

On the seasonal and interannual migrations of the transition zone chlorophyll front

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[1] High-resolution satellite measurements of ocean color and surface winds, along with historical in situ data, are used to explore interannual variability in the annual migration of the transition zone chlorophyll front (TZCF) in the central North Pacific Ocean. Significant variations in frontal position and annual range were observed, including a significant southerly displacement during El Niño events. This displacement, apparently forced by enhanced surface convergence and vertical mixing in the transition zone, creates vast regions of anomalously high wintertime surface chlorophyll in the North Pacific Subtropical Gyre. A remarkably close correspondence between the positions of the TZCF and the 18°C surface isotherm over a portion of the central North Pacific allows historical temperature data to be used as a proxy for TZCF position. This surface temperature proxy has revealed decadal-scale variability in frontal position, with greater (lesser) annual range and southerly extent following the large-scale 1976–77 (1998–99) climate shifts. Interannual variations in TZCF position could have important implications for the distribution and survival of a number of apex predators that utilize the TZCF as a migratory and foraging habitat.

INDEX TERMS: 4223 Oceanography: General: Descriptive and regional oceanography; 4215 Oceanography: General: Climate and interannual variability (3309); 4227 Oceanography: General: Diurnal, seasonal, and annual cycles; 4528 Oceanography: Physical: Fronts and jets; 4899 Oceanography: Biological and Chemical: General or miscellaneous. **Citation:** Bograd, S. J., D. G. Foley, F. B. Schwing, C. Wilson, R. M. Laurs, J. J. Polovina, E. A. Howell, and R. E. Brainard (2004), On the seasonal and interannual migrations of the transition zone chlorophyll front, *Geophys. Res. Lett.*, 31, L17204, doi:10.1029/2004GL020637.

1. Introduction

[2] A sharp transition in surface chlorophyll concentration separates the subpolar and subtropical gyres in most major ocean basins [Laurs *et al.*, 1984; Lewis *et al.*, 1988; Yoder *et al.*, 1993; Glover *et al.*, 1994]. In the North Pacific

Ocean, this feature has been called the transition zone chlorophyll front (TZCF), and is defined operationally as the 0.2 mg m⁻³ surface chlorophyll-*a* isopleth [Polovina *et al.*, 2001]. The TZCF has been identified as an important migratory and foraging habitat for a number of apex predators [Polovina *et al.*, 2000, 2001], and corresponds to an area of accumulation of marine debris [Kubota, 1994].

[3] The TZCF is observed to migrate more than 1000 km latitudinally over the course of an annual cycle, varying between 30–35°N in winter and 40–45°N in summer (Figure 1a). It is thought that this migration is due primarily to seasonal variations in vertical mixing, mixed layer depth, nutricline depth, and zooplankton grazing [Glover *et al.*, 1994; Chai *et al.*, 2003]. Interannual differences in frontal position have also been observed. Using the first three years of ocean color data from Sea-viewing Wide Field-of-view Sensor (SeaWiFS), Polovina *et al.* [2001] observed a greater southerly (northerly) extent of the TZCF in winter (summer) of 1998 during a strong El Niño event, and a much reduced latitudinal range during the strong La Niña event the following year (Figure 1b).

[4] More than six years of SeaWiFS ocean color data now allow for a more substantial characterization of the seasonal cycle of the TZCF as well as its interannual variability. The objectives of this paper are to (1) describe the seasonal and interannual migrations of the TZCF over the period 1997–2003; (2) examine the observed variability in frontal position in the context of the strong climate signals over this period; and (3) identify environmental proxies by which to extend the record back in time and explore decadal variability in TZCF position.

2. Data Sets

[5] Surface chlorophyll concentration data were obtained for the period September 1997 to October 2003 from SeaWiFS, version 4 data. The Pathfinder sea surface temperature (SST) data from the Advanced Very High Resolution Radiometers (AVHRR) were obtained from the NASA Jet Propulsion Laboratory. Both of these data sets were assembled with temporal resolution of 8-days and spatial resolution of 0.1° in both directions. COADS monthly

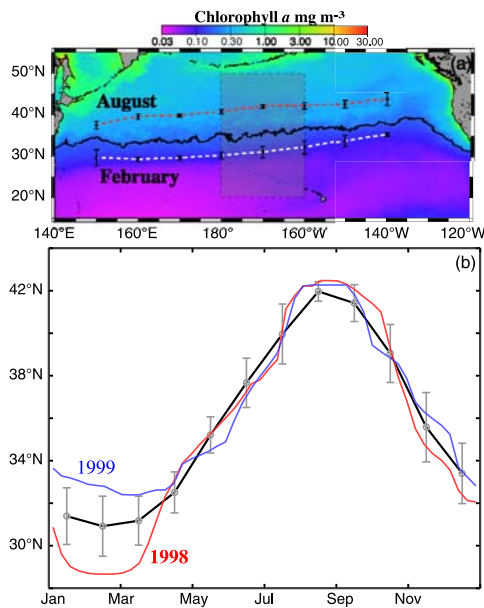


Figure 1. (a) Climatological mean surface chlorophyll-*a* in the North Pacific Ocean from SeaWiFS ocean color over the period September 1997 to October 2003. The 0.2 mg m⁻³ surface chlorophyll-*a* isopleth (black contour), which defines the TZCF, is overlaid. The mean positions of the TZCF and standard deviations in 10°-longitude zones between 140°W–150°E are shown for February (white dashed) and August (red dashed). The region explored in this study, bounded by 160°W–180°, is shown in the translucent box. (b) Monthly mean and standard deviation of the latitude of the 0.2 mg m⁻³ chlorophyll-*a* isopleth over the region bounded by 160°W–180° in the North Pacific Ocean. Data from 1998 (El Niño; red) and 1999 (La Niña; blue) are overlaid. Tic marks are at the beginning of the month.

surface temperature fields from 1960–97 were obtained with 1° spatial resolution. Ocean surface vector wind data were obtained at 8-day temporal resolution from the WOCE Mean Wind Fields produced by IFREMER/CERSAT. The data used here incorporate measurements from Advanced Microwave Instruments (AMI) on ERS-1 (1991–96) and ERS-2 (1996–99) and the SeaWinds sensor on QuikSCAT (2000–03). The spatial resolution of the ERS data is 1° in both directions while that of QuikSCAT is 0.5°. The fields include wind stress derived with a variable drag coefficient [Smith, 1988].

3. Results and Discussion

3.1. Seasonal and Interannual Variability of the TZCF

[6] The TZCF follows a consistent seasonal migration over most of the Pacific basin. Its climatological position based on six years of SeaWiFS data varies by only a few degrees latitude across the basin, but slopes sharply south-eastward east of 130°W (Figure 1a). The climatological winter (February) and summer (August) positions show a latitudinal range of approximately 10° over the annual cycle, consistent across most of the basin. The standard deviations of the winter and summer positions are considerably lower than the annual range at all locations between

150°E–140°W. At the eastern boundary, TZCF frontal position is more variable (not shown).

[7] The region of the central North Pacific bounded by 160°W–180° is representative of the seasonal and interannual variations in TZCF position over the open North Pacific. This is also a region with significant interactions between long-line fisheries and endangered species such as the loggerhead turtle [Polovina *et al.*, 2000]. The ~10° annual range and monthly standard deviations of ~2° are seen throughout the year (Figure 1b). Strong year-to-year variations in frontal position over this period are evident primarily in January–March. The winter position of the TZCF varied by nearly 500 km between 1998 and 1999, being greater than one standard deviation further south (north) in the El Niño (La Niña) year. During the weaker El Niño of 2002–03, the TZCF was also observed more than one standard deviation south of its climatological winter position.

[8] SeaWiFS-derived surface chlorophyll gradients can be used to monitor the position and strength of the TZCF. For most of the period of study, the maximum meridional gradient of surface chlorophyll follows the 0.2 mg m⁻³ surface chlorophyll isopleth (Figure 2), confirming the latter as a valid operational definition for the TZCF. But the strength of the frontal gradient varies both seasonally and interannually. The southernmost extent of the TZCF each winter is associated with the strongest meridional chlorophyll gradients, which usually weaken each spring. The particularly strong front at 36–38°N through spring and early summer of 1999 is an exception, but the TZCF tracked this gradient nonetheless. In summer, the strongest chlorophyll gradients tend to be north of the TZCF position, within the high chlorophyll regions of the subarctic gyre. The front strengthens again through autumn into winter.

3.2. Mechanisms of TZCF Variability

[9] The latitudinal variation of the TZCF results from a combination of several factors, including: (1) the amount of light and inorganic nutrients available for primary production at a given latitude; (2) the mixed layer depth, which controls light and nutrient levels in surface waters; and (3) grazing pressure on the phytoplankton [Glover *et al.*, 1994; Chai *et al.*, 2003]. The annual march of changing

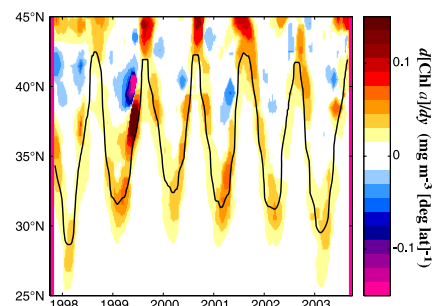


Figure 2. Time series (September 1997 to October 2003) of the meridional gradient of surface chlorophyll-*a* from the 8-day mean SeaWiFS ocean color averaged over the region bounded by 160°W–180° in the North Pacific Ocean. The latitude of the 0.2 mg m⁻³ surface chlorophyll-*a* isopleth (black contour) is overlaid. Tic marks are at the beginning of the year.

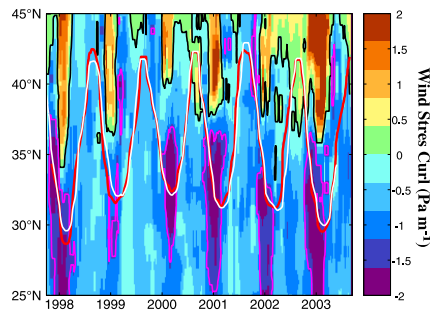


Figure 3. Time series (September 1997 to October 2003) of wind stress curl averaged over the region bounded by 160°W–180° in the North Pacific Ocean. The latitude of the 0.2 mg m⁻³ surface chlorophyll-*a* isopleth (red contour) and of the 18°C sea surface temperature isopleth (white contour) are overlaid. Tic marks are at the beginning of the year.

light levels will have some impact on the seasonal migration of the TZCF, but a strong pycnocline and relatively shallow mixed layers retain phytoplankton within layers of sufficient near-surface light levels throughout most of the mid-latitude North Pacific. Thus, variations in light level should not significantly impact TZCF seasonal range. Model results [Lewis *et al.*, 1988; Chai *et al.*, 2003] suggest that the processes that bring nutrients into the euphotic zone are most critical in determining frontal position. In regions just to the north of the front, the more nutrient-rich surface layer maintains relatively high chlorophyll levels throughout the year. South of the front more light is available for phytoplankton growth but the surface layer is seasonally nutrient-limited [Eppley *et al.*, 1973]. Thus, the extent to which nutrients can be upwelled into the surface layer, or advected laterally from higher latitudes, will determine the southern extent of the TZCF each winter.

[10] In addition to wintertime turbulent mixing, among the most persistent and effective mechanisms for redistributing nutrients in surface waters are Ekman pumping and horizontal divergence driven by the wind stress curl (WSC). Strong westerlies in mid-latitudes and easterlies in the tropics result in broad regions of negative WSC and a net convergence of surface waters in the North Pacific Subtropical Gyre. The winter TZCF position sits atop the regions of maximum negative WSC (Figure 3), implying that the equatorward advection of nutrient-rich waters is important in determining the southern extent of the TZCF. The lateral Ekman flux of nutrients has also been found to contribute strongly to new production in the subtropical North Atlantic [Williams and Follows, 1998]. Interannual variability in the timing, strength, total area, and location of the positive and negative WSC regions correspond to year-to-year differences in TZCF winter position. Large regions of both strongly negative and positive WSC, displaced further south than usual, were seen in the El Niño winters of 1997–98 and 2002–03. This was a result of both a larger meridional gradient of zonal wind stress, reflected in the WSC, and higher overall wind speeds (not shown), which would also act to more efficiently mix nutrients into the surface layer. The TZCF tracks these changes in surface winds, extending further south in El Niño winters and covering a large portion of the subtropical North Pacific

with anomalously high surface chlorophyll. The temporal lags between the seasonal development of the negative WSC regions and the southerly migrations of the TZCF (Figure 3) are consistent with the equatorward Ekman drift of nutrient-rich surface waters driving TZCF winter position.

3.3. A Proxy for Historical TZCF Position

[11] The same mechanisms that seasonally lift the nutricline will also lift the thermocline, reflecting a tight relationship between temperature and nutrient concentration that exists over much of the global ocean [Kamykowski, 1987]. Thus, surface isotherms can be expected to follow a similar seasonal migration to that of the TZCF. In fact, the 18°C surface isotherm closely follows the temporal evolution of the TZCF over the period 1997–2003 (Figure 3). Mean positions of the TZCF and the 18°C surface isotherm over the region bounded by 160°W–180° are highly correlated for all seasons (Figure 4).

[12] Since reliable historical records of SST in the North Pacific are readily available, the position of the 18°C surface isotherm across the basin provides a good proxy for describing long-term variability in TZCF position. A 44-year record of mean latitude of the 18°C surface isotherm across the central North Pacific reveals significant variability in both its winter and summer extent and seasonal range (Figure 5). Maximum northern extent varied between 40.4–43.3°N and averaged 42.0°N; maximum southern extent varied between 28.5–31.5°N and averaged 30.1°N. The greatest seasonal range and southern extent occurred in the early 1980s, which is consistent with a greater southerly extent of the TZCF estimated from Coastal Zone Color Scanner surface chlorophyll over the period 1979–86 (not shown) and from model results [Chai *et al.*, 2003].

[13] Differences in 18°C isotherm position are apparent on interannual and decadal time scales. Seasonal range increased by ~1° latitude following the 1976–77 climate shift [Miller *et al.*, 1994], primarily due to a greater southward encroachment in winter (Figure 5). This climate

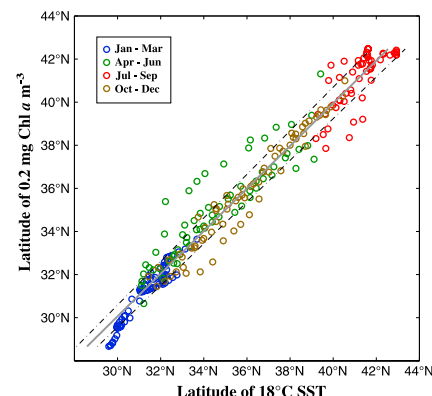


Figure 4. Latitude of the 0.2 mg m⁻³ surface chlorophyll-*a* isopleth vs. latitude of the 18°C sea surface temperature isopleth for 8-day means averaged over the region bounded by 160°W–180° in the North Pacific Ocean, September 1997 to October 2003. The linear regression ($r = 0.98$) and 95% confidence intervals are overlaid. Data are color-coded by season: winter (blue; $r = 0.96$), spring (green; $r = 0.91$), summer (red; $r = 0.78$), and autumn (yellow; $r = 0.98$).

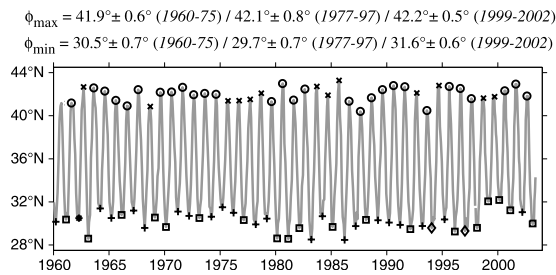


Figure 5. Time series of the latitude of the 18°C sea surface temperature isopleth from COADS (1960–1997) and Pathfinder (1998–October 2003) averaged over the region bounded by 160°W–180° in the North Pacific Ocean. The months of each year's summer maximum [August (circles); September (crosses)] and winter minimum [January (diamonds); February (squares); March (plusses); April (stars)] are marked. Tic marks are at the beginning of the year.

shift was characterized by a stronger Aleutian Low and a southward shift of the westerly winds [Peterson and Schwing, 2003], as well as greater mixed layer depths [Polovina *et al.*, 1995], enhanced wintertime equatorward Ekman transport, and stronger turbulent mixing [Parrish *et al.*, 2000] in the transition zone of the North Pacific. A more southerly winter excursion of the TZCF, as seen in the surface temperature proxy, is consistent with these large-scale environmental changes, and also validates similar results from the coupled physical-biological model of Chai *et al.* [2003]. These results are also consistent with observations of increased post-1976 euphotic zone chlorophyll concentrations in the Climax region [26.5–31.0°N, 150.5–158.0°W; Venrick *et al.*, 1987] and Hawaii Ocean Time Series site [Karl *et al.*, 2001] in the subtropical North Pacific.

[14] During the period 1999–2002, following another proposed climate shift characterized by a northeastward displacement of the Aleutian Low [Peterson and Schwing, 2003; Bond *et al.*, 2003], the TZCF (and 18°C surface isotherm) was never observed south of 30°N in the central North Pacific (Figures 3 and 5). This implies a reduced level of primary production in the southern portion of the transition zone in recent years. The timing of the seasonal extrema in isotherm position has also varied over the 44 years, especially for winter southerly extent (Figure 5). Since 1996, the 18°C surface isotherm (and TZCF) has tended to reach its southern apex earlier than in previous periods.

4. Conclusions

[15] Interannual variability in the annual migration of the TZCF has been observed from high-resolution satellite ocean color data. Maximum southerly extent of the front in the central North Pacific varied by as much as 500 km from one year to the next. Enhanced surface convergence and vertical mixing in the transition zone during El Niño events draw the TZCF further south than usual, creating vast regions of anomalously high wintertime surface chlorophyll in the North Pacific Subtropical Gyre. A surface temperature proxy for TZCF position has revealed decadal-

scale variability in frontal position, with greater (lesser) annual range and southerly extent following the large-scale 1976–77 (1998–99) climate shifts. Interannual variations in TZCF position and strength could have important implications for the distribution and survival of a number of apex predators that utilize the TZCF as a migratory and foraging habitat. This variability could also impact the insular coral reef ecosystems of the northern portion of the Northwestern Hawaiian Islands, which the TZCF reaches in some years but not others.

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References

- Bond, N. A., J. E. Overland, M. Spillane, and P. Stabenro (2003), Recent shifts in the state of the North Pacific, *Geophys. Res. Lett.*, **30**(23), 2183, doi:10.1029/2003GL018597.
- Chai, F., M. Jiang, R. T. Barber, R. C. Dugdale, and Y. Chao (2003), Interdecadal variation of the transition zone chlorophyll front, a physical-biological model simulation between 1960 and 1990, *J. Oceanogr.*, **59**, 461–475.
- Eppey, R. W., E. H. Renger, E. L. Venrick, and M. M. Mullin (1973), A study of plankton dynamics and nutrient cycling in the central gyre of the North Pacific Ocean, *Limnol. Oceanogr.*, **18**, 534–551.
- Glover, D. M., J. S. Wroblewski, and C. R. McClain (1994), Dynamics of the transition zone in coastal zone color scanner-sensed ocean color in the North Pacific during oceanographic spring, *J. Geophys. Res.*, **99**, 7501–7511.
- Kamykowski, D. (1987), A preliminary biophysical model of the relationship between temperature and plant nutrients in the upper ocean, *Deep Sea Res., Part I*, **34**, 1067–1079.
- Karl, D. M., R. R. Bidigare, and R. M. Letelier (2001), Long-term changes in plankton community structure and productivity in the North Pacific Subtropical Gyre: The domain shift hypothesis, *Deep Sea Res., Part II*, **48**, 1449–1470.
- Kubota, M. (1994), A mechanism for the accumulation of floating marine debris north of Hawaii, *J. Phys. Oceanogr.*, **24**, 1059–1064.
- Laurs, R. M., P. C. Fiedler, and D. C. Montgomery (1984), Albacore tuna catch distributions relative to environmental features observed from satellite, *Deep Sea Res., Part I*, **31**, 1085–1099.
- Lewis, M. R., N. Kuring, and C. Yentsch (1988), Global patterns of ocean transparency: Implications for the new production of the open ocean, *J. Geophys. Res.*, **93**, 6847–6856.
- Miller, A. J., D. R. Cayan, T. P. Barnett, N. E. Graham, and J. M. Oberhuber (1994), The 1976–77 climate shift of the Pacific Ocean, *Oceanography*, **7**, 21–26.
- Parrish, R. H., F. B. Schwing, and R. Mendelssohn (2000), Mid-latitude wind stress: The energy source for climatic shifts in the North Pacific Ocean, *Fish. Oceanogr.*, **9**, 224–238.
- Peterson, W. T., and F. B. Schwing (2003), A new climate regime in northeast Pacific ecosystems, *Geophys. Res. Lett.*, **30**(17), 1896, doi:10.1029/2003GL017528.
- Polovina, J. J., G. T. Mitchum, and G. T. Evans (1995), Decadal and basin-scale variation in mixed layer depth and the impact on biological production in the Central and North Pacific, 1960–88, *Deep Sea Res., Part I*, **42**, 1701–1716.
- Polovina, J. J., D. R. Kobayashi, D. M. Parker, M. P. Seki, and G. H. Balazs (2000), Turtles on the edge: Movement of loggerhead turtles (*Caretta caretta*) along oceanic fronts spanning longline fishing grounds in the central North Pacific, 1997–1998, *Fish. Oceanogr.*, **9**, 1–13.
- Polovina, J. J., E. Howell, D. R. Kobayashi, and M. P. Seki (2001), The transition zone chlorophyll front, a dynamic global feature defining migration and forage habitat for marine resources, *Prog. Oceanogr.*, **49**, 469–483.
- Smith, S. D. (1988), Coefficients for sea surface wind stress, heat flux and wind profiles as a function of wind speed and temperature, *J. Geophys. Res.*, **93**, 15,467–15,472.
- Venrick, E. L., J. A. McGowan, D. R. Cayan, and T. L. Hayward (1987), Climate and chlorophyll *a*: Long-term trends in the central North Pacific Ocean, *Science*, **238**, 70–72.
- Williams, R. G., and M. J. Follows (1998), The Ekman transfer of nutrients and maintenance of new production over the North Atlantic, *Deep Sea Res., Part I*, **45**, 461–489.

Yoder, J. A., C. R. McClain, G. C. Feldman, and W. E. Esaias (1993). Annual cycles of phytoplankton chlorophyll concentrations in the global ocean: A satellite view, *Global Biogeochem. Cycles*, 7, 181–194.

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