef Symp.* (ed. Taylor, D. L.) 383–389 (Univ. Miami, 1977).
8. Sverdrup, H. U. *J. Cons. perm. Int. Explor. Mer.* **18,** 287–295 (1953).
9. Levitus, S. *Climatological Atlas of the World Ocean* (NOAA Prof. Pap. 13, US Dept of Commerce, Washington DC, 1982).
10. Venrick, E. L. *Limnol. Oceanogr.* **38,** 1135–1149 (1993).
11. Guzman, H. M., Cortes, J., Glynn, P. W. & Richmond, R. H. *Mar. Ecol. Prog. Ser.* **60,** 299–303 (1990).
12. Jokiel, P. L. & Coles, S. L. *Mar. Biol.* **43,** 201–208 (1977).
13. Luhr, J. F. *Nature* **354,** 104–105 (1991).
14. Bluth, G. J. S., Doiron, S. D., Schnetzler, C. C., Krueger, A. J. & Walter, L. S. *Geophys. Res. Lett.* **19,** 151–154 (1992).
15. Graf, H.-F., Kirchner, I., Robock, A. & Schult, I. *Clim. Dyn.* **9,** 81–93 (1993).
16. Hansen, J., Lacis, A., Ruedy, R. & Sato, M. *Geophys. Res. Lett.* **19,** 215–218 (1992).
17. Dutton, E. & Christy, J. R. *Geophys. Res. Lett.* **19,** 2313–2316 (1992).
18. Graf, H.-F., Perlwitz, J. & Kirchner, I. *Beitr. Phys. Atmosph.* **67,** 3–13 (1994).
19. Brenner, S., Rosentraub, Z., Bishop, J. & Krom, M. *Dyn. Atmos. Oceans* **15,** 457–476 (1991).
20. Halpert, M. S. *et al. Eos* **74,** 433–438 (1993).
21. Kirchner, I. & Graf, H.-F. *Max-Planck-Inst. Mete. Rep.* **121,** 1–57 (1993).
22. Dunbar, R. B. & Wellington, G. M. *Nature* **293,** 453–455 (1981).
23. Shen, G. T. *et al. Paleoceanography* **7,** 563–588 (1992).
24. Venrick, E. L., & Hayward, T. L. *Calif. Coop. ocean. Fish. Invest Rep.* **25,** 74–79 (1984).
25. Wyrtki, K. *J. geophys. Res.* **70,** 4547–4559 (1965).
26. Kondo, J. *Bound. Layer Met.* **9,** 91–112 (1975).
27. Mellor, G. L. & Yamada, T. *J. Atmos. Sci.* **31,** 1791–1806 (1974).
28. Mellor, G. L. & Yamada, T. *Rev. Geophys. Space Phys.* **20,** 851–875 (1982).

ACKNOWLEDGEMENTS. We thank I. Ayalon, D. Lindell, T. Liberman, G. Gal, J. Erez, J. Bowman, A. Nehorai, T. Neubirth, L. Karp, M. Kiflawi, S. Ramati, P. Sulkis, R. Grossman and H. Borochov for assistance; the Israel Meteorological Service for air-temperature data; the Interuniversity Institute of Eilat for logistic support; the Coral-World Observatory for access to their pier and Station Zoologique, Villefranche-sur-mer; and L. Haury, P. Falkowsky, D. Genin, H.-F. Graf and G. Gorsky for comments. A.G. was supported by the France–Israel Arc en Ciel during the writing of this Letter. This work was supported by the Wolfson Foundation, the Israel Academy of Sciences, the US-Israel Binational Science Foundation, the Bat Sheva de Rothschild Fund, and the Israel Ministry for the Environment.

A 20,000-year record of ocean circulation and climate change from the Santa Barbara basin

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MUCH of the evidence for climate-driven fluctuations in ocean circulation during the past 20,000 years has come from studies of the North Atlantic region[1–6]. The extent to which such interactions have occurred in other ocean basins, and any associated teleconnections between basins, is poorly understood. Here we present high-resolution palaeoclimate and palaeoceanographic records from a 20,000-year sedimentary sequence from the Santa Barbara basin, on the eastern margin of the North Pacific Ocean. The sequence shows oscillations of the benthic environment between low-oxygen conditions (laminated sediments) during periods of warm climate, and higher-oxygen conditions (non-laminated, bioturbated sediments) during cool intervals. Age differences between coexisting benthic and planktonic foraminifers indicate climate-related changes in the age and source—and, hence, oxygen content—of basin bottom waters. Relatively young bottom waters are associated with the cooler intervals and are considered to reflect high proportions of intermediate waters derived from proximal sources. Conversely, older bottom waters are associated with the warmer

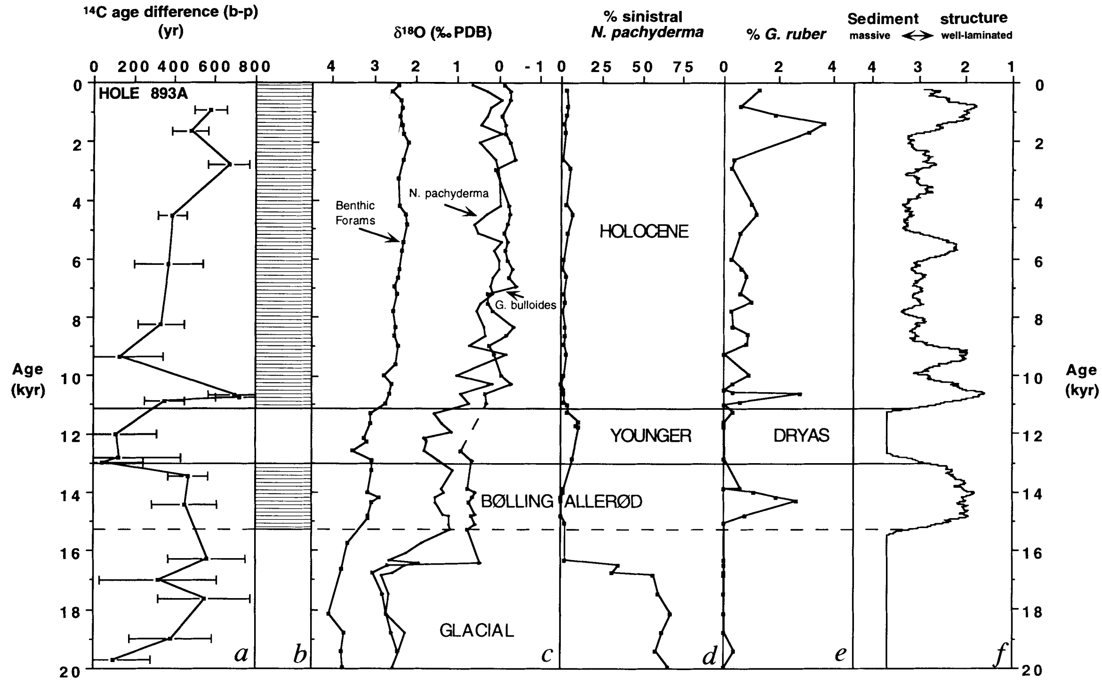

FIG. 1 Relations between inferred age of basinal waters (*a*), sediment structures (*b*, *f*) and palaeoclimate proxies (*c*, *d*, *e*) for the past 20 kyr (upper 32 m of Ocean Drilling Program Hole 893A, Santa Barbara basin, southern California. *a*, Benthic–planktonic (b–p) age differences (^{14}C) are considered to represent changes in age of basinal bottom waters[12]. Reservoir-age correction used throughout is 825 yr (ref. 12); insufficient information exists to adjust for probable changes in reservoir ages with time. Chronology after refs 12 and 16. *b*, Presence (shaded) or absence (unshaded) of sediment laminations. *f*, Visual description of sediment structures at one-cm resolution for the entire sequence[26], grading from well-laminated to massive (1, well-laminated; 2, indistinctly laminated; 3, trace laminations; 4, massive or non-laminated). The curve in *f* was smoothed using a 99-cm running average to reduce variability and exhibit more clearly the major events addressed here. Smoothing has little effect on age of sediment facies transitions. These changes have been correlated to changes in degree of oxygenation and sediment facies of Santa Barbara basin[26]. Palaeoclimate proxies shown are: c, $\delta^{18}O$ records of the deeper-dwelling planktonic foraminifer *Neogloboquadrina pachyderma*, the shallower-dwelling form *Globigerina bulloides* and benthic foraminifers[16,25]; *d*, *e*, percentage frequency changes in planktonic foraminifer assemblages (>150 μm) of the cool-water planktonic foraminifer, sinistral *N. pachyderma* (*d*) and the warm subtropical form, *Globigerinoides ruber*[17] (*e*). No single benthic species occurs throughout the sequence due to large palaeoenvironmental changes[16]. We isotopically analysed *Bolivina argentea*, *B. spissa*, *B. tumida*; *Uvigerina peregrina curticosta* and *Bulliminella tenuata*. We made interspecific oxygen-isotope comparisons and corrected *B. tumida* and *B. tenuata* by +0.25‰ relative to the other species[16]. Changes in palaeoclimate proxies[25] reveal the palaeoclimate episodes: Last Glacial Maximum, Bølling/Allerød interstadial, Younger Dryas cooling and Holocene warming.

intervals and were derived from more distal sources. These climate-driven variations in ocean circulation appear to be synchronous with the main ocean–climate fluctuations in the North Atlantic region[1–6], suggesting that a tight coupling mechanism operates between the two basins.

The study site was continuously cored at 34° 17.25′ N, 120° 2.2′ W, 20 km south of the Santa Barbara coastline in a water depth of 588 m. We recovered a ~160-kyr, 196.5-m sequence of alternating laminated and homogenous (non-laminated) sediments[8]. This sequence is especially valuable for palaeoclimate analyses because of persistently high sedimentation rates (>120 cm kyr^{-1}) and near-continuous foraminifer assemblages abundant enough to generate high-quality stable-isotope, faunal census and radiocarbon age data. Santa Barbara basin is partially isolated within the California Borderland Province[7]. The present-day western sill (475 m) is at a depth of low oxygen concentrations (<0.8 ml l^{-1}) in the upper part of the oxygen-minimum zone[9]. Organic material derived from highly productive surface waters further reduces basin oxygen levels, and waters below ~500 m depth are anaerobic (<0.1 ml l^{-1} oxygen), inhibiting bioturbation and preserving annual sediment laminations[8,9]. In contrast, recent and Holocene sediments lack laminations at similar depths on the California Margin outside the Borderland Province[10,11]. The past 20 kyr was dated using 32 accelerator mass spectrometry (AMS) radiocarbon ages of mixed planktonic foraminifera from 27 stratigraphic levels[12]. We calibrated radiocarbon ages to calendar years using two different methods[13–15], assuming a reservoir age of 825 yr (ref. 12). The samples range in age from 1,025 to 17,555 yr BP (reservoir-corrected) at intervals of ⩽1,000 yr, yielding[12] a near-linear sedimentation rate of 160 cm kyr^{-1}.

A total of 70 samples were taken for stable-isotope and faunal analyses at ~50-cm intervals in the upper 25 m (the last 16.6 kyr) and at ~100-cm intervals for the remaining core down to 32 m. Variations in $\delta^{18}O$ values of both planktonic and benthic foraminifers[16] and planktonic foraminiferal census data[17] exhibit climate oscillations characteristic of the Last Glacial Maximum and throughout the Holocene (Fig. 1) Four distinct climate episodes correlate with the well-known climate succession in northern Europe ranging from the Last Glacial Maximum (20–16 kyr) to the Bølling/Allerød interstadial (~15–13 kyr), to the Younger Dryas cooling (13–11.2 kyr), to the Holocene (<11.2 kyr). High $\delta^{18}O$ values mark the Last Glacial Maximum. Warming associated with termination 1A, the most recent deglacial episode, is diachronous between several proxies from 16.5 to 15 kyr. Initial warming is shown by a rapid switch at 16.5 kyr in planktonic foraminiferal faunas dominated by sinistral-coiled *Neogloboquadrina pachyderma* to warmer faunas with a high proportion of dextral-coiled *N. pachyderma*[17] (Fig. 1). The coiling ratio of *N. pachyderma* also switched from >90% sinistral to <10% sinistral. This was immediately followed by a 1.75‰ decrease in $\delta^{18}O$ in the deeper-dwelling planktonic foraminifer

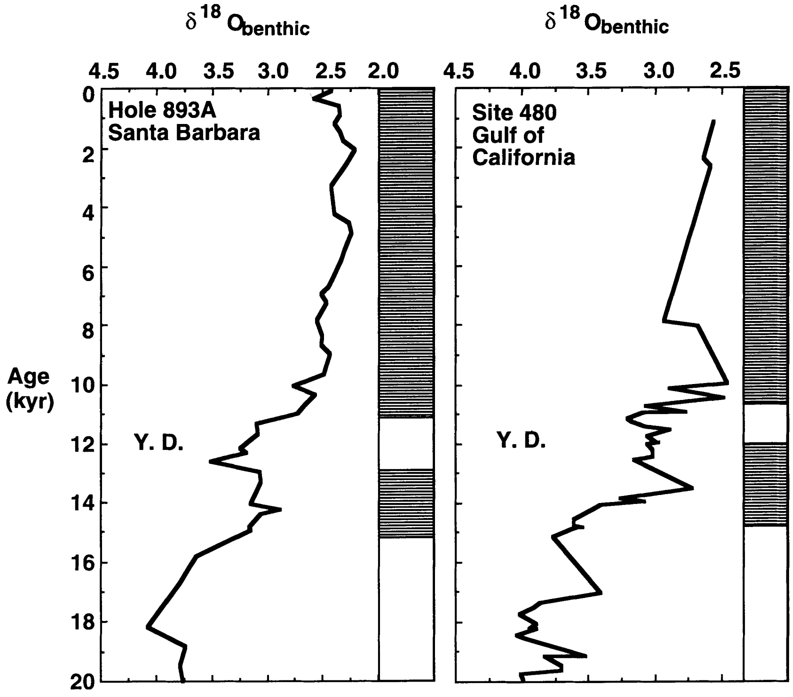

FIG. 2 Comparison of $\delta^{18}O$ records of benthic foraminifers and presence (shading) and absence (no shading) of sediment laminations between ODP Hole 893A (Santa Barbara basin) and DSDP Site 480 (Guaymas basin, Gulf of California)[29]. Younger Dryas (YD) cooling is indicated by reversal towards increased $\delta^{18}O$ values in both sequences. Data plotted in calendar years. Calendar years for Site 480 calculated using a reservoir-age correction[13] of 920 yr for original radiocarbon ages[29]. Reservoir-age correction for Hole 893A is[12] 825 yr.

N. pachyderma. These changes indicate climate warming associated with northward migration of surface waters warmer than ~10 °C across the site during termination 1A.

Relatively low $\delta^{18}O$ values between 15 and ~13 kyr record warmth correlated with the Bølling/Allerød interstadial. Further increases in warm-water planktonic foraminifers such as *Globigerinoides ruber* (Fig. 1), reflect increasing influence of the warm-subtropical countercurrent from the south. The Bølling/Allerød interstadial was succeeded by moderate cooling between 13 and 11.2 kyr that correlates closely with the Younger Dryas cooling episode of northern Europe[18]. In Hole 893A, Younger Dryas cooling is indicated by increases of $\delta^{18}O$ values in benthic foraminifers (0.5‰) and *N. pachyderma* (0.75‰), and increased relative abundances of cool-water planktonic foraminifers. Sinistral *N. pachyderma* exhibits a tenfold increase, and the coiling ratio of this species changed from ~98 to 80% dextral[17]. Full glacial conditions did not return during the Younger Dryas but are similar to those that mark the mid-point of termination 1A. Average sea surface temperatures remained above, but close to, 10 °C.

We compare our ages of the Younger Dryas with those elsewhere using either a "reservoir-corrected" age (^{14}C age, 825 yr) and a calendar age (a calibrated ^{14}C age). In Hole 893A, the reservoir-corrected ^{14}C age for the beginning of the Younger Dryas is[12] 11,155 yr BP which is within 45–150 yr of radiocarbon ages for this event in North Atlantic sediment cores[19,20] and New Zealand glacial deposits[21]. The calendar-year (calibrated) age of 12,970 BP at Santa Barbara basin[12] is remarkably close to an age of 12,940±260 BP based on annual layer counts in the Greenland ice sheet[22,23]. The reservoir-corrected (uncalibrated) age for the end of the Younger Dryas in Hole 893A is[12] 9,805 yr BP compared with the commonly accepted age[24] of ~10,000 radiocarbon years. The calibrated age is poorly constrained because of its occurrence during a 1,300-yr plateau on the calibration curve[15] which occurred[12] between 10,900 and 12,300 yr BP. The calendar age for the end of the Younger Dryas in the Greenland ice sheet is[22,23] 11,640±250 yr BP. For Hole 893A, we use a calendar age of 11,200 yr BP based on interpolation between dated samples[12,25].

The end of Younger Dryas cooling is marked by decreased $\delta^{18}O$ values in benthic and planktonic foraminifers. Oxygen-isotope values in benthic foraminifers and *N. pachyderma* decreased during the early Holocene until ~6 kyr (Fig. 1) in contrast to *Globigerina bulloides* which exhibits similar, although fluctuating Holocene $\delta^{18}O$ values after 10.3 kyr. The reappearance of warm-water planktonic foraminifers marks the end of the Younger Dryas (Fig. 1). Warm subtropical forms persisted during the remainder of the Holocene[17], and indicating continuous influence of a poleward-flowing, warm countercurrent[7]. Throughout the Holocene (Fig. 1), climate fluctuated but was always warmer than the Younger Dryas[17,25]. Basinal sediment facies is closely associated with climate change (Fig. 1), and the presence, absence and degree of biological disruption of laminations is controlled by changes in oxygen concentration of bottom waters[8,26]. Laminations remain preserved when oxygen concentrations fall below 0.1–0.2 ml l^{-1} (ref. 27). The degree of preservation of laminations is inferred to be related to small changes in bottom-water oxygen concentrations and different degrees (millimetre-scale) of meiofaunal bioturbation[26], with non-laminated (massive) sediments indicating complete bioturbation beneath oxygenated waters. Benthic foraminiferal assemblages in laminated sediments are dominated by taxa typical of low-oxygen environments (*Bolivina* ⩽90%; *Suggrunda* present). Non-laminated sediments contain a more diverse assemblage associated with well-oxygenated bottom waters[25]. Assemblages from the massive Last Glacial Maximum sediments are usually dominated by *Epistominella* with *Nonionellina*, *Nonionella* and *Cassidulina*. During the Younger Dryas (non-laminated), the diversity of benthic foraminifers is intermediate between assemblages of the glacial maximum and the Holocene. Faunas are dominated by *Epistominella*, and *Uvigerina* is a consistent and sometimes dominant element. During non-laminated intervals, the low-oxygen-tolerant forms *Bolivina* and *Suggrunda* are absent-to-rare and echinoid spines (spatangids) are present. These spines are absent in laminated sediments. In the modern basin, echinoid spines are only present beneath waters with oxygen concentrations of ⩾0.3 ml l^{-1} (interpreted from ref. 27).

Hole 893A records four major oscillations between laminated and non-laminated sediments during the past 20 kyr that have been semi-quantitatively described at one-cm resolution[26] (Fig. 1). The switch to laminated sediments at 15.2 kyr is closely associated with the beginning of the Bølling/Allerød interstadial (Fig. 1). Likewise, the climate changes at the beginning and end of the Younger Dryas are almost synchronous with the switches between laminated and non-laminated sediments (Fig. 1). The stratigraphic resolution of our data is such that if any lag exists between climate and ventilation change, it would be $\leqslant 200$ yr (Fig. 1). The Bølling/Allerød interstadial is marked by consistently strong laminations; in contrast, the Holocene sediments oscillate between weak and strong laminations[26] (Fig. 1). Our data is thus consistent with the hypothesis of synchronous to near-synchronous changes between global climate and ventilation in Santa Barbara basin.

The Santa Barbara basin sequence is not merely a record of local climate and oceanographic change, neither was it primarily controlled by local bathymetric configuration. Changes in basin ventilation are unlikely to have resulted directly from sea-level change because Younger Dryas cooling was associated with a pause rather than a fall in sea level[28]. Furthermore, ventilation changes in Santa Barbara basin do not primarily record varying geographical or bathymetric isolation from the productive California Current because the facies changes during the Younger Dryas occurred without significant eustatic sea-level-controlled changes in basin configuration. Similar oxygenation-related facies changes are also recorded in DSDP Site 480 in Guaymas basin, Gulf of California[29] (Fig. 2) suggesting widespread changes in more isolated northeastern Pacific margin basins during the past 20 kyr. DSDP Site 480 (655 m water depth), on the slope of Guaymas basin, is not isolated by a sill at shallow depths and hence was not affected by changes in sea level[28]. Poor ventilation in Santa Barbara basin seems to result from a delicate balance between warm deep waters, sufficiently high plankton productivity and relatively low oxygen concentrations of upper intermediate waters spilling into the basin from the Pacific. No consistent evidence exists from micropalaeontological[17,30] and organic-carbon accumulation rate[31,32] data for higher productivity during deposition of laminated versus non-laminated intervals. Furthermore, the benthic δ^{18}O record (Fig. 1) indicates the absence of any simple relations between changes in bottom-water temperatures and oxygenation in the basin. We believe that the cycles of oxygenation/dysaerobia in the basin were largely controlled by changes in oxygen concentrations of upper intermediate waters entering the basin, as has been suggested for the Gulf of California[29]. Any restriction in circulation enhances small palaeo-oxygen changes to a greater degree than at locations on open continental margin slopes[10,11]. Thus we suggest that relatively small changes in dissolved oxygen content in upper intermediate waters in the northeastern Pacific during the late Quaternary are amplified in Santa Barbara basin. Oxygen-poor intermediate water presently flowing into Santa Barbara basin over the western sill has[33] a radiocarbon age of 1,200–1,300 yr compared with 825 yr for the surface mixed layer, which is a mix of California Current and upwelled Pacific Intermediate Water[12]. Water as old as 1,980 yr has been reported at depths of ~3,000 m in the North Pacific[34].

If the sediment cycles in Santa Barbara and Guaymas basins were closely linked with changes in ocean circulation during the past 20 kyr (Fig. 1), relations should exist between climate change and age of basin bottom waters. Changes in relative ages of bottom waters are inferred from intrasample differences in radiocarbon ages (uncorrected) between planktonic and benthic foraminifers (Fig. 1). The radiocarbon age differences vary between 40 and 740 radiocarbon years, with the smallest average ^{14}C age differences (~100 yr) associated with the Last Glacial Maximum (19.6 kyr) and the Younger Dryas (13–11.2 kyr). In contrast, the average age difference for sample pairs from the Bølling/Allerød interstadial (16.5–13 kyr) is 493 yr. Likewise, during Holocene warmth, radiocarbon age differences, although variable, are relatively high (average 453 yr). Changes in sediment facies are generally correlated to the radiocarbon age differences of basin bottom waters during the past 20 kyr, especially at the beginning and end of the Younger Dryas. In contrast, an increase in radiocarbon age difference (to ~550 yr) of bottom waters at the end of the last glacial episode preceded by ~3 kyr the preservation of laminations during the Bølling/Allerød interstadial (Fig. 1).

We believe that the close relations between changes in climate and age of basin bottom waters indicate that the sediment cycles were largely caused by changes in source and age of intermediate waters in the North Pacific. Also, the changes in oxygenation in Santa Barbara basin appear to be nearly synchronous with major changes in North Atlantic Deep Water (NADW) circulation[1–6]. The Santa Barbara basin was relatively well oxygenated during the Last Glacial Maximum and the Younger Dryas when NADW production was reduced[1–6] and/or when its southward flow through the Atlantic was at shallower depths[35,36]. Conversely, when the flux of NADW in the Atlantic was high during the Holocene and Bølling/Allerød warm intervals[1–6] and/or when it flowed at greater depths[35,36], the basin was poorly oxygenated. The stratigraphic records indicate that the switches between the modes were closely synchronized between the North Pacific and North Atlantic.

We suggest two possible mechanism(s) to explain the near-synchronism of changes between the two oceans. The most likely hypothesis is that the inter-ocean palaeoceanographic changes were linked directly through global climate change transmitted through the atmosphere. In this case, cooling during the Last Glacial Maximum and the Younger Dryas increased ventilation of North Pacific Intermediate Waters[37–40]. This provided a greater contribution of proximal young, well-oxygenated waters to the northeastern Pacific coast margin. The second hypothesis indirectly implicates changes in the strength of thermohaline circulation originating as NADW in the North Atlantic—the so-called oceanic conveyor circulation[41]. Although the flow of water within the conveyor from the North Atlantic to North Pacific takes ~1,000 yr, changes in flux would be propagated rapidly to the North Pacific (J. McWilliams, personal communication). In this model, reduced contributions to the North Pacific of relatively old[34] waters originally derived from the North Atlantic led to greater proportions of younger, more oxygenated waters entering Santa Barbara basin. Variations in the age of basin waters (Fig. 1) would result from mixing of the two sources. Irrespective of the process(es) involved, our data indicate the existence of a tight coupling (of circumpolar extent in the Northern Hemisphere) between changes in the atmosphere–ocean–cryosphere during the latest Quaternary. □

Received 16 January; accepted 30 August 1995.

1. Boyle, E. A. & Keigwin, L. D. *Science* **218,** 784–787 (1982).
2. Curry, W. B. & Lohmann, G. P. *Quat. Res.* **18,** 218–235 (1982).
3. Broecker, W. S., Peteet, D. M. & Rind, D. *Nature* **315,** 21–26 (1985).
4. Boyle, E. A. & Keigwin, L. D. *Nature* **330,** 35–40 (1987).
5. Keigwin, L. D., Jones, G. A., Lehman, S. J. & Boyle, E. A. *J. geophys. Res.* **96,** 16811–16826 (1991).
6. Lehman, S. J. & Keigwin, L. D. *Nature* **356,** 757–762 (1992).
7. Emery, K. O. *The Sea Off Southern California: A Modern Habitat of Petroleum* (Wiley, New York, 1962).
8. Kennett, J. P. *et al. Init. Rep. ODP Leg* **146,** Part 2 (1994).
9. CalCOFI Data Report. Cruise 9401 and 9403 (*Univ. California Scripps Institution of Oceanography Ref.* 94-21, 1994).
10. Gardner, J. V. & Hemphill-Haley, E. *Geology* **14,** 691–694 (1986).
11. Hemphill-Haley, E. & Gardner, J. V. *Quat. Res.* **41,** 311–335 (1994).
12. Ingram, B. L. & Kennett, J. P. *Proc. ODP Sci. Res.* **146** (pt 2), 19–27 (1995).
13. Stuiver, M. & Braziunas, T. F. *Radiocarbon* **35,** 137–189 (1993).
14. Bard, E., Hamelin, B., Fairbanks, R. G. & Zindler, A. *Nature* **345,** 405–410 (1990).
15. Bard, E., Arnold, M., Fairbanks, R. G. & Hamelin, B. *Radiocarbon* **35,** 191–199 (1993).
16. Kennett, J. P. *Proc. ODP Sci. Res.* **146** (pt 2), 3–18 (1995).
17. Kennett, J. P. & Venz, K. *Proc. ODP Sci. Res.* **146** (pt 2), 281–293 (1995).
18. Mangerud, J., Anderson, S. T., Berglund, B. E. & Donner, J. J. *Boreas* **3,** 109–128 (1974).
19. Duplessy, J. C. in *Global Changes of the Past* (ed. Bradley, R. S.) 341–355 (Nat. Center Atmos. Res., Interdisc. Earth Stud., Boulder, CO, 1989).
20. Lehman, S. J. & Keigwin, L. D. *Nature* **356,** 757–762 (1992).
21. Denton, G. H. & Hendy, C. H. *Science* **264,** 1434–1437 (1994).
22. Alley, R. B. *et al. Nature* **362,** 527–529 (1993).
23. Johnson, S. J. *et al. Nature* **359,** 311 (1992).
24. Kennett, J. P. *Paleoceanography* **5,** 891–895 (1990).
25. Kennett, J. P. & Ingram, B. L. *Proc. ODP Sci. Res.* **146** (pt 2), 309–325 (1995).
26. Behl, R. J. *Proc. ODP Scient. Res.* **146** (pt 2), 295–308 (1995).
27. Savrda, C. E., Bottjer, D. J. & Gorsline, D. S. *Bull. Am. Ass. Petrol. Geol.* **68,** 1179–1192 (1984).
28. Fairbanks, R. G. *Nature* **342,** 637–642 (1989).
29. Keigwin, L. D. & Jones, G. A. *Paleoceanography* **5,** 1009–1023 (1990).

30. Hemphill-Haley, E. & Fourtanier, E. *Proc. ODP Sci. Res.* **146** (pt 2), 233–248 (1995).
31. Stein, R. & Rack, F. R. *Proc ODP, Sci. Res.* **146** (pt 2), 125–138 (1995).
32. Gardner, J. V. & Dartnell, P. *Proc. ODP Sci. Res.* **146** (pt 2), 103–124 (1995).
33. Ostlund, H. G. & Stuiver, M. *Radiocarbon* **22,** 25–53 (1980).
34. Broecker, W. S. & Peng, T. H. *Tracers in the Sea* 690 (Lamont-Doherty Geol. Observ., Palisades, NY, 1982).
35. Rahmstorf, S. *Nature* **372,** 82–85 (1994).
36. Boyle, E. & Weaver, A. *Nature* **372,** 41–42 (1994).
37. Keigwin, L. D. *Nature* **330,** 362–364 (1987).
38. Keigwin, L. D., Jones, G. A. & Froelich, P. N. *Earth planet. Sci. Lett.* **111,** 425–440 (1992).
39. Van Scoy, K. A., Olsen, D. B. & Fine, R. A. *J. geophys. Res.* **96,** 16801–16810 (1991).
40. Van Scoy, K. A. & Druffel, E. R. M. *J. geophys. Res.* **98,** 18083–18088 (1993).
41. Broecker, W. S. & Denton, G. H. *Geochim. cosmochim. Acta* **53,** 2465–2501 (1989).

ACKNOWLEDGEMENTS. We thank several colleagues for making useful suggestions, including R. Behl, L. Keigwin, E. Boyle, M. Kashgarian and J. Southon; R. Behl for criticizing the manuscript; and K. Thompson, S. Richard, I. Hendy and H. Berg for technical assistance. J.P.K. was supported by the US NSF, the US DOE NIGEC-WESTGEC and JOI/USSAC. B.L.I. was supported by US DOE and the Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory.

Glacial–interglacial changes in nutrient utilization in the equatorial Pacific Ocean

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THE eastern equatorial Pacific Ocean (EEP) today sustains up to 30% of global marine productivity[1], and the region is one of the largest and most variable marine sources of CO_2 to the atmosphere[2]. This variability is largely controlled by the balance between the physical input of nutrients to the surface ocean and their removal by biological assimilation—the relative nutrient utilization—but the spatial and temporal variability of this balance are poorly understood. Here we use the $^{15}N/^{14}N$ ratio in sedimentary marine organic matter to show strong spatial gradients in relative nitrate utilization throughout the modern EEP. We interpret down-core decline in this ratio through the Last Glacial Maximum (12–24 kyr ago) as a decrease in relative nitrate utilization; the increase in nitrate supply to surface waters due to upwelling during this period was greater than the apparent increase in nitrogen removal by organic matter export out of surface waters. This interpretation is consistent with cooler sea surface temperatures[3] and a higher CO_2 flux to the atmosphere[4,5] during the Last Glacial Maximum, indicating that the EEP surface waters have remained enriched in nutrients, and have not acted as a net sink for CO_2, for at least the past 30,000 years.

A promising method for tracking past changes in relative NO_3^- utilization is to determine variations in the ratios of the stable isotopes of nitrogen in sedimentary organic matter[6,7]. The efficiency with which phytoplankton assimilate NO_3^- from near-surface waters is reflected in the $\delta^{15}N$ values (in ‰ relative to atmospheric N_2: see Fig. 1) of particulate organic matter (POM)[8,9], some fraction of which is ultimately preserved in deep-sea sediments. This isotope proxy is based on the observation that POM $\delta^{15}N$ depends on the $\delta^{15}N$ of the nitrogenous substrate and on isotope fractionation that occurs during nitrogen uptake by phytoplankton[10]. Deep-water NO_3^- is the principal source of new nitrogen in the surface ocean. It is the dominant nitrogenous species for new production in the EEP[11] where equatorial divergence and strong coastal upwelling commonly result in concentrations in excess of 5 µM (refs 12, 13). In much of this region, upwelled NO_3^- has[14] an initial $\delta^{15}N$ of ~5 to 6‰. But in certain low-oxygen regions within the northern and southern tropical eastern Pacific, the $\delta^{15}N$ values can exceed 18‰ because denitrification enriches the remaining NO_3^- in ^{15}N (refs 14, 15).

After NO_3^- is delivered to the photic zone, there is preferential uptake of $^{14}NO_3^-$ by phytoplankton, giving rise to a photosynthate enriched in the light isotope, and to a residual dissolved NO_3^- pool that is correspondingly and progressively enriched in ^{15}N. This has led to the observation of an inverse relationship between the near-surface $[NO_3^-]$ and the $\delta^{15}N$ of POM in the

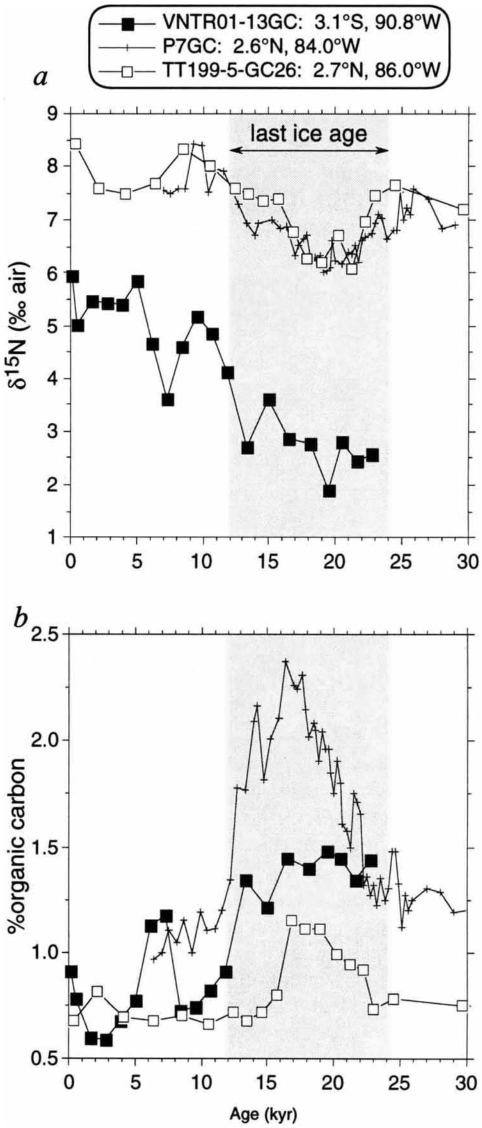

FIG. 1 *a*, *b*, Spatial distribution of surface (Holocene) sediment $\delta^{15}N$ values (*a*) compared with surface ocean $[NO_3^-]$ (*b*; ref. 13) in the eastern tropical Pacific Ocean. $\delta^{15}N$ values are from this study (available via e-mail from J.W.F. at jfarrell@brook.edu) and ref. 7. Bulk sediment samples were prepared for $\delta^{15}N$ analysis by freeze-drying and homogenizing by grinding. Sediments were combusted in an online Fisons NA 1500 element analyser and the evolved N_2 was passed to a VG PRISM isotope-ratio mass spectrometer in a continuous flow of He. Results are reported in the δ notation, $\delta^{15}N = [(^{15}N/^{14}N)_{sample}/(^{15}N/^{14}N)_{standard} - 1]$ per mil, relative to atmospheric N_2 and the measurement precision is better than ±0.3‰. We have assumed that the isotopic composition of the total nitrogen primarily reflects that of marine organic matter, rather than inorganic nitrogen (ammonium) within clay minerals and adsorbed atmospheric N_2. Initial results from Panama basin sediments indicate that inorganic nitrogen constitutes a minor proportion of the total. As drawn, the $\delta^{15}N$ isopleths violate 5% of samples by >0.3‰.

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