PALEOCEANOGRAPHY, VOL. 28, 453–467, doi:10.1002/palo.20043, 2013

Quantified intermediate water oxygenation history of the NE Pacific: A new benthic foraminiferal record from Santa Barbara basin

K. Ohkushi,[1] J. P. Kennett,[2] C. M. Zeleski,[3] S. E. Moffitt,[4] T. M. Hill,[5] C. Robert,[6]
L. Beaufort,[6] and R. J. Behl[3]

Received 3 May 2012; revised 22 July 2013; accepted 26 July 2013; published 30 August 2013.

[1] The oxygen minimum zone (OMZ) of the late Quaternary California margin experienced abrupt and dramatic changes in strength and depth in response to changes in intermediate water ventilation, ocean productivity, and climate at orbital through millennial time scales. Expansion and contraction of the OMZ is exhibited at high temporal resolution (107–126 year) by quantitative benthic foraminiferal assemblage changes in two piston cores forming a vertical profile in Santa Barbara Basin (569 m, basin floor; 481 m, near sill depth) to 34 and 24 ka, respectively. Variation in the OMZ is quantified by new benthic foraminiferal groupings and new dissolved oxygen index based on documented relations between species and water-mass oxygen concentrations. Foraminiferal-based paleoenvironmental assessments are integrated with principal component analysis, bioturbation, grain size, $CaCO_3$, total organic carbon, and $\delta^{13}C$ to reconstruct basin oxygenation history. Fauna responded similarly between the two sites, although with somewhat different magnitude and taxonomic expression. During cool episodes (Younger Dryas and stadials), the water column was well oxygenated, most strongly near the end of the glacial episode (17–16 ka; Heinrich 1). In contrast, the OMZ was strong during warm episodes (Bølling/Allerød, interstadials, and Pre-Boreal). During the Bølling/Allerød, the OMZ shoaled to <360 m of contemporaneous sea level, its greatest vertical expansion of the last glacial cycle. Assemblages were then dominated by *Bolivina tumida*, reflecting high concentrations of dissolved methane in bottom waters. Short decadal intervals were so severely oxygen-depleted that no benthic foraminifera were present. The middle to late Holocene (6–0 ka) was less dysoxic than the early Holocene.

Citation: Ohkushi, K., J. P. Kennett, C. M. Zeleski, S. E. Moffitt, T. M. Hill, C. Robert, L. Beaufort, and R. J. Behl (2013), Quantified intermediate water oxygenation history of the NE Pacific: A new benthic foraminiferal record from Santa Barbara basin, *Paleoceanography*, *28*, 453–467, doi:10.1002/palo.20043.

1. Introduction

[2] Widespread shifts in oceanic oxygenation occurred across the Pacific Ocean basin during the last glacial cycle in concert with climatic and oceanographic change [*Keigwin and Jones*, 1990; *Behl and Kennett*, 1996; *van Geen et al.*, 2003; *Cook et al.*, 2005; *Davies et al.*, 2011]. In particular, hypoxia was widely associated with the deglacial warming episodes following Termination 1A and 1B [*Zheng et al.*, 2000], and in more restricted locations, with the preceding Dansgaard/Oeschger climatic oscillations. These deoxygenation events have been especially well documented in the northeastern Pacific, from the Mexican margin to the Gulf of Alaska [*Keigwin and Jones*, 1990; *Behl and Kennett*, 1996; *Zheng et al.*, 2000; *Crusius et al.*, 2004]. The cause of the hypoxia is controversial and several explanations have been proposed, although these may not be mutually exclusive. These hypotheses include: changes in pathways and composition of intermediate water [*Behl and Kennett*, 1996; *van Geen et al.*, 1996; *Keigwin*, 1998; *Zheng et al.*, 2000], basinal changes in water column stratification [*Schmittner et al.*, 2007], changes in productivity triggered by strength of upwelling and related productivity [*Crusius et al.*, 2004; *Ortiz et al.*, 2004; *Barron et al.*, 2009], composition of upwelled water, or availability of limiting micronutrients [*Mix et al.*, 1999; *Davies et al.*, 2011].

Additional supporting information may be found in the online version of this article.

[1] Graduate School of Human Development and Environment, Kobe University, Kobe, Japan.

[2] Department of Earth Sciences and Marine Science Institute, University of California, Santa Barbara, California, USA.

[3] Department of Geological Sciences, California State University Long Beach, Long Beach, California, USA.

[4] Graduate Group in Evolution and Ecology and Bodega Marine Laboratory, University of California, Davis, California, USA.

[5] Department of Geology and Bodega Marine Laboratory, University of California, Davis, California, USA.

[6] CEREGE-CNRS, Aix Marseille University, Aix-en-Provence, France.

Corresponding author: R. J. Behl, Department of Geological Sciences, California State University, 1250 Bellflower Blvd., Long Beach, CA 90840, USA. (behl@csulb.edu)

0883-8305/13/10.1002/palo.20043

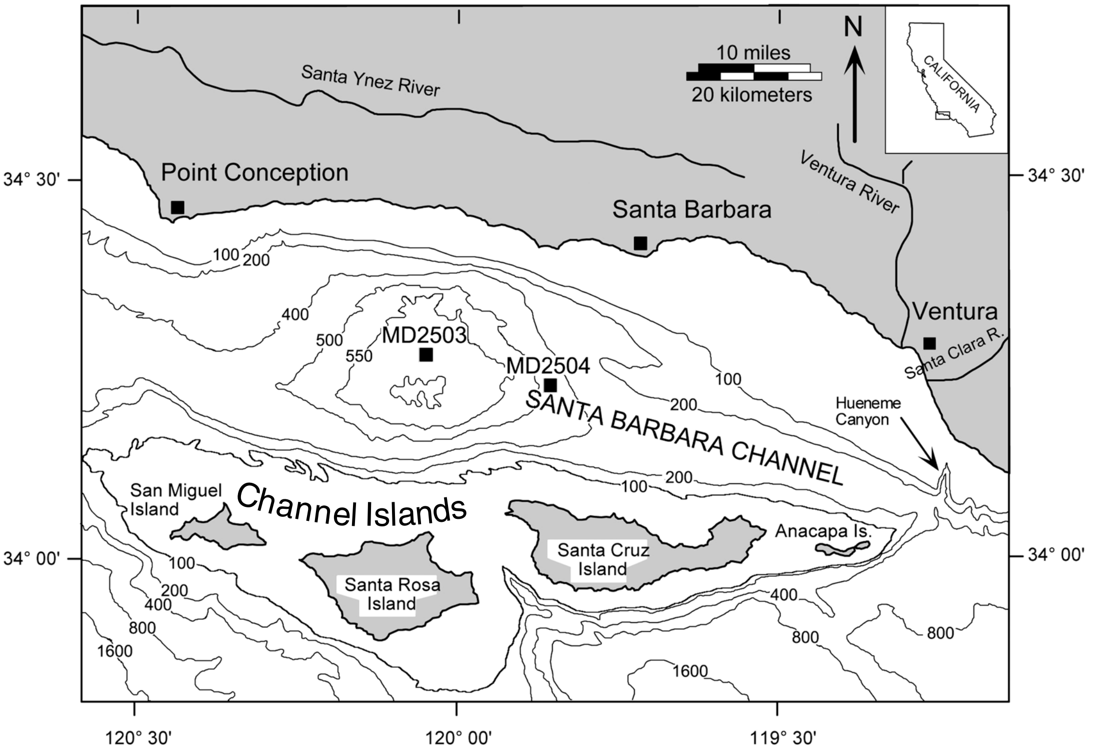

Figure 1. Locations of cores MD02-2503 and MD02-2504 in Santa Barbara Basin, to the north of the Channel Islands. Bathymetry in meters.

[3] Santa Barbara Basin (SBB), located on the northeast Pacific margin, is the northernmost basin of the California Borderland [*Emery*, 1960]. The ~600 m deep basin, bounded by the North America continent to the north and Channel Islands to the south (Figure 1), opens to upper North Pacific intermediate water (NPIW) to the west over a 475 m deep sill and to the east over a 230 m deep sill [*Emery*, 1960; *Reid*, 1965; *Talley*, 1993]. Currently, low-oxygen water (to <0.8mL/L O_2, average 0.5 mL/L; 2008–2011) entering the basin over the western sill is derived from the oxygen minimum zone (OMZ) (California Cooperative of Fisheries Investigations) [*Soutar and Crill*, 1977; *van Geen et al.*, 1996]. Basin water is further depleted of oxygen by degradation of organic matter from highly productive surface waters. Dysoxic conditions (<0.1 mL/L) in the deep basin below 525 m exclude benthic macrofauna, thus preventing bioturbation [*Emery and Hulsemann*, 1962; *Harman*, 1964; *Douglas*, 1981] and preserving annual sedimentary laminations [*Thunell et al.*, 1995; *Behl and Kennett*, 1996; *Nederbragt and Thurow*, 2005]. During the Last Glacial Maximum (LGM), sea level was about 125 m below that of the present day and four of the Channel Islands were merged as a single, larger island, further restricting circulation with the open Pacific Ocean.

[4] The linkage between oxygenation history and climate and water column change is well documented in Santa Barbara Basin, where a ~200 m core was recovered from Ocean Drilling Program (ODP) Site 893 in 1992 [*Kennett*, 1995]. Because of the exceedingly high sedimentation rate (>1 m/kyr) and exquisite preservation of minimally bioturbated or laminated sediment [*Behl*, 1995], numerous climatic and paleoceanographic investigations of the late Quaternary have been conducted at high temporal resolution in SBB (ODP 893A and subsequent piston cores; 577 m water depth) [*Behl and Kennett*, 1996; *Hendy et al.*, 2002; *Hendy and Kennett*, 1999, 2000, 2003; *Kennett and Ingram*, 1995; *Kennett et al.*, 2000, 2003; *Hill et al.*, 2006a]. These studies provided paleoclimatic records at resolution similar to those from the Greenland Ice Sheet (Greenland Ice Sheet Project (GISP2) and Greenland Ice Core Project records) [e.g., *Dansgaard et al.*, 1993; *Stuvier and Grootes*, 2000]. They established that Northeastern Pacific climate fluctuations closely matched that of Greenland with the well-known sequence of Dansgaard/Oeschger (D/O) cycles, exhibiting large (up to 8°C) average sea surface temperature shifts on the California margin, between stadial/interstadial and glacial/interglacial episodes [*Hendy and Kennett*, 2000; *Pak et al.*, 2012].

[5] The D/O climatic cycles recorded in northeastern Pacific continental margin sediments are intimately associated with expansions and contractions of the oxygen minimum zone (OMZ) due to changes in surface ocean productivity and ventilation of upper intermediate waters at depths shallower than ~1000 m. Warm intervals are marked by poorly oxygenated, older, nutrient-rich with warmer intermediate waters; colder intervals by well-oxygenated, younger, nutrient-poor with colder intermediate water [*Cannariato et al.*, 1999; *Hendy and Kennett*, 2003; *Murphy and Thomas*, 2010]. The effects of ventilation switching in upper intermediate waters are highly amplified in SBB and reflected by large changes in sedimentary fabric [*Behl and Kennett*, 1996] and benthic foraminiferal assemblages [*Cannariato et al.*, 1999]. The record from ODP Site 1017, deeper and further offshore than SBB, suggests that ventilation switching extends to 1000 m in this sector of the California margin [*Cannariato and Kennett*, 1999].

[6] In this investigation, we have quantitatively examined benthic foraminiferal assemblages in two late Quaternary cores retrieved from SBB (Figure 1) and analyzed the same cores for total organic carbon (TOC), $CaCO_3$, and grain size. Core MD02-2503 is from the central SBB (569 m), adjacent to ODP Site 893, and provides a sediment sequence from the most oxygen-depleted depths. The second core (MD02-2504) was taken from a shallower depth (481 m) to study variations in the upper OMZ, close to the depth of the western sill (~475 m). Foraminiferal changes are compared with records of climatic change and sediment composition and texture, allowing evaluation of other potential environmental factors known to influence benthic foraminiferal assemblages. Collectively with ODP Site 1017, these three records provide a 500 m vertical depth transect to better understand the changing strength of the OMZ in response to orbital through submillennial climatic oscillations.

[7] Benthic foraminifera are particularly suitable for monitoring abrupt climate change because many species are known to be opportunistic and rapidly respond to environmental change, including that related to seafloor oxygenation, organic matter flux, and current strength [*Bernhard and Reimers*, 1991; *Corliss and Silva*, 1993; *Sen Gupta and Machain-Castillo*, 1993; *Jorissen et al.*, 1995, 2007; *Bernhard*, 1992; *Bernhard et al.*, 1997; *Koho et al.*, 2008]. Most relevant to this investigation, *Cannariato and Kennett* [1999] and *Cannariato et al.* [1999] earlier demonstrated the sensitivity of benthic foraminiferal species and assemblages for monitoring changes in oxygen levels and consequently the extent and the strength of the OMZ on the California margin. The current investigation in SBB extends this work in five ways: (1) quantification of benthic assemblages at much higher temporal resolution than previously published, (2) comprehensive quantitative analyses of assemblages to include all benthic foraminiferal taxa, (3) employment of hydrographic data to establish environmental tolerances of taxa to oxygenation concentrations, (4) upward extension of the depth transect to sill depth by the addition of core MD02-2504, and (5) comparison of changes in the foraminiferal assemblages with geochemical and grain-size data to better understand basin ventilation and the potential contribution of other environmental variations on the benthic faunal record. With this investigation, we aim to better understand changes in basin oxygenation and in the intensity and vertical distribution of the OMZ and their relations with orbital through millennial climate change.

2. Materials and Methods

[8] The benthic foraminiferal assemblage record was generated through a census of 311 samples in MD02-2503 and 193 samples in MD02-2504 with a minimum sampling resolution of 20 cm intervals throughout both cores. For the foraminiferal census, sediment samples were disaggregated in water, washed over a 63 mm sieve, and oven dried at 50°C. Foraminiferal assemblages were quantified by typically counting 100–400 specimens at the species level. To assess change in benthic oxygen concentration, we use only 19 taxa of 87 counted with >1% abundance with published hydrographic data showing their environmental tolerances and niches (Table S4). Seven species are associated with dysoxic conditions (<0.1 mL/L), nine species with suboxic conditions (0.1–0.5 mL/L), and three species with weakly hypoxic-oxic conditions (>1.5 mL/L). The ratio of planktic/benthic foraminiferal abundance was also calculated for a portion of each core. Species diversity (H) was calculated from the benthic foraminiferal data using the Shannon index [*Gage and Tyler*, 1991]. Q-mode Principal Component Analysis was conducted for statistical comparisons of benthic foraminiferal assemblage changes.

[9] Splits from the same 20 cm spaced samples were used for the analysis of weight percent total organic carbon (TOC) and calcium carbonate ($CaCO_3$) using a CE-440 Exeter Analytical CHN (carbon, hydrogen, nitrogen) analyzer. Textural variation of the <63 μm sediment fraction was analyzed using a Micromeritics Sedigraph 5100 to determine the median grain size and silt percentage. Detailed methods for foraminiferal assemblage identification, census counts, species diversity analyses, and geochemical and grain-size analysis are presented in the supporting information.

3. Age Model

[10] Age models for cores MD02-2503 and MD02-2504 have previously been established by correlating foraminiferal δ^{18}O with the well-dated ODP Site 893 [*Hill et al.*, 2006a, 2006b]. A chronology for ODP Site 893 was developed using 70 ^{14}C ages for the last 30 kyr [*Hendy et al.*, 2002]. The δ^{18}O records are so similar between ODP 893 and the Greenland Ice Sheet (GISP2) that major climate transitions are considered to have been synchronous between the SBB and Greenland via changes transmitted through the atmosphere [*Behl and Kennett*, 1996; *Hendy and Kennett*, 1999; *Hendy et al.*, 2002; *Hill et al.*, 2006a]. Another, ultra-high-resolution study employing 430 additional ^{14}C dates between 18 and 14 ka in ODP 893 provides an independent chronology based on a sequence of radiocarbon plateaus for the early deglacial interval [*Sarnthein et al.*, 2007] that is largely consistent with previous age models and makes ODP 893 one of the best-dated marine successions anywhere.

[11] As expected, a strong similarity exists between the oxygen isotopic records for MD02-2503 and ODP 893 [*Hill et al.*, 2006a,2006b] and correlations are made directly to the ODP 893 time scale [*Hendy et al.*, 2002; *Hill et al.*, 2006a, 2006b]. Six primary tie points were selected based upon similarities in the δ^{18}O and sediment lamination records between closely adjacent MD02-2503 and ODP 893: Termination IB (11.7 ka), the onset of the Younger Dryas (12.9 ka), Termination IA (14.7 ka), Interstadial 2 (23.4 ka), Interstadial 4 (29.1 ka), and Interstadial 5 (32.4 ka). Cores MD02-2503 and 2504 were correlated using planktic foraminiferal δ^{8}O records [*Hill et al.*, 2006a, 2006b] and by tie points in the magnetic susceptibility and geochemical records [*Zeleski*, 2004], a revision to previously published age models. The identification of a distinct dysoxic event associated with Interstadial 2 in the benthic foraminiferal record of MD02-2504 provides an additional tie point between MD02-2503/ODP 893 and MD02-2504 for the lower part of the studied interval. The resulting chronology indicates that MD02-2503 extends 0 to 34 ka and MD02-2504 from 0 to 24 ka, with average sedimentation rates of ~135 and 159 cm/kyr, respectively.

4. Results

4.1. Core Descriptions

[12] Core MD02-2503 (34.2862°N; 120.0365°W; 569 m water depth) recovered 46.8 m of sediment from near the center of the basin (Figure 1), immediately adjacent to ODP Site 893. The sediment is predominantly silty clay containing abundant microfossils and nannofossils. The core is mostly laminated from the core top to 15 mbsf and from 18.5 to 21.5 mbsf. It is mostly massive below except for five thin laminated intervals between 33 and 46 mbsf.

[13] Core MD02-2504 (34.2333°N; 119.868°W; 481 m water depth; Figure 1) recovered 38.7 m of sediment. The dominant sediment consists of silty clay to clayey silt containing nannofossils and foraminifers. The sediment is generally homogeneous throughout, with a few clearly burrowed intervals. Laminated sediment in this core occurs in only two intervals from 15.1 to 19.5 mbsf and 35.6 to 36.0 mbsf. A 40 cm thick interval at 9.5 mbsf was disturbed by coring.

4.2. Benthic Foraminiferal Assemblages

[14] Changes in the relative abundances (frequency abundance) of benthic foraminiferal species in the late Quaternary sequences are shown in Figures S1–S3, as are their relationships to TOC, $CaCO_3$, and grain size. The relative abundances of benthic foraminifera in the two cores (Figure S1) show strong similarity in trends even though each sequence contains different relative abundances of taxa. Assemblages in MD02-2503 reflect changes in basin-floor conditions during the last 34 kyr, whereas assemblages in MD02-2504 record environments close to sill depth (currently ~475 m) over the last 24 kyr.

[15] Benthic foraminiferal assemblages in both cores changed dynamically on decadal through orbital timescales, and both the general and detailed patterns of change are well correlated between the cores (Figures S2 and S3). Benthic assemblages are distinctly different between warm and cold intervals. Frequent faunal changes during the Holocene are unassociated with major climate change and not matched in the relatively stable patterns of $\delta^{18}O$ exhibited in the SBB [*Hendy and Kennett*, 1999] and Greenland records [*Dansgaard et al.*, 1993; *Stuvier and Grootes*, 2000] (Figures 4 and 5). Blackman-Tukey spectral analysis of both cores shows weak cyclicity in % dysoxic species with 1350–1380 year periodicity during the Holocene (Figure S4).

[16] The numbers of counted specimens of benthic foraminifera are shown in Figure 2a and listed in Tables S1 and S2. Ideally, 300 specimens should be counted for each sample. However, several laminated levels contain few to no foraminifera, and consequently, 109 of the 504 total samples with <50 specimens (Tables A1 and A2) were included in the census. Many samples with low benthic abundances contain well-preserved planktic foraminifera, indicating that postdepositional benthic dissolution was not a significant factor and that the numbers reflect past seafloor environmental conditions rather than poor preservation (Figure S5). The overall pattern of changing benthic foraminiferal abundance (number of specimens) is similar between the two cores (Figure 2a) and also corresponds well to variation in the weight % $CaCO_3$ (Figure 2b), reflecting the predominance of benthic foraminifera in the sedimentary carbonate fraction. There is strong similarity of major trends between the ODP Site 893 and MD02-2503 records.

[17] The number of species counted (simple diversity) and the species diversity (Shannon index) in each sample are shown in Figures 2c and 2d. The time series shows large fluctuations in diversity ($H = 0$ to 3). These are well matched between the two cores except during the early Holocene (9.5 to 8 ka; Figure 2c) and in parts of the LGM. Thus, it appears that these data reflect widespread changes in simple species diversity in SBB during the latest Quaternary. With the exception of a few intervals in the Holocene, higher diversity and greater species number are generally associated with cooler intervals and lower diversity and number with warmer intervals. The species diversity at the deeper site (MD02-2503) is low in the early Holocene, B/A, and IS3-IS6 but especially from 11 to 9 ka. However, in the shallower site close to sill depth (MD02-2504), diversity values are high (~2) in the same interval. This reflects strong vertical differences in diversity at this time with low species diversity at the more dysoxic deeper site. After 9 ka, species diversity gradually increased with similar trends exhibited in each of the cores in response to increased basin oxygenation.

[18] Planktic/benthic foraminiferal ratios (P/B) peaked in the Pre-Boreal and B/A (Figure 2e), reflecting decreased production of benthic foraminifera. Intervals of high P/B ratio are most closely related to intervals of relative warmth and high carbonate content (Figure 2). However, several negative spikes of P/B ratio are exhibited at the onset of abrupt warming episodes IS2-6, B/A, and Pre-Boreal. These spikes may reflect major reduction in planktic foraminiferal production during the onset of abrupt warming and/or increase in sea surface salinity.

[19] For better understanding of the history of basin oxygenation in MD02-2503 and MD02-2504, we grouped species into assemblages ranked according to how they reflect bottom water oxygenation (Table S4 and Figures 3 and 5) following the example of recent studies that have moved toward quantitative interpretations of foraminiferal assemblages [e.g., *Jorissen et al.*, 2007]. These rankings were established using available hydrographic data showing oxygen ranges and tolerance for each species (Table S4 and references therein; *Moffitt et al.*, in review). These groupings build on (and broadly agree with) rankings previously established by *Cannariato and Kennett* [1999] and *Cannariato et al.* [1999] and are further supported by Principle Components Analysis, discussed below.

[20] The Dysoxic assemblage contains *Nonionella stella*, *Bolivina tumida*, *Buliminella tenuata*, *Bolivina pacifica*, *Bolivina pseudobeyrichi*, *Cassidulina delicata*, and *Cassidulina limbata*. This assemblage is considered to be associated with severely dysoxic conditions ($O_2 = <0.1$ mL/L), even more so than in the modern basin ($O_2 = 0.1–<2$ mL/L), in which *Bolivina tumida* is not observed and *N. stella* is the dominant foraminifera [*Bernhard and Bowser*, 1999]. This assemblage grouping is based upon a combination of lowest reported oxygen concentration where the individual species were observed (0.1 to 0.02 ml/L, depending upon species [*Harman*, 1964; *Douglas and Heitman*, 1979; *Mackensen and Douglas*, 1989; *Jannink et al.*, 1998; *Bernhard et al.*, 2001; *Bernhard et al.*, 2010]) and associations with methane-rich environments, preserved laminations, and symbionts or sequestered plastids [*Bernhard and Bowser*, 1999; *Cannariato et al.*, 1999; *Bernhard et al.*, 2000; *Bernhard et al.*, 2001; *Stott et al.*, 2002; *Hill et al.*, 2003]. This grouping replaces the Dysoxic I and II assemblages of *Cannariato and Kennett* [1999].

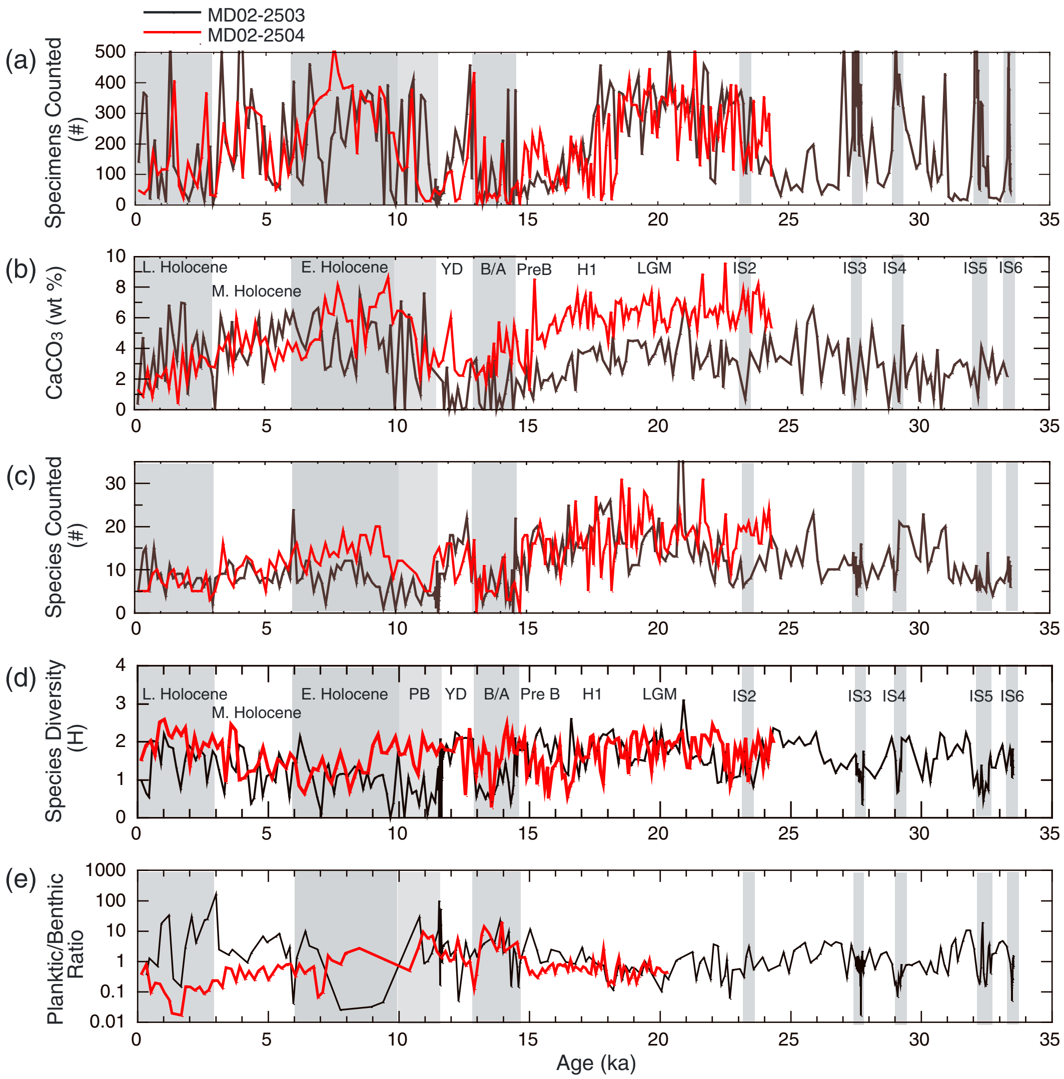

Figure 2. (a) Numbers of counted specimens, (b) weight % calcium carbonate, (c) number of counted species of benthic foraminifera in samples from cores MD02-2503 and MD02-2504, plotted by age, (d) the Shannon index (*H*) of benthic foraminifera, and (e) the abundance ratio of planktic foraminifera/benthic foraminifera. Climatic intervals indicated as L. Holocene: Late Holocene; M. Holocene: Middle Holocene; E. Holocene: Early Holocene; PB: Pre-Boreal warming of earliest Holocene; B/A: Bølling-Allerød warming; preB: pre-Bølling warming; H1: Heinrich cold event 1; LGM: Last Glacial Maximum, and IS: Interstadial warm episodes 6 through 2.

[21] The suboxic assemblage ($O_2 = 0.1$ to 0.5 mL/L) contains *Globobulimina* spp., *Bolivina argentea*, *Bolivina spissa*, *Cancris* sp., *Cassidulina carinata*, *Epistominella pacifica*, *Epistominella smithi*, *Uvigerina peregrina curticosta*, and *Valvulineria araucana*. This grouping is based upon the lowest reported oxygen concentration where individual species were observed (from 0.2 to 0.5 mL/L for most species [*Smith*, 1964; *Blake*, 1976; *Douglas and Heitman*, 1979; *Quinterno and Gardner*, 1987; *Mackensen and Douglas*, 1989; *Mullins et al.*, 1985; *Maas*, 2000]), in addition to documented associations with the open slope OMZ, methane-rich environments, laminated sediments, and/or presence of symbionts or sequestered plastids [*Ingle and Keller*, 1980; *Douglas*, 1981; *Quinterno and Gardner*, 1987; *Bernhard et al.*, 1997; *Cannariato et al.*, 1999; *Rathburn et al.*, 2000; *Bernhard et al.*, 2001; *Hill et al.*, 2003]. This grouping replaces Suboxic I and II groups found in *Cannariato et al.* [1999].

[22] The Weakly Hypoxic-Oxic assemblage contains *Nonionella labradorica*, *Pyrgo* spp., and *Quinqueloculina* spp. We consider this assemblage to represent O_2 concentrations >1.5 mL/L, based upon lowest reported oxygen concentrations for these three species [*Alve*, 1990; *Cedhagen*, 1991; *Kaiho*, 1994] and the observation that these are thick-walled, predation-resistant, motile fauna more likely to occur outside of the OMZ [*Linke and Lutze*, 1993; *Kaiho*, 1994].

[23] To assist in differentiating the taxonomic groupings making up these communities, we conducted a principal component analysis using 87 species with abundances of more than 1% in at least one sample. Principal component analysis indicates that 56% of the total variance can be expressed by three principal components (PC) (supporting information appendix).

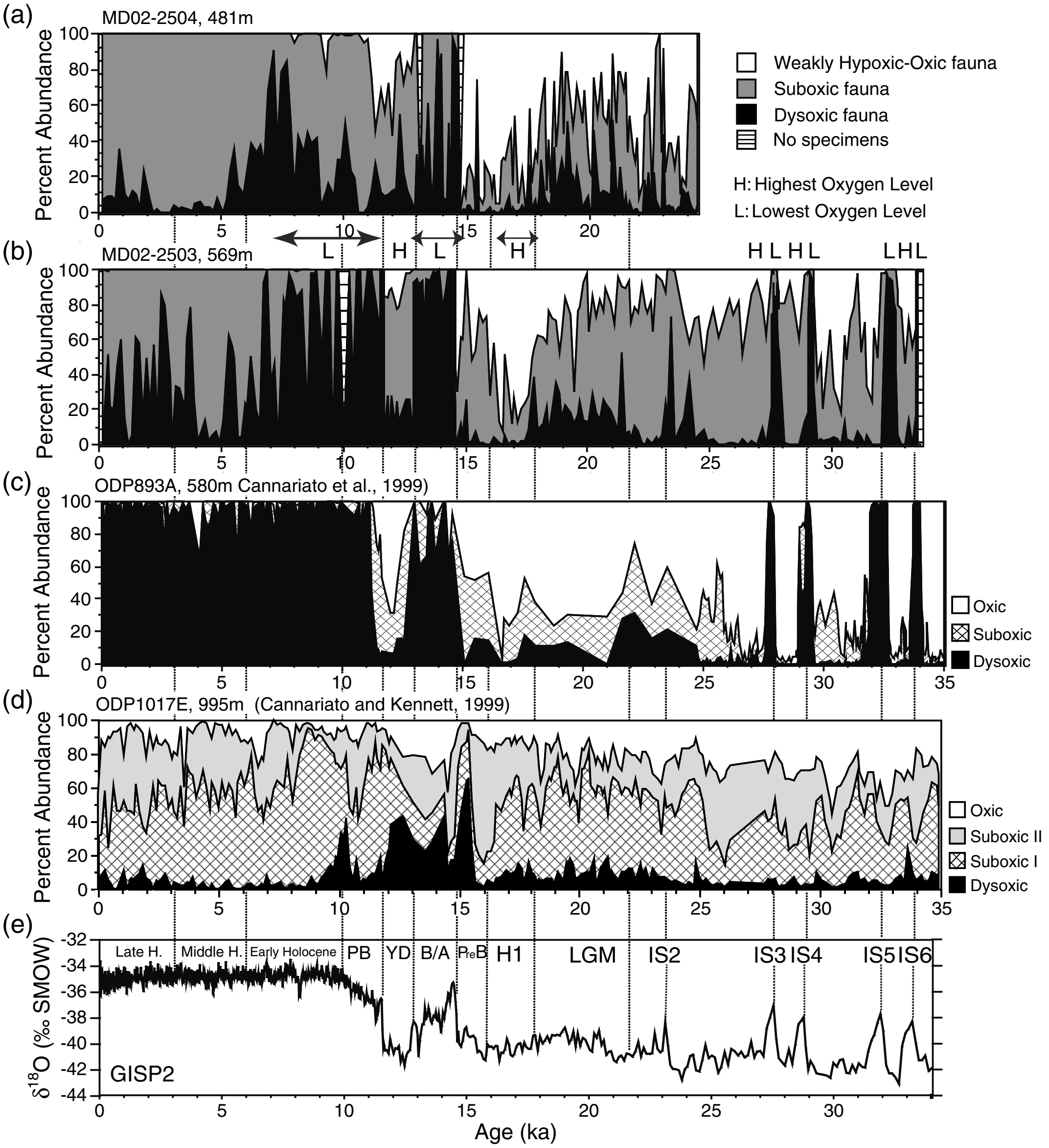

Figure 3. Changes in relative percent abundances of benthic foraminifera as three oxygenated-ranked groups: Weakly hypoxic-oxic, Suboxic, and Dysoxic in late Quaternary cores (a) MD02-2504 and (b) MD02-2503, compared with benthic foraminiferal assemblages in (c) ODP 893A in the SBB [*Cannariato et al.*, 1999] and (d) ODP 1017E off Point Conception [*Cannariato and Kennett*, 1999], and (e) Greenland (GISP2 $\delta^{18}O$) climatic record. Stratigraphic distribution of individual species in both MD cores is shown at high resolution in Figures S2 and S3. Abbreviations as in Figure 2.

PC1 explains 22% of the variance with *Bolivina tumida* displaying the highest loadings, supported by other high loadings in *Bolivina argentea*, *Buliminella tenuata*, and *Uvigerina peregrina curticosta* (Table S5). Thus, PC1 reflects variations in the Dysoxic and Suboxic assemblages. Trends in PC1 are similar between the two cores with especially high values in MD02-2503 during the early Holocene, B/A, and IS3 to 6, reflecting the most extreme, poorly oxygenated conditions in this sequence (Figures 4 and 5).

[24] PC2 explains 19% of the variance, with the highest loadings exhibited by *Nonionellina labradorica*, supported by high loadings in *Epistominella pacifica* and *U. peregrina curticosta*. These associations indicate that PC2 reflects suboxic, weakly hypoxic, and oxic environments, with highest values exhibited during cold episodes including the LGM, H1, and YD (Figure 4). PC2 exhibits similar values in the LGM, H1, YD, and the late early to early middle Holocene between the two cores. However, PC2 in MD02-2504 reveals higher values during the B/A, PB, and late Holocene compared with MD02-2503, implying the presence of higher oxygen levels at the shallower sill-depth site.

[25] PC3 explains 15% of the variance, with the highest loadings in *Bolivina argentea*, supported by high loadings in *U. peregrina curticosta*, and low loadings in *Nonionellina labradorica*. Thus, PC3 reflects a transitional assemblage between dysoxic and suboxic but is lacking the extreme dysoxic indicator species, *B. tumida*. This assemblage likely reflects conditions similar to the modern OMZ in SBB. PC3

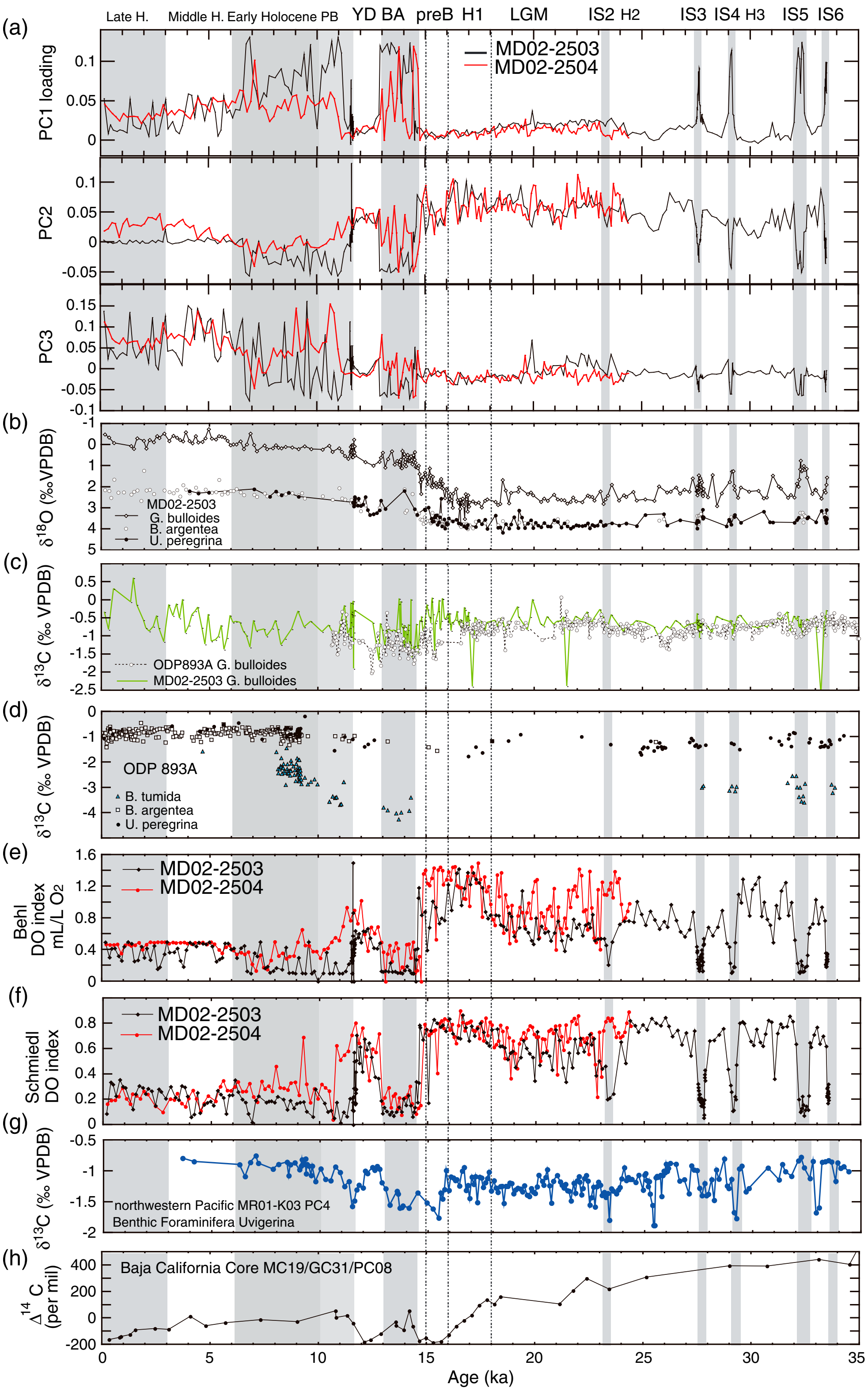

Figure 4. (a) Time series changes in score values of three PC in principal component analysis of benthic foraminifera in cores MD02-2503 and MD02-2504 compared with foraminiferal $\delta^{18}O$ records of (b) MD02-2503 [*Hill et al.*, 2006a], (c) planktic foraminiferal $\delta^{13}C$ records of ODP 893A and MD02-2503 [*Hill et al.*, 2006b; *Hendy and Kennett*, 2003], (d) benthic foraminiferal $\delta^{13}C$ records of ODP 893A [*Kennett et al.*, 2000], (e) Behl dissolved oxygen (DO) index in cores MD02-2503 and MD02-2504, (f) Schmiedl DO index in cores MD02-2503 and MD02-2504, (g) benthic foraminiferal $\delta^{13}C$ records in core MR01-K03 PC4 in northeastern Pacific [*Hoshiba et al.*, 2006], and (h) $\Delta^{14}C$ intermediate waters off Baja California [*Marchitto et al.*, 2007]. Abbreviations as in Figure 2.

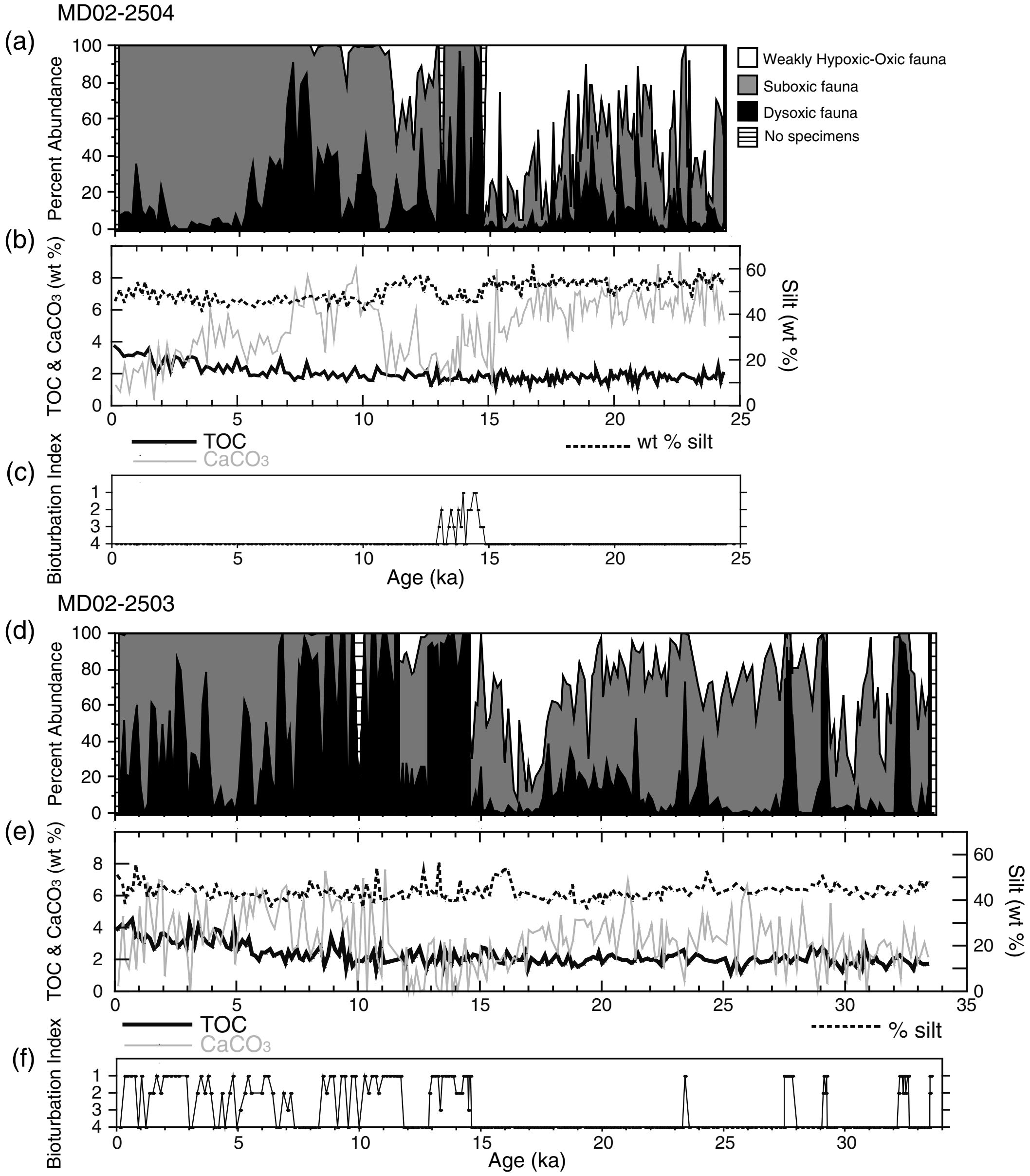

Figure 5. Changes in relative percent abundances of benthic foraminiferal assemblages grouped into oxygenated-ranked groups for cores (a) MD02-2504 and (d) MD02-2503. Patterns and abbreviations as in Figure 3. Percentages total organic carbon (TOC), calcium carbonate ($CaCO_3$), silt (grain size) for cores (e) MD02-2503 and (b) MD02-2504 and bioturbation index for (f) MD02-2503 and (c) MD02-2504: 1 = well-laminated sediments, 2 = indistinctly laminated sediments, 3 = sediments with trace laminations, 4 = bioturbated sediments [*Behl*, 1995].

also reveals intermittent especially high Holocene values in MD02-2503, reflecting millennial and possibly centennial-scale dysoxic episodes in the deep basin during this interval (Figure 4).

4.3. TOC, $CaCO_3$, Silt Content, and Bioturbation Index

[26] Sedimentary geochemistry and texture also vary through the record, with similarities between the two cores (Figure 5). TOC (1.1 to 4.5 wt.% for MD02-2503, 1.1 to 3.7 wt.% for MD02-2504) displays a very similar pattern in both cores, gradually increasing from the last glacial episode though the late Holocene. No distinct fluctuations in TOC are evident between stadial and interstadial intervals, or at glacial terminations; a slight and gradual increase in TOC is observed toward the late Holocene (Figure 5). $CaCO_3$, likewise, is quite similar in both cores (0 to 7.6 wt.% for MD02-2503, 0.4 to 9.6 wt.% for MD02-2504) but displays more variation in relation to climatic change (Figures 2 and 5). $CaCO_3$ is more abundant through the entire glacial interval to about 17 ka, followed by a decrease to a minimum between 15 and 11 ka (from Termination 1A to 1B), increasing again to values greater than the glacial episode by 10 ka and persisting at higher levels through the early Holocene. The two cores differ somewhat through the remainder of the Holocene, with carbonate remaining high in MD02-2503, but decreasing to the present in core MD02-2504. As with

TOC, $CaCO_3$ does not evidently vary with stadial-interstadial oscillations during the last glacial interval (Figures 2 and 5).

[27] Grain size of the siliciclastic portion of the sediment is quite different between the two core locations (Figure 5). In MD02-2503, grain size of the <63 µm fraction (37% to 57% silt) is relatively constant for the past 34 kyr but displays short-lived peaks (<few hundred years), generally during cold intervals at approximately 24, 16, 13, 12, 10, and 1 ka. In contrast, at shallower site MD02-2504, the sediment is slightly coarser (41% to 61% silt) and displays a distinct pattern related to climatic changes. The sediment is consistently coarser during the colder intervals associated with the entire glacial and YD intervals from 24 to 14.5 ka and 12.9 to ~11 ka (Figure 5). Sediment becomes distinctly finer during the B/A from 14.7 to 12.9 ka and throughout the Holocene from ~11 ka to present. In some cases, individual benthic foraminiferal species become more abundant in association with increased grain size during colder intervals that likely reflected lower sea level in the shallower and more proximal core MD02-2504 (Tables S2 and S3 and Figure 5).

[28] The presence and relative strength of lamination/bioturbation was semiquantitatively described according to the bioturbation index of *Behl* [1995]. Bioturbation index 1 represents well-laminated sediments, index 2 represents indistinctly laminated sediments, index 3 represents sediments with trace laminae, and index 4 represents massive or strongly bioturbated sediments. Bioturbation index 1 represents deposition and preservation of fine laminations under 0–0.1 mL/L O_2 and index 2 between ~0.1 and 0.2 mL/L O_2 [*Savrda and Bottjer*, 1991; *Diego and Douglas*, 1999]. In the basin-center core MD02-2503, well-laminated sediments are associated with warm intervals including the Holocene, B/A, and IS2 to 6. Strongly bioturbated sediments are associated with cold episodes including the LGM, H1, and YD. At the shallower core site MD02-2504, distinct laminations are only preserved during the B/A and IS2. The remainder of the core is massive to distinctly burrowed.

5. Discussion

[29] We interpret that changes in the benthic foraminiferal assemblages and sediment fabric during the late Quaternary in Santa Barbara Basin have been primarily controlled by oscillations in basin oxygenation. However, foraminiferal assemblages are indicators of a suite of environmental and hydrographic conditions including benthic ecosystem oxygenation, export productivity, and changing deep-water masses [*Jorissen et al*., 2007], all of which can be genetically related. Assemblages can respond to both changing influence of food source *and* seafloor oxygenation and is difficult, and maybe impossible, to distinguish the influence of the physiochemical parameter of oxygenation from the resource parameter of organic matter flux [*Jorissen et al*., 2007] in many cases. To prevent mischaracterization of seafloor conditions, we chose a marker species approach wherein a suite of species were selected because of known associations with modern hydrographic data determined during live faunal collections (Table S4). In addition, we assessed the marker species associations with modern marginal and depositional environments and key physiological characteristics (e.g., wall-thickness, motility, the presence of symbionts or the presence of sequestered plastids).

[30] We note that there are no clear fluctuations in TOC across stadial/interstadial or glacial/interglacial transitions in these cores; thus, it is unlikely that foraminifera responded solely to increases in organic carbon export to the seafloor. Although organic matter concentration is not equivalent to organic matter flux, numerous well-dated studies concluded that late Quaternary changes in sedimentation rate in SBB were minor [*Behl and Kennett*, 1996; *Hendy et al*., 2002; *Roark et al*., 2003; *Sarnthein et al*., 2007] thus consistency in TOC would reflect relatively steady organic matter flux or preservation. This interpretation is supported by trace elemental and isotopic data, indicating that bottom water oxygenation in SBB was dominantly controlled by changes in ventilation, not productivity across the last deglaciation. This evidence further suggests that significant increases in productivity for the last glacial cycle began within the Holocene [*Ivanochko and Pedersen*, 2004].

[31] We used two methods to quantitatively estimate relative changes of O_2 concentrations: one new and one previously published. First, we estimated O_2 concentrations by using a weighted average of the concentration for each oxygen category as 0.1 mL/L for Dysoxic, 0.5 mL/L for Suboxic, and 1.5 mL/L for Weakly hypoxic-oxic.

$$\text{Behl DO index} = (\text{Dysoxic}\% \times 0.1 + \text{Suboxic}\% \times 0.5 + \text{Weakly hypoxic} - \text{oxic}\% \times 1.5)/100.$$

[32] In the case of higher percentages of the Weakly hypoxic-oxic category, the oxygenation values may be underestimated. In the case of higher percentages of the Dysoxic category, oxygenation values may be overestimated.

[33] For comparison, we also employ the DO index of *Schmiedl et al.* [2003] based on the following formula: Schmiedl index = (HO/(HO + LO) + Div) × 0.5, with HO = relative abundance of high oxygen indicators, LO = relative abundance of low-oxygen indicators, and Div = normalized value of species diversity (H). In the present study, HO represents the Weakly hypoxic-oxic assemblage and LO the Dysoxic assemblage. The resulting trends in oxygen levels are similar using the two methods, although small differences are evident (Figures 4e and 4f), most notably that the Behl DO index exhibits higher values during the most oxygenated intervals.

5.1. Oxygen Levels on the SBB Basin Floor

[34] The foraminiferal and sedimentary results exhibit four events of extreme dysoxia between 34 and 26 ka that are associated with Interstadials 6 to 3. In each, oxygen levels fell rapidly to ~0.1–0.2 mL/L; then, after a few centuries, oxygen levels abruptly increased and persisted through the intervening stadial intervals (Figures 3 and 4). Some of these transitions took place in as little as 70 years. Variability was more moderate from 26 to 19 ka, when oxygen levels fluctuated in the suboxic range except for a transient dysoxic event at IS2 (23.4 ka). In the later half of this interval (LGM), oxygen concentration slightly decreased. During H1 (18–16 ka), some of the highest extended oxygenation levels of the record occurred (>1.5 mL/L) leaving no evidence for the OMZ in SBB. This was followed by increases in suboxic fauna indicating two pulses of deoxygenation during the pre-Bølling, before a brief return to weakly hypoxic-oxic conditions at 15 ka.

[35] An abrupt decrease in oxygen at the onset of the B/A marks the most dramatic and sudden change in benthic environment in the entire record. During the B/A (14.7–12.9 ka), the basin floor was marked by severe dysoxia (0–0.1 mL/L O_2) that was continuous for nearly 2000 years. Rapid oxygenation at the onset of the YD increased oxygen levels to >0.5 mL/L for the duration of this cool interval. The abrupt increase in the Dysoxic fauna during the Pre-Boreal (11.6–10 ka) marks the initiation of continuous dysoxia on the Holocene basin floor (0–0.2 mL/L O_2). Early Holocene (10 to 6 ka) assemblages were dominated by the Dysoxic fauna but display submillennial-scale fluctuations in oxygenation level. During the middle Holocene, from 6 to 3 ka, the predominance of the Suboxic fauna indicates stable, slightly more oxygenated conditions of 0.1–0.5 mL/L O_2. This increase in oxygen is well expressed in the Behl DO index (Figure 4) but not in the Schmiedl index (due to lack of intermediate faunal abundances in the Schmiedl equation). During the last 3 kyr of the late Holocene, the return of Dysoxic assemblages indicate that oxygen concentrations likely were similar to those of the late early to early middle Holocene (~7–5 ka), although an associated decrease in oxygen levels is not shown in either DO index nor in PC loading (Figure 4). Spectral analysis of % Dysoxic species reveals a weak 1380 year periodicity (Figure S4), indicating that the millennial-scale tempo of climate change exhibited by the D/O cycles of MIS 3 also extends through the Holocene. This observation suggests the existence of a background sub-Milankovitch forcing of climate change throughout the late Quaternary, including glacials and interglacials, although the amplitude of the response varied depending on the average climate state. PC analysis reveals a higher frequency variation in oxygenation expressed as intermittent peaks in the PC1 and PC3 fauna, also clearly exhibited by the Behl DO index (Figure 4).

[36] Absence to near absence of benthic foraminifera in some intervals appears to have resulted from extended periods of anoxia (O_2 = near 0 mL/L) in the deep basin rather than poor preservation. This conclusion is supported by many samples with low benthic abundances that contain well-preserved planktic foraminifera, particularly during the transition from YD to Pre-Boreal (PB). In a number of samples within the B/A and PB episodes, no benthic foraminifera were found (Figures 2a and S5). Experiments on living *Stainforthia fusiformis* have shown that this species, at least, can tolerate anoxic conditions for several months [*Alve and Bernhard*, 1995]. However, because each 1 cm thick sample represents approximately ~6–7 years duration, the extreme anoxic events resulting in complete absence of benthic species during the B/A and PB episodes must have persisted at least on a decadal scale without even intermittently reverting to oxygenated conditions long enough to permit benthic repopulation.

[37] The Dysoxic assemblage of the B/A, interstadials, and the PB on the basin floor was almost completely dominated by a single species—*B. tumida*. Previous work has shown that this species exhibits depleted carbon isotopic (δ^{13}C) values interpreted to reflect high concentrations of dissolved methane in the benthic environment [*Kennett et al.*, 2000; *Hill et al.*, 2003; 2004; 2006b]. Thus, our findings suggest localized increase in methane flux to the ocean floor during the interstadials, the three distinct episodes of the B/A, and the early Holocene. These interpretations are consistent with highly negative δ^{13}C values during this time (*B. tumida*, −4‰, Figure 4).

5.2. Oxygen Levels at Sill Depth

[38] The foraminiferal results from MD02-2504 provide an oxygen history of the seafloor at sill depth in SBB (Figure 3). Clear similarity of the foraminiferal assemblages from 24 to 15 ka to those at MD02-2503 indicates that the two depths experienced corresponding oxygenation and trends during this interval. Like the basin floor, the sill depth underwent moderate fluctuation with a gradual decrease in oxygen to suboxic levels (0.5–1.5 mL/L O_2) from 24 to 18 ka (Figures 3 and 4). At this shallower depth, IS2 (23.4 ka) is not a strong dysoxic event and actually records higher oxygenation levels as indicated by the two DO indices (Figures 4e and 4f). Through H1 and the pre-Bølling (18–15 ka), oxygen levels increased to >1.5 mL/L. Oxygen decreased dramatically in the Bølling /Allerød interval. In the B/A, three peaks of dysoxia are represented by oscillations between the Dysoxic and Suboxic faunas and also as maxima in PC1 loading. As on the basin floor, at sill depth, the onset of the YD was marked by an abrupt increase in oxygenation and terminated by an abrupt increase of Suboxic fauna at the PB, signaling the initiation of continuous low oxygen that continued throughout the Holocene. The early Holocene at sill depth was distinctly less dysoxic (~0.5–1.0 mL/L) than at the basin floor as indicated by dominance of Suboxic faunas and as suggested by both DO indices (Figures 3 and 4). Oxygenation decreased at the end of the early Holocene to levels similar to the B/A, then gradually increased during the middle Holocene from 6 to 3 ka. From ~5 ka to present (middle to late Holocene), oxygen levels (0.1–0.5 mL/L) remained relatively consistent at moderate levels as also indicated by the dominance of Suboxic faunas, PC loading, the two DO indices, and an absence of preserved laminations.

5.3. Oxygen Minimum Zone Fluctuations

[39] We have reconstructed a history of past variations in the vertical position of the OMZ by high temporal resolution correlation between records from cores MD02-2503, -2504, ODP Site 893, and 1017 (Figure 3) to form vertical transect spanning ~480–1000 m of the water column. The modern average water depth range of the OMZ in the northeastern Pacific is ~500 to 1000 m [*Cartapanis et al.*, 2011]; thus, the low-oxygen water currently entering the basin at sill depth (MD02-2504) is close to the upper boundary of the OMZ. ODP Site 1017 is near the lower limit of the modern OMZ on the open slope. In spite of the similarities between cores, there are differences in relative abundances between species and differences in taxonomic composition that reflects environmental gradients with depth and location (Figures 3 and S4) and consequently the structure of the OMZ.

[40] All three PC's (PC1, PC2, and PC3) have similar loadings during cold intervals, namely, the LGM, H1, and YD cold episodes. This implies an absence of a vertical oxygen gradient in Santa Barbara Basin during cold episodes. This behavior contrasts with the warmer episodes, where differential loading of PCs indicates existence of a vertical oxygenation gradient (Figures 4a, 4b, and 4c).

[41] From 34 ka (MIS 3) through the pre-Bølling, the OMZ was relatively weak and SBB was mostly well oxygenated except for brief intervals of dysoxia. The DO indices indicate that oxygen concentrations were >0.6 mL/L and frequently at or above 1.5 mL/L. The prevalence of the Weakly hypoxic-oxic

assemblage at basin floor and at sill depth, and the dominance of Suboxic assemblages at the 1000 m slope setting, indicates that the OMZ was deeper than SBB or was very weak. This interval was marked by five strong, albeit brief, dysoxic events within the basin associated with Interstadials 2 to 6 when O_2 levels decreased to 0.0–0.2 mL/L (Figure 4). These events are only weakly and partially expressed on the open slope reflecting the predominantly upward expansion of the OMZ. Increased oxygenation occurred at all water depths during H1 as the OMZ weakened and contracted, becoming entirely absent in SBB.

[42] Dysoxic conditions were especially strong during the PB episode (basin floor) through early Holocene (basin floor and sill depth), but all depths became increasingly suboxic during the middle Holocene from 6 to 3 ka. Finally, during the late Holocene, Dysoxic assemblages became increasingly dominant on the basin floor as sill depth remained more oxygenated. Late Holocene oxygenation at sill depth probably resulted from increased ventilation by NPIW related to regional cooling in the northern Pacific during the middle to late Holocene [*Mix et al.*, 1999]. Based upon these records and others, the early Holocene, marked by relatively high sea surface temperatures in SBB, was much more poorly oxygenated than the cooler late Holocene [*Kennett et al.*, 2007a, 2007b; *Kennett and Kennett*, 2007].

[43] Evidence from benthic foraminiferal assemblages, laminations, and $\delta^{13}C_{\text{foraminifera}}$ composition (see below) in four cores suggests that the vertical expansion of the OMZ was most extensive during the B/A, consistent with previous reports of OMZ expansion over even broader areas of the northeast Pacific margin [*Quinterno and Gardner*, 1987; *Cannariato and Kennett*, 1999; *Zheng et al.*, 2000; *van Geen et al.*, 2003; *Davies et al.*, 2011]. With sea level lower by ~100 m, the upper boundary of the OMZ must have been as shallow as 360 m below the paleo sea surface. Greatest expansion of the OMZ at the onset of the B/A is documented by transitory dysoxic conditions (0.0–0.1 mL/L O_2) on the 1000 m open slope and the first and strongest of the three dysoxic peaks at the shallowest location (Figure 5). This three-part climatic/oceanographic subdivision of the B/A (also weakly expressed on the basin floor) appears to correlate to climatic subdivisions identifiable in the Cariaco Basin [*Hughen et al.*, 1996] and Greenland ice core records [*Dansgaard et al.*, 1993]. At these B/A peaks, oxygen depletion extending from ~360 to >900 m paleo water depth was stronger than at the core of the modern OMZ.

[44] The Younger Dryas in SBB returned to a weakly hypoxic-oxic condition (>0.5 mL/L O_2) that was less oxygenated than during the LGM, H1, and pre-Bølling. At the same time, the 1000 m open slope remained under suboxic conditions, indicating weakening and deepening of the OMZ. An abrupt strengthening and expansion of the OMZ in the Pre-Boreal is recorded at all depths and marks the onset of near-continuous low-oxygen conditions throughout the Holocene. Strongest dysoxia occurred at the SBB basin floor, with somewhat weaker dysoxic to suboxic conditions at sill depth and on the open slope.

[45] Superimposed on these broader Holocene trends in oxygenation inferred from the benthic foraminifera are relatively abrupt, short-term fluctuations indicated by the presence of laminations [*Behl*, 1995] the Dysoxic fauna, and the DO indices. To determine if patterns exist in the Holocene benthic foraminiferal time series, we conducted spectral analyses of % Dysoxic assemblages from 12 ka to present with AnalySeries [*Paillard et al.*, 1996] using the Blackman-Tukey method. These analyses reveal relatively weak spectral peaks in the Holocene with nearly identical periods of 1.38 kyr in MD02-2503 and 1.35 kyr in MD02-2504 (Figure S4). The power of the millennial-scale cyclicity is twice as strong in the basin-center record (MD02-2503). As with the D/O oscillations, these cycles in oxygenation likely reflect the influence of climate changes with time scales similar to millennial Atlantic meridional overturning circulation and centennial solar activity [*Bond et al.*, 1997; Schulz, 2002]. Changes in relative abundances of dominant Holocene species (*B. argentea* and *S. eckisi*) suggest a higher frequency cycle on time scales of a few hundred years, but such details cannot be statistically resolved without higher chronological resolution.

5.4. Paleoceanographic Implications of Benthic Foraminiferal Trends

[46] It is possible that SBB sensitivity to OMZ fluctuations was in part modulated by sea level influence on intermediate water circulation in the semienclosed basin, given restriction by shallow sill depths. However, the data refute the significance of sea level on basin ventilation. The abrupt oxygenation changes evident in the benthic foraminiferal assemblages at both basin floor and sill depth do not resemble the gradual, secular trend of ~120 m of global sea level rise spanning the LGM to middle Holocene. Thus, we continue to conclude that the oxygenation changes in SBB are mainly controlled by the combined effects of changing intermediate water ventilation and surface ocean biological productivity outside of the basin, along the California margin [*Behl and Kennett*, 1996].

[47] We interpret peaks in the Dysoxic assemblage, preserved laminations, and decreased benthic foraminiferal $\delta^{13}C$ as combined evidence for expansion and strengthening of the OMZ (Figures 3–5). These indicate that the expansion of the OMZ was largest during the B/A, reaching from the shallowest (MD02-2504) to the deepest site (ODP 1017E) and occurred simultaneously with the most negative $\delta^{13}C$ values recorded in the surface water-dwelling planktic foraminifer *G. bulloides* in this basin [*Hendy and Kennett*, 2003; *Hill et al.*, 2006b] (Figure 4). Negative $\delta^{13}C$ values of this species are known to result from upwelling of deeper waters on the modern California margin [*Pak and Kennett*, 2002], implying shallowing or northward shift of nutrient-rich deep or intermediate waters of southern origin during the B/A [*Hendy and Kennett*, 2003; *Hendy et al.*, 2004]. A similar B/A negative excursion in $\delta^{13}C$ is observed in the northwest Pacific, near the source of NPIW [*Hoshiba et al.*, 2006] (Figure 4g), suggesting a simultaneous decrease in formation of northern sourced intermediate waters. Although there is no evidence for increased organic matter productivity in SBB during the B/A (Figure 5), increased organic carbon flux at ODP Site 1017 near Point Conception [*Hendy et al.*, 2004] supports increased production at a major upwelling center just outside of the basin on the open margin. This suggests that the record of paleo-oxygenation below sill depth in SBB reflects a strengthening of the regional OMZ and the associated signal of intermediate waters advected into the basin, rather than predominantly local basin productivity changes.

[48] The more oxygenated conditions during cooler episodes (H1 and YD) indicate active intermediate water ventilation of the basin, likely related to increased NPIW production at high latitudes in the north Pacific [*Okazaki et al*., 2010], restricted biological productivity [*Jaccard et al*., 2005; *Narita et al*., 2002], and consequent weakening of the OMZ (Figure 4).

[49] The oxygenation history at SBB is consistent with other records from around the north Pacific that reported hypoxic events associated with the deglacial warming episodes following Termination 1A and 1B, although not in as close agreement farther southward and equatorward [*Keigwin and Jones*, 1990; *Behl and Kennett*, 1996; *Zheng et al*., 2000; *Crusius et al*., 2004]. The SBB and these other oxygenation records are not congruent (Figures 4e–4h), however, with evidence of upward transport of radiocarbon-depleted, deep-water to intermediate and surface depths [*Marchitto et al*., 2007] by deep convection in the Southern Ocean [*Spero and Lea*, 2002]. Decreased radiocarbon activity during H1 to pre-Bølling and YD offshore of Baja California implies addition of very old waters, which is inconsistent with the well-oxygenated conditions at these times in sequences to the north. Furthermore, widespread oxygen depletion around the northeastern Pacific Rim during the B/A is associated with greater radiocarbon activity, reflecting a decreased flux of old carbon transported from the deep sea to shallower depths. Even other records in close proximity to the core off Baja California analyzed by *Marchitto et al.* [2007] provide evidence for oxygenated conditions during the times of admixed older water and, conversely, oxygen depletion during the B/A when radiocarbon activity increased [*Keigwin and Jones*, 1990; *Zheng et al*., 2000; *Crusius et al*., 2004]. This may be partially explained by more localized deep- to intermediate depth Δ^{14}C anomalies [*Hain et al*., 2011]. These relationships suggest that upward flux of old, deep radiocarbon-depleted water during the last deglaciation that was significant enough to cause a major shift in the Δ^{14}C of marine bicarbonate did not substantially influence the oxygenation state of the east Pacific margin. Instead, changing ventilation of the North Pacific by NPIW and Antarctic Intermediate Water and variations in surface productivity in upwelling centers likely played a more critical role (Figures 4g and 4h).

[50] Widespread shifts in oceanic oxygenation that occurred around the northern circum-Pacific Ocean during the last glacial require explanations that relate changes in intermediate water circulation, upwelling-driven local productivity, and water column stratification. The cause of hypoxia is controversial and separate explanations have been proposed related to age and path of intermediate water, as well as surface productivity [*Behl and Kennett*, 1996; *van Geen et al*., 1996; *Keigwin*, 1998; *Mix et al*., 1999; *Zheng et al*., 2000; *Crusius et al*., 2004; *Ortiz et al*., 2004; *Schmittner et al*., 2007; *Barron et al*., 2009; *Davies et al*., 2011; *Rella et al*., 2012]. The North Pacific High influences surface water temperature stability and upwelling intensity off western North America such that changes in regional productivity may impact oxygenation of intermediate waters and hence ventilation in SBB. The position and strength of the North Pacific High can also influence subarctic and subtropical gyral circulation, cross-gyre transport, and rate of transport of North Pacific Intermediate Water from the Okhotsk Sea and Kuroshio-Oyashio interfrontal zone [*Yasuda*, 2004; *Hoshiba et al*., 2006] (Figure 4g) and the Bering Sea [*Rella et al*., 2012]. Paleoclimatic records from Cariaco Basin, Venezuela document a southward shift in the Intertropical Convergence Zone (ITCZ) during the Holocene in response to changes in insolation associated with orbital forcing [*Haug et al*., 2001] and the latitudinal shift of the ITCZ influences the position and intensity of the North Pacific High. Diatom, sediment, and geochemical data suggest that productivity in the northeastern Pacific was higher during the warmer early Holocene compared with the cooler late Holocene [*Barron et al*., 2003; *Ortiz et al*., 2004; *Roark et al*., 2003; *Ivanochko and Pedersen*, 2004]. Thus, the gradual Holocene oxygenation trend may have resulted from intensity changes in the North Pacific High, possibly driven by decreased summer insolation. In this manner, the paleo-oxygenation changes recorded in great detail in SBB during the entire late Quaternary appear to reflect Pacific basin-wide climatic changes, a finding that is supported by another recent study in SBB (*Moffitt et al*., in review).

6. Conclusions

[51] We have produced quantitative benthic foraminiferal assemblage data for two late Quaternary cores (last 34 ka) providing a vertical profile in SBB used for reconstruction of millennial-scale basin expansions and contractions of the oxygen minimum zone (OMZ). The resulting time series is at exceptionally high temporal resolution (107–126 year) that is amongst the highest available for such a biotic record. In addition to the foraminiferal record, we have generated equally high-resolution records in TOC (a proxy for organic matter flux), $CaCO_3$ (a proxy for carbonate productivity), and grain size (a proxy for bottom current strength). Lack of relationships between sedimentary, geochemical, and textural parameters with benthic foraminiferal assemblages indicates that changes in the faunal record at these low-oxygen concentrations do not reflect variation in organic matter flux nor current strength but rather changes in benthic oxygenation.

[52] Quantitative assessment of variation in oxygen concentrations and strength of the OMZ is accomplished by development of new benthic foraminiferal categories based on documented relations between species and water-mass oxygen concentrations. We use 19 species to identify Dysoxic (0.0–0.1 mL/L O_2), Suboxic (0.1–0.5 mL/L O_2), and Weakly Suboxic/Oxic (>0.5 mL/L O_2) assemblages. We then utilize these to create a new quantitative index of bottom water oxygenation. The complete absence of foraminifera during brief episodes of extreme anoxia in SBB indicates that no species of benthic foraminifera survived extended periods without oxygen, i.e., there are no known truly anoxic benthic foraminifera in the Quaternary of California.

[53] Changes in foraminiferal assemblages in these records were often dynamic and abrupt, with strong ties to global climatic events. The oxygen-sensitive behavior of assemblages at sill depth is consistent with the interpretation that low-oxygen waters in the basin are derived from the OMZ (upper intermediate waters derived from the open Pacific margin) and represent a history of broad-scale changes outside the basin, albeit amplified because of additional basinal restriction.

[54] The OMZ expanded during warmer and contracted during cooler intervals. It was especially well developed during the B/A, interstadial episodes, and during the earliest

Holocene. Subtle shifts in the foraminiferal record reveal three distinct peaks in dysoxia during the B/A and strong dysoxia during the early Holocene compared with the late Holocene. In contrast, the OMZ appears to have completely disappeared from within SBB near the end of the last glacial ~17.5–16 ka during the H1 event. Intervals representing strongest OMZ development are often dominated by high abundances of a single taxon—*B. tumida*. The dominance and abundance of this species may have been in response to increased methane expulsions especially during the B/A, interstadials, and earliest Holocene (PB) with further reduction of oxygen levels through methane oxidation.

[55] **Acknowledgments.** The cores used in this investigation were collected during the MONA (Margins Of North America) Cruise of the French research vessel "Marion Dufresne" between 27 May and 2 June 2002. The MONA cruise was organized within the framework of the international IMAGES program (we acknowledge the contributions of IPEV). We thank the captain, crew, and scientific staff of the MONA Cruise for their technical expertise and kind cooperation in successfully collecting these two long piston cores. Laboratory research was supported by the U.S. National Science Foundation (grants OCE-0825322 and 0825366). CMZ was supported by a Boeing-CNSM Fellowship. We thank Karen Thompson for technical assistance. The scope and focus of this manuscript was greatly improved by the insightful comments of the three anonymous reviewers and the editors.

References

Alve, E. (1990), Variations in estuarine foraminiferal biofacies with diminishing oxygen conditions in Drammensfjord, SE Norway, in *Paleoecology, Biostratigraphy, Paleoceanography and Taxonomy of Agglutinated Foraminifera*, edited by C. Hemleben, M. Kaminski, W. Kuhnt, and D. B. Scott, Kluwer Academic, Amsterdam, pp. 661–694.

Alve, E., and J. M. Bernhard (1995), Vertical migratory response of benthic foraminifera to controlled oxygen concentrations in an experimental mesocosm, *Mar. Ecol. Prog. Ser.*, *116*, 137–151.

Barron, J. A., L. Heusser, T. Herbert, and M. Lyle (2003), High-resolution climate evolution of coastal northern California during the past 16000 years, *Paleoceanography*, *18*, 1020, doi:10.1019/2002PA000768.

Barron, J. A., D. Bukry, W. E. Dean, J. A. Addison, and B. Finney (2009), Paleoceanography of the Gulf of Alaska during the past 15,000 years: Results from diatoms, silicoflagellates, and geochemistry, *Mar. Micropaleontol.*, *72*, 176–195.

Behl, R. J. (1995), Sedimentary facies and sedimentology of the late Quaternary Santa Barbara Basin, Site 893, in *Proc. ODP, Sci. Results*, edited by J. P. Kennett, J. G. Baldauf, and M. Lyle, Ocean Drilling Program, College Station, TX, vol. *144*, pp. 295–308.

Behl, R. J., and J. P. Kennett (1996), Brief interstadial events in the Santa Barbara basin, NE Pacific, during the past 60 kyr, *Nature*, *379*, 243–246.

Bernhard, J. M. (1992), Benthic foraminiferal distribution and biomass related to pore-water oxygen content: California continental slope and rise, *Deep-Sea Res.*, *39*, 585–605.

Bernhard, J. M., and S. S. Bowser (1999), Benthic foraminifera of dysoxic sediments: Chloroplast sequestration and functional morphology, *Earth Sci. Rev.*, *46*, 149–165.

Bernhard, J. M., and C. E. Reimers (1991), Benthic foraminiferal population fluctuations related to anoxia: Santa Barbara Basin, *Biogeochemistry*, *15*, 127–149.

Bernhard, J. M., B. K. Sen Gupta, and P. F. Borne (1997), Benthic foraminiferal proxy to estimate dysoxic bottom-water oxygen concentrations: Santa Barbara Basin, U.S. Pacific continental margin, *J. Foraminiferal Res.*, *27*, 301–310.

Bernhard, J. M., K. R. Buck, M. A. Farmer, and S. S. Bowser (2000), The Santa Barbara Basin is a symbiosis oasis, *Nature*, *403*, 77–80.

Bernhard, J. M., K. R. Buck, and J. P. Barry (2001), Monterey Bay cold-seep biota: Assemblages, abundance and ultrastructure of living foraminifera, *Deep Sea Res., Part I*, *48*, 2233–2249.

Bernhard, J. M., S. T. Goldstein, and S. S. Bowser (2010), An ectobiont-bearing foraminiferan, *Bolivina pacifica*, that inhabits microbic pore waters: Cell-biological and pale oceanographic insights, *Environ. Microbiol.*, *12*, 2107–2119.

Blake, G. H. (1976), The distribution of benthic foraminifera in the outer borderland and its relationship to Pleistocene facies. M.S. thesis, Univ. of Southern California, Los Angeles, USA.

Bond, G., W. Showers, M. Cheseby, R. Lotti, P. Almasi, P. deMenocal, P. Priore, H. Cullen, I. Hajdas, and G. Bonani (1997), A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates, *Science*, *278*, 1257–1266, doi:10.1126/science.278.5341.1257.

California Cooperative Oceanic Fisheries Investigations, CalCOFI Cruises Hydrographic Data Reports, http://data.calcofi.org/bottle-data/cruise-hydrographic-data/hydrographic-data.html.

Cannariato, K. G., and J. P. Kennett (1999), Climatically related millennial-scale fluctuations in strength of California margin oxygen-minimum zone during the past 60 k.y., *Geology*, *27*(*11*), 975–978, doi:10.1130/0091-7613(1999)027<0975:CRMSFI>2.3.CO;2.

Cannariato, K. G., J. P. Kennett, and R. J. Behl (1999), Biotic response to late Quaternary rapid climate switches in Santa Barbara Basin: Ecological and evolutionary implications, *Geology*, *27*(1), 63–66, doi:10.1130/0091-7613 (1999)027<0063:BRTLQR>2.3.CO;2.

Cartapanis, O., K. Tachikawa, and E. Bard (2011), Northeastern Pacific oxygen minimum zone variability over the past 70 kyr: Impact of biological production and oceanic ventilation, *Paleoceanography*, *26*, PA4208, doi:10.1029/2011PA002126.

Cedhagen, T. (1991), Retention of chloroplasts and bathymetric distribution in the sublittoral foraminiferan *Nonionellina labradorica*, *Ophelia*, *33*, 17–30.

Cook, M. S., L. D. Keigwin, and C. A. Sancetta (2005), The deglacial history of surface and intermediate water of the Bering Sea, *Deep Sea Res., Part II*, *52*(16–18), 2163–2173, doi:10.1016/j.dsr2.2005.07.004.

Corliss, B., and K. A. Silva (1993), Rapid growth of deep-sea benthic foraminifera, *Geology*, *21*(11), 991–994.

Crusius, J., T. F. Pedersen, S. Kienast, L. Keigwin, and L. Labeyrie (2004), Influence of northwest Pacific productivity on North Pacific Intermediate Water oxygen concentrations during the Bølling-Allerød interval (14.7-12.9 ka), *Geology*, *32*(7), 633–636, doi:10.1130/G20508.1.

Dansgaard, W., et al. (1993), Evidence for general instability of past climate from a 250-kyr ice-core record, *Nature*, *364*(6434), 218–220.

Davies, M. H., A. C. Mix, J. S. Stoner, J. A. Addison, J. Jaeger, B. Finney, and J. Wiest (2011), The deglacial transition on the SE Alaskan Margin: Meltwater input, sea level rise, marine productivity, and sedimentary anoxia. *Paleoceanography*, *26*, PA2223, doi:10.1029/2010PA002051.

Diego, T. D., and R. G. Douglas (1999), Oxygen-related sediment microfabrics in modern "black shales", Gulf of California, Mexico, *J. Foram. Res.*, *29*(4), 453–464.

Douglas, R. G. (1981), Paleoecology of continental margin basins: A modern case history from the Borderland of southern California, in *Depositional Systems of Active Continental Margin Basins, Pacific Section Society Economic Paleontologists Mineralogists (SEPM), Short Course Notes*, edited by R. G. Douglas, I. P. Colburn, and D. S. Gorsline, Los Angeles, pp. 121–156.

Douglas, R. G., and H. L. Heitman (1979), Slope and basin benthic foraminifera of the California Borderland, *Society of Economic Paleontologists and Mineralogists*, *27*, 231–246.

Emery, K. O. (1960), *The Sea Off Southern California: A Modern Habitat of Petroleum*, John Wiley, New York.

Emery, K. O., and J. Hulsemann (1962), The relationships of sediments, life and water in a marine basin, *Deep-Sea Res.*, *8*, 165–180.

Gage, J. D., and P. A. Tyler (Eds) (1991), *Deep-Sea Biology: A Natural History of Organisms at the Deep-Sea Floor*, Cambridge Univ. Press, Cambridge, U. K.

Hain, M. P., D. M. Sigman, and G. H. Haug (2011) Shortcomings of the isolated abyssal reservoir model for deglacial radiocarbon changes in the mid-depth Indo-Pacific Ocean, *Geophys. Res. Lett.*, *38*, L04604, doi:10.1029/2010GL046158.

Harman, R. A. (1964), Distribution of foraminifera in the Santa Barbara Basin, California, *Micropaleontology*, *10*(1), 81–96.

Haug, G. H., K. A. Hughen, D. M. Sigman, L. C. Peterson, and U. Rohl (2001), Southward migration of the intertropical convergence zone through the Holocene, *Science*, *293*, 1304–1308.

Hendy, I. L., and J. P. Kennett (1999), Latest Quaternary North Pacific surface-water responses imply atmosphere-driven climate instability, *Geology*, *27*(4), 291–294, doi:10.1130/0091-7613(1999)027<0291: LQNPSW>2.3.CO;2.

Hendy, I. L., and J. P. Kennett (2000), Dansgaard-Oeschger cycles and the California current system: Planktonic foraminiferal response to rapid climate change in Santa Barbara Basin, Ocean Drilling Program hole 893A, *Paleoceanography*, *15*(1), 30–42, doi:10.1029/1999PA000413.

Hendy, I. L., and J. P. Kennett (2003), Tropical forcing of North Pacific intermediate water distribution during Late Quaternary rapid climate change?, *Quat. Sci. Rev.*, *22*(5–7), 673–689, doi:10.1016/S0277-3791 (02)00186-5.

Hendy, I. L., J. P. Kennett, E. B. Roark, and B. L. Ingram (2002), Apparent synchroneity of submillennial scale climate events between Greenland and Santa Barbara Basin, California from 30–10 ka, *Quat. Sci. Rev.*, *21*(10), 1167–1184, doi:10.1016/S0277-3791(01)00138-X.

Hendy, I. L., T. F. Pedersen, J. P. Kennett, and R. Tada (2004), Intermittent existence of a southern Californian upwelling cell during submillennial climate change of the last 60 kyr, *Paleoceanography*, *19*, PA3007, doi:10.1029/2003PA000965.

Hill, T. M., J. P. Kennett, and H. J. Spero (2003), Foraminifera as indicators of methane-rich environments: A study of modern methane seeps in Santa Barbara Channel, California, *Mar. Micropaleontol.*, *49*, 123–138.

Hill, T. M., J. P. Kennett, and H. J. Spero (2004), High-resolution records of methane hydrate dissociation: ODP Site 893, Santa Barbara Basin, *Earth Planet. Sci. Lett.*, *223*, 127–140.

Hill, T. M., J. P. Kennett, D. K. Pak, R. J. Behl, C. Robert, and L. Beaufort (2006a), Pre-Bolling warming in Santa Barbara Basin, California: Surface and intermediate water records of early deglacial warmth, *Quat. Sci. Rev.*, *25*, 2835–2845.

Hill, T. M., J. P. Kennett, D. L. Valentine, Z. Yang, C. M. Reddy, R. K. Nelson, R. J. Behl, C. Robert, and L. Beaufort (2006b), Climatically driven emissions of hydrocarbons from marine sediments during deglaciation, *Proc. Nat. Acad. Sci. U.S.A.*, *103*, 13,570–13,574.

Hoshiba, M., N. Ahagon, K. Ohkushi, M. Uchida, I. Motoyama, and A. Nishimura (2006), Foraminiferal oxygen and carbon isotopes during the last 34 kyr off northern Japan, northwestern Pacific, *Mar. Micropaleontol.*, *61*, 196–208.

Hughen, K. A., J. T. Overpeck, L. C. Peterson, and S. Trumbore (1996), Rapid climate changes in the tropical Atlantic region during the last deglaciation, *Nature*, *380*, 51–54.

Ingle, J. C., and G. Keller (1980), Benthic foraminifera biofacies of the eastern Pacific margin between 40°S and 32°N, in *Quaternary Depositional Environments of the Pacific Coast*, edited by M. E. Field et al., pp. 341–355, *Pac. Sect., Soc. of Econ. Paleontol. and Mineral.*, Los Angeles, Calif.

Ivanochko, T. S., and T. F. Pedersen (2004), Determining the influences of Late Quaternary ventilation and productivity variations on Santa Barbara Basin sedimentary oxygenation: A multi-proxy approach, *Quat. Sci. Rev.*, *23*, 467–480.

Jaccard, S. L., G. H. Haug, D. M. Sigman, T. F. Pedersen, H. R. Thierstein, and U. Röhl (2005), Glacial/interglacial changes in subarctic North Pacific stratification, *Science*, *308*(5724), 1003–1006, doi:10.1126/science.1108696.

Jannink, N. T., W. J. Zachariasse, and G. J. Van der Zwaan (1998), Living (Rose Bengal stained) benthic foraminifera from the Pakistan continental margin (northern Arabian Sea), *Deep-Sea Res. I: Oceanogr. Res. Pap.*, *45*, 1483–1513, doi:10.1016/S0967-0637(98)00027-2.

Jorissen, F. J., H. C. De Stigter, and J. G. V. Widmark (1995), A conceptual model explaining benthic foraminiferal microhabitats, *Mar. Micropaleontol.*, *26*, 3–15.

Jorissen, F. J., C. Fontanier, and E. Thomas (2007), Paleoceanographical proxies based on deep-sea benthic foraminiferal assemblage characteristics, in *Developments in Marine Geology*, edited by C. Hillaire–Marcel and A. De Vernal, Elsevier, Amsterdam, vol. *1*, pp. 263–325.

Kaiho, K. (1994), Benthic foraminiferal dissolved-oxygen index and dissolved-oxygen levels in the modern ocean, *Geology*, *22*(8), 719–722, doi:10.1130/0091-7613(1994)022<0719:BFDOIA>2.3.CO;2.

Keigwin, L. D. (1998), Glacial-age hydrography of the far northwest Pacific Ocean, *Paleoceanography*, *13*, 323–333.

Keigwin, L. D., and G. Jones (1990), Deglacial climatic oscillations in the Gulf of California, *Paleoceanography*, *5*, 1009–1023.

Kennett, J. P. (1995), Latest Quaternary benthic oxygen and carbon isotope stratigraphy: Hole 893A, Santa Barbara Basin, California, in *Proc. ODP, Sci. Results*, edited by J. P. Kennett, J. G. Baldauf, and M. Lyle, Ocean Drilling Program, College Station, TX, vol. *146*, pp. 3–18.

Kennett, J. P., and B. L. Ingram (1995), A 20,000-year record of ocean circulation and climate change from the Santa Barbara basin, *Nature*, *377*, 510–513, doi:10.1038/377510a0.

Kennett, D. J., and J. P. Kennett (2007), Influence of Holocene marine transgression and climate change on cultural evolution in southern Mesopotamia, in *Climate Change and Climate Dynamics: A Global Perspective on Mid-Holocene Transitions*, edited by D. G. Anderson, K. A. Maasch, and D. H. Sandweiss, Elsevier, Amsterdam, pp. 229–264.

Kennett, J. P., K. G. Cannariato, I. L. Hendy, and R. J. Behl (2000), Carbon isotopic evidence for methane hydrate instability during Quaternary interstadials, *Science*, *288*, 128–133, doi:10.1126/science.288.5463.128.

Kennett, J. P., K. G. Cannariato, I. L. Hendy, and R. J. Behl (2003), *Methane Hydrates in Quaternary Climate Change: The Clathrate Gun Hypothesis*, AGU, Washington D. C.

Kennett, D. J., J. P. Kennett, J. M. Erlandson, and K. G. Cannariato (2007a), Human responses to middle Holocene climate change on California's Channel Islands, *Quat. Sci. Rev.*, *26*, 351–367.

Kennett, D. J., B. J. Culleton, J. P. Kennett, J. M. Erlandson, and K. G. Cannariato (2007b), Middle Holocene climate change and human population dispersal in western North America, in *Climate Change and Cultural Dynamics: A Global Perspective on Mid-Holocene Transitions*, edited by D. G. Anderson, K. A. Maasch, and D. H. Sandweiss, Elsevier, Amsterdam, pp. 531–557.

Koho, K. A., R. García, H. C. de Stigter, E. Epping, E. Koning, T. J. Kouwenhoven, and G. J. van der Zwaan (2008), Sedimentary labile organic carbon and pore water redox control on species distribution of benthic foraminifera: A case study from Lisbon–Setúbal Canyon (southern Portugal), *Prog. Oceanogr.*, *79*, 55–82.

Linke, P., and G. F. Lutze (1993), Microhabitats preferences of benthic foraminifera – A static concept or a dynamic adaptation to optimise food acquisition? In: M. R. Langer (Ed.), Foraminiferal Microhabitats, *Mar. Micropaleontol.*, *20*, 215–234.

Maas, M. (2000), Verteilung lebendgefärbter bentischer foraminiferen in einer intensivierten sauerstoffminimumzone, Indo-Pacifischer kontinentalrand, nordliches Arabisches mee, *Meyniana*, *52*, 101–129.

Mackensen, A., and R. G. Douglas (1989), Down-core distribution of live and dead deep-water benthic foraminifera in box cores from the Weddell Sea and the California continental borderland, *Deep-Sea Res.*, *36*, 879–900.

Marchitto, T. M., S. J. Lehman, J. D. Ortiz, J. Fluckiger, and A. van Geen (2007), Marine radiocarbon evidence for the mechanism of deglacial atmospheric CO_2 rise, *Science*, *316*, 1456–1459.

Mix, A. C., D. C. Lund, N. G. Pisias, P. Bodén, L. Bornmalm, M. Lyle, and J. Pike (1999), Rapid climate oscillations in the Northeast Pacific during the last deglaciation reflect northern and southern hemisphere sources, in *Mechanisms of Global Climate Change at Millennial Time Scales*, *Geophys. Monogr. Ser.*, edited by P. U. Clark et al., AGU, Washington, D. C., vol. *112*, pp. 127–148.

Mullins, H. T., J. B. Thompson, K. McDougall, and T. L. Vercoutere (1985), Oxygen-minimum zone edge effects: Evidence from central California coastal upwelling system, *Geology*, *13*, 491–494.

Murphy, D. P., and D. J. Thomas (2010), The negligible role of intermediate water circulation in stadial-interstadial oxygenation variations along the southern California margin: Evidence from Nd isotopes, *Quat. Sci. Rev.*, *29*, 2442–2450.

Narita, H., M. Sato, S. Tsunogai, M. Murayama, M. Ikehara, T. Nakatsuka, M. Wakatsuchi, N. Harada, and Y. Ujiie (2002), Biogenic opal indicating less productive northwestern North Pacific during the glacial ages, *Geophys. Res. Lett.*, *29*(15), 1732, doi:10.1029/2001GL014320.

Nederbragt, A. J., and J. Thurow (2005), Geographic coherence of millennial-scale climate cycles during the Holocene, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *221*, 313–324.

Okazaki, Y., A. Timmermann, L. Menviel, N. Harada, A. Abe-Ouchi, M. O. Chikamoto, A. Mouchet, and H. Asahi (2010), Deepwater formation in the North Pacific during the last glacial termination, *Science*, *329*, 200–204.

Ortiz, J. D., S. E. O'Connell, J. DelViscio, W. Dean, J. D. Carriquiry, T. J. Marchitto, Y. Zheng, and A. van Geen (2004), Enhanced marine productivity off western North America during warm climate intervals of the past 52 k.y, *Geology*, *32*(6), 521–524, doi:10.1130/G20234.1.

Paillard, D., L. Labeyrie, and P. Yiou (1996), Macintosh Program performs time-series analysis, *Eos Trans. AGU*, *77*, 379.

Pak, D. K., and J. P. Kennett (2002), A foraminiferal isotropic proxy for upper water mass stratification, *J. Foraminiferal Res.*, *32*, 319–327, doi:10.2113/32.3.319.

Pak, D. K., Lea, D. W., and J. P. Kennett (2012), Millennial scale changes in sea surface temperature and ocean circulation in the northeast Pacific, 10–60 kyr BP, *Paleoceanography*, *27*, PA1212, doi:10.1029/2011PA002238.

Quinterno, P., and J. V. Gardner (1987), Benthic foraminifers on the continental shelf and upper slope, Russian River area, northern California, *J. Foraminiferal Res.*, *17*, 132–152.

Rathburn, A. E., L. A. Levin, Z. Held, and K. C. Lohmann (2000), Benthic foraminifera associated with cold methane seeps on the northern California margin: Ecology and stable isotope composition, *Mar. Micropalentol.*, *38*, 247–266.

Reid, J. L. (1965), *Intermediate Waters of the Pacific Ocean*, Johns Hopkins Press, Baltimore, Maryland.

Rella, S. F., R. Tada, K. Nagashima, M. Ikehara, T. Itaki, K. Ohkushi, T. Sakamoto, N. Harada, and M. Uchida (2012), Abrupt changes of intermediate water properties on the northeastern slope of the Bering Sea during the last glacial and deglacial period, *Paleoceanography 27*, PA3203, doi:10.1029/2011PA002205.

Roark, E. B., B. L. Ingram, J. Southon, and J. P. Kennett (2003), Holocene foraminiferal radiocarbon record of paleocirculation in the Santa Barbara Basin, *Geology*, *31*, 379–382.

Sarnthein, M., P. M. Grootes, J. P. Kennett, and M.-J. Nadeau (2007), ^{14}C reservoir ages show deglacial changes in ocean currents and carbon cycle, in *Ocean Circulation: Mechanisms and Impacts*, *Geophys. Monogr. Ser.*, edited by A. Schmittner, J. C. H. Chiang, and S. R. Hemming, AGU, Washington D. C, vol. *173*, pp. 175–196.

Savrda, C. E., and D. J. Bottjer (1991), in *Modern and Ancient Continental Shelf Anoxia*, *Geol. Soc. Spec. Publ.*, *58*, 201–219.

Schmiedl, G., A. Mitschele, S. Beck, K.-C. Emeis, C. Hemleben, H. Schulz, M. Sperling, and S. Weldeab (2003), Benthic foraminiferal record of ecosystem variability in the eastern Mediterranean Sea during times of sapropel S5 and S6 formation, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *190*, 139–164.

Schmittner, A., E. D. Galbraith, S. W. Hostetler, T. F. Pedersen, and R. Zhang (2007), Large fluctuations of dissolved oxygen in the Indian and Pacific oceans during Dansgaard-Oeschger oscillations caused by variations of North Atlantic Deep Water subduction. *Paleoceanography*, *22*, PA3207, doi:10.1029/2006PA001384.

Schulz, M. (2002), On the 1470-year pacing of Dansgaard-Oeschger warm events. *Paleoceanography*, *17*(2), 1014, doi:10.1029/2000PA000571.

Sen Gupta, B. K., and M. L. Machain-Castillo (1993), Benthic foraminifera in oxygen-poor habitats, *Mar. Micropaleontol.*, *20*, 183–201.

Smith, P. B. (1964), Recent foraminifera off Central America. Ecology of benthic species, *USGS Professional Paper*, *429B*, 1–55.

Soutar, A., and P. A. Crill (1977), Sedimentation and climatic patterns in the Santa Barbara Basin during the 19th and 20th centuries, *Geol. Soc. Am. Bull.*, *88*, 1161–1172.

Spero, H. J., and D. W. Lea (2002), The cause of carbon isotope minimum events on glacial terminations, *Science*, *296*, 522–525.

Stott, L. D., T. Bunn, M. Prokopenko, C. Mahn, J. Gieskes, and J. M. Bernhard (2002), Does the oxidation of methane leave an isotopic fingerprint in the geologic record?, *Geochem. Geophys. Geosyst.*, *3*, doi:10.1029/2001GC000196.

Stuvier, M., and P. M. Grootes (2000), GISP2 oxygen isotopic ratios, *Quat. Res.*, *53*, 277–283.

Talley, L. D. (1993), Distribution and formation of North Pacific intermediate water, *J. Phys. Oceanogr.*, *23*, 517–537.

Thunell, R. C., E. Tappa, and D. M. Anderson (1995), Sediment fluxes and varve formation in Santa Barbara Basin, offshore California, *Geology*, *23*, 1083–1086.

van Geen, A., R. G. Fairbanks, P. Dartnell, M. McGann, J. V. Gardner, and M. Kashgarian (1996), Ventilation changes in the northeast Pacific during the last deglaciation, *Paleoceanography*, *11*(5), 519–528, doi:10.1029/96PA01860.

van Geen, A., J. Bernhard, K. Cannariato, J. D. Carriquiry, W. Dean, B. Eakins, J. Pike, and Y. Zheng (2003), On the preservation of laminations along the western margin of North America, *Paleoceanography*, *18*(4), 1098, doi:10.1029/2003PA000911.

Yasuda, I. (2004), North Pacific Intermediate Water: Progress in SAGE (SubArctic Gyre Experiment) and Related Projects, *J. Oceanogr.*, *60*, 385–395.

Zeleski, C. M. (2004), Climatic and environmental controls of organic carbon accumulation on the California continental margin: A comparison of deep and shallow sites in the Late Quaternary Santa Barbara Basin, M.S. thesis, Calif. State Univ., Long Beach.

Zheng, Y., A. van Geen, R. F. Anderson, J. V. Gardner, and W. E. Dean (2000), Intensification of the northeast Pacific oxygen minimum zone during the Bolling–Allerod warm period, *Paleoceanography*, *15*(5), 528–536, doi:10.1029/1999PA000473.