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Persistently declining oxygen levels in the interior waters of the eastern subarctic Pacific

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Abstract

Fifty years of measurements at Ocean Station Papa (OSP, 50°N, 145°W) show trends in the interior waters of the subarctic Pacific that are both impacted by short term (few years to bi-decadal) atmospheric or ocean circulation oscillations and by persistent climate trends. Between 1956 and 2006, waters below the ocean mixed layer to a depth of at least 1000 m have been warming and losing oxygen. On density surfaces found in the depth range 100–400 m ($\sigma_\theta = 26.3$–27.0), the ocean is warming at 0.005–0.012 °C y^{-1}, whereas oxygen is declining at 0.39–0.70 μmol kg^{-1} y^{-1} or at an integrated rate of 123 mmol m^{-2} y^{-1} (decrease of 22% over 50 years). During this time, the hypoxic boundary (defined as 60 μmol O_2 kg^{-1}) has shoaled from ~400 to 300 m. In the Alaska Gyre, the 26.2 isopycnal occasionally ventilates, whereas at OSP $26.0\sigma_\theta$ has not been seen at the ocean surface since 1971 as the upper ocean continues to stratify. To interpret the 50 year record at OSP, the isopycnal transport of oxygenated waters within the interior of the subarctic Pacific is assessed by using a slightly modified "NO" parameter [Broecker, W., 1974. "NO" a conservative water-mass tracer. Earth and Planetary Science Letters 23, 100–107]. The highest nitrate–oxygen signature in interior waters of the North Pacific is found in the Bering Sea Gyre, Western Subarctic Gyre and East Kamchatka Current region as a consequence of winter mixing to the ~26.6 isopycnal. By mixing with low NO waters found in the subtropics and Okhotsk Sea, this signature is diluted as waters flow eastward across the Pacific. Evidence of low NO waters flowing north from California is seen along the coasts of British Columbia and SE Alaska. Oxygen in the subsurface waters of the Alaskan Gyre was supplied ~60% by subarctic and 40% by subtropical waters during WOCE surveys, whereas such estimates are shown to periodically vary by 20% at OSP. Other features discernable in the OSP data include periods of increased ventilation of deeper isopycnals on an ~18 year cycle and strong, short term (few month) variability caused by passing mesoscale eddies. The potential impacts of declining oxygen on coastal ecosystems are discussed.
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Keywords: Dissolved oxygen; Oxygen consumption; Hypoxia; Nitrates; Water temperature; Ocean stratification; Subarctic Pacific Ocean; Alaska Gyre; Ocean Station P

1. Introduction

Animal life in our oceans requires oxygen. Various organisms adapt to a wide range of oxic conditions, some requiring the saturated conditions of the surface ocean to support their rapid metabolism, while others

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tolerate brief periods of anoxia (Davis, 1975; Gray et al., 2002). A range of effects occur as oxygen levels decline, progressing from slowed growth rates, to metabolic impairments, to death. Since oxygen is continuously consumed in seawater via the remineralization of organic matter, the ocean depends on continuous inputs to the intermediate and deep ocean from the atmosphere. However, the subarctic Pacific is a strongly stratified ocean realm because of the freshness (low salinity) of the surface layer. Over much of this area, the ocean is mixed to about 100–125 m during winter storms (Freeland et al., 1997; Watanabe et al., 2001). Only in the mixed layer are oxygen levels found near 100% saturation. Below this, oxygen levels rapidly decline, reaching saturations of less than 10% below ~500 m (Aydin et al., 2004).

As the atmosphere of our planet warms, so do its oceans. This is evident in the eastern subarctic Pacific (Freeland et al., 1997) where both near shore and open ocean data records show ocean warming rates similar to those being measured in the atmosphere. Possibly, the poleward transport of freshwater from the subtropics is also stabilizing the surface layer of the subarctic Pacific (Wong et al., 2001; Freeland et al., 1997), as has been observed in the North Atlantic (Curry et al., 2003). As a result, the upper ocean is becoming more buoyant and stratification is strengthening. In surveys conducted over the past half century, oxygen declines are being observed in all areas of the subarctic Pacific. A 30 year record in the Oyashio Current off the north coast of Japan (Fig. 1) shows declining and oscillating trends in waters between ~190 and 630 m, on density surfaces from 26.7 to 27.2 (Ono et al., 2001). Rates of oxygen decline varied between 0.6 and 1.3 μmol kg^{-1} y^{-1} over the period 1968–1999. Concomitant with this trend was a tendency for the density of the upper ocean to decline and these waters to warm at about 0.007 °C y^{-1}.

Other regions of the western Pacific see similar trends. A 50 year decline of 0.4–0.8 μmol O_2 kg^{-1} y^{-1} was found on the 26.9–27.0 isopycnal surfaces in the Western Subarctic Gyre and Sea of Okhotsk (Andreev and Kusakabe, 2001). Also in the adjoining East/Japan Sea, observed oxygen decreases of between 0.5 and 2.0 μmol kg^{-1} y^{-1} were observed in deep waters between 1932 and 2000 (Watanabe et al., 2003; Kang et al., 2004).

World Ocean Circulation Experiment (WOCE) surveys have been repeated in most regions of the subarctic Pacific (Fig. 2). One of the persistent trends coming out of this work is the striking rate of oxygen decline over about a decade. Watanabe et al. (2001) used WOCE sections along 47°N (1985 and 1999) and 165°E (1987 and 2000) to calculate apparent oxygen utilization (AOU, the estimated amount of oxygen consumed since waters were last in equilibrium with the atmosphere) increases of 6 μmol kg^{-1} y^{-1}. In general, they found oxygen levels decreased by 18 mol m^{-2} over 13 years and concluded that the formation rate of intermediate ocean waters had substantially declined over a 15 year period, perhaps longer based on results of Ono et al. (2001). These

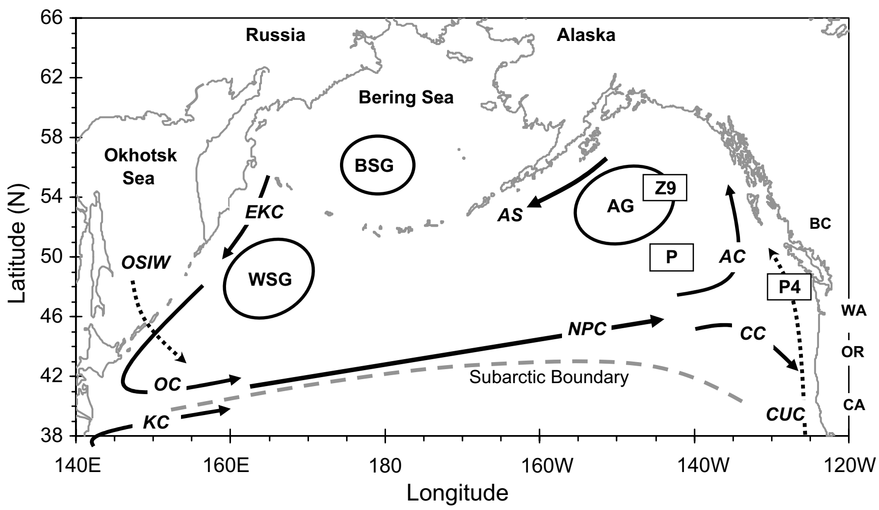

Fig. 1. Map of the North Pacific Ocean showing major currents (EKC, East Kamchatka Current; OSIW, Okhotsk Sea Intermediate Water; OC, Oyashio Current; KC, Kuroshio Current; NPC, North Pacific Current; CC, California Current; CUC, California Undercurrent; AC, Alaska Current; AS, Alaska Stream) and gyres (BSG, Bering Sea gyre, WSG, Western Subarctic gyre; AG, Alaska gyre). Also identified are Ocean Station P (P) at 50°N, 145°W, station P4 at 48.66°N, 126.67°W, British Columbia (BC), Washington (WA), Oregon (OR) and California (CA).

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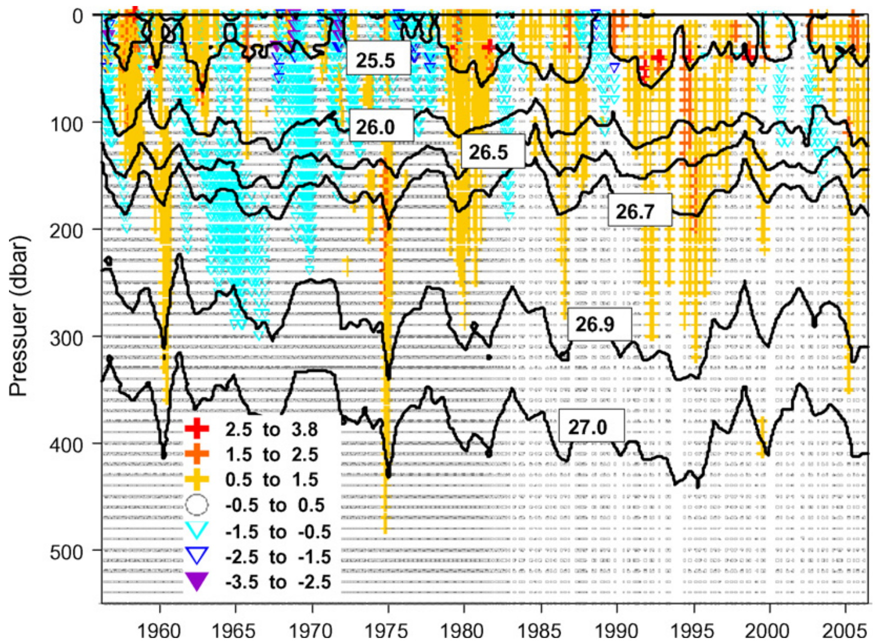

Fig. 2. Temperature anomaly in °C (monthly anomalies computed against 1956–1991 averages) and seawater density as sigma theta (monthly averages) at Ocean Station Papa from 1956 to 2005. Two mesoscale eddies are evident in: (1) 1960 and (2) 1974, these waters being characterized by depressed isopycnal surfaces, elevated temperature and low oxygen. After 1981, sampling was too infrequent to reliably resolve eddies.

findings were expanded by adding data from an eastern Pacific WOCE repeat section (Emerson et al., 2001; Emerson et al., 2004). These authors reported oxygen decreases of up to 60–80 μmol kg^{-1} over the 13–14 year period between surveys, but averaging in the 10–40 μmol kg^{-1} range for waters in the 33–34 salinity range (densities of ~26.5–27.0). Kumamoto et al. (2004) reported very similar findings along WOCE section P17 through the Alaska Gyre. Again, oxygen declined by as much as 60 μmol kg^{-1} between the early 1990s and 2001 in the density range 26.2–27.0 (200–600 m). Feely et al. (2004) used CFCs to age waters then calculated oxygen utilization rates (OUR) for various ocean basins. For the subarctic Pacific, they estimate an OUR of 4.1 μmol kg^{-1} y^{-1} for the depth range 200–900 m.

The subarctic Pacific is an ocean basin bordered by eastern Asia and western North America (Fig. 1), and by a distinct subarctic boundary that separates the warm, saline waters of the subtropics from the cool, fresh waters of the subarctic (Dodimead et al., 1963). Within the subarctic region, ocean circulation is generally cyclonic, with the eastward flowing North Pacific Current (NPC) bifurcating into subtropical (California Current, CC) and subarctic (Alaska Current, AC) branches as it approaches shore. The AC and buoyant coastal currents feed into the Alaskan Stream (AS) north of the Alaskan Gyre, with some portion of this water flowing southward around the western side of the Alaskan Gyre (AG). Flow through the Bering Sea is not shown but results in a strong southward flowing East Kamchatka Current (EKC), which is modified by outflow of Sea of Okhotsk Intermediate Water (SOIW) into the Oyashio Current (OC). The warm Kuroshio Current (KC) and the OC create the NPC. The subarctic boundary is imbedded within the NPC, separating subarctic OC waters from subtropical KC waters. Three cyclonic gyres form in the subarctic Pacific and Bering Sea, the Western Subarctic Gyre (WSG), the Alaskan Gyre (AG) and the Bering Sea Gyre (BSG).

Only in the western subarctic Pacific can isopycnal surfaces in the 26.5–26.7 range exchange gases with the atmosphere. The 26.6 isopycnal is seen outcropping within the domain of the East Kamchatka and Oyashio Currents (Ono et al., 2001; Mecking et al., 2006). From this region, winter ventilated waters submerge into the interior of the Pacific. You (2005) followed the path of the North Pacific Intermediate Waters (NPIW), centred on the 26.8 isopycnal surface, eastward from the coast of Japan. He suggests mixing of Okhotsk Intermediate Water and Gulf of Alaska Intermediate Water along the North Pacific Current produces NPIW waters with the appropriate CFC age. Mixing of various waters results in a product with increased density (cabbeling), with the result that these waters sink into the subtropics. Ueno and Yasuda (2003) show that waters on the

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26.5–26.8 isopycnal surface flow almost directly eastward, taking 5–7 years to travel from the edge of the eastern subarctic ventilation region (155°E) to Ocean Station Papa (OSP) at 50°N, 145°W. Their model estimates a ~15 year transit time on the 27.2 isopycnal.

Waters in the North Pacific Current acquire heat and salt from the subtropics as they cross the ocean (Mecking et al., 2006). These waters contain more oxygen than those within the Alaska Gyre, and are the oxygen source for the interior waters of the eastern subarctic Pacific. Between WOCE sections near 180 and 145°W, integrated oxygen utilization rates were calculated to be 2.1 ± 0.4 M m^{-2} y^{-1} for waters found between the 26.5 and 27.2 isopycnals (the 132–706 m range; Aydin et al., 2004).

Much of what has recently been discovered about the circulation of intermediate waters in the subarctic Pacific is based on WOCE data sets. Almost all of these open ocean results are dependent on repeat surveys, separated by periods of 10–15 years. Detail is provided for the ventilation site off the coast of Japan (Ono et al., 2001) otherwise not much open ocean time-series data is available except at OSP. Tabata (1989) reviewed data from this site for the period 1956–1983 and concluded declines in oxygen levels were mostly not significant. The trend on the 26.8 isopycnal surface was for oxygen to decrease by 0.7 μmol kg^{-1} y^{-1}. We now have 50 years of data at OSP and revisit this data set to discuss patterns of oxygen decline in the subarctic Pacific. Our analysis of oxygen trends begins with a review of the OSP data set, next we define areas where interior waters of the subarctic Pacific exchange gases with the atmosphere, then use a conservative property of seawater that is based on a combination of oxygen and nitrate (Broecker, 1974) to trace oxygen transport into the NE subarctic Pacific, and finally suggest reasons for oxygen decline and impacts this trend may have on biota.

2. Methods and data sources

Three major data sets are used to describe oxygen distribution and transport in the subarctic Pacific; the WOCE (World Ocean Circulation Experiment) one time surveys (http://whpo.ucsd.edu/about.htm), the OSP time series data (http://www.pac.dfo-mpo.gc.ca/sci/osap/projects/linepdata/default_e.htm) and Argo (http://www.argo.net/). A brief description of data reliability for each source follows.

WOCE data go through quality control steps which compare survey sections with both historical data and with other WOCE sections that occupy the same stations. For routine parameters such as temperature, salinity and oxygen, these data are of excellent quality. For example, the Institute of Ocean Sciences contributed WOCE data from sections P15 and P1W, the latter in collaboration with Pacific Oceanological Institute of Vladivostok, Russia. During IOS surveys, data accuracy was 0.002 °C for temperature, 0.003 for salinity and 1.02 μmol kg^{-1} for oxygen (Whitney and Barwell-Clarke, 1997). A repeat survey of WOCE line P1 (Fukasawa et al., 2004) suggested that temperature differences as low as 0.0007 °C might be detectable.

Argo temperature and salinity data are collected from profiling autonomous floats, typically from 2000 m to the surface every 10 days. These data then go through quality control procedures set up by each country contributing to the program. Data are compared to historical values or recent CTD casts throughout the ocean to identify floats that may have drifting sensors. Generally, salinity data are accurate to better than 0.01. Temperature sensors are believed to have relatively slow drift from factory calibrations over time, therefore errors are thought to be considerably less than 0.01 °C. When these data are used to calculate density, the error will be less than 0.01. Since we use Argo data only to show areas of winter outcropping of the 26.5 isopycnal surface, this potential error is trivial.

The data record from OSP consists of an era from 1956 to 1981 when Weatherships occupied this station almost continually and the period from 1981 to 2006 during which research vessels transited there 2–5 times annually. During the Weathership era, casts for temperature, salinity and oxygen were taken as frequently as several times per week.

Sampling and analytical procedures for samples collected along Line P (a survey line running from southern British Columbia to Ocean Station Papa) and at OSP have evolved over the 50 years of this program. Early sampling relied on unprotected reversing thermometers to provide depth measurements (estimated to be accurate to ±15 m at 4000 m depth), Loran C for position (navigators could be 10 nautical miles off station), reversing thermometers for temperature (accurate to ±0.02 °C) and older style water samplers for salinity and oxygen. OSP temperature and salinity data were carefully edited to remove bad data from leaking

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samplers, misfires, poor sample handling and analyses (data sets reviewed by Tabata and Peart, 1985; Tabata and Brown, 1994). During the WOCE era, IOS improved its sampling procedures by switching from hydro casts (sampling bottles on wire) to rosette casts. Depth accuracy of pressure sensors on various CTDs (conductivity–temperature–depth probes) improved to better than 2 m at full ocean depth, temperature was generally accurate to 0.002 °C, and salinity analyses were accurate to 0.002 with the use of IAPSO standard seawater and Guildline Autosal salinometers. Oxygen procedures have been consistent throughout the 50 year period with all samples being analyzed by the Winkler method (Carpenter, 1965). Still, a large number of analysts (perhaps >100) carried out this work. Therefore, this data set was hand edited to remove obvious outliers. These were primarily identified as single anomalous points in a profile. Nutrient methods are those described by Barwell-Clarke and Whitney (1996) and rely on modified Technicon procedures. Nitrate is accurate to $\pm 0.2\ \mu mol\ kg^{-1}$. Since 1987 nutrients have been analyzed at sea. Prior to this, samples were frozen and results contain uncertainty large enough to make data unreliable for the analysis of ventilation trends.

To track changes in temperature over 50 years, monthly anomalies have been computed. Steps taken to produce a monthly climatology for OSP include: (1) for months when more than one cast was made, data were averaged so that years with many casts in a month do not bias the climatology; (2) all data for various months between 1956 and 1991 were used to calculate monthly averages for this period; (3) monthly averages were then used to derive monthly anomalies. Only 13 stations were samples along Line P up to 1981, whereas 27 stations were sampled following the Weathership era.

Data for the three isopycnal surfaces of interest included all data within certain ranges (26.45–26.55, 26.65–26.75 and 26.88–26.92). When data were scarce in a geographic region or within a time period, interpolation of values to the desired density surface was carried out by assuming linear change between data points above and below the density surface in a profile. Potential densities (σ_θ) are implied throughout this paper whenever we refer to isopycnal surfaces. Differences between temperature and potential temperature to depths of 600 m are on the order of 0.04 °C, creating only a minor difference (~0.003) between density and potential density (σ_t and σ_θ).

When discussing ocean circulation, oceanographers commonly use oxygen changes in units not easily applied to biological studies. The following conversions represent a level commonly used to define the hypoxic threshold (Gray et al., 2002):

$$2.0\ mg\ l^{-1} = 1.40\ ml\ l^{-1} \sim 60\ \mu mol\ kg^{-1}.$$

In the 3–5 °C temperature and between salinities of 33–34, Apparent Oxygen Utilization (AOU) levels for hypoxia would be $6.0 \pm 0.2\ ml\ l^{-1}$ ($\sim 260\ \mu mol\ kg^{-1}$) which is equivalent to an oxygen saturation of ~23%.

3. Results

3.1. The 50 year record at Ocean Station P

Ocean Station P, at 50°N, 145°W, is located between the southern edge of the Alaska Gyre (AG) and the North Pacific Current (NPC) within the subarctic waters of the NE Pacific (Fig. 1). It is the terminal station of a 1400 km long survey line (Line P) that starts on the south coast of British Columbia. P4, a Line P station at which oxygen and nutrients have been routinely measured since 1987, sits on the continental slope in 1300 m of water. Because Weatherships occupied OSP for most of the period from 1956 to 1981 and research cruises have continued to sample a few times per year since, an extensive temperature, salinity and oxygen data set has been collected with which to observe impacts of climate on the subarctic ocean.

Temperature anomalies for OSP (Fig. 2) highlight some of the processes creating variability in the oxygen record. In general, the entire upper water column was cool in the decade from 1962 to 1972, with waters being >2 °C below the 1956–1991 average for much of this period. Since 1972, the T anomaly has been generally positive for the upper 500 m. Cool events of several months are associated with La Niñas in late 1983, 1989 and 1999. Maximum subsurface temperatures, occasionally exceeding the climatology by >3 °C, were recorded in association with El Niños of 1958, 1983, 1992, 1994–1995 and 1998, also with mesoscale eddies that passed through OSP in 1960 and 1974. These eddies, containing warm, low oxygen core waters of coastal origin, are characterized by strongly depressed isopycnal surfaces (Whitney and Robert, 2002).

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Density surfaces in intermediate waters (Fig. 2) fluctuate in response to short (few months) to medium length (several years) events. In monthly-averaged data, the 26.0 isopycnal surface does not outcrop at OSP, although the data record does show surface waters briefly reaching this density in the winters of 1959, 1961, 1965 and 1971. In the past 35 years, the 26.0 surface has not been observed at the ocean surface at OSP, even with Argo profilers frequently sampling waters in this region since 2001. Waters of the Alaska Gyre (centred at about 53°N, 150°W; see Fig. 3) ventilate to the 26.2 isopycnal during cool winters such as 2006 and 2007. The 26.5 and 26.7 surfaces have remained at fairly constant depths of 137 ± 17 and 168 ± 17 m. However $\sigma_\theta = 26.9$ and 27.0 show a deepening of ~30 m over 50 years, with the result that a greater volume of water now resides between the 26.7 and 26.9 surfaces.

It is not immediately apparent, when describing temporal trends in the ocean interior, whether it is appropriate to track constant depth, density or salinity surfaces. Each has its merits: depth indicates changes in habitat range for marine organisms, density is essential in describing transport within the ocean, and salinity follows the most conservative property of oceanic waters. By choosing to follow isopycnal (density) surfaces, changes in the interior ocean will deviate slightly from salinity. For example, a temperature increase at OSP on the 26.7 isopycnal of 0.6 °C over 50 years (Fig. 4) requires waters to become more saline by 0.01, causing a density surface deepening of ~1.5 m (a trivial amount) compared with salinity.

Rates of T and O_2 change are computed for depth, density and salinity surfaces at OSP (Table 1; density trends shown in Fig. 4). Trends listed were found significant at the 95% confidence level, with confidence levels being estimated using a standard *t*-test but with the number of degrees of freedom reduced to allow for autocorrelation as outlined in von Storch and Zwiers (2002). Warming rates are as high as 0.018 °C y^{-1} at 200 m, and oxygen decreases by as much as 0.96 μmol kg^{-1} y^{-1} at 150 m. Rates of change diminish below 200 m, but remain detectable at 1000 m.

It is interesting that if we restrict ourselves to the occasional ship-board sampling at OSP then the warming trend in the mixed layer reported by Freeland et al. (1997) is no longer significant at the 95% confidence level. However, if we examine the so-called Reynolds OI.v2 data set for this region (Reynolds and Marsico, 1993) then their monthly data from 1982 to present produces a warming trend in mid-winter (January through March) SST observations of 0.020 °C y^{-1} (95% significance level of 0.008 °C y^{-1}). The Reynolds data show a rather weaker and less significant warming trend in mid-summer SSTs. This would suggest that a warming trend is present even though it may not be convincingly apparent in direct observations at OSP. Crawford et al. (2007) observe a very similar warming trend along the survey line from OSP to the British Columbia coast of 0.9 °C over 48 years, but note that it is only significant at the 90% confidence level. Because both the upper and intermediate ocean are warming at similar rates, an increasing stratification of the upper ocean is the result of a freshening mixed layer (salinity decrease of 0.0036 y^{-1} between 1956 and 2006, similar to 0.0043 y^{-1} previously reported for OSP by Freeland et al., 1997).

Oxygen trends in the mixed layer cannot be reliably assessed because strong seasonality, driven by both solar radiation and biology, produces an annual cycle of 70–80 μmol kg^{-1}. Our data suggest oxygen is decreasing at

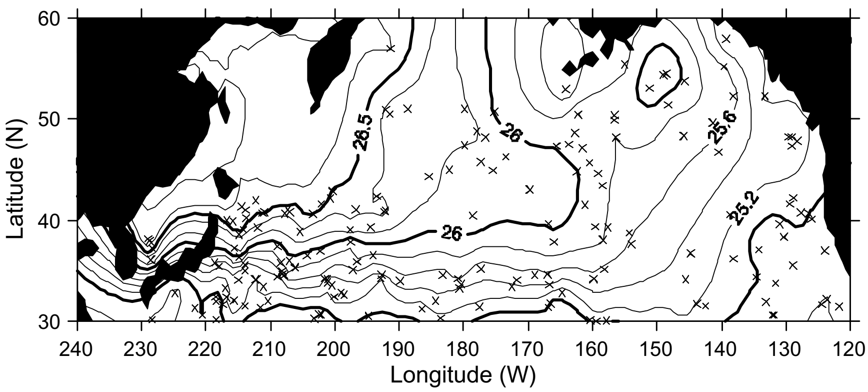

Fig. 3. Surface density (σ_θ) in March 2006, × marking the locations of reporting Argo profilers. The region with densities higher than 26.5 is found along the western margin of the subarctic Pacific. Density contours are every 0.2, except 26.5 has been added. Note that there are few or no floats in marginal seas of the North Pacific, therefore contours should be disregarded in these regions.

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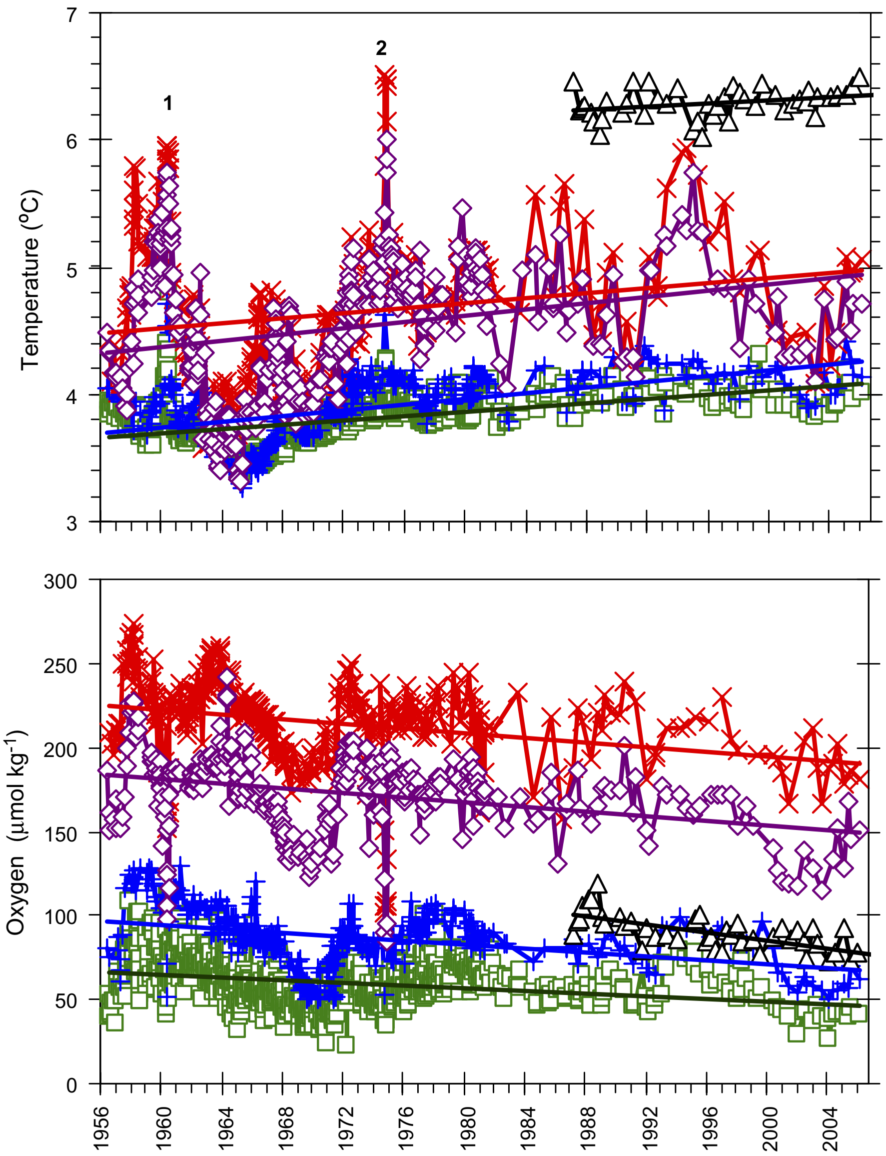

Fig. 4. Temperature and oxygen trends at Ocean Station P on the 26.5 (×), 26.7 (◇), 26.9 (+) and 27.0 (□) isopycnal surfaces and at station P4 (△) on the 26.7 surface. T and oxygen trends from linear regressions are provided in Table 1. Depth ranges (average and standard deviation) are 140 ± 15 m, 168 ± 17 m, 278 ± 27 m and 370 ± 44 m on the 26.5, 26.7, 26.9 and 27.0 isopycnals, respectively. P4 is warming at 0.0084 °C y^{-1}, with O_2 declining at 1.22 µmol kg^{-1} y^{-1}. Two mesoscale eddies are labelled 1 and 2.

Table 1
Rates of change on various depth, density and salinity (33.25–34.00) surfaces at OSP based on linear regressions for data between 1956 and 2006

Surface	T (°C y^{-1})	O_2 (µmol kg^{-1} y^{-1})	Surface	T (°C y^{-1})	O_2 (µmol kg^{-1} y^{-1})
ML	ns	ns	26.3σ_θ	0.000	−0.54
150 m	+0.016	−0.96	26.5σ_θ	0.009	−0.70
200 m	+0.018	−0.57	26.7σ_θ	0.012	−0.68
300 m	+0.013	−0.18	26.9σ_θ	0.011	−0.60
400 m	+0.010	−0.07	27.0σ_θ	0.008	−0.39
600 m	+0.0059	−0.07	33.25	0.0083	−0.57
800 m	+0.0029	−0.18	33.50	0.0076	−0.64
1000 m	+0.0020	−0.22	33.75	0.016	−0.07
4000 m	ns	−0.08	34.00	0.0094	−0.10

Non-significant (ns) trends are found for surface temperature due to high seasonal variability and for 4000 m temperature due to the low precision (0.01 °C) of early data. Salinity trend for the mixed layer (ML) is −0.0036 y^{-1}.

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0.17 μmol m^{-2} y^{-1}, but this analysis is biased by a lack of late winter data (March–May) since 1982. A declining trend might be anticipated due to the reduced solubility of oxygen in warming waters.

Below the mixed layer, seasonality has little effect on temperature or oxygen. At all depths below the mixed layer to at least 1000 m, oxygen is declining and temperature increasing at OSP (Table 1). Oxygen is decreasing more quickly at 800–1000 m than in waters in the 400–600 m range, suggesting more organic carbon remineralization in this layer. At 4000 m, a weak trend towards lower oxygen is evident whereas temperature data are not precise enough to detect a trend (pre 1991 T data being reported only to 0.01 °C). Whitney and Freeland (1999) previously reported that OSP waters were warming to at least 1000 m, with the strongest warming occurring at ~200 m.

Four isopycnal surfaces at OSP are selected for trend analyses (Fig. 4). The 26.5 and 26.7 surfaces show very similar patterns over time, with temperatures rising from 4.4 °C in the early record to ~5.0 °C at present. Linear regressions through all data indicate a warming rate of 0.009 and 0.012 °C y^{-1} on these two density surfaces, at ocean depths of between 120 and 200 m. The 26.9 and 27.0 isopycnals, found at depths between 230 and 420 m, show warming trends of 0.011 and 0.008 °C y^{-1}. These trends are strongly influenced by the cool (less saline) 1960s, so much so that warming is not apparent since 1972. Also plotted are $26.7\sigma_\theta$ data from a Line P station (P4) on the continental slope, these results showing ocean warming by 0.008 °C y^{-1} and oxygen declines of 1.22 μmol kg^{-1} y^{-1}. The California Undercurrent is a major component of these slope waters (Mackas et al., 1987), so trends will be strongly influenced by variability to the south. Temperature and oxygen at P4 are similar to values observed for mesoscale eddies at OSP in 1960 and 1974.

Oxygen levels at OSP decline over time by rates ranging from 0.39 to 0.70 μmol kg^{-1} y^{-1} (Fig. 4). Lower oxygen levels are evident during the cool periods of the 1960s and 1999–2002, and brief periods of oxygen increase are occasionally associated with abrupt warming, (late 1950s, mid 1970s and 2004). In the eastern subarctic Pacific, oxygen levels increase within warm, saline waters from the subtropics, decrease in cool, fresh waters from the Alaska Gyre, and strongly decrease in warm, fresh waters carried by mesoscale eddies.

For the period 1987–2005, OSP oxygen (O_2) and nitrate (NO_3) were integrated between 100 and 600 m (Fig. 5). As expected, their trends are mirror images of each other. Between 1994 and 2003, oxygen declined at a rate of −2.4 mol m^{-2} y^{-1} as nitrate increased at 0.26 mol m^{-2} y^{-1}, with the oxygen decline to nitrate increase ratio being 9.2. Between 2003 and 2005, the trend reversed with oxygen increasing (6.7 mol m^{-2} y^{-1}) and nitrate declining (−0.66 mol m^{-2} y^{-1}) yielding an O_2/NO_3 ratio of 10.2. Average depth integrated values (±1 SD) are 19.2 ± 0.7 mol NO_3 m^{-2} and 45.8 ± 6.5 mol O_2 m^{-2}, with oxygen concentrations averaging from 290 μmol kg^{-1} at 100 m to 30 μmol kg^{-1} at 600 m and nitrate varying between 17 μmol kg^{-1} at 100 m and 44 μmol kg^{-1} at 600 m.

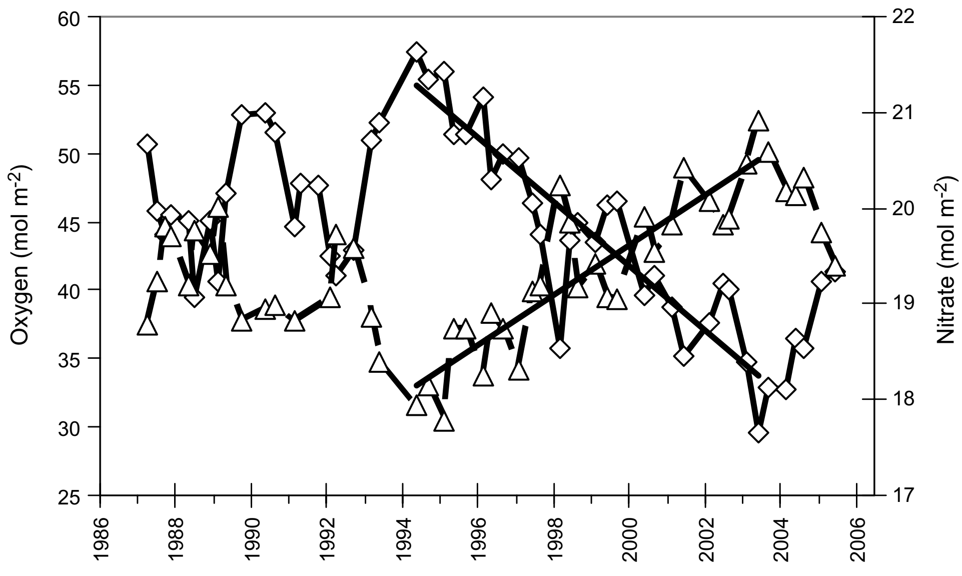

Fig. 5. Nitrate (△) and oxygen (◇) integrated between 100 and 600 m at Ocean Station P. Rates of change from 1994 to 2003 are +0.26 mol NO_3 m^{-2} y^{-1} and −2.4 mol O_2 m^{-2} y^{-1}. Between 2003 and 2006, rates are −0.66 mol NO_3 m^{-2} y^{-1} and +6.7 mol O_2 m^{-2} y^{-1}.

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3.2. Oxygen distribution on isopycnal surfaces from WOCE one time surveys

To interpret trends at OSP, it is necessary to identify the regions in the North Pacific where the interior waters of the subarctic region exchange gases with the atmosphere. WOCE completed its surveys of the subarctic Pacific (Fig. 6) between 1985 (P1) and 1994 (P15). These data are used to characterize waters on the 26.5, 26.7 and 26.9 isopycnal surfaces poleward of 35°N, the lightest of these layers commonly outcropping in the western Pacific during winter (Mecking et al., 2006, Fig. 3). As well as showing WOCE stations, Fig. 6 also identifies sites at which oxygen saturation exceeded 95% on the 26.5 isopycnal, these locations being areas where winter ventilation of interior subarctic waters is inferred. Three larger coastal regions are also encircled, identifying areas that will subsequently be shown to have low.

Each of the selected density surfaces deepen from west to east in the subarctic (Fig. 7), the densest layer being ~200 m in the vicinity of 160°E and ~460 m near North America. As subarctic waters flow across the Pacific in the NPC, they warm and lose oxygen. Coldest waters on the 26.5 isopycnal (−1.5 to 1.5 °C, data from Okhotsk not plotted) are found in the Sea of Okhotsk and along the coasts of northern Japan and Russia. South of the Alaska Gyre NPC waters are between 6.8 and 8 °C. Waters >8.0 are found south of 42°N in the subtropical domain. Near the North American coast, waters have warmed by 2–6 °C and lost 35–200 μmol kg^{-1} oxygen in their ~6000 km journey across the Pacific. The lowest oxygen levels are found in coastal waters of the California Undercurrent, decreasing to 45 μmol kg^{-1} on the 26.9 isopycnal at a depth of 460 m.

On the 26.5 isopycnal surface oxygen levels >95% saturation are broadly found in Western Subarctic Gyre (Figs. 6 and 8), also on the shelf of the Okhotsk Sea, in the Bering Sea Gyre and near the Russian coast in the East Kamchatka Current. Many of the WSG values come from the 1985 survey of P1, not from the warm 1990s when other lines were surveyed. Each of the WOCE surveys was carried out in summer when thermal stratification prevented the 26.5 isopycnal from reaching the ocean surface.

Oxygen decreases to saturation levels of ~65% in the eastern waters of the NPC. On the 26.7 (26.9) isopycnals, oxygen declines from >80% (50%) saturation in the Okhotsk Sea and near Japan to <30% (~10%) along the North American coast from California to British Columbia. Low oxygen and high nitrate occur in coastal regions along: (1) the Aleutian Islands on section P14, (2) the south coast of Alaska and (3) the California to British Columbia coast (Fig. 8). Region 2 includes the waters of the Alaska Gyre. Between region 2 and the North Pacific Current directly to the south on the 26.5 isopycnal, oxygen increases from ~150 to ~250 μmol kg^{-1} as nitrate decreases from ~32 to ~20 μmol kg^{-1} (O_2/NO_3 ratio of 9.2). OSP lies in the midst of this gradient.

Normally, nitrate distribution will be the inverse of oxygen, since oxygen is consumed in a fairly constant ratio as nitrate is remineralized from organic N in seawater. In general, this is evident in Fig. 8. High oxygen

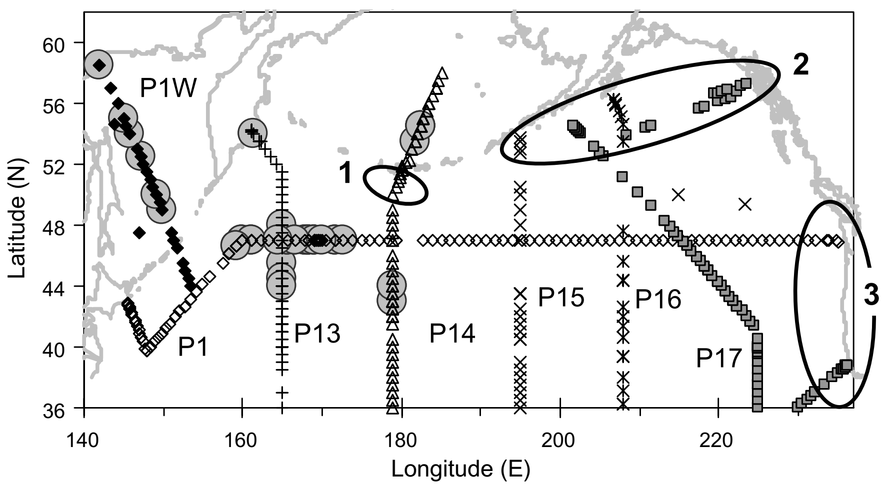

Fig. 6. Station locations of WOCE one-time surveys in the subarctic Pacific (lines P1 through P17). Regions of low oxygen (ovals numbered 1–3) are referred to in subsequent figures. Stations highlighted by grey circles in the western subarctic region indicate locations at which oxygen saturation exceeds 95% on the 26.5 isopycnal.

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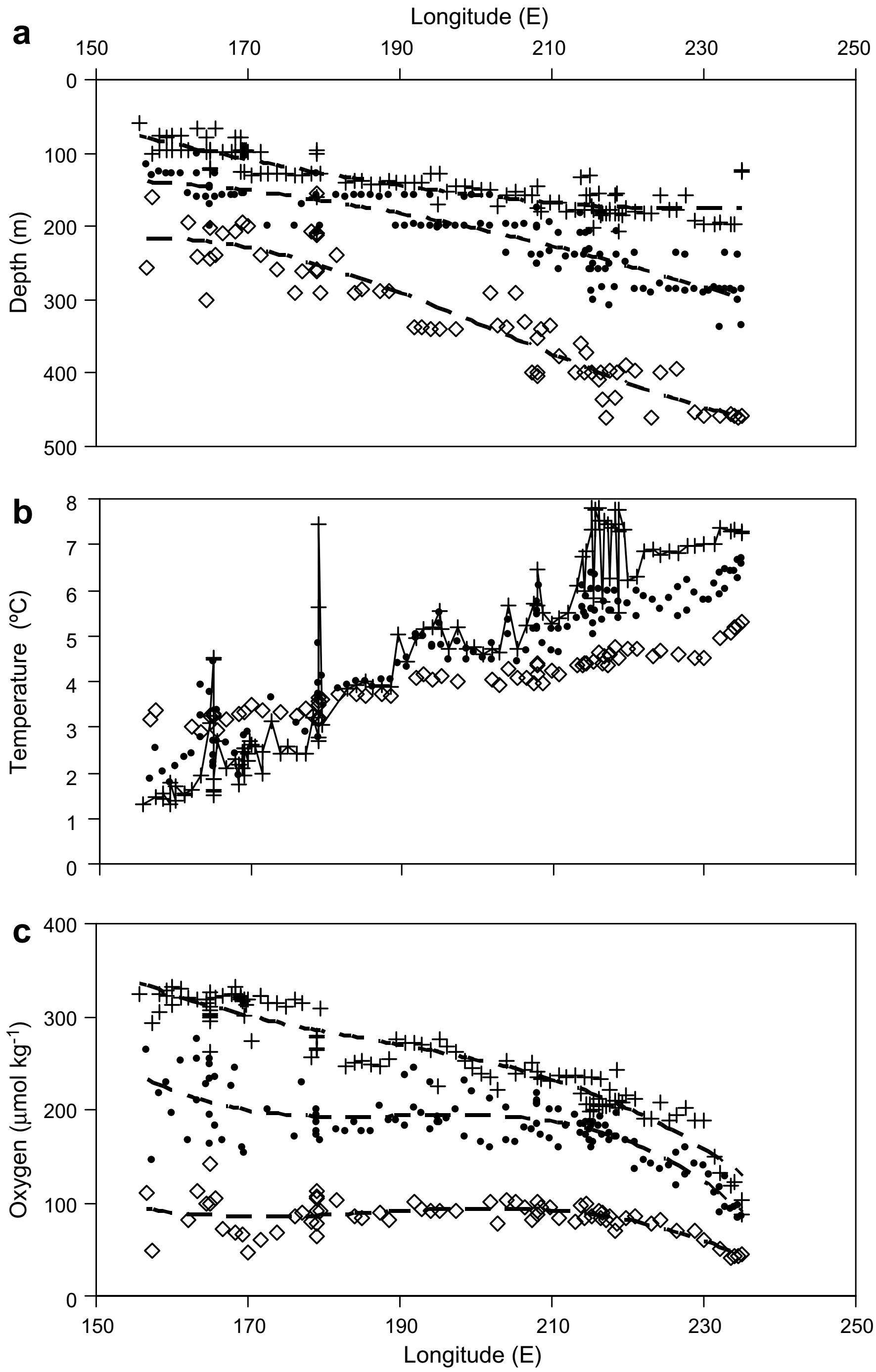

Fig. 7. Plots of: (a) depth, (b) temperature and (c) oxygen on the 26.5 (+), 26.7 (●) and 26.9 (◇) isopycnal surfaces between 45 and 48°N, 155°E and 125°W from WOCE surveys. Data in panels a and c are fit with a third order polynomial to show trends (dashed lines).

and low nitrate is found in waters near the Asian coast. However, nitrate is not highest off the coast of California, despite oxygen levels being lowest for the entire subarctic Pacific. Except for one high nitrate value off the BC coast, nitrate reaches maximum concentrations of 33 µmol kg^{-1} along the south Alaska coast. The high BC value (37 µmol kg^{-1}) is the result of estuarine outflow from a coastal channel (Whitney et al., 2005). Between P13 and P14, nitrate drops from ~18 to 14 µmol kg^{-1} in the latitudinal band 42–45°N, its lowest concentration in the subarctic Pacific, excluding the Sea of Okhotsk.

The covariance of nitrate and oxygen is seen more clearly in Fig. 9. Within the scatter of data, regional patterns emerge. Data are bound by dashed lines with O_2/NO_3 slopes of 9.2, the same ratio observed along

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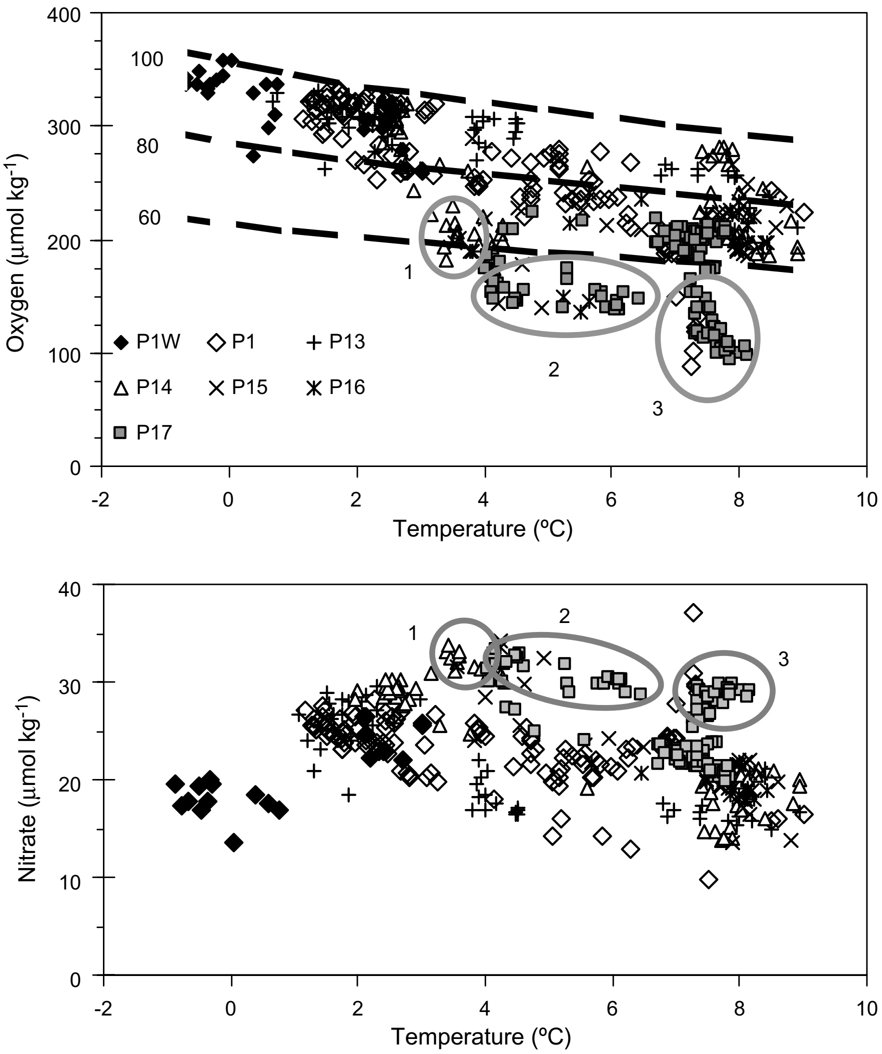

Fig. 8. Oxygen vs. temperature (top) and nitrate vs. temperature (bottom) for all data on the 26.5 isopycnal surface from WOCE survey lines. Oxygen (salinity 33) is shown for 60%, 80% and 100% saturation (dashed lines). Clusters of oxygen and nitrate data are circled (low O_2) and their locations shown in Fig. 6.

the 26.5 isopycnal in the eastern subarctic. Data along the California and southern BC coast cluster along the lower line (oval 3). Successive data clusters off the south coast of Alaska and through the Aleutian Islands (ovals 1 and 2) show higher oxygen and nitrate as waters take on more subarctic characteristics. Data clusters with the lowest nitrate levels outside the Sea of Okhotsk are found in subtropical waters on WOCE sections P13 and P14.

4. Discussion

4.1. Oxygen and temperature in the interior ocean at OSP

A freshening of the surface layer and shoaling of the winter mixed layer from 120 to 100 m between the 1960s and 1990s (Freeland et al., 1997) are symptoms of continually increasing stratification of the upper ocean at OSP. Royer and Grosch (2006) also find that upper ocean stratification is enhancing ocean stratification along the Alaska coast. As a result, nutrient transport to the mixed layer weakened along Line P between the 1970s and 1990s (Whitney et al., 1998). The 50 years of observations at OSP suggests upper ocean stratification is also reducing oxygen transport to the ocean interior. The most rapid rates of O_2 decrease are

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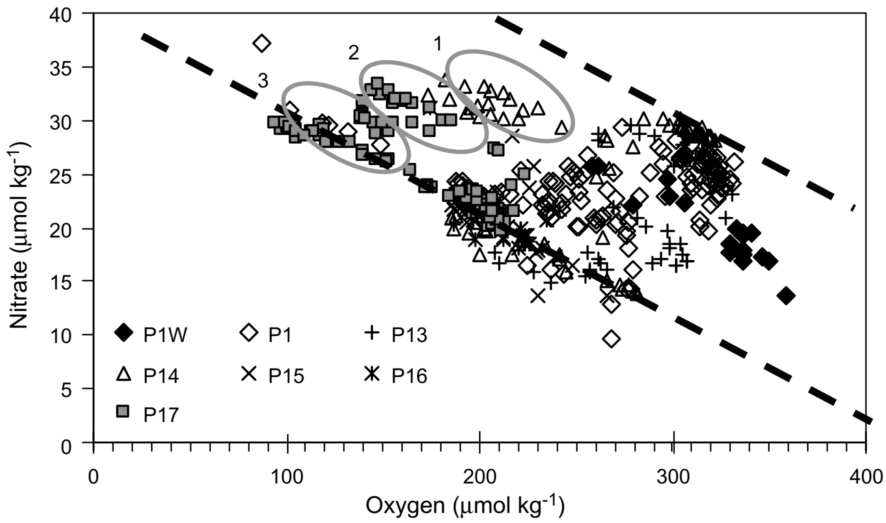

Fig. 9. Nitrate vs. oxygen for WOCE data on the 26.5 isopycnal. Dashed lines show an O_2:NO_3 ratio of 9.2 for waters of subtropical (lower) and subarctic (upper) nature. Three circled data groups are the same as those in Figs. 6 and 8, being found: (1) near the Aleutian Islands, (2) along the Alaska south coast and (3) along the BC to California coast.

observed between 100 and 400 m, on density surfaces of ~26.3–27.0. Over this range, oxygen is decreasing on average by 123 mmol m^{-2} y^{-1}, equivalent to a loss of 22% over 50 years. This loss of oxygen has resulted in the "hypoxic" boundary (considered to be 60 µmol kg^{-1}; Gray et al., 2002) shoaling from ~400 to 300 m between the 1950s and the present.

Oxygen declines are detectable to at least 1000 m at OSP. The vertical transport of particulate organic carbon at this site is 6.6 g C m^{-2} y^{-1} to 200 m and 2.7 g C m^{-2} y^{-1} to 1000 m (Wong et al., 1999). Carbon loss between these depths (2.9 g C = 240 mmol C) is greater than the 50 year average integrated rate of oxygen loss (165 mmol O_2 m^{-2} y^{-1} from Table 1). Reduced ocean ventilation results in carbon dioxide being stored in the ocean interior at a rate we calculate to be 130 mmol m^{-2} y^{-1} in the vicinity of OSP (O_2 consumption/1.3 to convert to CO_2; Anderson, 1995). This accumulation rate for CO_2 is considerably less than the average carbon remineralization rate of ~2000 mmol m^{-2} y^{-1} estimated for subsurface waters (200–900 m) of the subarctic Pacific by Feely et al. (2004) or for the integrated oxygen consumption rate (150–800 m) derived by Imasoto et al. (2000) of 1600 mmol m^{-2} y^{-1} for the same region, and suggests perhaps 5–10% of the CO_2 being produced by remineralization of organic detritus is being stored in the ocean interior. Without oxygen transport to OSP, the approximately 34 mol O_2 m^{-2} found between 200 and 1000 m (Table 1) would be consumed within 15–20 years.

Warming is evident in subsurface waters over the 1956–2006 period, but without the strong cool period in the 1960s, this trend would weaken considerably. During the cool 1960s and from 1995 to 2003, oxygen levels diminish relatively rapidly. Andreev and Baturina (2006) suggested this was the result of an intensified winter Aleutian Low pressure system, causing a strengthening and expansion of the Alaska Gyre. However, Crawford et al. (2007) show that increased southward transport of cool waters in this region also occur when the positioning of the Aleutian Low and North Pacific High create stronger westerly winds towards the BC coast. The warm period in the early 1990s, associated with a series of El Niño events, slowed the rate of oxygen decline on the 26.5 and 26.7 isopycnals but had little effect on deeper waters.

Only in the past decade has it been clear that mesoscale eddies transport coastal waters westward against prevailing currents in the NE Pacific (Whitney and Robert, 2002). Along the BC and SE Alaska coasts, waters are warmer and contain less oxygen than in the Alaska Gyre, due both to the influence of the California Undercurrent and to the increased oxygen demand fuelled by high primary productivity. During the Weathership era, water properties were measured frequently enough to resolve the passage of eddies through OSP. Both in 1960 and 1974, depressed isopycnals, elevated temperature and low oxygen signal the presence of these anticyclonic eddies at OSP for 4–6 months. In these eddies, temperature and oxygen levels on the 26.7 isopycnal are similar to those measured at station P4 near the BC coast (Fig. 4), indicating waters of coastal origin

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that contain a substantial CUC component. If these short duration anomalies were due to enhanced recirculation of waters from the coast of Alaska to OSP (as described in Whitney et al., 2005), depressed isopycnal surfaces coincident with the low oxygen/high temperature would not accompany the anomaly, nor would the temperature anomaly be as strong (Fig. 8). Some of the variability seen over the past couple of decades could also have resulted from such eddies.

Oxygen has been observed oscillating in the western waters of the subarctic Pacific on an 18–20 year cycle (Ono et al., 2001; Andreev and Kusakabe, 2001; Watanabe et al., 2003). A recent suggestion is that this oscillation is driven by the 18.6 year lunar nodal cycle (Yasuda et al., 2006; Andreev and Baturina, 2006) which causes diurnal tidal amplitude to vary by 20% and consequently gives rise to strong tidal mixing variability in the narrow passes between Okhotsk Sea and the Pacific Ocean. On the 26.9 and 27.0 isopycnals at OSP, such an oscillation can be detected, with oxygen maxima appearing in 1959, 1978 and 1995. These peaks occur 6–7 years after those observed off Japan and 9–10 years after the nodal maxima (1951, 1969 and 1988). Such oscillations make the interpretation of climate trends difficult with short term data records.

4.2. Oxygen transport within the subarctic Pacific

Oxygen enters the interior ocean in areas where dense waters exchange gases with the atmosphere. In the subarctic Pacific, the densest isopycnal surfaces (~26.6) are mixed to the surface in the cold waters along the Asian coast and in the Bering Sea. In the eastern Pacific, the 26.2 isopycnal has outcropped occasionally in the Alaska Gyre in the past several years but at OSP, the 26.0σ_θ has not been observed at the ocean surface since 1971 despite the frequent measurements that have being taken in this region by profiling Argo floats since 2001 (Freeland and Cummins, 2005).

Rapid declines in oxygen (~60 μmol kg^{-1} over ~10 years; Watanabe et al., 2001; Kumamoto et al., 2004) have been reported for the interior waters of the subarctic Pacific in recent years from repeat survey sections. These surveys sampled through a period of persistently weak ventilation which resulted in high apparent oxygen utilization (Ono et al., 2001; Mecking et al., 2006; Andreev and Baturina, 2006, Fig. 4). Such periods provide a good opportunity to estimate oxygen consumption rates, since periods between sampling can be similar to the aging of these water masses. Estimates of oxygen utilization rates vary from 2.1 M m^{-2} y^{-1} between 26.5 and 27.2σ_θ in the NPC (Aydin et al., 2004) to 4.1 M m^{-2} y^{-1} over the depth range 200–900 m as an average for the subarctic Pacific (Feely et al., 2004). In comparison, the observed rate of oxygen decline at OSP was 2.4 M m^{-2} y^{-1} (100–600 m, Fig. 5) through the low ventilation period of 1994–2003, but is only 0.14 M m^{-2} y^{-1} over the same depth range (Table 1) when averaged over 50 years. Over the entire observation period, annual oxygen consumption only slightly exceeded transport, but resulted in a 22% decline in oxygen levels.

Circulation models developed in recent years clarify transport processes in the interior subarctic Pacific, especially within the North Pacific Current. The strong Kuroshio and Oyashio Currents initiate the NPC near the Asian coast. Ueno and Yasuda (2003) picked the 26.7 and 27.2 isopycnals to show how flow and mixing along these surfaces transports oxygen from the ventilation regions along the Asian coast and heat from the subtropics towards North America. Their model provided estimates of transport times from the edge of ventilated waters to OSP of 7 years on the 26.7 surface and 15 years on 27.2σ_θ. A 10 year lag was observed in a high oxygen, low temperature signal crossing from the WSG to the AG in the 1960s (Andreev and Watanabe, 2002). These results are consistent with the increase in age of NPC waters observed over a shorter section of the NPC (between WOCE lines P14 and P17, Fig. 6) of from 4 years on 27.0σ_θ to 6 and 8 years on the 27.1 and 27.2 isopycnals (Aydin et al., 2004).

Oxygen declines can be due to either increased biological demand or changes in physical processes. In modelling oxygen transport within the North Pacific, Deutsch et al. (2006) found little evidence of increased biological activity, but instead showed that changes in ocean circulation and ventilation accounted for both episodic oxygen increases in the subtropics and declining oxygen in the subarctic Pacific. Reduced ventilation of the 26.6 isopycnal was thought to be the major reason for decreasing oxygen levels in the subarctic region. A freshening upper ocean (Freeland et al., 1997; Wong et al., 2001; Royer and Grosch, 2006), something we continue to see at OSP, may be exacerbating this trend. Over ~50 years, Andreev and Watanabe (2002) observed a surface salinity decrease in the Bering Sea of 0.32 per century which is close to the 0.36 per century

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we calculate for OSP. Surface freshwater transport from the eastern to western Pacific, which is strengthened during periods of intense Aleutian Lows, appears to be an essential modulator of ocean ventilation throughout the subarctic region (Andreev and Baturina, 2006).

4.3. Using NO to characterize water masses influencing the subarctic Pacific

Broecker (1974) derived a conservative tracer using nitrate and oxygen to track water masses in the Atlantic, using the formulation $9NO_3 + O_2 = NO$. Based on rates of change in oxygen and nitrate at OSP through a period of persistent oxygen decline (Fig. 5) we slightly modify the NO parameter to be $9.2NO_3 + O_2$. NO is a water mass signature that is carried from its formation region; waters exchanging gases with the atmosphere in cold regions with high nitrate levels will have high NO whereas warm waters of low nitrate produce a low NO value. As waters sink away from their formation regions, remineralization of organic matter consumes oxygen and produces nitrate in a fairly constant ratio as long as oxygen is available. Biotic processes in oxygen rich waters do not alter the NO of a water mass, hence this parameter is considered conservative (does not alter with time). When oxygen nears zero, denitrification by bacteria will begin to remove nitrate, producing anomalously low NO.

Based on Redfield ratios of major elements in marine algae, the O_2/NO_3 ratio is 8.5 during oxidation of organic matter (e.g. Anderson, 1995). However, the carbon to nitrogen ratio of marine detritus at OSP increases from 6.5 in the suspended particles of the photic zone to 8.0 in particles sinking to 100 m (Wong et al., 1999), potentially increasing the O_2/NO_3 ratio to 9.5 for complete oxidation of this organic material. For subsurface waters, a ratio of 9.2 fits well with the composition of its detrital material.

In the North Pacific, NO provides insights to water mass transport and mixing that temperature and salinity cannot. Strong regional differences are seen throughout the North Pacific (Fig. 10), with values up to 580 μmol kg^{-1} being found on the 26.5 isopycnal in the Bering Sea during the WOCE survey in 1992 and 1993. Unpublished data from surveys we carried out along 165°W in 1991 and 1992 also show NO values approaching 580 μmol kg^{-1} in the Western Subarctic Gyre (Whitney, unpublished). Minima for NO occur across the subtropics and in the California Undercurrent. Okhotsk Sea Intermediate Waters differ from those

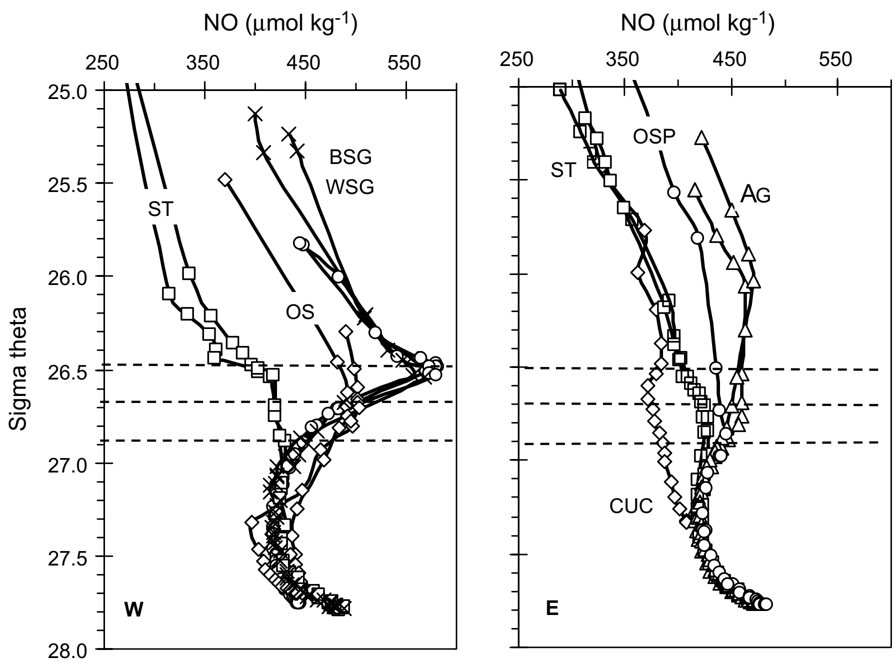

Fig. 10. NO vs. density for representative profiles in the western (W) and eastern (E) Pacific. Western water types include subtropical (ST), Okhotsk Sea (OS) where waters flow into the Pacific, Bering Sea (BSG) and Western Subarctic Gyres (WSG). Eastern water types are the California Undercurrent (CUC), subtropics (ST), Ocean Station P (OSP) and Alaska Gyre (AG). Dashed horizontal lines highlight isopycnal surfaces of interest.

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in the EKC and WSG either because the 26.5 and 26.7 isopycnals are close enough to the surface on the northern shelf that nitrate can be utilized by phytoplankton or due to denitrification within sediments of the broad shelf (Chen et al., 2004; Yoshikawa et al., 2006). Perhaps other processes such as brine rejection during ice formation, and incomplete ventilation of waters under winter ice contribute to the low NO signature of OSIW. In the eastern subarctic region, Alaska Gyre waters contain the highest NO values, although these waters only ventilate to the ~26.2 isopycnal. No subsurface maximum indicative of winter ventilation to depth is seen in the profiles selected from 1993 WOCE surveys. Situated between the NPC and the gyre, OSP has an intermediate nitrate–oxygen value compared with AG and the subtropics. Low NO in the California Undercurrent is the result of denitrification further south (Castro et al., 2001) and can be used to track its influence up the coast of BC and Alaska. Especially the denitrification signatures of Okhotsk and the CUC are lost when trying to use temperature or salinity to track oxygen transport and consumption in the interior ocean.

Other NO values from the North Pacific covary with temperature (Fig. 11). Trends are similar on three isopycnals, except there is less evidence of the well ventilated BSG/EKC water below the 26.5 isopycnal. Cold OSIW is still common on the 26.7σ_θ, however Okhotsk Sea waters are warmer (~1.4 °C) and have lower NO (~465 μmol kg^{-1}) on the 26.9 surface. Such low values at depth suggest a strong shelf component to these waters (Wong et al., 1998) and less contribution from the EKC. Assuming 100% oxygen levels during water formation, preformed nitrate concentrations vary between 7 and 27 μmol kg^{-1} north of 35°N in the Pacific. Data from OSP over the period 1987–2006 all fall in the centre of the mixing line for 26.7σ_θ waters, whereas NO from the near coastal station P4 shows similarity to the CUC. The distinct characteristics of various contributing water masses to the NPC make them useful in defining their influence on this trans-Pacific current and within the Gulf of Alaska.

Water mass characteristics are summarized in Table 2 and mapped out for the 26.7 isopycnal in Fig. 12. There is a clear mixing line (Fig. 11) between the cool, NO rich waters of the western subarctic (2 °C and

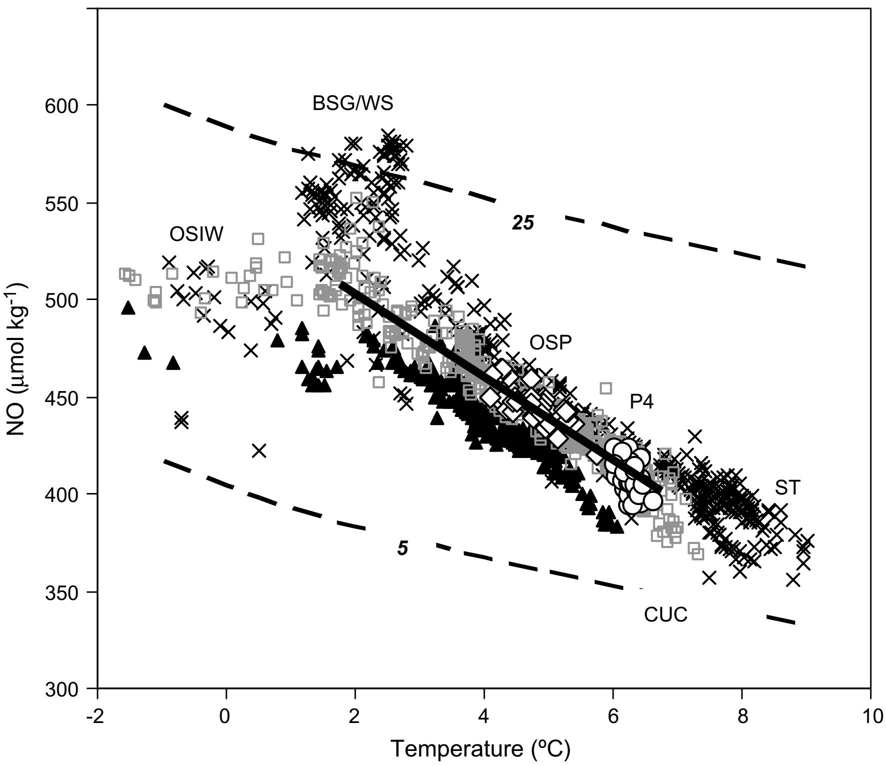

Fig. 11. NO vs. temperature on the 26.5 (×), 26.7 (□) and 26.9 (▲) isopycnal surfaces. Dashed lines represent preformed NO at 100% oxygen saturation and nitrate concentrations of 5 and 25 μmol kg^{-1} at a salinity of 33. Waters of the Bering Sea Gyre (BSG) and East Kamchatka Current (EKC) show elevated NO values, whereas lowest values identify subtropical (ST) waters. Okhotsk Sea Intermediate water (OSIW) has a slightly reduced NO signature. The CUC waters along the California coast show weak evidence of denitrification on all density surfaces. Also shown are NO values for OSP (◇) and P4 (○) on the 26.7 isopycnal for the period 1987–2006.

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Table 2
Water mass characteristics (±SD of data in defined regions) on the 26.5 and 26.7 isopycnal surfaces from WOCE data

Water type	26.5 isopycnal			26.7 isopycnal		
	T (°C)	Salinity	NO (μmol kg^{-1})	T (°C)	Salinity	NO (μmol kg^{-1})
Bering Gyre, 53–56°N	2.6 ± 0.1	33.22 ± 0.02	575 ± 5	3.64 ± 0.08	33.61 ± 0.03	482 ± 7
WSG at 165°E, 47–52°N	2.1 ± 0.5	33.2 ± 0.1	545 ± 5	2.9 ± 0.7	33.52 ± 0.09	490 ± 20
Okhotsk, 49–52°N	0.0 ± 0.5	33.03 ± 0.02	504 ± 12	0.35 ± 0.45	33.27 ± 0.06	508 ± 11
NPC at 179°E, 45–48°N	3.0 ± 0.3	33.27 ± 0.10	505 ± 15	3.8 ± 0.7	33.64 ± 0.07	470 ± 16
NPC at 165°E, 41–46°N	4.1 ± 0.2	33.4 ± 0.1	460 ± 10	4.4 ± 0.7	33.69 ± 0.11	450 ± 15
NPC/AG at ~150°W 48–54°N	4.3 ± 0.2	33.4 ± 0.1	450 ± 5	4.4 ± 0.4	33.70 ± 0.06	451 ± 10
Alaska coast, 135–140°W 56–57°N	6.0 ± 0.2	33.65 ± 0.05	415 ± 5	6.0 ± 0.2	33.91 ± 0.02	417 ± 5
Subtropics at 165°E 35–41°N	8.2 ± 0.2	34.08 ± 0.05	380 ± 10	6.0 ± 0.5	33.93 ± 0.07	412 ± 10
CUC at ~124°W, 38°N	7.83 ± 0.08	34.00 ± 0.05	371 ± 2	6.9 ± 0.3	34.08 ± 0.04	380 ± 8

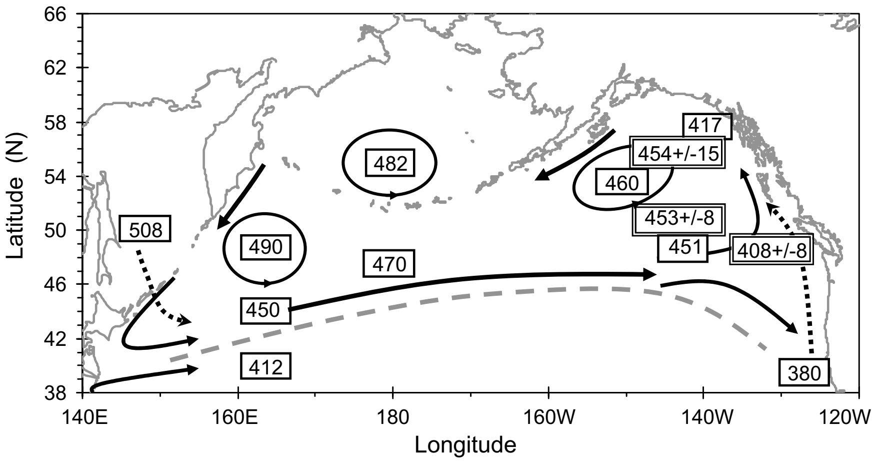

Fig. 12. NO characteristics (μmol kg^{-1}) of water masses (identified in Table 2) on the 26.7 isopycnal surface in the Northern Pacific Ocean. Arrows show the direction of flow of major currents and gyres. Standard deviations are provided for three time-series stations in the NE Pacific, all other values are based on one-time WOCE surveys.

NO ~ 500 μmol kg^{-1}) and those of the subtropics (6 °C, NO = 412 μmol kg^{-1}), although how the cool end member forms is not obvious. Its contributors will be OSIW (NO ~ 508 μmol kg^{-1}) and the WSG (NO ~ 490 μmol kg^{-1}). To estimate the influence of the western subarctic on the region around Ocean Station P, we use an NO of 500 μmol kg^{-1} to type this source region with an uncertainty of 10 μmol kg^{-1}. The subtropics component is well defined along 165°E. These waters will be somewhat contaminated by the subarctic as North Pacific Intermediate Water forms along the NPC (You, 2005). Still, they are the waters that supply the subtropical component of the NPC.

We now use NO to estimate contributions of western subtropical and subarctic waters to the makeup of water masses in the eastern subarctic. Table 3 summarizes these estimates and shows that, for the WOCE data set of the 1990s, waters of the Alaska Gyre were ~60% from the WSG region and 40% from the subtropics on the 26.7 isopycnal. On the edge of AG, stations Z9 and OSP contained slightly more water from the subtropics (Fig. 12), although here variability could be assessed. With fewer samplings in a period of high variability due to a strong El Niño (1997/98) and La Niña (1999), the subtropical component at Z9 varied between 25% and 65%. At OSP, the variability (assessed as 1SD) was less, with the subtropics contributing 37–57%. An "outlier" for OSP in February 1995 (Fig. 13) suggested waters were ~90% subtropical, with their abnormally warm 5.7 °C temperature supporting this estimate.

An important issue along the coast of Oregon, Washington and BC in the past several years has been hypoxia of shelf waters (Grantham et al., 2004). Understanding which waters (subarctic or subtropical) are being upwelled onto the shelf in summer is core to determining why these events are now occurring when they had not before. NO is a good parameter to track influences of the California Undercurrent or AG waters in

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Table 3
Estimated contributors of oxygen along the 26.7 isopycnal surface in the NE Pacific

Water mass	% Subarctic	% Subtropical	Comments
WSG	100	0	1992
ST	0	100	1992
AG	60	40	1993
Z9	55 ± 20	45 ± 20	1991–1999, $n = 8$
OSP	53 ± 10	47 ± 10	1987–2006, $n = 40$
	% AG	% CUC	Comments
P4	35 ± 10	65 ± 10	1987–2006, $n = 44$
SE Alaska	54	46	1993

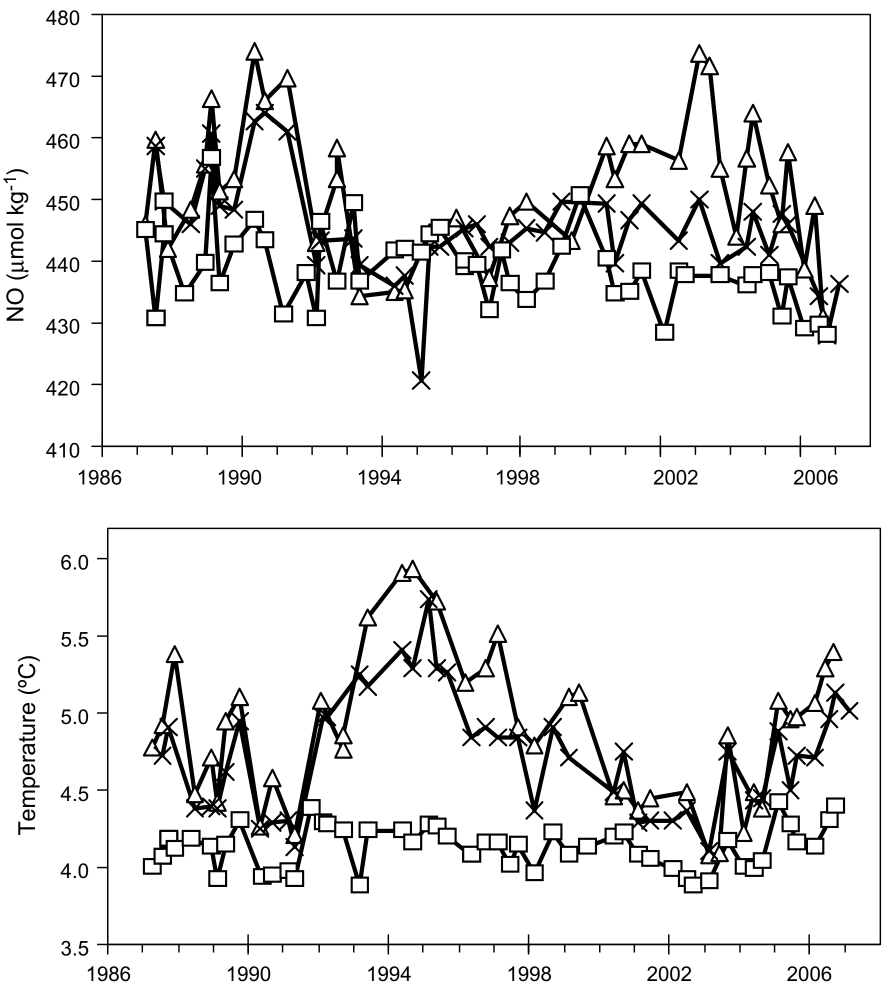

Fig. 13. Changes in NO and temperature at OSP on the 26.5 (△), 26.7 (×) and 26.9 (□) isopycnal surfaces between 1987 and 2006.

this region. WOCE surveys suggest waters of the SE Alaska coast are a near equal mixture of CUC and AG waters in 1993. At station P4 on the southern BC coast, CUC has comprised ~55–75% of the slope waters on the 26.7 isopycnal in the past 20 years. This is similar to the findings of Mackas et al. (1987) for summer 1980 when CUC formed a large fraction of the waters below ~100 m along the continental slope of southern BC.

The interannual variability of NO and temperature at OSP are plotted for the period 1987–2006 (Fig 13). Both parameters clearly show the influence of the serial El Niños in the mid 1990s. Temperatures on the 26.5 and 26.7 isopycnals rose to almost 6 °C in 1995 as NO decreased as low as 420 µmol kg^{-1}. Assuming contributing water mass properties changed little through this period, this low NO suggests waters at OSP were 90%

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subtropical. Such a strong change in water properties requires a substantial northward movement of the NPC. With the advent of the Argo array of profiling floats, it has become possible to monitor shifts in large-scale ocean current systems. Freeland and Cummins (2005) and Freeland (2007) demonstrated methods that allow such current tracking to be executed and show that since the initiation of the Argo array in the NE Pacific in 2001 some large shifts (several degrees latitude) in the location of the NPC have indeed taken place. Because Argo has been deployed for only a few years, the statistics available from such mappings are insufficient to allow any shifts to be correlated with the Line P time series. However, the idea that changes in water properties might be associated with spatial variability in major ocean currents is supportable as shifts in the NPC are now known to take place.

4.4. Implications of oxygen decline

Understanding ocean ventilation provides great insight into the circulation of the interior ocean. Because ventilation sites are rare, processes within them have the potential to control biological activity over broad areas and time scales. Past periods of weak ventilation appear to be the cause of near anoxic conditions in regions affected by the subarctic Pacific. Core records from Baja California show periods of near anoxia thousands of years ago (Van Geen et al., 2006), the suggested cause being weakened ocean ventilation in the western subarctic Pacific. WOCE repeat sections consistently point to a rapid decline in oxygen throughout the subarctic Pacific between the late 1980s and 2000, reporting oxygen declines of 40–80 μmol kg^{-1} in large patches of the subarctic over periods of 6–14 years (Watanabe et al., 2001; Emerson et al., 2004; Kumamoto et al., 2004; Mecking et al., 2006). Such rates of decline, if persisting, could turn broad regions of the subarctic Pacific hypoxic within a decade. However, observations at OSP show there are multi-year periods of virtually no ventilation such as those seen in the late 1960 and 1990s (Fig. 4), followed by periods of oxygen enrichment that may be associated with increased tidal mixing amongst the Kuril and Aleutian Islands (Yasuda et al., 2006; Andreev and Baturina, 2006). Enhanced mixing at nodal tidal maxima in 1969 and 1988 match periods of minimum AOU (maximum oxygen concentration) in Oyashio waters off northern Japan (Ono et al., 2001). With a lag of 6–9 years, this signal is evident on the 26.9 isopycnal at OSP and perhaps on shallower surfaces as well (Fig. 4). The broad outcropping of the 26.5 and 26.6 isopycnals in March 2006 (Fig. 3) may be the result of increased tidal mixing as the next nodal maximum is approached. Andreev and Kusakabe (2001) attempted to link oscillations in oxygen levels in the Western Subarctic Gyre and Sea of Okhotsk with atmospheric forcing. They observed oxygen maxima on the 27.0 isopycnal surface in WSG near 1951, 1969 and 1988, each a nodal tidal maximum.

In the depth range 125–300 m at OSP, oxygen decreased between 17% and 30% (20–40 μmol kg^{-1}) from 1956 to 2006. The oscillations recorded at OSP have resulted in the hypoxic boundary (~60 μmol kg^{-1}) varying between ~400 and 250 m depth. Such changes must affect the distribution of organisms intolerant of low oxygen. Little research appears to have been done on the distribution of oceanic species within oxygen gradients, however quite a sharp oxygen boundary of 3.4 mg l^{-1} (~100 μmol kg^{-1} or 28% O_2 saturation) was noted for Atlantic cod in the Gulf of St Lawrence (D'Amours, 1993), demonstrating oxygen is a crucial parameter in defining ocean habitat. Marine animal life depends on oxygen. The impacts of oxygen declines need to be understood for both open ocean and coastal ecosystems. Low oxygen is of broad concern because of human impacts on continental margins where increased waste discharges place demands on the oxygen supply. This can arise from increased nutrient inputs such has been observed in the Gulf of Mexico (Turner and Rabalais, 1994) or localized discharges from coastal industry and municipal sewage treatment plants that may affect areas of restricted circulation. More ominously, hypoxia may increase in the subarctic Pacific due to global warming as upper ocean stratification strengthens (Freeland et al., 1997; Royer and Grosch, 2006). Bakun (1990) theorized that a warming world would establish stronger thermal gradients between land and ocean, resulting in stronger along shore winds. This would enhance upwelling, increasing nutrient supply to surface waters and placing higher demands on bottom water oxygen as an increased flux of detritus is consumed. Not imagined by Bakun was the decline in oxygen and increase in nutrient levels in the ocean interior. Increased upwelling in his scenario will be accompanied by transport of waters predisposed to cause hypoxia or anoxia. Fish and crab kills in the Oregon upwelling region in the past few years (Grantham et al., 2004) suggest this is occurring to some extent already. As oxygen declines in the subarctic Pacific, the hypoxic threshold will shoal.

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For example on the 26.7 isopycnal at OSP, oxygen levels have declined from ~187 to 150 μmol kg^{-1} over 50 years of observations. With no ventilation and oxygen consumption rates of ~4 μmol kg^{-1} y^{-1} (Feely et al., 2004), it would take a little more than 20 years to create hypoxia (60 μmol kg^{-1}) in these waters. This isopycnal is less than 250 m deep near the North American coast, a region where increased oxygen demand from productive waters places a high demand on oxygen.

On the Oregon to British Columbia coast, upwelling draws waters from depths of 100 to >250 m (Freeland and Denman, 1982; Whitney et al., 2005). If coastal upwelling does strengthen, and with some assurance that oxygen levels will continue to decline (and nutrient levels increase) in the subarctic waters of the Pacific, it is reasonable to project that shelf and slope ecosystems will lose oxygenated habitat. The few fish species such as sablefish and some rock fishes that tolerate low oxygen may expand their territory, but in general mid water organisms will be forced to find shallower habitat or perish. This will increase competition for resources and may expose some species to greater predation from above. In coastal basins and fjords whose basin waters are rejuvenated with periodic replacement from the ocean, declining oxygen levels in the subarctic Pacific could lead to serious hypoxia resulting in widespread mortality of benthic species.

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