

North Pacific Intermediate Water response to a modern climate warming shift

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[1] Oceanic observations and an isopycnal ocean model simulation are used to investigate the response of North Pacific Intermediate Water (NPIW) to atmospheric forcing associated with the well-known 1976–1977 climate regime shift to a warm regime. The model reproduces numerous features of NPIW including distribution, depth, temperature, and salinity. Changes in NPIW associated with the climate shift in the California coastal region were strongly influenced by an anomalous poleward flow at depth (300–1100 m). This current transports old, high salinity, low oxygen intermediate waters from the northern tropics to the midlatitudes. For depths below the mixed layer, the model reproduces observed changes in salinity, nitrates, and, to some extent, oxygen, thus suggesting that advective/diffusive processes are dominant in determining their concentrations below 300 m, isolated from the surface effects of direct atmospheric forcing and biological processes. These changes are structurally similar to those induced by much larger, abrupt climate changes at the end of the last glacial episode. *INDEX TERMS*: 1635 Global Change: Oceans (4203); 4267 Oceanography: General: Paleoceanography; 4255 Oceanography: General: Numerical modeling; 4283 Oceanography: General: Water masses; *KEYWORDS*: intermediate water, climate change, global warming, ice age, coastal productivity off California, California Undercurrent

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1. Introduction

[2] Variability in oceanic productivity depends on a large number of factors that are often poorly understood. For instance, it was speculated [Roemmich and McGowan, 1995] that a dramatic decline in zooplankton off California since the late 1950s was in response to surface water changes associated with climatic warming. Furthermore, it has been suggested that these oceanic changes were also linked to the transitions from sardine to anchovy dominance in the California Current [Chavez *et al.*, 2003].

[3] It remains unclear how oceanic climate changes at different water depths are controlled by local atmospheric forcing and remote forcing by oceanic waves or advection. Hendy and Kennett [1999] reported a rapid response in oceanic surface conditions of the California Current System (CCS) to decadal-scale atmospheric forcing changes, while Weinheimer *et al.* [1999] documented abrupt changes at thermocline depths. The findings of Cannariato and

Kennett [1999] recorded an apparent disconnection between shallow and intermediate waters. Deep changes induced by advection in North Pacific Intermediate Waters (NPIW) [Talley, 1993] have been implicated by Kennett and Ingram [1995] in driving deep changes of different nutrients. Growing evidence suggests that NPIW is sensitive to atmospheric climate changes [e.g., Banks *et al.*, 2000] and influences the intermediate circulation in the California Current System in general [Cannariato and Kennett, 1999], and the Santa Barbara Basin in particular [Kennett *et al.*, 2000].

[4] Our goal is to determine the approximate water depth below which climate changes driven by advection of NPIW are dominant over those driven by local surface effects associated with atmospheric or biological processes. We will also determine how NPIW responds to atmospheric climate changes and its consequent effect on the ocean off California. Given that atmospheric fields of wind stress and surface heat fluxes are only available with adequate temporal and spatial resolution since the 1950s, we use one recent climate shift event as a case study. The 1976–1977 climate shift [Trenberth and Hurrell, 1994] is recent enough to have

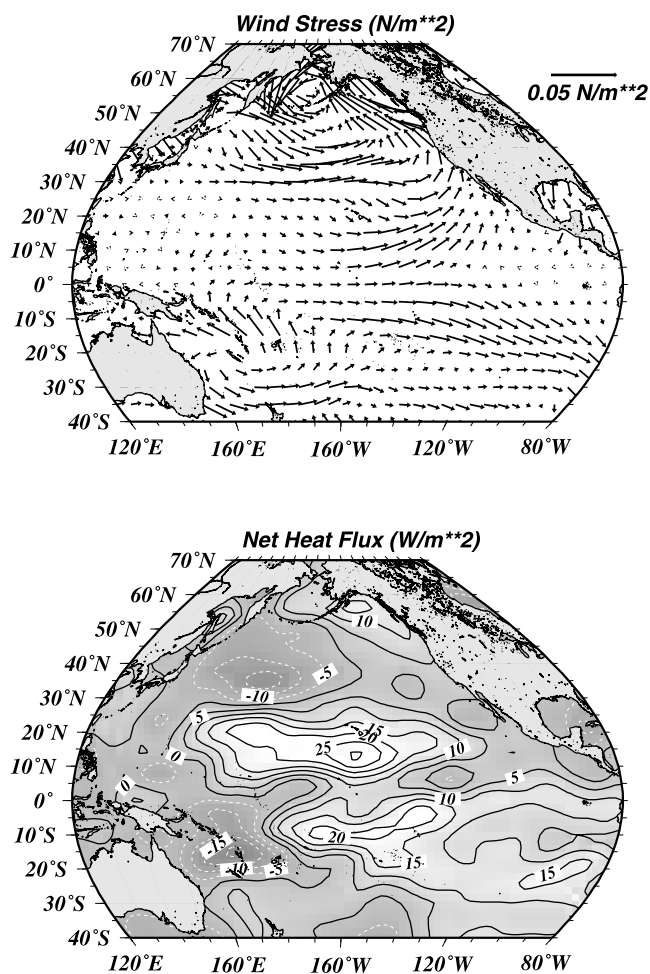


Figure 1. Climatological differences (1982/1977–1976/1971) for (top) wind stress and (bottom) surface heat fluxes.

been captured by the available observational data and serves here as an analogue for other studies of paleoclimate change on different timescales. Although relatively small (an average increase in SST off California of no more than 1°C) compared to the largest inferred climate changes on record of about 5°–6°C [e.g., *Hendy and Kennett, 1999; Cannariato et al., 1999; Kennett et al. 2000; Behl and Kennett, 1996*], it clearly affected the intermediate and shallow Pacific Ocean both biologically and physically [*Ebbesmeyer et al., 1991; Beamish and Bouillon, 1993; Miller et al., 1994; Miller and Schneider, 2000*].

2. Model

[5] We use a primitive equation ocean model to simulate changes in Pacific basin-scale oceanic conditions associated with this climate shift. The model, known as OPYC (Oberhuber's isoPYCnal model), was developed by *Oberhuber [1993]* and successfully applied to study seasonal, ENSO, decadal, and interdecadal variability in the Pacific Ocean [*Miller et al., 1994; Auad et al., 1998a, 1998b; Auad et al., 2001; White et al., 2001*]. The grid extends from 119°E to 70°W and from 67.5°S to 67.5°N. The model is constructed using ten isopycnal layers (each with nearly constant potential density but variable thickness, temperature, and salinity)

that are fully coupled to a bulk surface mixed-layer model. The resolution is 1.5° in midlatitudes, gradually increasing to 0.65° in the meridional direction only, and within 10° of the equator.

[6] The monthly mean seasonal cycle climatology of total surface heat flux is computed during spin-up (no anomalous forcing) by evaluating bulk formulae that use model SST with ECMWF-derived atmospheric fields (air temperature, humidity, cloudiness, etc.). The daily mean seasonal cycle is then saved (averaged over the last 10 years of a 140-year spin-up) and subsequently employed during anomalously forced runs, while the mixed layer salinity, unlike deeper layers, is relaxed to the model's climatological values (*Oberhuber's climatologies*) which are based on the COADS data set. Near the equator, anomalous heat fluxes are both poorly known and generally serve as a damping mechanism. Thus, Newtonian damping is employed within a 6° e-folding scale around the equator, where the SST anomalies are damped back to model climatology on 1- to 4-month timescales [*Barnett et al., 1991*].

[7] Two runs were performed with different forcing, but identical initial conditions. In the first run, only climatological monthly mean seasonal cycle forcing was used for 10 years (after a 140-year spin-up run). In the second run, anomalous fields of wind stress, total surface heat flux, turbulent kinetic energy and freshwater fluxes (NCEP/NCAR) were added to the respective climatic seasonal cycles. Anomalous forcing is constructed for all four fields by subtracting the monthly mean climatologies of the 1971/1976 period from those of the 1977/1982 period. The model was then run for 10 years from the same initial conditions as the first run. Then the anomalous ocean model response is obtained by subtracting the annual mean model variables of the climate run for year ten from those of the anomalous run for year ten. The equilibrium state reached in year six is virtually the same as that one reached in year ten. Moreover, this latter comparison is a good indicator of statistical equilibrium, while the comparisons with observations, for both deep and shallow fields, described below in section 4, will be an ad hoc source of confidence when determining the overall validity of our results.

3. Observations

[8] Oxygen and nitrate observations from the National Oceanographic Data Center (NODC SD2 Station data set) have been used to initialize the ocean model and to compare with the model response. The data were first subjected to quality control and then binned in five boxes. To compare the model with these observations, annual mean shift-induced anomalies of oxygen and nitrate were computed by subtracting the 6-year pre-shift annual means from the 6-year post-shift annual means.

4. Results

4.1. Model Response

[9] The signature of the 1976–1977 shift in NCEP wind stress and surface heat flux fields is exhibited in Figure 1. A positive curl is clearly seen centered at about 50°N with increased warming (or decreased cooling) by surface heat fluxes all along the eastern boundary and along 20°N. These features are similar to those described by *Miller et al. [1994]*

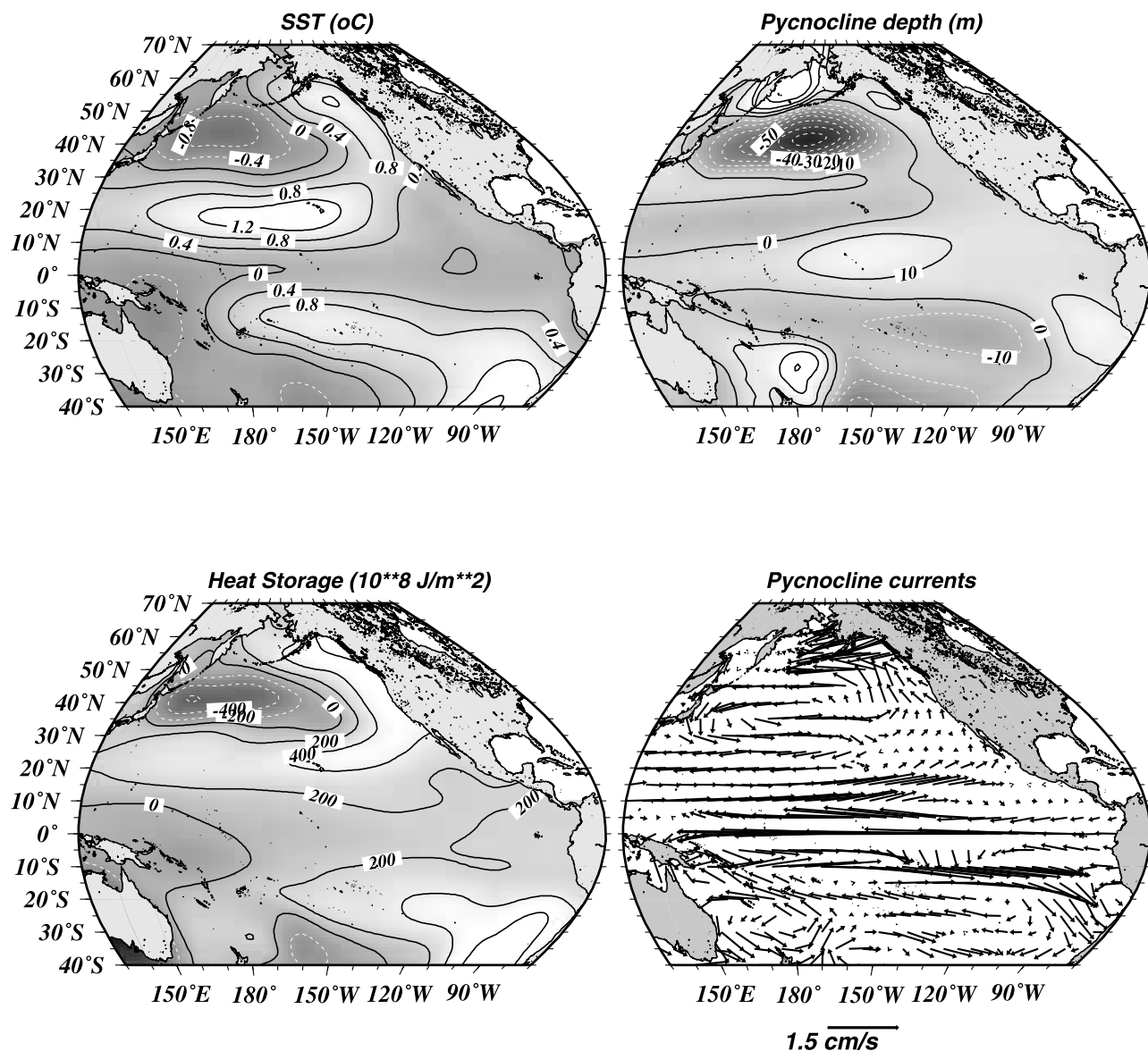


Figure 2. Model response for (top left) SST, (top right) pycnocline depth, (bottom left) heat storage, and (bottom right) pycnocline currents. These fields are obtained after reaching equilibrium and forcing the model with the fields summarized in Figure 1. They are also computed as the difference between the average computed in the 1982/1977 period minus the average computed in the 1976/1971 period.

who forced a coarser resolution version of OPYC using COADS observations. The model response to this atmospheric change is shown in Figure 2 for SST, heat storage, pycnocline depth, and pycnocline currents. The SST and heat storage patterns are similar to previous descriptions of the 1976–1977 regime shift, [e.g., Miller *et al.*, 1994] and to modal SST patterns [e.g., Mantua *et al.*, 1997], which describe the Pacific Decadal Oscillation (PDO). The deep poleward flow along the eastern boundary north of 20°N increases the speed of the California Undercurrent, which is associated with shallower pycnocline depths at about 45°N in direct response to the positive anomaly in the wind stress curl at that location [Miller *et al.*, 1998]. This anomalous northward circulation is evident in the model at depths from about 300 m down to 1300 m, with amplitude decaying with depth. While salinity is a dynamic variable of OPYC,

nitrate and oxygen were incorporated as passive tracers. The validity of this assumption does not hold for surface waters because of biogeochemical influences, but is more appropriate at depth as indicated by the degree of correspondence there between the model and observations.

[10] The tracer fields will be on the 300-m surface because this is shallow enough to be exposed to the influence of uppermost NPIW, yet deep enough (well below the mixed layer) to escape the effects of direct atmospheric forcing. In this respect, the upper 400 m of the model temperature field compares well with observations [Miller *et al.*, 1998], and as seen in the next section, deeper isopycnal layers also compare well with data over the large spatial scales relevant to climate.

[11] We will compare these model results with observations and with the results of Kennett *et al.* [2000], for Santa

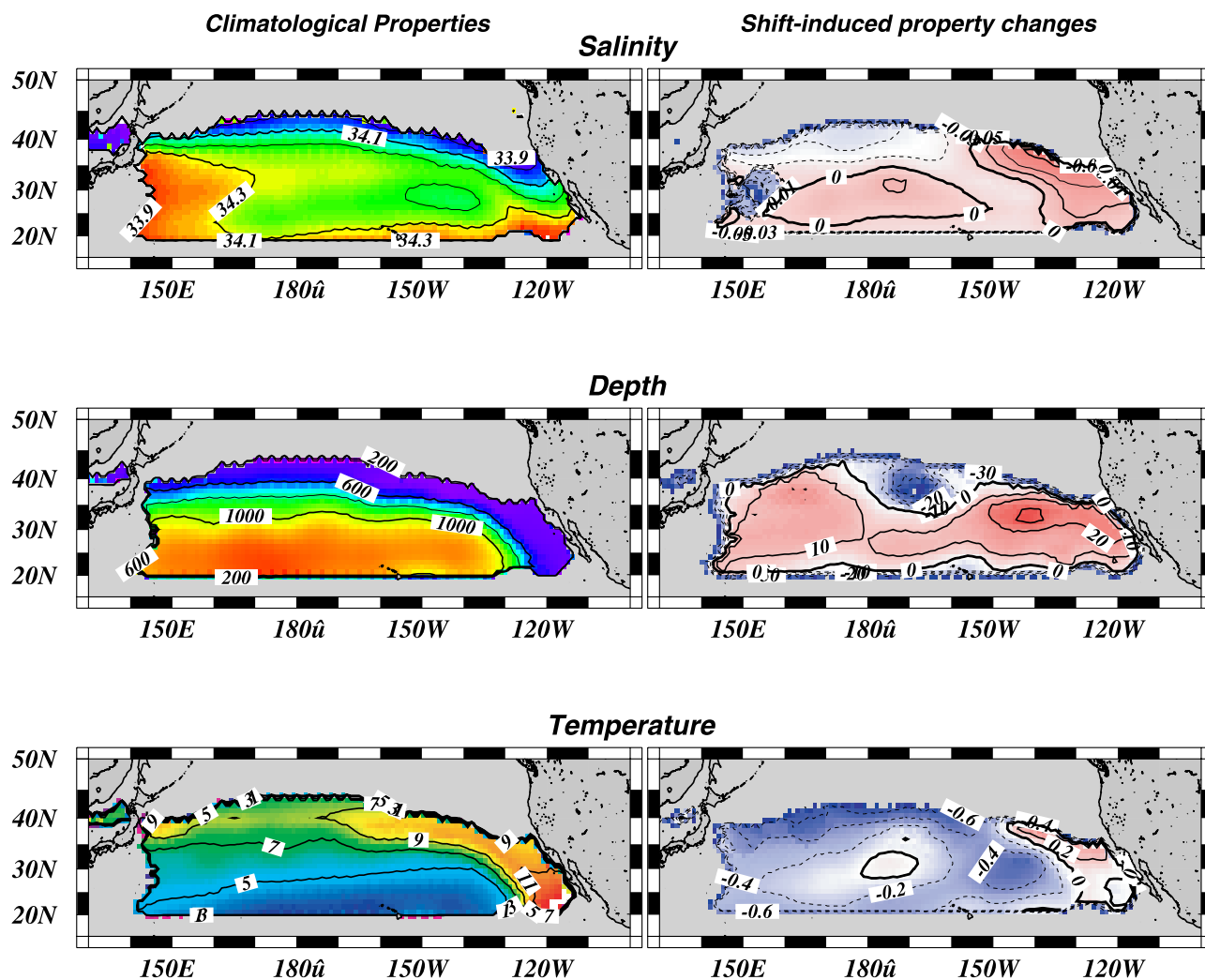


Figure 3. Mean properties (salinity, depth, and temperature) (left) for the annual mean of the model NPIW and (right) for the climatological change in those conditions after the sudden 1976–1977 climate shift. Differences were computed identically as in Figures 1 and 2.

Barbara Basin, which has a sill depth close to 480 m. Above this depth (their Figure 4), water from the California Undercurrent enters the Santa Barbara Basin. According to them, this current brings on occasion, old intermediate water from the tropics [Kennett and Ingram, 1995; Behl and Kennett, 1996; Cannariato and Kennett, 1999; Hendy and Kennett, 2003]. This is relevant to the local ecology since the basin below sill depth becomes almost anoxic.

4.2. Model NPIW

[12] The definition of NPIW varies among authors. Sverdrup *et al.* [1942] define this water mass as a salinity minimum in the range 300–700 m, while Reid [1965] defines the NPIW as the salinity minimum. More recently, Van Scoy *et al.* [1991] referred to the isopycnal $\sigma_\theta = 26.8$ as NPIW. In this study, we define the NPIW to be the salinity minimum within the 33.9- to 34.4-ppt salinity range, which characterizes the NPIW field obtained from observations by Talley [1993]. Given the model architecture, we obtained interpolated profiles for potential density, salinity and potential temperature. Figure 3 shows the model

annual means of salinity, temperature, and NPIW depth for climatological forcing and for the differences between the 1971–1976 and 1977–1982 periods. The mean salinity field resembles that of Talley [1993] with increasing values from the northeast toward the southwest; the values are slightly higher than the average observed, but within the observed range.

[13] The depth of NPIW ranges from 300 m to 1100 m in the model, where that maximum depth (occurring along its southern boundary) is somewhat too deep compared with observations. This problem arises because the ocean model has only 11 layers and has maximum resolution in the upper ocean, while the deeper layers can become as thick as 500–1200 m. The model NPIW potential temperature also shows a reasonable agreement with the observations of Talley [1993, Figure 16], which she considers a “classic” NPIW profile and is located at 32°2′N and 142°23′E. The observed salinity minimum there has an associated potential temperature of 6.8°C, which corresponds well not only with the model’s potential temperature ($\sim 7^\circ\text{C}$), but also with latitude as shown in Talley’s [1993] Figure 14 (solid inverted

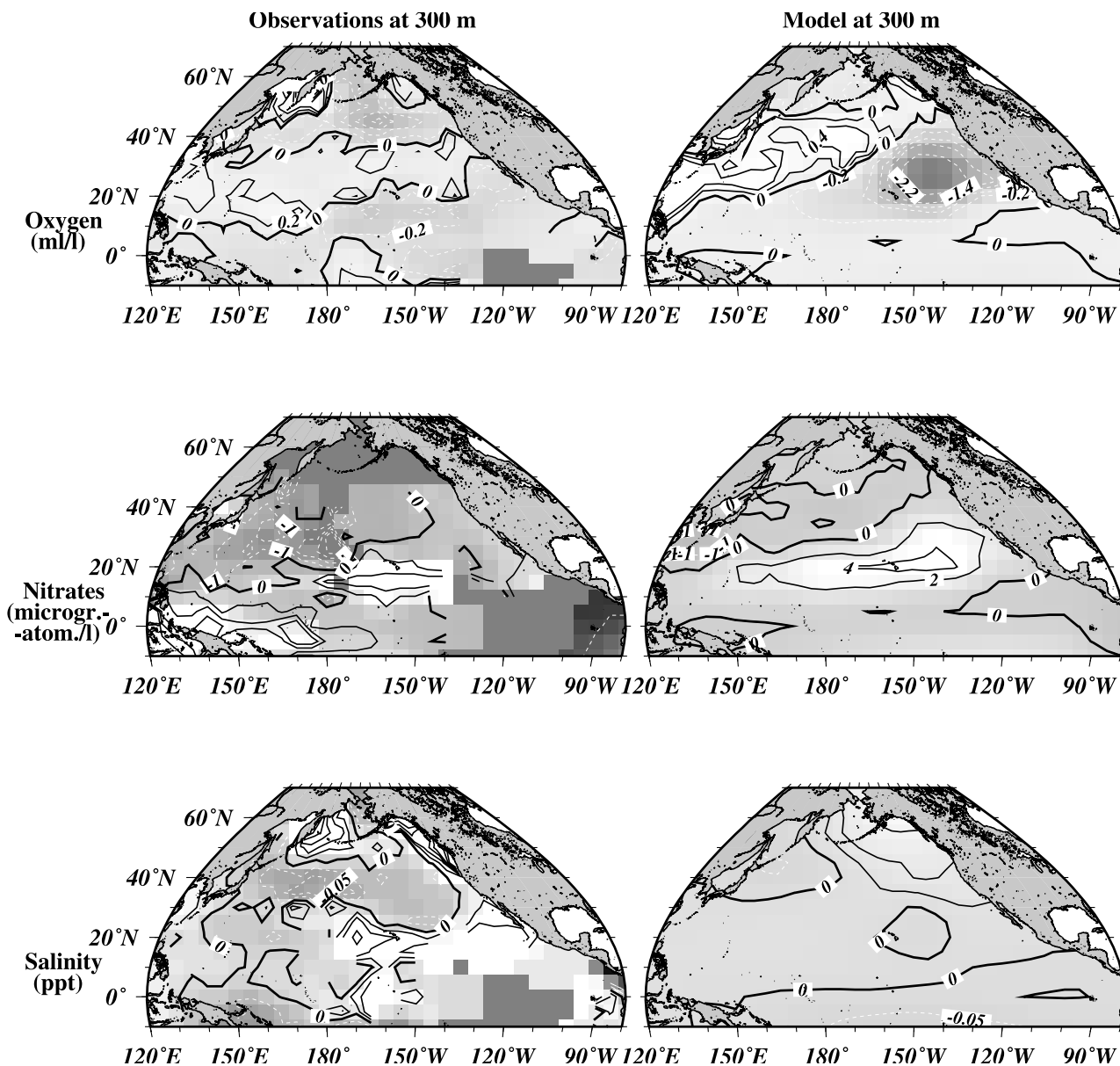


Figure 4. Shift-induced changes in the concentration of (top) oxygen, (middle) nitrates, and (bottom) salinity for (left) observations and (right) model. The differences were identically computed as in Figures 1, 2, and 3. Large negative to large positive values are shaded from dark to light, respectively.

triangles located around 33°N). Geographically, there is also good agreement with the observations of *Talley* [1993], as the model NPIW lies roughly between 20°N and 40°N (compare to her Figure 4). However, the model NPIW is shifted more toward the east with respect to her geographical definition of NPIW, which, unlike ours, requires a salinity minimum in a given density range. The fact that the model NPIW domain is located close to the basin's eastern boundary is of particular interest to us, and we will dedicate one section below to this area.

[14] The property changes (Figure 3, right-hand panels) were obtained after forcing the ocean model with anomalous wind stresses and heat fluxes associated with the 1976/1977 climate shift. The salinity change shows decreased values on the northern boundary west of 150°W and increased values along its eastern boundary. Both changes would result from

anomalous currents advecting mean properties (Figure 2, bottom right-hand panel). This is in line with the findings of *Talley* [1993] and of *Kennett et al.* [2003], who note that the Okhotsk Sea is the main producer of NPIW, and with the findings of *Hendy and Kennett* [2003], who studied the tropical forcing of the NPIW. Thus the high mean salinities that characterize the eastern tropical Pacific are advected by the poleward flow shown in Figure 2. In similar fashion, the southeastward anomalous flow located southeastward of the Okhotsk Sea (Figure 2) brings cold, low salinity, high oxygen waters into the NPIW domain of Figure 3.

[15] NPIW depth changes result from diabatic and adiabatic processes. Both seem to be present and competing in the model NPIW, although with different strengths at different locations. In general, it is seen (Figure 3) both that the model NPIW experiences depth changes every-

where and that at those locations, the potential temperature and salinity anomalies have non-zero values. Thus, diabatic processes seem to play some role in affecting the variability of the NPIW. The only area where adiabatic Ekman pumping is likely having an important contribution to NPIW depth changes is around 37°N – 170°E . This area, which is ~ 700 m deep, experienced a depth change (sinking) of the same sign as that one of the main pycnocline (Figure 2, top right-hand panel), which at this location both occurs at ~ 300 m above the model NPIW and is the largest one in the North Pacific. Most importantly, the different changes observed between the southeastern and northwestern corners of model NPIW are reminiscent of the competition between old and new intermediate waters mentioned by Kennett *et al.* [2003] at these same locations.

4.3. Shift-Induced Nutrient Distribution Changes

[16] To better understand how post-shift conditions affect the distribution of nutrients in the model, oxygen and nitrate fields were computed as independent passive tracers for pre-shift and post-shift conditions. The initial three-dimensional oxygen and nitrate fields were computed as a 6-year pre-shift January mean from the NODC data set. The 10-year climatological run and the 10-year anomalously forced (post-shift) run, both started from this initial state. The tenth years of these runs were then used to compute annual mean difference fields, indicative of the changes in nutrient distributions induced by the shift (Figure 4). For comparison, salinity is also shown in Figure 4, although it is a dynamic model variable that is conserved except at the surface, where fresh-water fluxes influence concentrations.

[17] Modeled and observed changes in the horizontal distribution of oxygen, nitrate, and salinity were compared at the 300-m level. This is deep enough to escape direct surface forcing effects, yet sufficiently shallow to include the upper part of NPIW, especially along the eastern boundary of the model's NPIW domain. We have chosen a given level, rather than an isopycnal surface or layer, to better identify the regions, mostly at the eastern boundary, where the NPIW reaches higher and more effectively up into the water column.

[18] Figure 4 shows the results of the model shift-induced changes for all three variables, along with observed changes based on the 6-year post-shift/pre-shift annual mean differences. The oxygen field exhibits an east-west band of positive values, which separates areas of negative values across the northern North Pacific (in both model and observations). However, in the model (top right-hand panel in Figure 4) this band of positive values is more in a northeast-southwest direction, compared with the east-west band seen in the observations (top left-hand panel in Figure 4). OPYC, like many other coarse resolution ocean models, tends to underestimate the amplitude of the mean flow at all depths [Auad, 2003], which may explain part of this discrepancy. Along most of the equator and in the subtropics, both model and observations show decreased concentration of oxygen, which is associated with the 1976–1977 climate shift.

[19] Nitrates show reasonable agreement between both model and observations (middle panels in Figure 4). An increase in concentration occurs along the midlatitudinal

subtropics and eastern boundary with maximum concentration at and near the Hawaiian islands. Off California, this increase is again likely the result of horizontal, rather than vertical, advection. Poleward currents (Figure 2) transport nitrates from the northern part of the tropics where their concentrations exhibit maximum values in the Pacific basin. East of Japan, both model and observations show decreased concentration of nitrates, with the model underestimating real ocean values.

[20] Salinity (bottom panels of Figure 4) exhibits increased concentrations along the eastern boundary of the midlatitudinal North Pacific, with largest values, in both model and observations, along the Canadian and Alaskan coasts. The strong currents of the Kuroshio Extension advect and spread areas of low salinity concentrations toward the east, more effectively in the real ocean than in the model, where its coarse resolution induces a weakened mean flow, and thus lower inferred advection.

[21] As noted above, we have intentionally excluded an analysis of the surface and near-surface levels of nutrients since the passive-tracer assumption breaks down due to direct atmospheric forcing and biological activity within the mixed layer. However, we note that the analysis of the NODC observations revealed (not shown) surface changes in salinity, oxygen, and nitrates off California that had opposite sign to those found at depth (i.e., increased surface oxygen and decreased surface nitrates and salinity).

4.4. Implications for Climate Change Off Western North America

[22] Figures 3 and 4 reveal that for both model and observations, there is an increase in the concentration of nitrates and salinities at depths between 300 m and 700 m following the 1976–1977 climate shift off California and Baja California. The temperatures at depth off California also increased following the climate shift in the model (Figure 3), while observed and model sea surface temperatures also increase. The oxygen concentration off Baja California decreased following the climate shift, in both model and observations, but north of about 34°N in Southern California, and including Santa Barbara Basin, the model and observed oxygen concentrations disagree due either to underestimation of the mean flow by OPYC or due to the effects of biological activity.

[23] These changes off California can be described as follows. Atmospheric changes in wind stress and heat fluxes after 1976–1977 led to surface warming and deepening of the main pycnocline in this region. In the central North Pacific the increased positive wind stress curl and diabatic forcing led to shallower pycnocline depths centered along 45°N . This, in turn, led to increased northward flow between 300 m and 1300 m in the model, which advects old, saline, low oxygen intermediate water from the northern tropics northward into the eastern North Pacific. This is in line with the ideas of Kennett *et al.* [2003] (see their Figure 25 and work by Hendy and Kennett [2003]), in the sense that the NPIW is sensitive to atmospheric climate changes. It is this mechanism [Kennett *et al.*, 2000] that brings low oxygen waters to the Southern California margin. Of course, other processes could be present in the spin up of the model (e.g., shift-induced Rossby waves), but we will not study them here as we are only concerned with the

equilibrium state, which is the response of the NPIW to an atmospheric climate change.

5. Discussion and Conclusions

[24] The midlatitude North Pacific ocean responded dynamically to the 1976–1977 climate shift in atmospheric conditions by adjusting its mass and current distributions, accompanied by changes in temperature, salinity, and nutrients. Old intermediate waters, as mentioned by *Kennett et al.* [2003] and *Hendy and Kennett* [2003], penetrate from the northern tropics into midlatitudes, leading to deeper NPIW near the eastern boundary of the North Pacific. From this modeling study, we find support for the idea that upper intermediate waters play an important role, and are strongly associated with rapid climate changes. Not only do we find analogous responses of the North Pacific Ocean to warming events that followed glacial and stadial episodes and the 1976–1977 shift (change in trade winds, increased SSTs, and poleward current developing on the eastern boundary of the North Pacific), but also in the dynamics of these events. This dynamics is marked by active competition between old (southeastern corner of the model NPIW) and new (northwestern corner of the model NPIW) intermediate waters, leading to crucial changes in the physical and chemical properties of the North Pacific ocean below the mixed layer. SST and surface nutrients are apparently insensitive to changes in upper intermediate waters, and their changes would be apparently dictated by variations in atmospheric conditions, which agrees with the findings of *Cannariato and Kennett* [1999].

[25] Thus, if there is any connection between deep and surface waters, it would reinforce the surface warming effect, which is mostly caused by atmospheric diabatic forcing, as part of the basin-scale climatic change. The assumption that oxygen and nitrates can be treated as passive tracers below the mixed layer is acceptable in some locations as off the coasts of Mexico and off the U.S. west coast south of Point Conception, given the similarities found between model and observations. A corollary of this, for those regions away from the surface (i.e., below 200–300 m) and where model and observations agree, is that the concentration of nutrients is dominated by advective/diffusive processes. In the near-surface waters, direct forcing and biological effects become more important than advection in controlling those concentrations.

[26] The patterns of ocean-atmosphere variations found in the present analysis of NPIW changes, due to the 1976–1977 climate shift, are consistent with those found by *Kennett et al.* [2003] and *Hendy and Kennett* [2003]. This suggests that similar physical processes control both small (events occurring in the twentieth century) and large (events occurring during the late Quaternary) abrupt climate changes. This analogy needs to be further explored with more complex models that include biogeochemical and physical processes and interactions.

[27] These changes have the potential to trigger further changes, such as the liberation of methane into the water column from sediments, and possibly into the atmosphere [*Kennett et al.*, 2003]. The 1976–1977 climate shift also exhibited a number of other features that marked Quaternary climate changes. For instance, a change in the trade winds,

such as that shown in Figure 1, preceded the termination of stadial and glacial episodes [*Hendy and Kennett*, 2003]. A poleward deep current (Figure 2) thus developed carrying tropical waters to midlatitudes. *Hendy and Kennett* [2003] place this anomalous flow between 300 m and 500 m in their studies of Quaternary climate change, in agreement with our results using the 1976–1977 climate shift as a prototype. It is this current that warms up the easternmost portion of the model NPIW, and, if sufficiently warm and sustained, could liberate methane from the seafloor through dissociation of gas hydrates. In this respect, and in the light of the findings reported by *Kennett et al.* [2003], the warming of waters off California by the 1976–1977 climate shift was not sufficiently long nor intense to destabilize gas hydrates. According to them, a sustained warming at the end of the last ice age, an increase of 1°C to 1.5°C which should last at least for a few decades in the bottom water temperature [see *Kennett et al.*, 2003, Figure 20b] below the 300-m level, resulted in dissociation of gas hydrates and release of methane into the oceanic-atmospheric system with potential climatic effects [*Kennett et al.*, 2003; *Hendy and Kennett*, 2003].

[28] Additional fundamental questions emerge from this study: Can the last glacial episode be characterized as a negative phase of the modern PDO index [*Mantua et al.*, 1997]? Abrupt warming events marked the end of both the last glacial episode some 15,000 years ago (large shift) and the cooling regime that dominated the Pacific Ocean until 1977 (small shift). During the latter, trade winds changes led to the development of a poleward undercurrent along the eastern boundary between 300 and 500 m. It needs to be determined if both large and small events are governed by similar physical processes such that paleoclimate variability may be described as a large-amplitude version of the PDO. A starting point could be the work of *Auad* [2003], who related the PDO to the dynamics proposed by *Liu* [1999] and to the observations shown by *Tourre et al.* [2001].

[29] In summary, this modeling effort is consistent with geological evidence that oceanic upper intermediate waters are dynamically associated with changes in global climate and heat transport. The role of change in intermediate waters in abrupt climate shifts is further reinforced by its effects on the strength of the oxygen minimum zone and in stability of methane hydrates and resulting changes in the atmospheric composition of the climatically important greenhouse gases nitrous oxide (N₂O) and methane (CH₄). Clearly there are a number of processes that can lead to abrupt ocean and climate change. Our modeling experiment provides support for only one such process that involves abrupt changes in upper intermediate waters in the North Pacific Ocean. In this model, changing strength of trade winds leads to warming, intensification of the poleward undercurrent along the eastern boundary, and associated changes in competition between new and old intermediate waters. Related to this, it will be relevant to further investigate if changes in SST and nutrients are proportional in amplitude to those of wind stress and surface heat fluxes. Are the magnitudes dependent on the initial state of the ocean and do preferred frequencies generate greater magnitude responses? Addressing these questions should assist with the understanding of the physical processes involved in the triggering and ending of abrupt events, thus providing firmer basis for anticipating future abrupt climate shifts.

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