Climatic forcing and phytoplankton phenology over the subarctic North Pacific from 1998 to 2006, as observed from ocean color data

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[1] We investigated phenological changes in phytoplankton in the subarctic North Pacific and the relationship to climatic forcing variability from 1998 to 2006, using ocean color satellite data combined with climatological data. The interannual to 9-year mean variability in the timing, magnitude, and duration of phytoplankton blooms were examined using SeaWiFS data. Based on 10-day composites of SeaWiFS data, the date of bloom initiation, peak chlorophyll *a* concentration, and bloom duration were estimated each year for every 2 × 2 degree grid within the range of 40–60°N and 130°E–120°W. The peak chlorophyll *a* concentration, date of bloom initiation, and bloom duration were derived by a Gaussian curve fitting technique. In addition, to investigate the geographic pattern of phytoplankton phenology, we classified the oceanic regions of the subarctic North Pacific into three groups based on K-means clustering: group A was distributed in the coastal regions of marginal seas, waters around the Kamchatka Peninsula, and along the Aleutian Islands; group B was mainly distributed offshore in western and central regions; and group C was in southeastern and central regions. The timing of spring blooms of group A was earlier in El Niño phase and later in La Niña phase, whereas the opposite pattern was seen in group B. Our study clearly revealed regional differences in phytoplankton seasonality, phenology patterns, and relationships between interannual variability of phytoplankton phenology and El Niño Southern Oscillation-scale variation. **Citation:** Sasaoka, K., S. Chiba, and T. Saino (2011), Climatic forcing and phytoplankton phenology over the subarctic North Pacific from 1998 to 2006, as observed from ocean color data, *Geophys. Res. Lett.*, *38*, L15609, doi:10.1029/2011GL048299.

1. Introduction

[2] Over the past few decades, many researchers have reported on climate variations in the North Pacific and discussed marine ecosystem responses to large-scale climatic forcing such as El Niño Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO), and Aleutian Low Pressure variations [e.g., *Beamish and Bouillon*, 1993; *Mantua et al.*, 1997; *Miller et al.*, 2004]. Climatic forcing with decadal variations has significant consequences on the oceanic ecosystem and has been recognized as an important factor affecting phytoplankton, zooplankton, and higher trophic levels of ecosystem dynamics [e.g., *Francis et al.*, 1998].

[3] Past investigations based on satellite remote sensing and ship measurements in the North Pacific reported the seasonality and interannual variability of chl-*a* concentrations and production in relation to changing oceanic conditions during ENSO events [e.g., *Goes et al.*, 2004; *Sasaoka et al.*, 2002]. These studies demonstrated that ocean color remote sensing is an essential tool to detect large-scale seasonality and interannual variability of phytoplankton biomass and productivity.

[4] Phytoplankton phenology (e.g., changes in initiation or peak of the spring bloom) is one of the critical factors regulating the seasonal cycle of marine ecosystems, because changes in phytoplankton phenology are likely to cause a "match-mismatch" between lower and higher trophic levels in the food web and affect fishery resources [e.g., *Edwards and Richardson*, 2004; *Koeller et al.*, 2009]. Based on ocean color data off the eastern continental shelf of Nova Scotia, Canada, *Platt et al.* [2003] concluded that the timing of the spring bloom is an important factor for interannual variations in fish stocks.

[5] The subarctic North Pacific represents one of the world's most biologically productive regions and strong air-sea interactions characterize the carbon cycle of this region. Investigating the phytoplankton phenology in the subarctic North Pacific is very important to understand the role of phytoplankton in linking climate change, zooplankton phenology, fish stocks and the capacity of biological carbon pump. Decadal phenological changes in the zooplankton community (e.g., changes in timing and peak of seasonal zooplankton production) in the North Pacific have been investigated using long-term historical data [e.g., *Chiba et al.*, 2008], but the role of phytoplankton phenology in linking climate variations and zooplankton phenology is still unclear, because long-term phytoplankton data were limited to only seasonal data and sampling coverage was inadequate for statistical analysis. In addition, *Chiba et al.* [2008] noted that phenological responses to PDO variation were different between local regions. Although several studies using ocean color data have investigated phenological characteristics of the annual to decadal cycles of phytoplankton biomass in the North Atlantic [e.g., *Siegel et al.*, 2002; *Platt and Sathyendranath*, 2008; *Henson et al.*, 2009; *Vargas et al.*, 2009], only a few remote-sensing-based studies on plankton phenology have been conducted in the subarctic North Pacific. Thus, we should examine the phytoplankton seasonality and phenology and clarify regional differences in the subarctic North Pacific using higher

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temporal and spatial resolution satellite-derived chl-*a* data. In this study, we investigated the interannual phenology of phytoplankton in the subarctic North Pacific from 1998 to 2006 using satellite ocean color data, and the relationships between phytoplankton phenology and climatic indices that the phytoplankton will respond to, such as the Southern Oscillation Index (SOI), PDO, North Pacific Index (NPI) and Aleutian Low Pressure Index (ALPI) etc. Here we discuss the regional differences in phytoplankton seasonality and phenological responses to climatic forcing by comparing two climate indices, SOI and PDO.

2. Materials and Methods

[6] The temporal and spatial variability of chl-*a* in the subarctic North Pacific was investigated using SeaWiFS Level-3 data for January 1998 to December 2006 obtained from NASA GSFC's Ocean Color website (http://oceancolor. gsfc.nasa.gov/). According to *Gregg and Casey* [2004], the accuracy of SeaWiFS data was in relatively good agreement with *in situ* data in the North Pacific. To detect the phytoplankton phenology, the timing of the annual phytoplankton spring bloom was estimated by a Gaussian curve fit [*Yamada and Ishizaka*, 2006] of satellite chl-*a* variation from February to August for each year and each grid. To apply a Gaussian curve fit, original SeaWiFS data with 9-km spatial and daily temporal resolution were regridded to 2×2 degree and 10-day resolution within the range of 40–60°N and 140°E–120°W, because we needed to solve the no data problem due to persistent cloud cover. The Gaussian function of chl-*a* concentration over time (t) was expressed by the following equation:

$$\text{Chl-}a(t) = Y_0 + \left\{ Y_\text{p} \times \exp\left[-0.5(t - t_\text{p})/\sigma^2\right] \right\}, \qquad (1)$$

where Y_0 is the baseline chl-*a* concentration, Y_p is the peak chl-*a* concentration, t_p is the peak timing, and σ is the standard deviation of the Gaussian curve. $t_\text{p} - 2\sigma$ and $t_\text{p} + 2\sigma$ were defined as initiation (bloom timing) and termination of a spring bloom. Consequently, bloom duration was defined as 4σ.

[7] The K-means clustering method was used to group the grids to investigate the geographic pattern of phytoplankton phenology over the subarctic North Pacific based on the 9-year mean of the four parameters, bloom timing, peak chl-*a* concentration, baseline chl-*a* concentration, and bloom duration. Each parameter was normalized as mean = 0 and standard deviation = 1 prior to analysis.

[8] After cluster analysis, we examined the interannual variation of timing and duration of phytoplankton blooms of each group and the relationship to ENSO variation, because most dramatic changes in meteorological and oceanographic conditions over the entire North Pacific are usually triggered by ENSO episodes in the tropics [e.g., *Emery and Hamilton*, 1985]. The PDO is a first mode empirical orthogonal function based on the Sea Surface Temperature (SST) anomaly [*Mantua et al.*, 1997]. The PDO is also related to the ENSO [e.g., *Newman et al.*, 2003], and the spatial SST anomaly pattern over the North Pacific derived by SOI corresponds well to that of the PDO [*Mantua et al.*, 1997]. Therefore, to understand the relationships between changes in phytoplankton phenology and ENSO variation, we compared the two climate indices and SST anomaly from

satellite data. SOI and PDO data were obtained from the following websites: http://www.cgd.ucar.edu/cas/catalog/climind/and http://jisao.washington.edu/pdo/. SST anomalies were calculated by monthly averaged, 4-km resolution, AVHRR Pathfinder Global SST data obtained from NASA JPL's PODAAC website (http://podaac.jpl.nasa.gov/DATA_CATALOG/avhrrinfo.html).

3. Results and Discussion

3.1. Mean Phytoplankton Seasonality During 1998 to 2006

[9] Figure 1 illustrates the regional characteristics of the 9-year mean phytoplankton seasonality calculated using annual values from 1998 to 2006. High peak chl-*a* concentrations during spring bloom (>2 mg m^{-3}; indicated by colors ranging from green to red) were found in the Okhotsk Sea and Bering Sea, waters adjacent to Japan, and the coastal region of Gulf of Alaska, whereas low peak chl-*a* concentrations (<2 mg m^{-3}; indicated by blue) were seen in the majority of offshore regions (Figure 1a). The timing of spring blooms in the eastern North Pacific, particularly offshore southeastern regions indicated by red to orange, was earlier (starting in March) than that in western regions (Figure 1b). Bloom duration in the central and eastern parts of the study region was longer (shown by orange to red) than that in the western offshore region and marginal seas (Figure 1c). The regions with longer blooms mostly corresponded to low peak chl-*a* regions (blue in Figure 1a). In contrast, the bloom duration tended to be shorter in the marginal seas as compared with other regions (Figure 1c).

[10] To investigate the geographic pattern of phytoplankton phenology, we classified oceanic regions in the North Pacific into five groups according to K-means clustering based on the 9-year means of four parameters (peak chl-*a*, baseline chl-*a*, peak timing, and standard deviation) derived from the Gaussian curve for each grid. We identified three major groups in our study region: group A (blue in Figure 2a) was seen in the coastal regions of marginal seas and the Gulf of Alaska, waters around the Kamchatka Peninsula, and along the Aleutian Islands; group B (green in Figure 2a) was mainly seen in western and eastern offshore regions and at the center of the marginal seas; and group C (red in Figure 2a) was mainly in the southeastern and central regions of the North Pacific. Groups D and E were omitted from further analysis, because they included only six and eight grids, respectively (Figure 2a).

[11] The 9-year mean phytoplankton seasonality of each group is shown in Figures 2b–2d. In group A (Figure 2b), the baseline and peak chl-*a* concentrations were highest (peak chl-*a* > 10 mg m^{-3}) and bloom duration was shortest among the three groups, and the bloom timing was earlier (late April to late May) than that of group B. In group B (Figure 2c), the baseline and peak chl-*a* concentrations were low, bloom timing was the latest (early April to middle June) among the three groups, and the bloom duration was longer than that of group A. In group C (Figure 2d), the baseline and peak chl-*a* concentrations were extremely low and it is no longer bloom. The peak timing was earliest (late January to early May) and the peak duration was longest among the three groups.

[12] These regional characteristics of phytoplankton seasonality are consistent with past remote-sensing studies
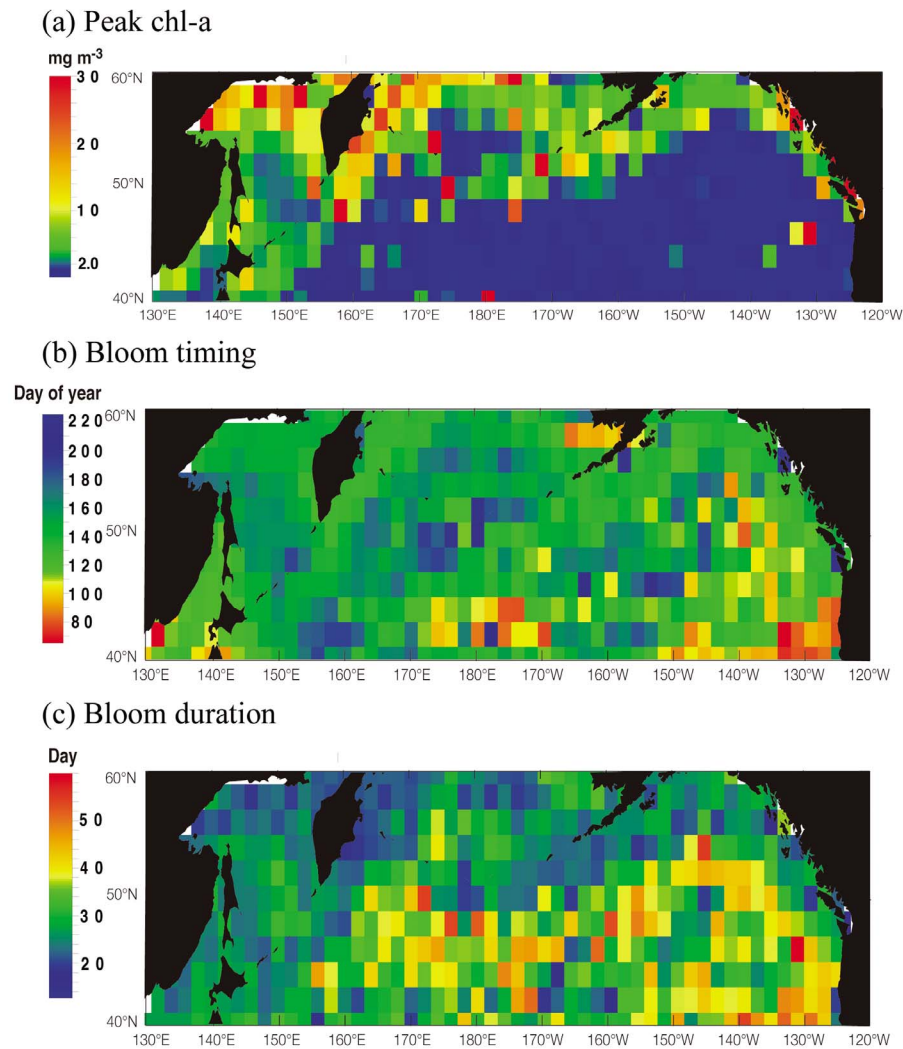
(a) Peak chl-a

(b) Bloom timing

(c) Bloom duration

Figure 1. Regional characteristics of the 9-year mean (a) peak chl-*a* concentrations (mg m^{-3}), (b) bloom timing (day of year), and (c) bloom duration (day), calculated using annual values for 1998 to 2006 derived from Gaussian curve fit.

conducted in the subarctic North Pacific. *Sasaoka et al.* [2002] reported that chl-*a* concentrations were consistently high (>10 mg m^{-3}) along the Kuril Islands and in the coastal waters around the Kamchatka Peninsula. High peak chl-*a* (>2 mg m^{-3}) during April and May was observed in the Oyashio region at the western edge of the subarctic North Pacific around Japan [*Goes et al.*, 2004]. *Ueno et al.* [2010] reported that high chl-*a* concentrations in spring extended south of the Aleutian Islands and the Alaskan Peninsula around Alaskan Stream eddies. Our finding of low peak chl-*a* concentrations in offshore regions was also in good agreement with satellite observations [*Sasaoka et al.*, 2002; *Goes et al.*, 2004; *Yoo et al.*, 2008]. These low chl-*a* regions correspond to the High-Nutrient, Low-Chlorophyll (HNLC) region reported by *Banse and English* [1999]. Within this region, *Goes et al.* [2004] noted that chl-*a* concentrations in spring were greater in the Western Subarctic Gyre than in the Alaskan Gyre, and chl-*a* concentrations in the central offshore region were consistently low (<1 mg m^{-3}). Our results show good agreement with these findings: The peak chl-*a* concentration of group B, corresponding to the Western Subarctic Gyre, was higher than that of group C,

corresponding to the Alaskan Gyre and central offshore region.

3.2. Relationships Between Phytoplankton Phenology and Climatic Forcing

[13] Here we describe the interannual variability of bloom timing and duration of groups A-C and the relationships of these parameters with two climate indices (SOI and PDO) and SST anomalies. *Chiba et al.* [2008] discussed that the wintertime hydrographic condition is the crucial factor controlling the timing of water column stratification associated with light availability of phytoplankton during spring season. Thus we compared bloom timing with winter-mean (December to March) SOI and SST anomalies. Figures 3a and 3b illustrate the interannual variability of bloom timing for group A versus winter-mean (December through March) SOI and annual PDO. Delayed (advanced) timing of the spring bloom in group A coincided with positive (negative) SOI ($r = 0.73$, $p < 0.05$; Figure 3a), whereas later (earlier) bloom timing coincided with negative (positive) PDO ($r = -0.80$, $p < 0.05$; Figure 3b). The interannual variability of winter-mean SST anomalies was similar pat-
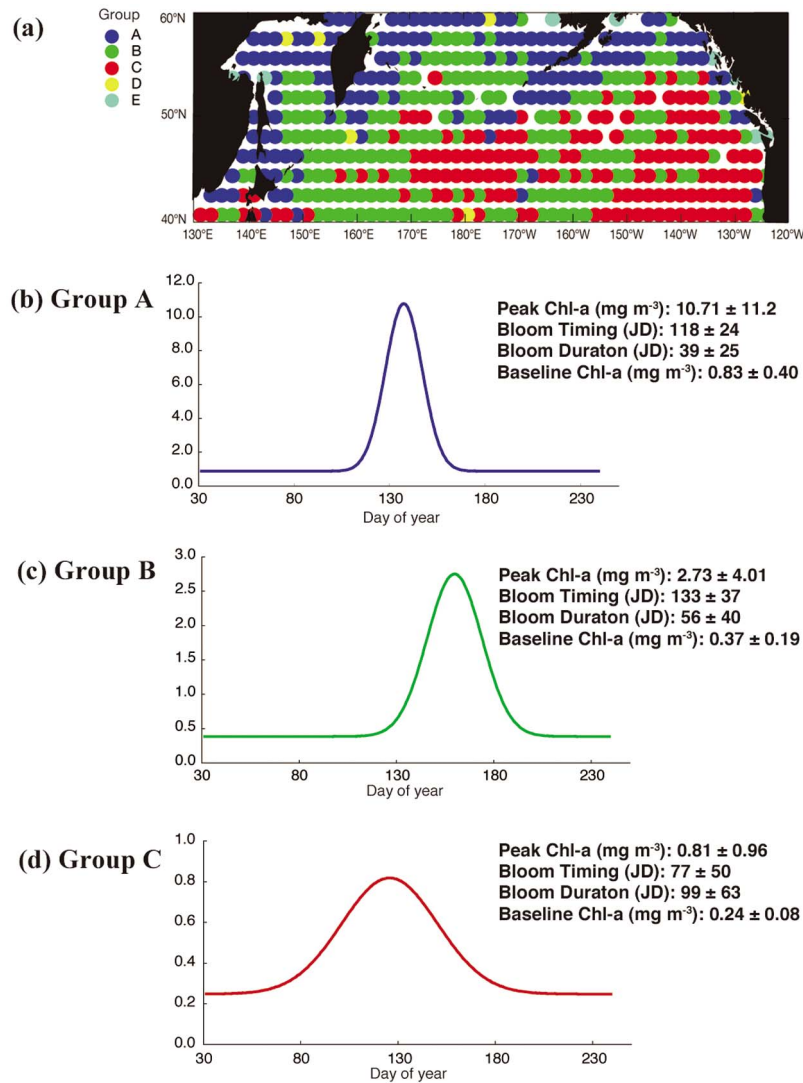
Figure 2. (a) Map of the five groups detected by K-means cluster analysis. The 9-year mean phytoplankton seasonality of (b) group A, (c) group B, and (d) group C from 1998 to 2006, with mean ± standard deviation of peak chl-*a*, bloom timing, bloom duration, and baseline chl-*a*.

tern among three groups; cooler (warmer) SST coincided with positive (negative) SOI and negative (positive) PDO (Figure 3e). This means that the spring bloom timing in group A was earlier in El Niño phase (Warmer SST) and later in La Niña phase (Cooler SST). Although interannual variability of SST anomalies of group B showed the similar pattern to that of group A (Figure 3e), the bloom timing of group B was corresponding to the opposite pattern to that of group A. In group B, later timing of the spring bloom coincided with negative SOI in the winter of the following year (SOI_{win} + 1 year) and earlier bloom timing coincided with positive SOI_{win} + 1 year ($r = -0.88, p < 0.05$; Figure 3c). Later (earlier) bloom timing of group B was corresponding to a warmer (cooler) SST from 2000 to 2006 (Figures 3c and 3e). In group C, we found only a weak correlation between duration of chl-*a* peak and SOI. Shorter peak duration of group C coincided with positive SOI and longer peak duration with negative SOI ($r = -0.61, p < 0.1$; Figure 3d). This means the peak duration of group C tends to be longer

in El Niño phase (Warmer SST) and shorter in La Niña phase (Cooler SST) (Figures 3d and 3e).

[14] Our findings indicate that significant correlations exist between interannual variability of bloom timing of groups A and B in the subarctic North Pacific and SOI and PDO, which are associated with ENSO-scale variation. Based on seasonal shipboard observation data, *Chiba et al.* [2008] suggested the importance of water column stratification for the timing of phytoplankton production in the Oyashio region, because water column stabilization increases light availability. They reported that cooler (warmer) conditions during winter induced weak (strong) stratification of the water column during spring, and consequently the timing of phytoplankton production was later (earlier). This previous study supports our satellite-based findings that later (earlier) bloom timing of group A corresponds to a cooler (warmer) phase. A part of the Okhotsk Sea and Bering Sea shelf were recognized as group A. In those regions, past works suggested that sea ice variability is significant factor for spring bloom [e.g., *Saitoh et al.*, 2002;
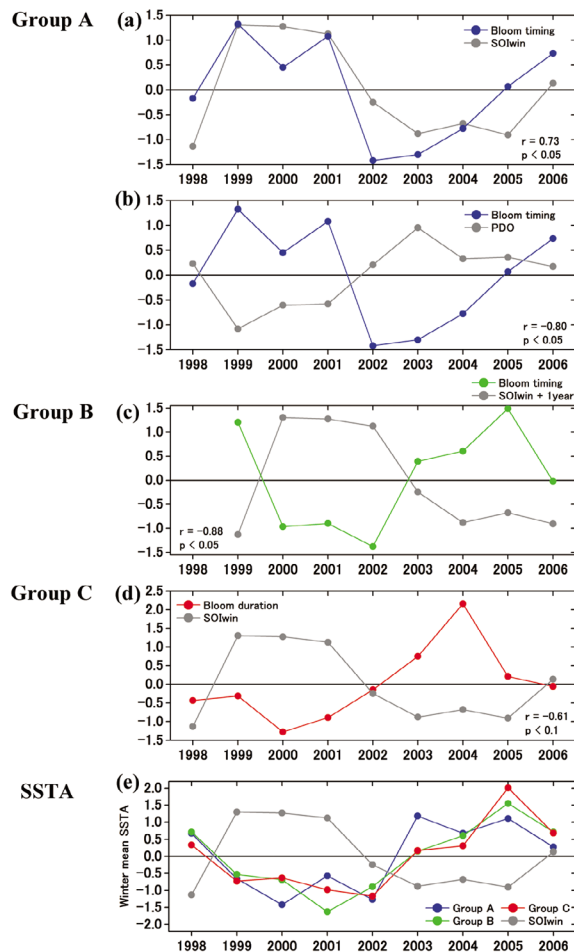
Figure 3. Interannual variability of group A bloom timing and (a) winter-mean (December through March) SOI and (b) annual PDO; (c) interannual variability of group B bloom timing and winter-mean SOI + 1 year; (d) interannual variability of group C bloom duration and winter-mean SOI; and (e) interannual variability of winter-mean (December through March) SST anomalies calculated by monthly averaged AVHRR data. Pearson correlation coefficient (*r*) and significance level (*p*) are shown in each graph. All plotted values of phenology parameters (bloom timing and duration) were normalized (mean = 0, standard deviation = 1). Climate indices (SOI and PDO) represent anomalous values during 1998–2006.

Jin et al., 2007]. *Saitoh et al.* [2002] reported that phytoplankton biomass in the Bering Sea depends on the timing of sea ice melting and tends to increase when the melting is delayed. They also suggested that wind stress is one of the important factors controlling the timing of the spring bloom. Although interannual variability of base line and peak chl-*a* concentrations of group A had no significant correlations with SOI and PDO, sea ice variability may impact the phytoplankton phenology of group A.

[15] The bloom timing of group B was corresponding to the opposite pattern to that of group A. Later (earlier) bloom timing of group B was corresponding to a warmer (cooler) SST. Group-A region is mostly continental shelf area where

might be dominated by iron-rich water (Figure 2a). On the other hand, group B regions were mainly located in offshore regions of the central to western North Pacific (Figure 2a) and well known as a typical HNLC region where primary production is limited by availability of iron [e.g., *Boyd et al.*, 2004]. Recent works have shown that a subsurface supply of iron-rich water between the continental margin and open ocean is important to productivity, particularly in the wintertime when dust flux is low and mixed layers are deep [e.g., *Whitney et al.*, 2005; *Lam and Bishop*, 2008]. It is noted that a part of group B is extending from iron-rich area to HNLC region (i.e., the Okhotsk Sea basin to the western subarctic gyre, the Bering Sea basin and Alaskan Stream to Gulf of Alaska), which is similar to the pathway of sub-surface iron-rich water transport. It is likely that earlier bloom timing of group B was caused by enhanced iron supply to surface layer or euphotic zone in group B region due to strong vertical mixing during cooler winter and horizontal advection from iron-rich area. Thus, our geo-graphical pattern of group A and B may reflect the differences of nutrient conditions.

[16] The duration of chl-*a* peak of group C was longer during El Niño phase (Warmer SST) and shorter during La Niña phase (Cooler SST). Group C was mainly located in not only California coastal area, but also the middle of North Pacific, eastern side of study area (Figure 2a). We assume that group C region is low-nutrient area; hence, cooler phase could induce shorter bloom due to weak stratification (i.e., relative strong mixing) and efficient nutrient supply or uptake. However, we couldn't understand the mechanisms controlling duration change of group C related to ENSO-scale variations. *Gregg and Casey* [2004] have reported the SeaWiFS algorithm having 31% errors and group C area is characterized as extremely low chl-*a* concentrations. Thus it might be difficult to evaluate chl-a variation in group C region and we have to require further investigation.

4. Concluding Remarks

[17] Phytoplankton phenology was investigated by using satellite ocean color data and we improved our under-standing of phytoplankton phenology and its regional to basin-scale differences in the subarctic North Pacific. Our study clearly revealed the timing, duration, magnitude of spring bloom, regional differences in the phenology pattern, and significant correlations between interannual variability of phytoplankton phenology and ENSO-scale variation. However, some of the mechanisms controlling phenological change remain unclear, such as why the bloom timing of group B lagged an El Niño or La Niña change by 1 year or what ENSO-scale variations control the variability of peak duration of group C. Further analysis will be needed to compare more precisely the region's phytoplankton phe-nology with ENSO-scale variation of other environmental factors (e.g., impact of atmospheric forcing, horizontal advection and sea ice variability). Additionally, ocean color observation in the North Pacific is subject to much coverage loss due to persistent cloud cover, particularly spring to summer season. Our analysis data in the period February through August observed about 3–15 times per month at each pixel. We must consider that this problem may impact our results.

References

Banse, K., and D. C. English (1999), Comparing phytoplankton seasonality in the eastern and western subarctic Pacific and the western Bering Sea, *Prog. Oceanogr.*, *43*, 235–287, doi:10.1016/S0079-6611(99)00010-5.

Beamish, R. J., and D. R. Bouillon (1993), Pacific salmon production trends in relation to climate, *Can. J. Fish. Aquat. Sci.*, *50*, 1002–1016, doi:10.1139/f93-116.

Boyd, P. W., et al. (2004), The decline and fate of an iron-induced subarctic phytoplankton bloom, *Nature*, *428*(6982), 549–553, doi:10.1038/nature02437.

Chiba, S., M. N. Aita, K. Tadokoro, T. Saino, H. Sugisaki, and K. Nakata (2008), From climate regime shifts to lower-trophic level phenology: Synthesis of recent progress in retrospective studies of the western North Pacific, *Prog. Oceanogr.*, *77*, 112–126, doi:10.1016/j.pocean.2008.03.004.

Edwards, M., and A. J. Richardson (2004), Impact of climate change on marine pelagic phenology and trophic mismatch, *Nature*, *430*, 881–884, doi:10.1038/nature02808.

Emery, W. J., and K. Hamilton (1985), Atmospheric forcing of interannual variability in the northeast Pacific Ocean: Connections with El Niño, *J. Geophys. Res.*, *90*, 857–868, doi:10.1029/JC090iC01p00857.

Francis, R. C., S. R. Hare, A. B. Hollowed, and W. S. Wooster (1998), Effects of interdecadal climate variability on the oceanic ecosystems of the NE Pacific, *Fish. Oceanogr.*, *7*, 1–21, doi:10.1046/j.1365-2419.1998.00052.x.

Goes, J. I., K. Sasaoka, H. R. Gomes, S. Saitoh, and T. Saino (2004), A comparison of the seasonality and interannual variability of phytoplankton biomass and production in the western and eastern gyres of the subarctic Pacific using multi-sensor satellite data, *J. Oceanogr.*, *60*, 75–91, doi:10.1023/B:JOCE.0000038320.94273.25.

Gregg, W. W., and N. W. Casey (2004), Global and regional evaluation of the SeaWiFS chlorophyll data set, *Remote Sens. Environ.*, *93*, 463–479, doi:10.1016/j.rse.2003.12.012.

Henson, S. A., J. P. Dunne, and J. L. Sarmiento (2009), Decadal variability in North Atlantic phytoplankton blooms, *J. Geophys. Res.*, *114*, C04013, doi:10.1029/2008JC005139.

Jin, M., C. Deal, J. Wang, V. Alexander, R. Gradinger, S. Saitoh, T. Iida, Z. Wan, and P. Stabeno (2007), Ice-associated phytoplankton blooms in the southeastern Bering Sea, *Geophys. Res. Lett.*, *34*, L06612, doi:10.1029/2006GL028849.

Koeller, P., et al. (2009), Basin-scale coherence in phenology of shrimps and phytoplankton in the North Atlantic Ocean, *Science*, *324*, 791–793, doi:10.1126/science.1170987.

Lam, P. J., and J. K. B. Bishop (2008), The continental margin is a key source of iron to the HNLC North Pacific Ocean, *Geophys. Res. Lett.*, *35*, L07608, doi:10.1029/2008GL033294.

Mantua, N. J., S. R. Hare, Y. Zhang, J. M. Wallace, and R. C. Francis (1997), A Pacific interdecadal climate oscillation with impacts on salmon production, *Bull. Am. Meteorol. Soc.*, *78*, 1069–1079, doi:10.1175/1520-0477(1997)078<1069:APICOW>2.0.CO;2.

Miller, A. J., F. Chai, S. Chiba, J. R. Moisan, and D. J. Neilson (2004), Decadal-scale climate and ecosystem interactions in the North Pacific Ocean, *J. Oceanogr.*, *60*, 163–188, doi:10.1023/B:JOCE.0000038325.36306.95.

Newman, M., G. P. Compo, and M. A. Alexander (2003), ENSO-forced variability of the Pacific Decadal Oscillation, *J. Clim.*, *16*, 3853–3857, doi:10.1175/1520-0442(2003)016<3853:EVOTPD>2.0.CO;2.

Platt, T., and S. Sathyendranath (2008), Ecological indicators for the pelagic zone of the ocean from remote sensing, *Remote Sens. Environ.*, *112*, 3426–3436, doi:10.1016/j.rse.2007.10.016.

Platt, T., C. Fuentes-Yaco, and K. T. Frank (2003), Spring algal bloom and larval fish survival, *Nature*, *423*, 398–399, doi:10.1038/423398b.

Saitoh, S., T. Iida, and K. Sasaoka (2002), A description of temporal and spatial variability in the Bering Sea spring phytoplankton blooms (1997–1999) using satellite multi-sensor remote sensing, *Prog. Oceanogr.*, *55*, 131–146, doi:10.1016/S0079-6611(02)00074-5.

Sasaoka, K., S. Saitoh, I. Asanuma, K. Imai, M. Honda, Y. Nojiri, and T. Saino (2002), Temporal and spatial variability of chlorophyll a in the western subarctic Pacific determined from satellite and ship observations from 1997 to 1999, *Deep Sea Res., Part II*, *49*, 5557–5576, doi:10.1016/S0967-0645(02)00206-0.

Siegel, D. A., S. C. Doney, and J. A. Yoder (2002), The North Atlantic spring phytoplankton bloom and Sverdrup's critical depth hypothesis, *Science*, *296*, 730–733, doi:10.1126/science.1069174.

Ueno, H., W. R. Crawford, and H. Onishi (2010), Impact of Alaskan stream eddies on chlorophyll distribution, *J. Oceanogr.*, *39*, 934–951.

Vargas, M., C. W. Brown, and M. R. P. Sapiano (2009), Phenology of marine phytoplankton from satellite ocean color measurements, *Geophys. Res. Lett.*, *36*, L01608, doi:10.1029/2008GL036006.

Whitney, F. A., W. R. Crawforda, and P. J. Harrison (2005), Physical processes that enhance nutrient transport and primary productivity in the coastal and open ocean of the subarctic NE Pacific, *Deep Sea Res., Part II*, *52*, 681–706, doi:10.1016/j.dsr2.2004.12.023.

Yamada, K., and J. Ishizaka (2006), Estimation of interdecadal change of spring bloom timing, in the case of the Japan Sea, *Geophys. Res. Lett.*, *33*, L02608, doi:10.1029/2005GL024792.

Yoo, S., H. P. Batchelder, W. T. Peterson, and W. J. Sydeman (2008), Seasonal, interannual and event scale variation in North Pacific ecosystems, *Prog. Oceanogr.*, *77*, 155–181, doi:10.1016/j.pocean.2008.03.013.

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