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¹ **Western Arctic Primary Productivity Regulated By**
² **Shelf-break Warm Eddies**

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Abstract

The response of phytoplankton to the Beaufort shelf-break eddies in the western Arctic Ocean is examined using the eddy-resolving coupled sea ice-ocean model including a lower-trophic marine ecosystem formulation. The regional model driven by the reanalysis 2003 atmospheric forcing from March to November captures the major spatial and temporal features of phytoplankton bloom following summertime sea ice retreat in the shallow Chukchi shelf and Barrow Canyon. The shelf-break warm eddies spawned north of the Barrow Canyon initially transport the Chukchi shelf water with high primary productivity toward the Canada Basin interior. In the eddy-developing period, the anti-cyclonic rotational flow along the outer edge of each eddy moving offshore occasionally traps the shelf water. The primary production inside the warm eddies is maintained by internal dynamics in the eddy-maturity period. In particular, the surface central area of an anti-cyclonic eddy acquires adequate light, nutrient, and warm environment for photosynthetic activity partly attributing to turbulent mixing with underlying nutrient-rich water. The simulated biogeochemical properties with the dominance of small-size phytoplankton inside the warm eddies are consistent with the observational findings in the western Arctic Ocean. It is also suggested that the light limitation before autumn sea ice freezing shuts down the primary production in the shelf-break eddies in spite of nutrient recovery. These results indicate that the time lag between the phytoplankton bloom in the shelf region following the summertime sea ice retreat and the eddy generation along

26 the Beaufort shelf break is an important index to determine biological regimes
27 in the Canada Basin.

28 Keyword :

29 phytoplankton bloom

30 eddy dynamics

31 shelf-basin exchange

32 Pacific-origin water

33 a lower-trophic marine ecosystem model

1. Introduction

34 The influence of abrupt sea ice retreat on the Arctic marine ecosystem is a crucial topic
35 in academic, commercial, and social communities [*Arrigo et al.*, 2008; *Pabi et al.*, 2008].
36 Recent satellite analysis indicated that high primary production region shifts northward
37 from the Bering Sea to the Chukchi Sea following sea ice reduction [*Grebmeier et al.*,
38 2010]. While the systematic field campaign conducted in the Western Arctic Shelf-Basin
39 Interaction (SBI) project has produced major important findings in the 2000s [*Grebmeier*
40 *and Harvey*, 2005; *Grebmeier et al.*, 2009], an impact of ocean dynamics on polar bi-
41 ological cycles still has a lot of uncertainties. The Pacific-origin water inflowing from
42 the Bering Strait is a predominant source of not only heat and fresh water, but also
43 nitrate, silicate, dissolved- and particulate organic materials in the Arctic Ocean. The
44 nutrient-rich Pacific water undergoes physical and biogeochemical modifications through
45 seasonal primary production and interactions with organic-rich bottom sediments in the
46 Chukchi shelf [*Weingartner et al.*, 2005]. The buoyant Pacific summer water flows along
47 the Alaskan northwestern coast as a surface-intensified current, and its significant part is
48 eventually transported into the Canada Basin by shelf-break eddies during late summer
49 and early autumn [*Pickart*, 2004; *Watanabe and Hasumi*, 2009].

50 In the northern Gulf of Alaska, anti-cyclonic eddies propagating along a continental
51 slope were frequently detected by satellite remote sensing [*Okkonen et al.*, 2003]. The
52 eddy-like features of high chlorophyll-*a* (CHL) concentration implied that mesoscale eddies
53 were carrying phytoplankton and nutrient from the outer shelf toward the deep basin. In
54 *Ueno et al.* [2010], the role of the Alaskan Stream eddies in the CHL distribution in the

55 central subarctic North Pacific was categorized into three types: 1) the lateral transport of
56 shelf-origin nutrient and biological organism via advection of eddy body and/or rotational
57 circulation around of each eddy, 2) nutricline shallowing accompanied by eddy decay, and
58 3) Ekman upwelling induced by eddy-wind interaction. The latter two processes work
59 on upward nutrient flux. Thus biological activities could be closely related to mesoscale
60 eddies. In the western Arctic, eddy contribution to phytoplankton activities has not been
61 fully evaluated yet. It is reported that the surface and halocline layers above 300 m depth
62 in the Canada Basin are full of a number of anti-cyclonic eddies [*Manley and Hunkins,*
63 1985]. A major part of anti-cyclonic cold-core eddies are considered to be generated
64 from a bottom-intensified current along the northern edge of Chukchi shelf via baroclinic
65 instability in early spring [*Spall et al., 2008*]. *Mathis et al.* [2007] and *Kadko et al.* [2008]
66 suggested that a cold eddy observed on the Chukchi Sea continental slope played a crucial
67 role in the transport of carbon, oxygen, and nutrient associated with the Pacific winter
68 water. However, the cold-eddy intrusion into the halocline layer of Canada Basin hardly
69 enforces primary production in the surface euphotic zone at once.

70 The surface-intensified warm-core eddies with the signal of the Pacific summer water
71 are occasionally detected in the vicinity of the Barrow Canyon from later summer to
72 early autumn [*Pickart, 2004*]. The detailed properties of a visible warm eddy observed by
73 the R/V *Mirai* in 2010 were reported in *Nishino et al.* [2011a]. The buoyant warm-core
74 eddy would efficiently supply the surface layer of basin interior with a great amount of
75 ammonium produced by the decomposition of organic matters on the bottom of Chukchi
76 Sea shelf during summer, although the annual nutrient transport of warm eddies might
77 be much smaller than the contribution of cold eddies [*Nishino et al., 2005*]. The eddy

78 transport of ammonium-rich shelf water could sustain locally enhanced growth of smaller-
79 size phytoplankton in the southern Canada Basin. However, it is still unclear whether
80 the eddy-like structures of biological signal attributed to the lateral advection of high
81 phytoplankton biomass from the Chukchi and Beaufort shelf region or were formed by
82 local biological primary production inside the eddies.

83 In recent years, numerical studies on the Arctic marine biology become activated. The
84 impact of Arctic sea ice decline on the marine plankton ecosystem was also addressed
85 using both global simulation products [*Popova et al.*, 2010; *Jin et al.*, 2011] and a pan-
86 Arctic regional modeling [*Zhang et al.*, 2010]. According to their analyses, the primary
87 productivity of phytoplankton over seasonal ice areas of the Arctic Ocean is enhanced
88 primarily by the increases in photosynthetically active radiation and nutrient availability
89 in the euphotic zone, and partially due to surface water warming. The strong mixing and
90 upwelling associated with summertime sea ice reduction are dynamic drivers for nutrient
91 retrieval. The spatial pattern of ice algae and the size distribution of pelagic phytoplank-
92 ton groups were also reported [*Jin et al.*, 2011]. These models successfully reproduced the
93 Arctic basin-scale features of CHL and primary production. On the other hand, several
94 kinds of complexity on the shelf process and shelf-basin exchange for the analysis have
95 commonly emerged owing to insufficient model resolution to explicitly resolve topographic
96 ocean current, waves, and mesoscale eddies.

97 *Watanabe and Hasumi* [2009] and *Watanabe* [2011] examined 1) mechanisms controlling
98 the hydrographic properties of the Beaufort shelf-break warm eddies and 2) the relation-
99 ship of Pacific water transport with summertime sea ice extent and shelf-wide wind fields
100 using an eddy-resolving coupled sea ice-ocean model covering the western Arctic Ocean.

101 These previous studies clarified that early sea ice retreat toward the Canada Basin and
102 westerly surface wind over the Chukchi shelf promoted the shelf-to-basin transport of the
103 Pacific summer water via enhanced eddy activities.

104 In this paper, the phytoplankton behaviors regulated by the shelf-break warm eddies
105 are addressed using the high-resolution model newly coupled with a lower-trophic marine
106 ecosystem formulation. We performed the model integration from March to November
107 in the same framework as *Watanabe* [2011] and investigated the contribution of local dy-
108 namics, such as shelf-water transportation, turbulent mixing, and upwelling, to primary
109 productivity following eddy life cycle in the Beaufort shelf-break region. The method of
110 modeling analyses is described in section 2. The seasonal transitions in physical oceanographic
111 fields and biological performance in the western Arctic Ocean are traced in section
112 3, and the eddy-related phytoplankton activities are then focused on in section 4. The
113 findings obtained in the present work are summarized in section 5.

2. Model and Experimental Design

114 The coupled sea ice-ocean model used in the present work is composed of a physical
115 ocean general circulation model named Center for Climate System Research Ocean Com-
116 ponent Model (COCO) version 3.4 [Hasumi, 2006] and a lower-trophic marine ecosystem
117 part named North Pacific Ecosystem Model for Understanding Regional Oceanography
118 (NEMURO) [Kishi *et al.*, 2007]. The model domain contains the entire Chukchi Sea and
119 the southern area of the Canada Basin (Figure 1a). The horizontal resolution is about 2.5
120 km, and there are 25 vertical levels. The physical oceanographic part of model and exper-
121 imental design is the same as the 2003 case in *Watanabe* [2011] except the Bering Strait
122 condition. In the previous experiments (i.e., *Watanabe and Hasumi* [2009]; *Watanabe*
123 [2011]), the temperature of Bering Strait throughflow was prescribed to an idealized sea-
124 sonal cycle, and the coldest water at the freezing point inflowed into the southern Chukchi
125 shelf even after sea ice retreat in early summer. To prevent this inconsistent situation,
126 the freezing period of inflow is shortened until May. The atmospheric forcing components
127 are obtained from the National Centers for Environmental Prediction / National Center
128 for Atmospheric Research reanalysis daily dataset in 2003 [Kalnay *et al.*, 1996].

129 The detailed configuration of NEMURO is described in *Kishi et al.* [2007]. The model
130 treats the concentration of 11 biogeochemical variables illustrated in Figure 1b. While
131 the NEMURO has been originally developed for the assessment of marine biology in
132 the North Pacific, its reasonable performance was recently confirmed even in the global
133 domain [Sumata *et al.*, 2010; Kishi *et al.*, 2011] and in the Arctic basin-scale primary
134 productivity [Zhang *et al.*, 2010]. In our experiments, the parameter values such as the
135 photosynthetic rate of phytoplankton and the grazing rate of zooplankton basically follow

136 *Zhang et al.* [2010]. Their modification of model structure from *Kishi et al.* [2007] (i.e., ZS
137 grazing on PL in Figure 1b) is not incorporated, because we do not still have established
138 findings on zooplankton behaviors in the Arctic Ocean. Instead, a sensitivity experiment
139 on zooplankton impact is conducted as an extreme case (see section 5). The sea ice
140 ecosystem with ice algae, nutrient exchange via ice-ocean interface, and light penetration
141 through ice column are neglected for simplicity.

142 As demonstrated in *Watanabe and Hasumi* [2009], the Pacific water transport across
143 the Chukchi and Beaufort shelf breaks reaches a maximum from late summer to early
144 autumn and becomes a minimum in mid-winter. In addition, we focus on summertime
145 biological activities, especially the impact of shelf-break warm eddies on phytoplankton
146 dynamics. The model is hence integrated for 9 months from March to November. The
147 experiment has no spin-up stage to minimize the disturbance of hydrographic structure in
148 the Canada Basin due to the closed lateral boundary of model domain. This integration
149 period is regarded to cover the annual bloom event. The model is initiated from tem-
150 perature and salinity fields in the Polar Science Center Hydrographic Climatology (PHC)
151 March data [*Steele et al.*, 2001], arbitrary sea ice thickness constructed by multiplying
152 a factor of 1 m to the climatological mean sea ice concentration [*National Ice Center*,
153 2006], no circulation of both sea ice and ocean following *Watanabe* [2011]. The monthly
154 climatology data of nitrate and silicate concentration derived from the World Ocean Atlas
155 2009 (WOA09) [*Garcia et al.*, 2010] are utilized for restoring along the lateral boundary
156 region of model domain, and its March values are given to the initial fields. Nutrient sup-
157 ply from river water discharge is not taken into account, because its influence on primary
158 productivity would be localized in the vicinity of each river mouth [*Wang et al.*, 2005].

159 To indirectly account for ammonium dissolution from sea bottom sediments, ammonium
160 concentration in the deepest layer on the entire shelf where the water depth is shallower
161 than 200 m is restored to seasonally varying amount with the annual average of $2 \mu\text{M}$
162 ($1 \mu\text{M} \equiv 1 \text{ mmolN m}^{-3}$). The idealized seasonal cycle with the maximum value of $4 \mu\text{M}$
163 in August and the minimum value of $0 \mu\text{M}$ in February is prescribed based on previous
164 observations [*Nishino et al.*, 2005]. Initial phytoplankton and zooplankton biomass is uni-
165 formly distributed in the upper 200 m of the entire domain. The initial concentration of
166 $0.02 \mu\text{M}$ is approximately two-order smaller than the simulated phytoplankton biomass
167 during the blooming period (see section 3). The initial value of other components is set
168 to quite low value. The experiment described above is named the 2003 case, hereafter.

169 The simulated sea ice concentration is compared with the Advanced Microwave Scan-
170 ning Radiometer for Earth Observing System (AMSR-E) data set, which is available at
171 the International Arctic Research Center of the University of Alaska Fairbanks (IARC)
172 - Japan Aerospace Exploration Agency (JAXA) Information System (IJIS) data archive.
173 The 2003 July average is shown in section 3.1. For the comparison of phytoplankton
174 distribution, the spatial features of summertime CHL concentration in the western Arc-
175 tic Ocean are checked using the monthly and 8-day composites of Moderate-Resolution
176 Imaging Spectroradiometer (MODIS). The scenes with a spatial resolution of 4 km are
177 downloaded from the National Aeronautics and Space Administration (NASA) ocean color
178 website [<http://oceancolor.gsfc.nasa.gov>] and are digitally processed using the NASA's
179 Sea-Viewing Wide Field-of-View Sensor (SeaWiFS) Data Analysis System (SeaDAS) soft-
180 ware as in *Watanabe* [2011]. Since the detection of local particular signals from satellite
181 ocean-color sensor is difficult owing to numerous missing values under cloud-cover condi-

182 tion in the western Arctic region, we pick up the 2003 July composite for the shelf field
183 in section 3.1 and two examples of 2003 and 2010 weekly composites for the eddy field
184 in section 4.1. In this paper, we assume $1 \text{ mmolN} \sim 80 \text{ mgC} \sim 1.6 \text{ mgChl}$ based on the
185 mass ratio C/CHL of 50 and the Redfield ratio C/N of 6.625 for comparison between the
186 model output and the observational estimate with different units, as in *Aita et al.* [2007].

3. Phytoplankton Activity in the Western Arctic Ocean

3.1. Model Performance on Phytoplankton Bloom Before Eddy Generation

187 The overall spatial pattern in the 2003 case is documented including comparison with
188 previous observational findings. As described in *Watanabe* [2011], the sea ice extent,
189 sea surface temperature (SST), sea surface height, shelf-wide and basin-scale sea ice and
190 ocean circulation fields are mostly reproduced. From March to May, the wide area of
191 model domain is covered by sea ice. In June, an initial phytoplankton bloom occurs
192 following the summertime sea ice retreat in the Chukchi shelf. The surface phytoplankton
193 concentration on July 1 captures the high biomass band along sea ice margin over the
194 northern edge of Chukchi Sea and north of the Alaskan northern coast, where the sea ice
195 extent is consistent with the AMSR-E scene (Figures 2a-b).

196 The quite low concentration appears in the northeastern Chukchi shelf from the Cape
197 Lisburne to the Barrow Canyon. Figure 3 displays the vertical profile of phytoplank-
198 ton and nitrate concentration along the meridional section from the Bering Strait to the
199 Northwind Ridge on the same day as Figure 2a. The profile is characterized by the subsur-
200 face maximum of phytoplankton biomass reaching $4 \mu\text{M}$ over the northern Chukchi shelf
201 ($70^\circ\text{N} - 73^\circ\text{N}$). The subsurface maximum is accompanied by the deepening of nutricline
202 and can be simply explained by a combination of nutrient limitation and light availability,
203 since the half saturation constant for nitrate is set to $0.7 \mu\text{M}$ for both small (PS) and large
204 (PL) phytoplankton following *Zhang et al.* [2010]. Actually, most of nitrate in the surface
205 euphotic zone of central Chukchi shelf is depleted owing to the initial bloom after sea ice
206 opening by the end of June. The depression of iso-nitrate contours in this region is cap-
207 tured by the WOA09 July climatology (Figure 3). In the same period, a significant part of

208 phytoplankton biomass diminished via northward advection [*Hill et al.*, 2005; *Watanabe*,
209 2011] and local biological process (i.e., mortality, grazing of zooplankton etc.). On the
210 other hand, the warm and nutrient-rich Pacific water inflowing from the Bering Strait
211 continuously sustains the phytoplankton bloom in the southern Chukchi shelf, where the
212 southward increase in surface CHL concentration was detected by the MODIS July com-
213 posite in 2003 (Figures 2a-b). The vertically integrated PS and PL biomass reaches 366
214 $\mu\text{M m}$ ($\sim 586 \text{ mgChl m}^{-2}$) at the Bering Strait in the 2003 case. In reality, the high nitrate
215 concentration over $10 \mu\text{M}$ and the CHL content up to 734 mgChl m^{-2} were observed at
216 the strait during the 2002 - 2004 cruises [*Lee et al.*, 2007].

217 Another feature in Figure 2a is lower phytoplankton biomass in the central Canada
218 Basin than that over the Chukchi Plateau, although both the regions are still covered
219 by sea ice packs. The remarkable deepening of nutricline via Ekman downwelling associ-
220 ated with the Beaufort High sea level pressure would account for the restricted primary
221 production in the central basin [*Nishino et al.*, 2011b; *Watanabe*, 2012]. The nitrate
222 concentration over the Northwind Ridge is somewhat lower than the WOA09 July clima-
223 tological composite (Figure 3), presumably because the wider sea ice opening than the
224 multi-decadal average promoted the nitrate uptake associated with primary production
225 for the recent years. In addition, we should consider that the original data of WOA09
226 with the coarse resolution of $1^\circ \times 1^\circ$ have difficulty in capturing local features. It is also
227 found that the initial bloom is dominated by PL in the entire domain, and the PS habitat
228 is limited to the Alaskan coastal region south of the Cape Lisburne at the beginning of
229 July (see yellow contours in Figure 2a).

3.2. Seasonal Transition in the Barrow Canyon

230 The Barrow Canyon is an important passage for the transport of Pacific summer water
231 from the Chukchi shelf to the Canada Basin. Since the Pacific summer water is a major
232 source of the Beaufort shelf-break warm eddies, the water mass properties in the canyon
233 would significantly control biological regimes in the basin. The physical and biogeochem-
234 ical properties in the canyon are investigated hereafter in this subsection. The Barrow
235 Canyon is dominated by strong northeastward ocean current, and the major variations in
236 plankton biomass and nutrient amount in the canyon represent the biological activity in
237 the upstream regions over the Chukchi shelf. It should be hence noted that the changes in
238 the Barrow Canyon are not immediately recovered even when water mass is continuously
239 replaced via advection.

240 The temporal variability of vertically cumulative phyto- and zooplankton biomass, SST,
241 surface light availability, sea ice concentration, and nutrient concentration are plotted in
242 Figure 4. The readers can check the exact formulation and parameter values of primary
243 production rate adopted in the 2003 case in Appendix. All the daily values are their
244 averages in the Barrow Canyon section (Figure 2a), which is defined by a line between a
245 location northwest of the canyon (156.6°W , 71.9°N) and Point Barrow (156.3°W , 71.3°N)
246 as same as *Watanabe* [2011]. First, it is found that there are remarkable differences in
247 the seasonal variations between PS and PL (Figure 4a). PL biomass initially breaks up
248 at the end of April and reaches its peak of $82 \mu\text{M m}$ ($\sim 6.6 \text{ gC m}^{-2}$) in the early period of
249 June. After then, PL biomass gradually decreases without its distinct peak until the finish
250 of model integration. PS biomass slowly increases from May, but overtakes PL biomass
251 at the end of July. The seasonal variation in ZL biomass follows these phytoplankton

252 behaviors via grazing process (Figure 4b). ZL biomass averaged in the Barrow Canyon
253 increases from June to July, and has a peak of $50 \mu\text{M m}$ ($\sim 4.0 \text{ gC m}^{-2}$) at the beginning
254 of August, when a significant amounts of PS and PL are both alive. To the contrary, ZS
255 biomass is quite low throughout the blooming season. ZP biomass slightly increases after
256 July, but the amount is evidently low relative to ZL. *Campbell et al.* [2009] estimated that
257 the total copepod biomass in the upper 100 m of northern Chukchi shelf region seasonally
258 changed from 0.6 gC m^{-2} in May - June to 1.9 gC m^{-2} in July - August, 2002. Since the
259 ZL biomass recalculated in the top 100 m is $45 \mu\text{M m}$ ($\sim 3.6 \text{ gC m}^{-2}$) in the 2003 case,
260 the model result may overestimate the ZL grazing on phytoplankton species. The impact
261 of zooplankton is discussed in section 5.

262 SST is kept at the freezing point until May and then suddenly rises up to 6°C within two
263 weeks in June (Figure 4c). The rapid warming corresponds to sea ice removal from the
264 canyon. The surface light availability (SLA) is also principally synchronized with sea ice
265 concentration over the canyon (Figure 4d). The non-zero value of SLA even with freezing
266 SST during the spring season means the partial exposure of open water area. The SLA
267 drop in early July reflects the photoinhibition of phytoplankton for the annual maximum
268 period of solar radiation. While the simulated light property depends on the parameter
269 values of P-E curve (see Appendix), the summertime photoinhibition of diatom group was
270 at least reported in the southeastern Beaufort Sea [*Palmer et al.*, 2011]. In August and
271 September, when the SLA fluctuates just below the maximum value of 1, photosynthesis
272 is not completely restricted by light intensity in the surface layer.

273 The vertical profile along the Barrow Canyon section on June 11, when the PL biomass
274 maximum is recorded, is shown in Figure 5. The LA term vertically lapses to 0.1 at 30 - 50

275 m depth and 0.01 at 60 - 100 m depth. The light extinction rate depends on self-shading
276 effect of phytoplankton biomass in addition to a water depth [*Kishi et al.*, 2007]. The
277 high PL biomass dominates the upper 30 m, where nitrate uptake has been considerably
278 progressed since sea ice opening. Recalling the half saturation constant of $0.7 \mu\text{M}$, the
279 profile indicates that nitrate is still available for photosynthesis even below the depth
280 of PL biomass maximum. Another feature is the east-west gradient of PL and nitrate
281 concentration below the 40 m depth. PL biomass substantially resides in the eastern side
282 of the canyon (i.e., the Alaskan coastal side). This distribution is accompanied by nutrient
283 consumed in the upstream region and reflects the pathway of Pacific summer water, as
284 shown in the ship-based measurements [*Walsh et al.*, 2011] and in the tracer experiment
285 [*Watanabe and Hasumi*, 2009].

286 Here, we consider the nutrient concentration multiplied by the light availability in each
287 horizontal and vertical grid in order to filter out the value in the dark layer (Figure 4e).
288 For example, light-weighted nutrient concentration is nearly zero due to full ice cover re-
289 gardless of nutrient amount in March. The multiplied value in each grid is also averaged
290 through the entire water column. The optimized nitrate concentration abruptly drops
291 in early May after its maximum of $0.29 \mu\text{M}$ due to phytoplankton bloom, while a part
292 of the depression is explained by the weakened light condition during May 5 - 8. In the
293 case of June 11, the value records $0.06 \mu\text{M}$ (cf. Figure 5). The ammonium concentration
294 is low until the end of May, but the gradual increase is then induced by the modeled
295 remineralization processes and the restoring in the deepest layer. The effect of bottom
296 flux is referred in section 5. The silicate variation traces similar features with nitrate
297 concentration before August and then becomes gradually reduced in contrast to nitrate

298 regeneration via nitrification in late summer (Figures 4e-f). The surface nitrate deple-
299 tion, the subsurface maximum of phytoplankton biomass above $2 \mu\text{M}$ ($\sim 3.2 \text{ mgChl m}^{-3}$)
300 located at the 20 - 40 m depth, and the dominance of larger-size diatom in the entire
301 canyon column after the early bloom are consistent with the SBI measurements [*Hill*
302 *et al.*, 2005; *Sukhanova et al.*, 2009], although the greatest value of CHL concentration up
303 to 12 mgChl m^{-3} observed in their cruises is not reproduced in our experiment.

304 Based on each time series described above, the phytoplankton growth in the Chukchi
305 shelf before the eddy generation period are summarized (see also Appendix). Solar input
306 with sea ice retreat initializes the phytoplankton bloom. The SST increase of 8°C provides
307 the Tmp term with 1.74 times larger contribution to photosynthesis following the so called
308 Q_{10} relationship. The delayed growth of PS relative to PL is explained by the difference
309 in parameter values of maximum photosynthetic rate (V_{max}) as long as enough nutrient
310 is available for uptake. The silicate depletion in the surface layer gradually inhibits the
311 PL growth in early summer, while PS can continue to grow up with the uptake of nitrate
312 and ammonium even in late summer. Actually, the similar seasonal shift of dominant
313 species to smaller-size phytoplankton groups was observed by *Hill et al.* [2005]. Thus
314 when the Beaufort shelf-break eddies are initially generated in late July, the changes in
315 PL and PS biomass in the Barrow Canyon are negative and positive, contrastively. The
316 gross primary production rate of PL (GppPL) and PS (GppPS) with the peak of nearly
317 $8 \mu\text{M m day}^{-1}$ ($\sim 0.64 \text{ gC m}^{-2} \text{ day}^{-1}$) shown in Figure 4g fundamentally changes in phase
318 with the biomass of PL and PS itself and is modified by nutrient environment includ-
319 ing the high efficiency of ammonium uptake (cf., characterized by lower half saturation
320 constant for ammonium). It is expected that these seasonal variations would be closely

³²¹ linked to the primary productivity in the Beaufort shelf-break warm eddies unless nutrient
³²² compensation occurs in the downstream region.

4. Phytoplankton Behaviors associated with Beaufort Shelf-break Eddies

4.1. Eddy-like Structure of Phytoplankton Biomass

323 In August, a few anti-cyclones spawned from the Barrow Canyon jet and eastward shelf-
324 break current intrude into the basin interior [*Watanabe, 2011*]. The spatial distribution of
325 combined PS and PL biomass illustrate the swirling structure north of the Barrow Canyon
326 (Figure 6a). The phytoplankton biomass is clearly higher in an individual eddy than the
327 surrounding basin area. We found that two examples of MODIS CHL distribution among
328 all weekly composites from 2003 to 2008 were most visible as introduced in section 2.
329 First, the eddy-like feature of high CHL concentration appeared north of the Beaufort
330 shelf break on September 6 - 13, 2003 (Figure 6b), when the warm eddies were detected
331 by the MODIS 11- μm brightness temperature [*Watanabe, 2011*]. The maximum value of
332 CHL concentration reaches $0.3 \mu\text{M}$ ($\sim 0.5 \text{ mgChl m}^{-3}$) at 145°W and 71°N . The other
333 feature of eddy-like CHL distribution was captured northwest of Point Barrow on August
334 21 - 28, 2010 (Figure 6c). The maximum CHL concentration records $0.9 \mu\text{M}$ (~ 1.5
335 mgChl m^{-3}) at 158°W and 74°N . This location is close to the distinct warm eddy tracked
336 by the R/V *Mirai* cruise during late summer and early autumn in the same year [*Nishino*
337 *et al.*, 2011a]. The maximum value of surface phytoplankton concentration on August 27
338 in the 2003 case corresponds to $0.4 \mu\text{M}$ (0.6 mgChl m^{-3}) (Figure 6a), which is in the same
339 order as the MODIS values and the observed CHL concentration up to 0.8 mgChl m^{-3} in
340 *Nishino et al.* [2011a] (Figure 3c in their paper).

4.2. Eddy Generation/Development Stage

341 The daily primary production rate in the Beaufort shelf-break region is tracked following
342 the eddy life stages (Figure 7). When the shelf-break warm eddies are initially generated
343 at the end of July, the eddy-like distribution of shelf-origin water with high primary
344 productivity arises north of the Barrow Canyon (Figure 7a). These anti-cyclonic eddies
345 migrate offshore with developing their sizes in August (Figure 7b). Even though several
346 weeks have passed from its generation, apparent primary production still continues at the
347 eddy center. In addition, the western July eddy (named JED) traps the shelf water along
348 the outer edge of eddy via clockwise rotational flow.

349 To estimate the relative importance of shelf-water transport on primary productivity
350 inside the shelf-break eddies, we conducted an idealized experiment. In the original 2003
351 case, a virtual passive tracer associated with the Pacific-origin water is provided at the
352 Bering Strait so that the tracer concentration is kept at 100% at the strait throughout the
353 integration period. The readers can confirm the overall distribution of Pacific water tracer
354 in *Watanabe* [2011] and the offshore intrusion accompanied with the JED in Figure 7a.
355 In the “No Basin Biogeochemical Value (NBBV)” case, all values of 11 biogeochemical
356 components of NEMURO (Figure 1b) are reset to zero on July 24 in the basin-side area
357 where the concentration of Pacific water tracer is below 0.1 in each vertical level on
358 the same day (see blue contours in Figure 7a). The modified integration from July 24 to
359 August 27 results in the similar spatial pattern of gross primary production rate including
360 the eddy-like maximum, although the magnitude is slightly smaller compared with the
361 2003 case (Figures 7b and 8). Therefore, it can be regarded that the primary productivity
362 inside the JED is maintained by consuming the residual of nutrient taken at the initial

363 eddy stage and that the exchange of biogeochemical properties with background basin
364 environment is not an essential process.

4.3. Eddy Maturity Stage

365 The primary production inside the JED is enhanced when the eddy activities become
 366 prominent in September (Figure 7c). It can be also detected that a newly generated eddy
 367 north of the Barrow Canyon (named SED) transports the Chukchi shelf water with high
 368 primary productivity. The vertical profiles of primary production rate, nitrate concentra-
 369 tion, and vertical diffusivity/velocity across the JED and SED are shown in Figure 9. The
 370 eddy properties are characterized by the lower nitrate concentration, which is below $1 \mu\text{M}$
 371 at the depth shallower than 80 m at the eddy center as observed in late summer 2010
 372 *Nishino et al.* [2011a]. It is sometimes reported that the depression of nutricline restricts
 373 primary production in an anti-cyclonic eddy [*Ueno et al.*, 2010]. On the other hand, the
 374 model result shows a hot spot of production in the inner uppermost layer.

375 GppPS at the center of JED in the surface layer is $6.86 \times 10^{-2} \mu\text{M day}^{-1}$ on September
 376 19, which is two order of magnitude greater than that of $0.01 \times 10^{-2} \mu\text{M day}^{-1}$ at the
 377 surface eddy edge (Figure 9c and Table 1). The relative contribution of nutrient uptake,
 378 temperature-dependent, and light availability terms to GppPS is compared between the
 379 center and edge of JED at the ocean surface. Evidently, the eddy center has higher
 380 nitrate, ammonium, and water temperature than the eddy edge. The Michaelis-Menten
 381 formulation with the gourmet term of ammonium intends that the higher anomalies of
 382 nitrate ($0.13 \mu\text{M}$) and ammonium ($0.07 \mu\text{M}$) result in an order greater contribution to
 383 GppPS. The Q_{10} relationship prescribes that the warm anomaly of 2°C is equivalent to
 384 20% larger increases in GppPS as well. The horizontal gradient of surface light availability
 385 is negligible for photosynthesis on the eddy scale of $O(10 \text{ km})$. While the upper column
 386 of eddy center has quasi-uniform nutrient and temperature condition, GppPS of $0.93 \times$

387 $10^{-2} \mu\text{M day}^{-1}$ in the subsurface layer is significantly smaller than at the ocean surface
388 (Figure 9c and Table 1). The vertical contrast of GppPS is simply explained by surface
389 light availability that is eight times higher than that at the 60 m depth. Therefore, the
390 surface central area of shelf-break warm eddies satisfies enough light, nitrate/ammonium,
391 and high-temperature environments, which are all suitable for PS photosynthetic activity.
392 A key point here is that the primary production of PL is constrained even at the surface
393 eddy center in contrast to the seasonal dominance in the basin subsurface layer (Figure 9d).
394 Since the light, nitrogenous nutrient (nitrate and ammonium), and temperature conditions
395 including parameter values for PS and PL are same, the lower GppPL is explained by
396 the limited silicate uptake. In fact, the residual silicate concentration is quite low on
397 the pathway of shelf-origin water through the Barrow Canyon (Figure 4f) and inside the
398 eddies (Table 1). The demonstrated water temperature and predominance of smaller-size
399 phytoplankton species are consistent with the in-situ measurements conducted by the
400 R/V Mirai 2010 cruise [*Nishino et al.*, 2011a].

401 The roles of internal eddy dynamics, such as vertical mixing and local up-
402 welling/downwelling, in nutrient redistribution and the consequent regulation of primary
403 productivity are analyzed. Figure 9e indicates that the enlarged vertical diffusivity up
404 to $2.7 \times 10^{-2} \text{m}^2 \text{s}^{-1}$ in the upper 80 m of JED center drives the exchange with under-
405 lying nutrient-rich water that is also shelf-origin. The localized high vertical diffusivity
406 is comparable with the tidal mixing of $O(10^{-2} \text{m}^2 \text{s}^{-1})$ reported around the Kuril Strait
407 in the North Pacific [*Kawasaki and Hasumi*, 2010]. In the COCO model, the vertical
408 diffusion coefficient is diagnosed using the turbulence closure scheme of *Noh and Kim*
409 [1999] at each time step. The turbulent kinetic energy inside the shelf-break eddies arises

410 according to the vertical shear of horizontal velocity and is maintained for the sake of
411 weakened stratification. Another features of eddy dynamics include the upward nutrient
412 shift induced by upwelling flow that reaches 6.2 m day^{-1} in the outer side of JED (Figure
413 9e). The upwelling/downwelling event is recognized as the perturbation of eddy activi-
414 ties, since the daily fluctuation of vertical velocity is not synchronized with the Ekman
415 convergence/divergence derived from surface wind stress (Figure 9b).

416 To evaluate the impact of these eddy processes on GppPS and GppPL, we conducted
417 sensitivity experiments where nitrate, ammonium, and silicate redistribution caused by
418 each of vertical diffusion and advection flux terms is excluded during the model integration
419 from August 27 to September 19. These idealized experiments are named “No Vertical
420 Nutrient Diffusion (NVND)” and “No Vertical Nutrient Advection (NVNA)” cases, re-
421 spectively. In the NVNA case, the vertical component of nutrient advection parts written
422 in the flux form is set to zero at all layer boundaries for the conservation of total nutrient
423 content in this calculation. The other model part including physical fields is identical
424 to the 2003 case. The simulated eddy fields in the NVND case are characterized by the
425 nutricline shoaling and the reduction of primary productivity in the central column of
426 JED (Figures 10a-b). The decrease in net upward nutrient flux accounts for the GppPS
427 (GppPL) decrease to 1.64×10^{-2} (0.21×10^{-2}) $\mu\text{M day}^{-1}$ at the surface eddy center. At
428 the same time, the increase rate of GppPS (GppPL) to 1.43×10^{-2} (0.18×10^{-2}) $\mu\text{M day}^{-1}$
429 accompanied by the positive anomaly of nutrient concentration is small due to limited
430 light availability at the subsurface eddy center (Table 1). Thus it is confirmed that the
431 primary production inside the shelf-break warm eddies is enhanced by turbulent mixing
432 even a few months after their generation, and the magnitude is comparable with the sea-

433 sonal subsurface maximum in the basin interior (Figures 9c-d). In contrast, the deviation
434 of GppPS and GppPL in the NVNA case from the 2003 case is small inside the JED
435 relative to that in the NVND case (Figures 10c-d, Table 1). The result suggests that the
436 nutrient redistribution due to local upwelling/downwelling event has a minor contribution
437 to the primary productivity of shelf-break eddies. The large negative anomaly of GppPL
438 at the nutricline depth of basin interior is expected to be explained in part by the lack
439 of wind-driven upwelling of underlying nitrate-rich water, while such basin processes are
440 out of the scope in this study.

4.4. Eddy Decay Stage

441 There is a few reports on the shoaling of nutricline inside the anti-cyclonic eddies ac-
442 cording to their decay and the corresponding restart of active primary production in the
443 North Pacific [*Ueno et al.*, 2010]. On the other hand, the model result in the 2003 case
444 shows weakened primary productivity in the southern Beaufort Sea in October (Figure
445 7d), when the vorticity of shelf-break warm eddies gradually shrinks due to lateral fric-
446 tion [*Watanabe*, 2011]. The AMSR-E sea ice monitor indicated that the ice freezing
447 period in the southern Beaufort Sea started during October and November every year in
448 the 2000s [<http://www.ijis.iarc.uaf.edu/cgi-bin/seaice-monitor.cgi>]. The autumn sea ice
449 spread normally precedes the disappearance of shelf-break warm eddies, and the further
450 eddy shrinking then occurs due to sea ice drag in addition to lateral friction. In the
451 2003 case, sea ice freezing prevents light absorption at the ocean surface and ceases the
452 primary production of pelagic phytoplankton species in the most shelf-break region in
453 mid-October. An exceptional area appears in the vicinity of JED centered at 153°W and
454 73°N (see the location in Figure 7d). The local delay of sea ice freezing until November 8
455 occurs attributing to ocean heat content inside the JED. To confirm whether open water
456 exposure over the shelf-break warm eddies can prolong the photosynthesis period or not in
457 the southern Beaufort Sea, the JED properties are compared between the eddy maturity
458 and decay periods (Figure 11).

459 The temporal changes are characterized by the remarkable decline of GppPS from
460 $6.86 \times 10^{-2} \mu\text{M day}^{-1}$ on September 19 to $0.18 \times 10^{-2} \mu\text{M day}^{-1}$ on October 31 at the
461 surface eddy center. The following values are also taken at the same place. During this
462 period, nitrate (ammonium) concentration recovers from 0.14 (0.08) μM to 0.55 (0.11) μM

463 due to the decomposition of organic matter and dynamic processes (Figure 11c). This
464 change corresponds to the increase in Nit (Amn) term from 0.15 (0.29) to 0.39 (0.36) and
465 has a positive contribution to GppPS. Similarly, the GppPL decrease from $0.57 \mu\text{M day}^{-1}$
466 to $0.02 \mu\text{M day}^{-1}$ is not explained by the Sil term from 0.05 to 0.07. The rate of decrease
467 in the Tmp term from 1.25 to 1.04, which derives from the cooling of ocean surface from
468 $3.27 \text{ }^\circ\text{C}$ to $0.61 \text{ }^\circ\text{C}$, is just 20%. The temperature profile higher than the freezing point of
469 about $-1.8 \text{ }^\circ\text{C}$ with remaining temperature maximum at 130 m depth certainly can afford
470 to maintain open water pool at the top of JED. The LA term of 0.32 associated with
471 the NCEP solar radiation of 10 W m^{-2} at the ocean surface assures the continuance of
472 photosynthesis in the available nutrient condition after when sea ice covers the shelf-break
473 region surrounding the JED on October 20. Actually, the LA term decreases from 0.93 on
474 September 19 to 0.06 on October 31 in the open water area. Sea ice concentration then
475 increases from 0 to 1 for November 1 to 9. The daily downward shortwave flux drops to
476 zero (i.e., polar night all the day) owing to solar incidence on November 8 at 73°N . Even
477 taken account for the decrease in PS (PL) biomass itself of 0.19 (0.04) μM to 0.06 (0.02)
478 μM , it is indicated that the principal factor for the termination of primary production in
479 spite of nutrient recovery inside the JED is light limitation due to solar incidence rather
480 than sea ice freezing, and surface cooling has secondary contribution.

5. Summary and Discussions

481 The response of phytoplankton to the Beaufort shelf-break eddies in the western Arctic
482 Ocean is examined using the eddy-resolving coupled sea ice-ocean model including a lower
483 trophic marine ecosystem formulation. In the integration from March to November, the
484 reasonable performance on seasonal phytoplankton bloom following sea ice retreat in the
485 shallow Chukchi shelf is achieved. The sea ice margin is located north of the shelf-basin
486 boundary, and several warm eddies are produced during late summer and early autumn.
487 The shelf-break warm eddies initially transports the Chukchi shelf water with high primary
488 productivity toward the Canada Basin interior. In the eddy-developing period, the anti-
489 cyclonic rotational flow along the outer edge of each eddy moving offshore occasionally
490 traps the shelf water. The primary production inside the warm eddies is maintained by
491 internal dynamics in the eddy-maturity period. In particular, the surface central area
492 of an anti-cyclonic eddy satisfies adequate light, nutrient, and warm environment for
493 photosynthetic activity partly attributing to turbulent mixing with underlying nutrient-
494 rich water. The simulated biogeochemical properties with the dominance of small-size
495 phytoplankton inside the warm eddies are consistent with the previous in-situ observation
496 in the western Arctic Ocean. It is also suggested that the light limitation before sea
497 ice freezing rather than nutrient depletion shuts down the primary production in the
498 shelf-break eddies. The findings provided in the present study indicate that the time lag
499 between the phytoplankton bloom in the shelf region following the summertime sea ice
500 retreat and the eddy generation along the Beaufort shelf break is an important index to
501 determine biological regimes in the Canada Basin (Figure 12). When the phytoplankton
502 bloom in the shelf occurs later or continues longer, the shelf-break eddies would further

503 enhance the primary productivity in the basin interior. Therefore, we propose that the
504 shelf-break eddies play a significant role in the extension of primary production period,
505 the migration of shelf species toward the basin interior, and the bottom-up process of
506 local food web.

507 There are a lot of uncertainties in zooplankton characteristics, presumably because of
508 insufficient zooplankton measurements. A few model studies incorporated the seasonal
509 vertical migration of *Calanus* species reported in the subarctic North Pacific [*Aita et al.*,
510 2003]. While *Calanus* is categorized as ZL in the NEMURO model, the period and depth
511 of its vertical migration in the Arctic Ocean are specifically unknown and might differ
512 from other regions. That is the reason why the vertical migration process is not included
513 in our experiment. This uncertainty may cause the overestimation of zooplankton biomass
514 described in section 3.2 and consequent high grazing pressure on phytoplankton species.
515 To broadly estimate the impact of zooplankton on the phytoplankton life cycle, we addi-
516 tionally conducted a sensitivity experiment where the initial zooplankton concentration
517 is set to zero. This extreme experiment, whose design is identical to the 2003 case except
518 the removal of zooplankton, is named the “No Zooplankton” (NZOO) case. As expected,
519 the primary productivity in the shelf-break eddies is considerably enhanced in the NZOO
520 case (Figure 13). Thus we regard that the phytoplankton performance is sensitive to
521 the zooplankton behaviors and that the detailed assessment of grazing process should be
522 addressed as our future work. In this connection, the vertically integrated total of GppPS
523 and GppPL at the JED center ranges from $2.30 \mu\text{M m day}^{-1}$ ($\sim 0.18 \text{ gC m}^{-2} \text{ day}^{-1}$) in
524 the 2003 case to $3.72 \mu\text{M m day}^{-1}$ ($\sim 0.30 \text{ gC m}^{-2} \text{ day}^{-1}$) in the NZOO case on September
525 19. The eddy-induced productivity estimated in our experiment is hence below the half of

526 observed summertime average of $0.78 \text{ gC m}^{-2} \text{ day}^{-1}$ in the Chukchi shelf [Hill and Cota,
527 2005] and might have just a minor impact on the pan-Arctic biogeochemical cycle.

528 The ship-based in-situ measurements have captured the highest ammonium concentra-
529 tion just above the shallow Chukchi shelf [Nishino *et al.*, 2011b]. The dissolution from sea
530 bottom sediments is considered to be a crucial nitrogen source in the shelf region and even
531 in the basin interior throughout the year, although the reliable sequential data has not
532 been acquired yet. Besides, the benthos communities are assumed to be representative
533 consumers of particulate and dissolved organic nitrogen [Grebmeier *et al.*, 2006]. The
534 benthos species might be an important reservoir of nitrogen and possibly modulate am-
535 monium concentration in the deep layer. The ammonium concentration over the Chukchi
536 shelf bottom is assumed to have a distinct seasonal cycle. Actually, the ammonium con-
537 centration more than $4 \mu\text{M}$ was observed along the northern edge of Chukchi shelf during
538 the summer cruise, while the winter value is below $2 \mu\text{M}$ [Nishino *et al.*, 2005]. Our ex-
539 periment hence applies the restoring method of ammonium concentration with a seasonal
540 cycle as described in section 2. The adopted restoring works on controlling the phase
541 of seasonal variation in the shelf, because we have checked that the experiment without
542 restoring could not reproduce the observed signal in Nishino *et al.* [2005, 2011a]. On the
543 other hand, silicate supply from the shelf bottom is not incorporated because of few obser-
544 vational evidences in the Chukchi Sea. Actually, a major silicate source is suggested to be
545 the Bering Sea rather than bottom sediments [S. Nishino, personal communication]. The
546 silicate inflow from the Bering Sea is indirectly represented by the restoring to WOA09
547 data at the Bering Strait in the present experiments. The geochemical fluxes from the

548 Chukchi shelf bottom, which have not been quantitatively estimated yet, would be useful
549 information for further model development.

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Appendix: Photosynthesis Formulation in NEMURO

The NEMURO model configuration assumes that the photosynthesis of PS and PL is a function of nitrate NO_3 , ammonium NH_4 , and silicate $Si(OH)_4$ concentration, water temperature T , light intensity I [Kishi *et al.*, 2007]. The formulation of gross primary production rate of PS (GppPS) and PL (GppPL) consists of nitrate, ammonium, silicate uptake terms (Nit, Amn, Sil, respectively), temperature-dependent term (Tmp), and light availability term (LA), and the biomass itself PS and PL as follows:

$$\begin{aligned}
 \text{GppPS} &= V_{\text{maxs}} \left(\underbrace{\frac{NO_3}{NO_3 + K_{NO_3}} \exp(-\Psi NH_4)}_{[\text{Nit}]} + \underbrace{\frac{NH_4}{NH_4 + K_{NH_4}}}_{[\text{Amn}]} \right) \\
 &\quad \underbrace{\exp(\kappa T)}_{[\text{Tmp}]} \underbrace{\frac{I}{I_{\text{opt}}} \exp\left(1 - \frac{I}{I_{\text{opt}}}\right)}_{[\text{LA}]} \underbrace{PS}_{[\text{PS}]}, \\
 \text{GppPL} &= V_{\text{maxl}} \min \left(\frac{NO_3}{NO_3 + K_{NO_3}} \exp(-\Psi NH_4) + \frac{NH_4}{NH_4 + K_{NH_4}}, \underbrace{\frac{Si(OH)_4}{Si(OH)_4 + K_{Si(OH)_4}} / \text{RSiN}}_{[\text{Sil}]} \right) \\
 &\quad \exp(\kappa T) \frac{I}{I_{\text{opt}}} \exp\left(1 - \frac{I}{I_{\text{opt}}}\right) \underbrace{PL}_{[\text{PL}]},
 \end{aligned}$$

where the parameter values of maximum photosynthetic rate at 0°C V_{maxs} (0.7 day⁻¹), V_{maxl} (2.0 day⁻¹), half saturation constant for nitrate K_{NO_3} (0.7 μM), ammonium K_{NH_4} (0.2 μM), silicate $K_{Si(OH)_4}$ (1.15 μM), ammonium inhibition coefficient Ψ (1 μM⁻¹), temperature coefficient for photosynthetic rate κ (0.0693 °C⁻¹), optimum light intensity constant I_{opt} (104.7 W m⁻²), and Si:N ratio RSiN of 2 are all given based on Zhang *et al.* [2010]. The nutrient uptake, Tmp, and LA terms are represented by traditional Michaelis-Menten, Q₁₀ relationship, and P-E curve formulation, respectively.

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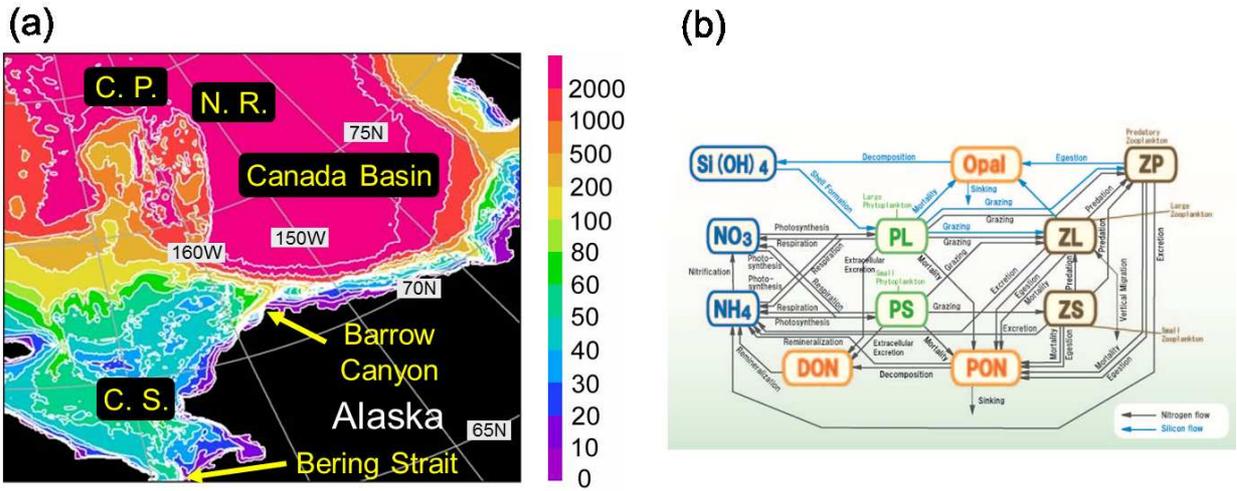


Figure 1. (a) Model bathymetry [m]. The locations include Chukchi Sea (C.S.), Chukchi Plateau (C.P.), and Northwind Ridge (N.R.). (b) NEMURO is composed of small phytoplankton (PS), large phytoplankton (PL), small zooplankton (ZS), large zooplankton (ZL), predatory zooplankton (ZP), nitrate (NO_3), ammonium (NH_4), silicate ($Si(OH)_4$), opal (Opal), particulate organic nitrogen (PON), and dissolved organic nitrogen (DON). PL and PS are assumed to represent diatom and other non-diatom group (e.g., flagellates, haptophytes), respectively.

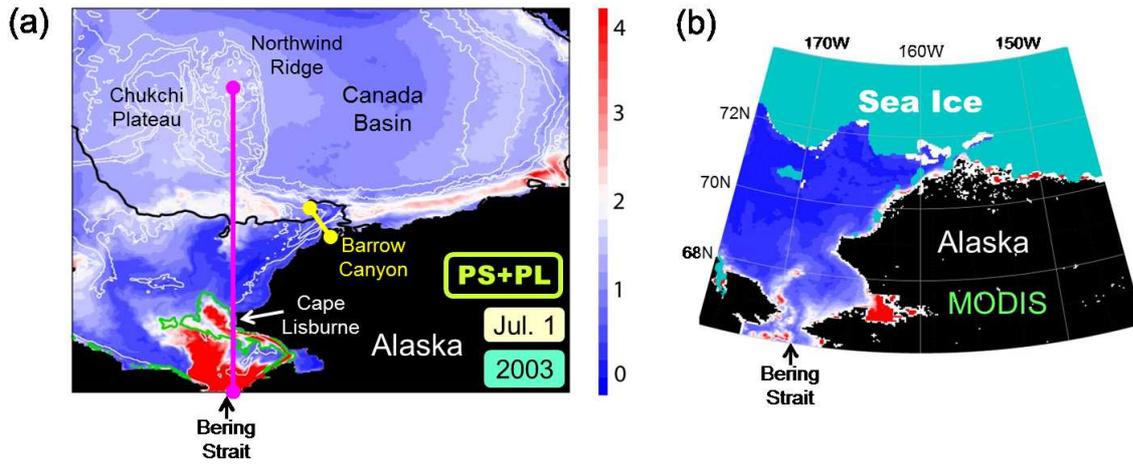


Figure 2. (a) Surface phytoplankton concentration on July 1 in the 2003 case [μM]. Sum of PS and PL is shaded, and PS concentration of $0.1 \mu\text{M}$ is plotted by green contours. Black contour shows the simulated sea ice concentration of 0.5 on the same day. White contours show bottom bathymetry. Pink line connected between the Bering Strait and the Northwind Ridge is referred in Figure 3. Yellow line across the Barrow Canyon is referred in Figures 4 and 5. (b) MODIS monthly composite of chlorophyll-*a* (CHL) concentration in July 2003. Scene ID is A20031822003212. MODIS values with the unit of mgChl m^{-3} are divided by 1.6 to directly compare the simulated PS and PL concentration. AMSR-E sea ice area is overlaid by sky-blue shade.

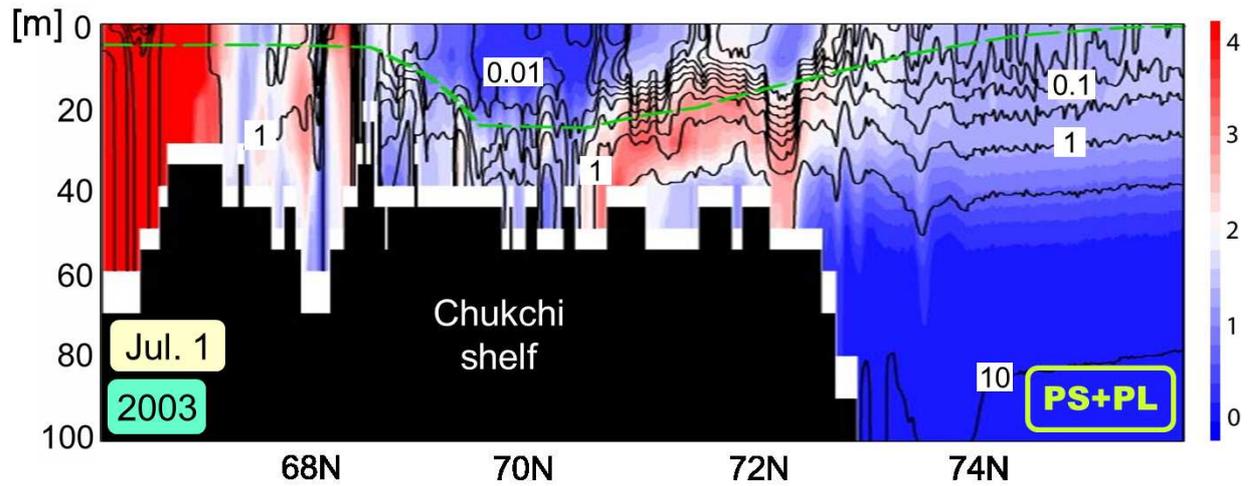


Figure 3. Vertical profile of combined PS and PL concentration along the meridional section from the Bering Strait to the Northwind Ridge (a pink line in Figure 2a) on July 1 in the 2003 case. Black contours show the simulated nitrate concentration with logarithm scale. Unit of both the properties is μM . WOA09 July climatology of nitrate concentration of $1 \mu\text{M}$ is overlaid by a green dashed line.

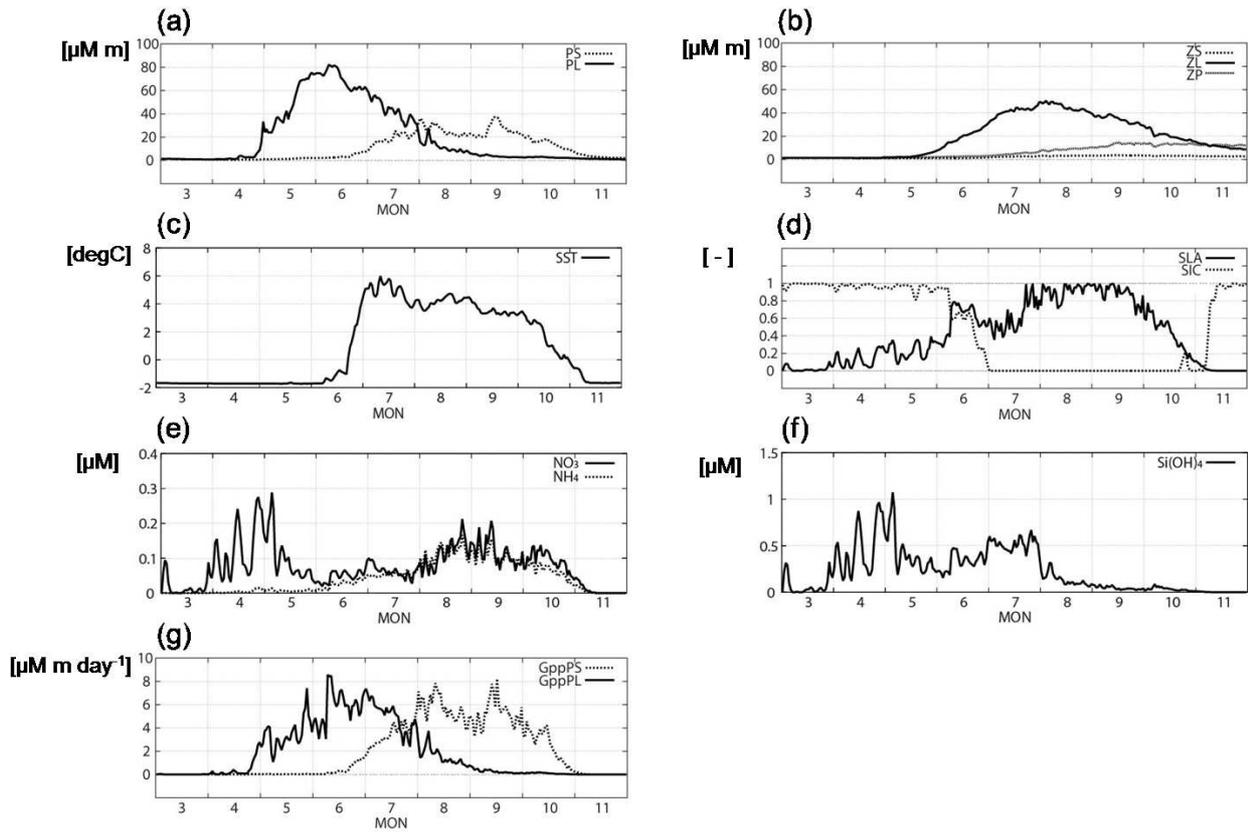


Figure 4. Seasonal variations in simulated biogeochemical and physical properties averaged in the Barrow Canyon section (a yellow line in Figure 2a) in the 2003 case. (a) PS, PL, (b) ZS, ZL, and ZP biomass integrated in the entire water column [$\mu\text{M m}$]. (c) Sea surface temperature (SST) [$^{\circ}\text{C}$]. (d) Surface light availability (SLA) and sea ice concentration (SIC) [non-dimension]. (e) NO_3^- , NH_4^+ , and (f) $\text{Si}(\text{OH})_4$ concentration weighted by light availability [μM]. As for the unit of $\text{Si}(\text{OH})_4$, $1 \mu\text{M} \equiv 1 \text{ mmolSi m}^{-3}$ is assumed. (g) Gross primary production rate of PS (GppPS) and PL (GppPL) [$\mu\text{M m day}^{-1}$]. See their definition in section 3.2.

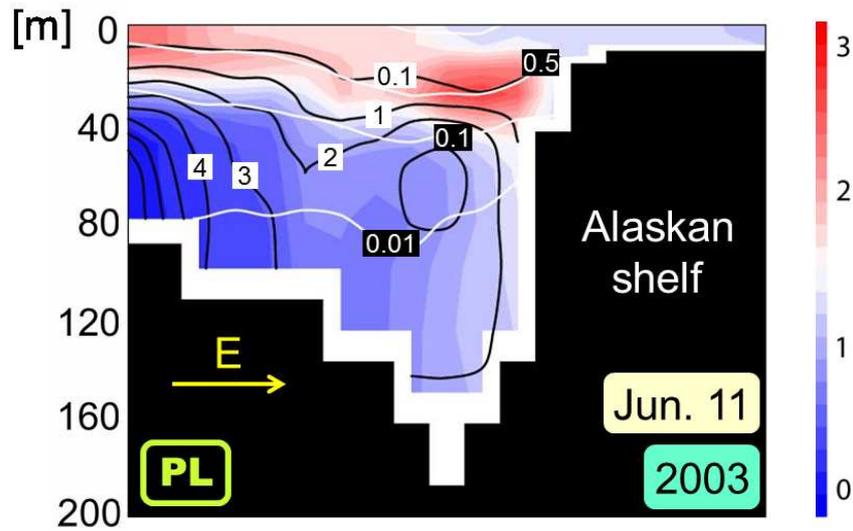


Figure 5. Color shade indicates the vertical profile of PL concentration along the Barrow Canyon section on June 11 in the 2003 case [μM]. Black and white contours show the simulated nitrate concentration [μM] and light availability [non-dimension], respectively.

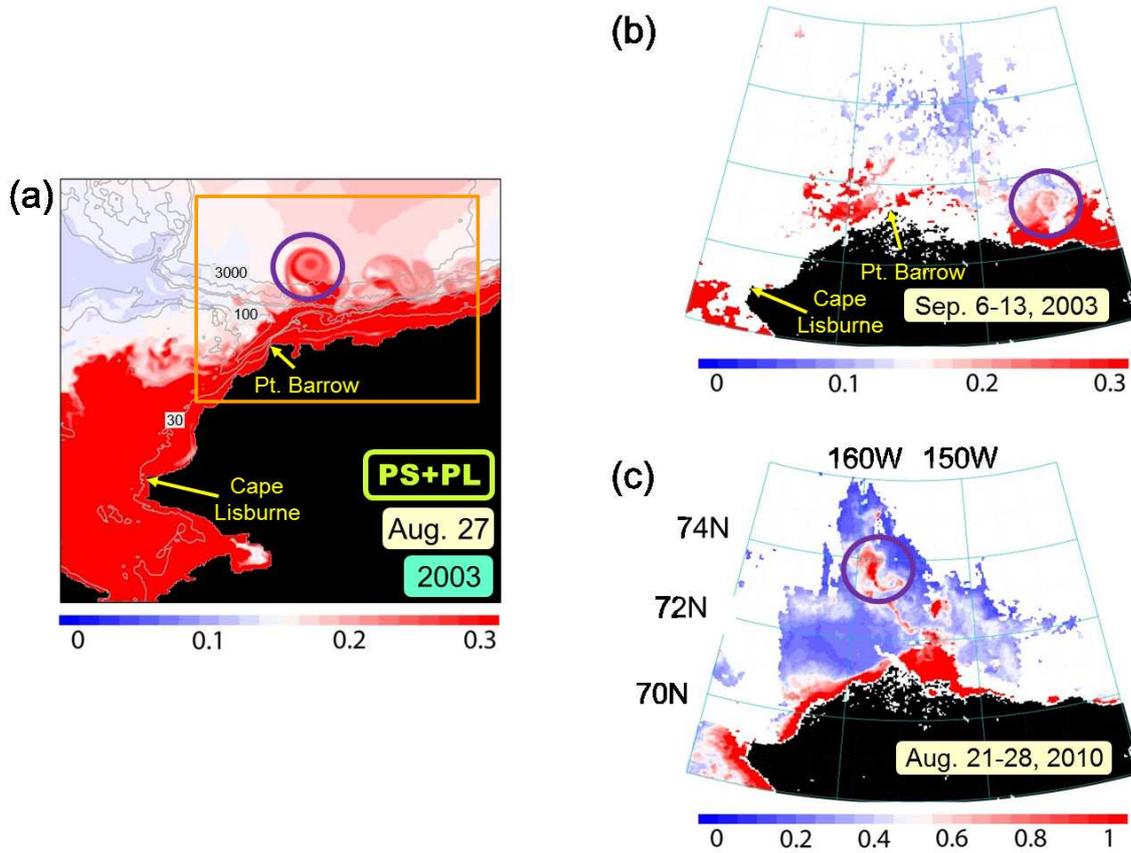


Figure 6. (a) Surface combined PS and PL concentration on August 27 in the 2003 case [μM]. Gray contours denote the water depths of 30, 60, 100, 500, 1000, 2000, and 3000 m. Orange rectangle is a target region in Figures 7, 8, and 13. MODIS 8-day composites of CHL concentration on (b) September 6 - 13, 2003 and (c) August 21 - 28, 2010. Scene ID is (b) A20032492003256 and (c) A20102332010240. MODIS values with the unit of mgChl m^{-3} are divided by 1.6 to directly compare the simulated PS and PL concentration. Eddy-like CHL patterns are captured in purple circles.

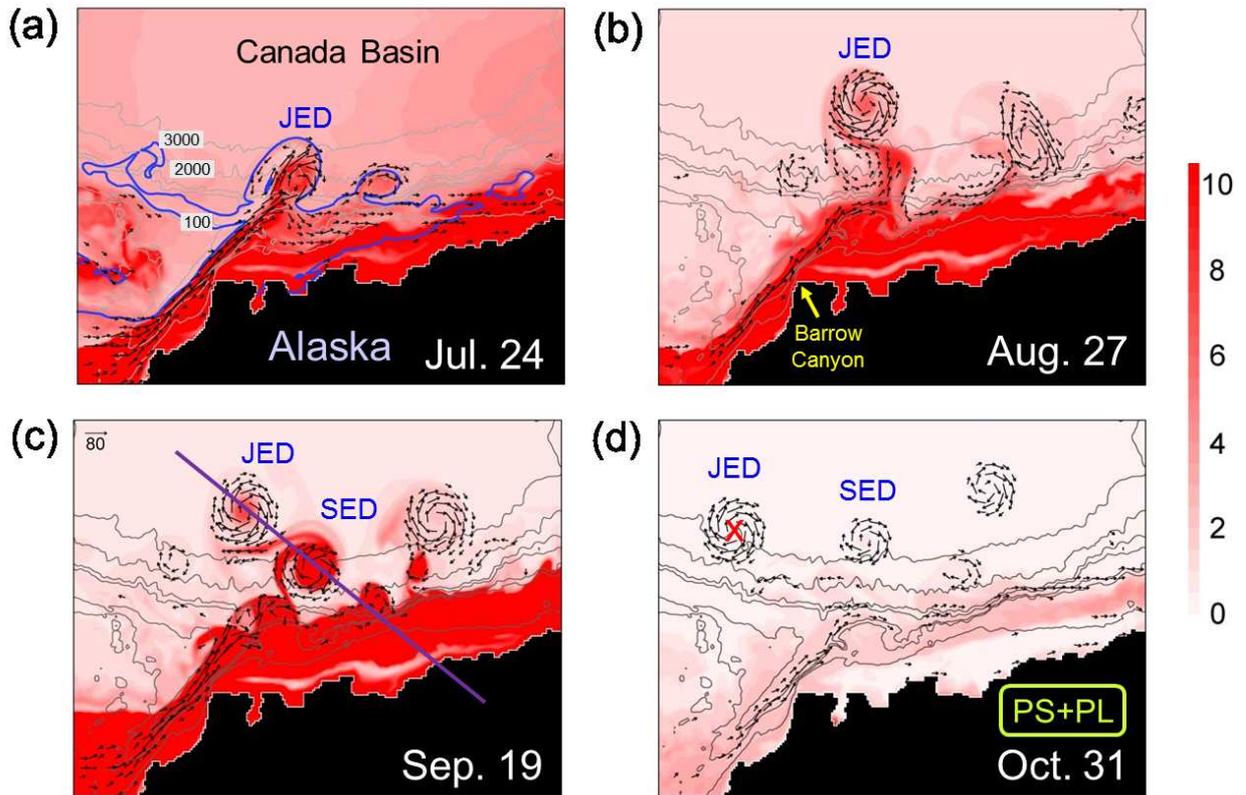


Figure 7. (Shade) Sum of G_{ppPS} and G_{ppPL} [$10^{-2} \mu\text{M day}^{-1}$] and (vectors) horizontal velocity at the ocean surface on (a) July 24, (b) August 27, (c) September 19, and (d) October 31. Unit vector of ocean velocity in (c) is 80 cm s^{-1} , and vectors of velocity below 20 cm s^{-1} are hidden. See Figure 6a for each displayed region and gray contours. Surface concentration of Pacific water tracer is 0.1 along blue contours in (a). Purple line across July Eddy (JED) and September Eddy (SED) in (c) is referred in Figures 9 and 10. Red cross in (d) is referred in Figure 11.

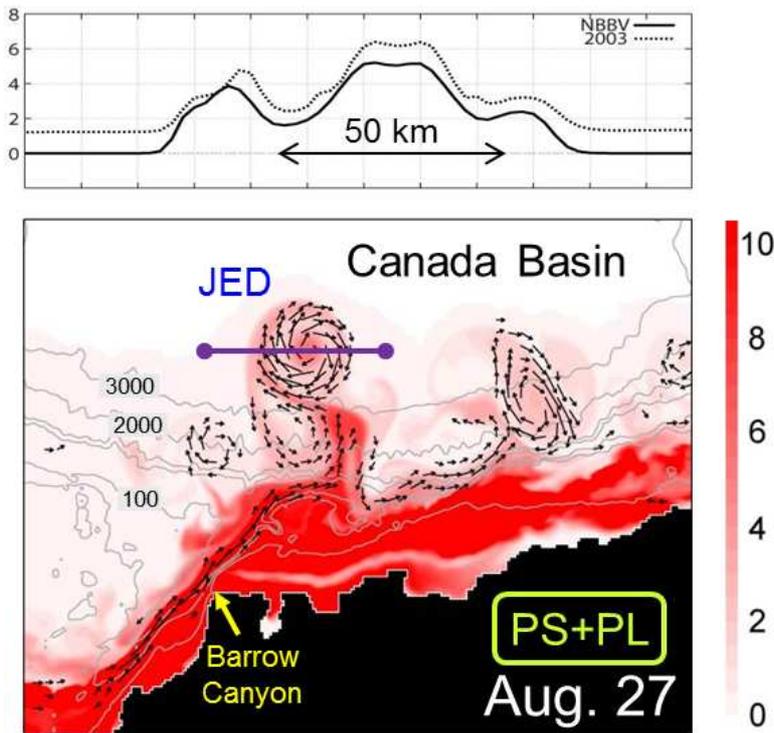


Figure 8. Same as Figure 7b, but in the NBBV case. Top figure shows the sum of surface G_{ppPS} and G_{ppPL} in the (solid line) 2003 and (dashed line) NBBV cases along a purple line in the bottom figure [$10^{-2} \mu\text{M day}^{-1}$].

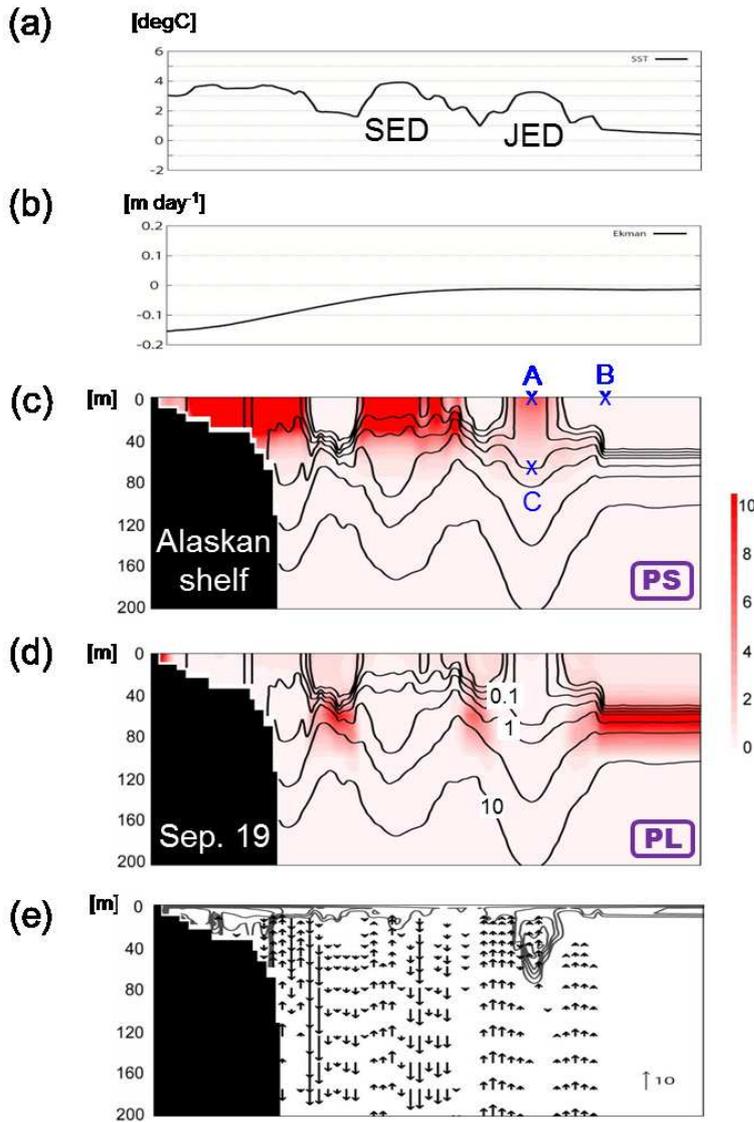


Figure 9. Profiles along a purple line in Figure 7c on September 19 in the 2003 case. (a) SST [$^{\circ}\text{C}$] and (b) Ekman upwelling/downwelling calculated from surface wind stress [m day^{-1}]. (c) GppPS and (d) GppPL [$10^{-2} \mu\text{M day}^{-1}$]. Nitrate concentration is overlaid by contours in (c-d) [μM]. (e) (contours) Vertical diffusivity diagnosed by the turbulence mixing scheme and (vectors) vertical velocity. Contour interval is $50 \text{ cm}^2 \text{ s}^{-1}$. Unit vector of velocity is 10 m day^{-1} , and vectors of velocity below 1 m day^{-1} are hidden. Locations of surface eddy center, surface eddy edge, and subsurface eddy center are labeled as A, B, and C, respectively.

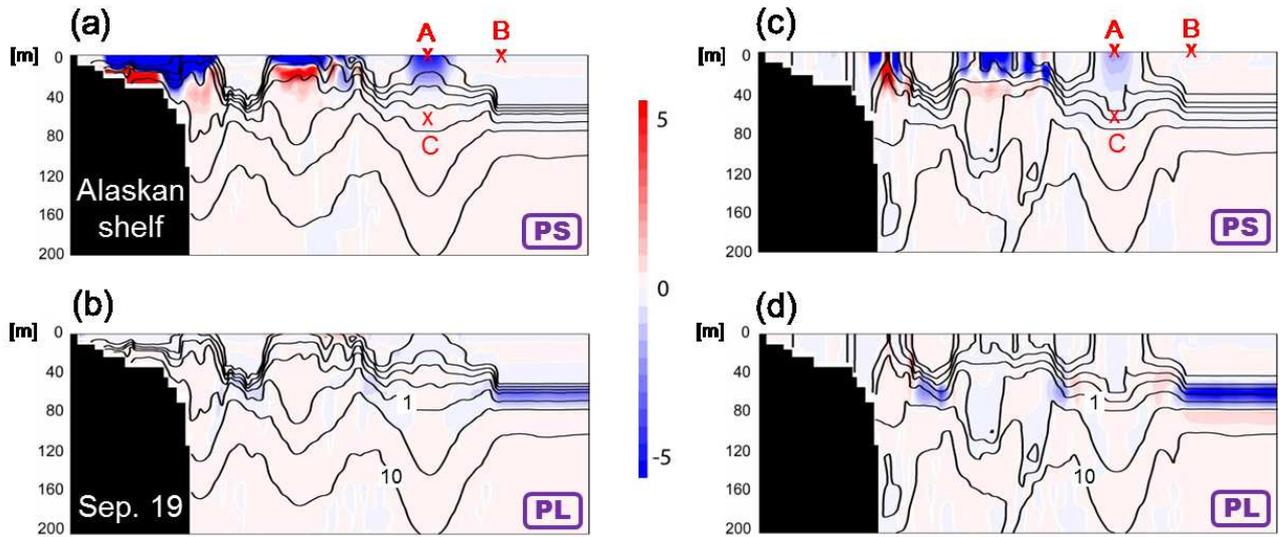


Figure 10. Same as (a,c) Figure 9c and (b,d) Figure 9d, but their anomalies in the (a-b) NVND and (c-d) NVNA cases, respectively, from the 2003 case.

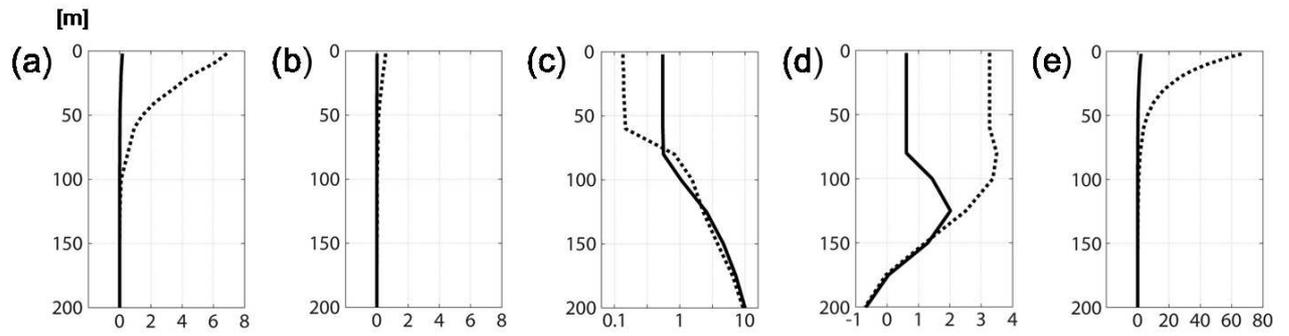


Figure 11. Vertical profiles at the JED center on (dashed line) September 19 and (solid line) October 31 in the 2003 case. See the exact location in Figure 7d. (a) GppPS and (b) GppPL [$10^{-2} \mu\text{M day}^{-1}$]. (c) Nitrate concentration with logarithm scale [μM], (d) water temperature [$^{\circ}\text{C}$], and (e) light intensity [W m^{-2}].

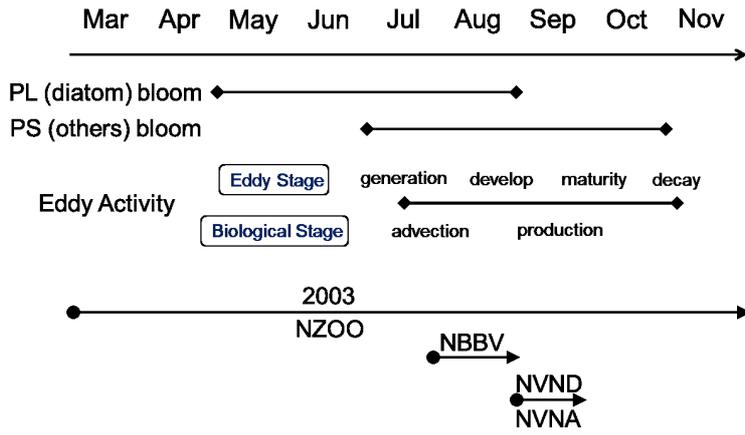


Figure 12. Time lag between phytoplankton bloom in the Chukchi shelf and eddy activity in the Beaufort shelf-break region. Integration periods in each experiment are also attached.

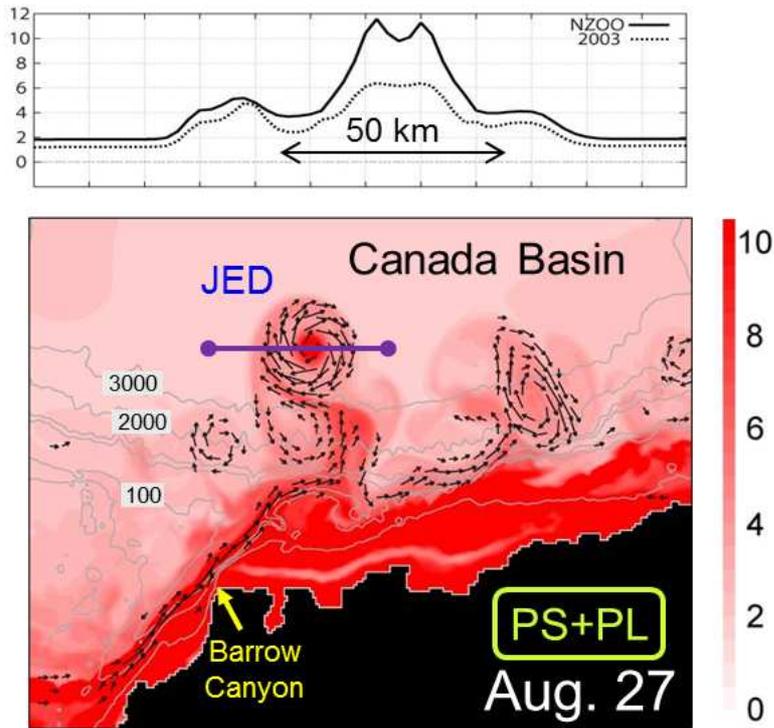


Figure 13. Same as Figure 8, but in the NZOO case.

Table 1. GppPS, GppPL [$10^{-2} \mu\text{M day}^{-1}$], PS, PL [μM], NO_3 , NH_4 , $\text{Si}(\text{OH})_4$ [$10^{-2} \mu\text{M}$], T [$^\circ\text{C}$], and I [W m^{-2}] at surface eddy center, surface eddy edge, and subsurface eddy center in Figures 9c-d and 10. Values of each term of photosynthesis formulation (Nit, Amn, Sil, Tmp, LA) are also presented.

| [2003] | | Surface Eddy Center | Surface Eddy Edge | Subsurface Eddy Center |
|--------------------------|-------|---------------------|-------------------|------------------------|
| GppPS | | 6.86 | 0.01 | 0.93 |
| GppPL | | 0.57 | 0.70 | 0.08 |
| PS | | 0.19 | 0.01 | 0.18 |
| PL | | 0.04 | 0.09 | 0.04 |
| NO_3 | (Nit) | 13.55 (0.15) | 0.20 (0.00) | 14.76 (0.16) |
| NH_4 | (Amn) | 8.06 (0.29) | 0.81 (0.04) | 9.01 (0.31) |
| $\text{Si}(\text{OH})_4$ | (Sil) | 6.59 (0.05) | 267.89 (0.70) | 6.92 (0.06) |
| T | (Tmp) | 3.27 (1.25) | 0.70 (1.05) | 3.27 (1.25) |
| I | (LA) | 66.26 (0.93) | 66.30 (0.93) | 3.81 (0.12) |

| [NVND] | | Surface Eddy Center | Surface Eddy Edge | Subsurface Eddy Center |
|--------------------------|-------|---------------------|-------------------|------------------------|
| GppPS | | 1.64 | 0.01 | 1.43 |
| GppPL | | 0.21 | 0.61 | 0.18 |
| PS | | 0.13 | 0.01 | 0.13 |
| PL | | 0.04 | 0.09 | 0.04 |
| NO_3 | (Nit) | 1.53 (0.02) | 0.14 (0.00) | 60.47 (0.36) |
| NH_4 | (Amn) | 3.19 (0.14) | 0.72 (0.03) | 24.96 (0.56) |
| $\text{Si}(\text{OH})_4$ | (Sil) | 2.64 (0.02) | 286.52 (0.71) | 16.01 (0.12) |
| T | (Tmp) | 3.27 (1.25) | 0.70 (1.05) | 3.27 (1.25) |
| I | (LA) | 66.63 (0.93) | 66.30 (0.93) | 4.44 (0.14) |

| [NVNA] | | Surface Eddy Center | Surface Eddy Edge | Subsurface Eddy Center |
|--------------------------|-------|---------------------|-------------------|------------------------|
| GppPS | | 5.12 | 0.01 | 0.74 |
| GppPL | | 0.44 | 0.71 | 0.06 |
| PS | | 0.17 | 0.01 | 0.16 |
| PL | | 0.04 | 0.09 | 0.04 |
| NO_3 | (Nit) | 8.95 (0.11) | 0.20 (0.00) | 9.53 (0.11) |
| NH_4 | (Amn) | 7.41 (0.27) | 0.82 (0.04) | 8.24 (0.29) |
| $\text{Si}(\text{OH})_4$ | (Sil) | 5.36 (0.04) | 260.37 (0.69) | 5.64 (0.05) |
| T | (Tmp) | 3.27 (1.25) | 0.70 (1.05) | 3.27 (1.25) |
| I | (LA) | 66.41 (0.93) | 66.30 (0.93) | 4.05 (0.13) |