Effects of storms on primary productivity and air-sea CO₂ exchange in the subarctic western North Pacific: a modeling study

M. Fujii¹,² and Y. Yamanaka²,³

¹Sustainability Governance Project, Center for Sustainability Science, Hokkaido University, N9W8, Kita-ku, Sapporo, Hokkaido 060-0809, Japan
²Graduate School of Environmental Science, Hokkaido University, N10W5, Kita-ku, Sapporo, Hokkaido 060-0810, Japan
³Frontier Research System for Global Change, 3173-25 Showa-machi, Kanazawa-ku, Yokohama, Kanagawa 236-0001, Japan

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Abstract. Biogeochemical responses of the open ocean to storms and their feedback to climate are still poorly understood. Using a marine ecosystem model, we examined biogeochemical responses to the storms in the subarctic western North Pacific. The storms in summer through early autumn enhance net community production by wind-induced nutrient injections into the surface waters while the storms in the other seasons reduce net community production by intensifying light limitation on the phytoplankton growth due to vertical dilution of the phytoplankton. The two compensating effects diminish the storm-induced annual change of net community production to only 1%. On the contrary, the storms reduce the annual oceanic uptake of the atmospheric CO₂ by 3%, resulting from storm-induced strong winds. Our results suggest that previous studies using climatological wind, sea level pressure, and CO₂ data probably overestimated the air-to-sea CO₂ influx during storms in the subarctic western North Pacific, and therefore, continuous high-frequency observations of these variables are required to reduce uncertainties in the global oceanic CO₂ uptake.

1 Introduction

Episodic atmospheric disturbances such as storms reduce incoming solar radiation, enhance wind-driven vertical mixing and upwelling, and decrease the temperature, in the surface waters (e.g. Iverson, 1977; Price, 1981; Greatbatch, 1985; Stramma et al., 1986; Cornillon et al., 1987; Sanford et al., 1987; Shay and Elsberry, 1987; Sakaida et al., 1998; Senjyu and Watanabe, 1999; Hong et al., 2003; Lee and Niller, 2003). Such physical changes are expected to impact marine biogeochemistry. For example, decreases in solar radiation and downward transport of phytoplankton out of the euphotic zone are likely to increase light limitation on the phytoplankton growth. On the other hand, wind-induced nutrient injections into the surface waters allow for potentially more growth of phytoplankton.

However, biogeochemical responses of the open ocean to storms are still poorly understood (e.g. Babin et al., 2004). In situ ship-based sampling has greatly enhanced our understanding of the biogeochemistry, but rough weather prevents sampling during storms. Only a few cruises and mooring buoys have serendipitously encountered episodic strong wind events (e.g. Marra et al., 1990; Bates et al., 1998a, b; Dickey et al., 1998; Nemoto et al., 1999; Wanninkhof et al., 2007). In most cases, sudden decrease in the sea surface temperature (SST) and abrupt change in the partial pressure of CO₂ in seawater (pCO₂sea) have been observed.

Recently satellite-derived estimates of wind speed, temperature, chlorophyll-a concentration, and nitrate concentration estimated from temperature in the surface waters have provided an opportunity to explore the physical and biological responses to episodic strong wind events (e.g. Stramma et al., 1986; Subrahmanyam et al., 2002; Lin et al., 2003; Babin et al., 2004; Davis and Yan, 2004; Platt et al., 2005; Son et al., 2006 and 2007). However, they cannot clarify whether the observed changes in chlorophyll-a concentration are the result of wind-induced nutrient injections or the result of horizontal advection of waters with high chlorophyll-a concentration. In addition, wind-induced change in carbon biogeochemistry, which is considered to greatly contribute to ecosystem dynamics and climate, cannot be estimated from observations by satellites at the present stage.
Several previous studies have demonstrated that ecosystem modeling allows us to estimate the episodic storm-induced biogeochemical responses at various spatial and temporal scales (e.g., McCreary et al., 2001; Soetaert et al., 2001; Kawamiya and Oschlies, 2004; Fujii et al., 2007; Wu et al., 2007). Therefore, the application of ecosystem models could significantly advance our understanding of the biogeochemical response to storm passage in the open ocean (Babin et al., 2004). However, the responses of the oceanic carbon cycling have not been discussed sufficiently because very few previous ecosystem models included the carbon cycling. Especially, the biogeochemical responses to storms have hardly been explored in the subarctic western North Pacific, which is considered one of the key regions impacting marine resources and uptake of anthropogenic CO\(_2\) (e.g., Sabine et al., 2004).

We therefore undertook this study to investigate the storm-induced biogeochemical responses in this region using a marine ecosystem model in which the carbon cycling is embedded. We applied the model to time-series Station KNOT (Kyodo North pacific Ocean Time-series; 44\(^\circ\) N, 155\(^\circ\) E) located in the subarctic western North Pacific. The subarctic western North Pacific is known as the oceanic region of highest incidence rate of explosively developing extra-tropical cyclones (e.g., Roebber, 1984; Yoshida and Asuma, 2004). The extra-tropical cyclones rapidly decrease the sea level pressure, sometimes as low as during typhoons and hurricanes, and possibly enhance abrupt physical processes in the ocean. In addition, Kataoka (2003) revealed that 41 typhoons reached to nearby Station KNOT for 30 years from 1961 to 1990. As a result, more than ten days are under storm events which are capable of deepening the mixed layer depth (MLD) and affecting the biogeochemistry on average each year at Station KNOT (Table 1). We investigate the biogeochemical responses to the storms, especially focusing on the impacts to the phytoplankton dynamics and the air-sea CO\(_2\) exchange.

### Table 1. Number of days on which the wind speed was more than 2\(\sigma\) or 3\(\sigma\) values from the 30-day running mean in each year from 1982 to 2000 at Station KNOT.

<table>
<thead>
<tr>
<th>Year</th>
<th>82</th>
<th>83</th>
<th>84</th>
<th>85</th>
<th>86</th>
<th>87</th>
<th>88</th>
<th>89</th>
<th>90</th>
<th>91</th>
<th>92</th>
<th>93</th>
<th>94</th>
<th>95</th>
<th>96</th>
<th>97</th>
<th>98</th>
<th>99</th>
<th>00</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt;2(\sigma)</td>
<td>10</td>
<td>14</td>
<td>15</td>
<td>8</td>
<td>22</td>
<td>19</td>
<td>13</td>
<td>12</td>
<td>12</td>
<td>12</td>
<td>14</td>
<td>15</td>
<td>10</td>
<td>11</td>
<td>15</td>
<td>9</td>
<td>10</td>
<td>16</td>
<td>249</td>
<td></td>
</tr>
<tr>
<td>&gt;3(\sigma)</td>
<td>0</td>
<td>4</td>
<td>2</td>
<td>3</td>
<td>5</td>
<td>1</td>
<td>2</td>
<td>2</td>
<td>1</td>
<td>3</td>
<td>0</td>
<td>0</td>
<td>3</td>
<td>1</td>
<td>2</td>
<td>0</td>
<td>0</td>
<td>2</td>
<td>34</td>
<td></td>
</tr>
</tbody>
</table>

2 Model description and experimental design

We use a 16-compartment (16-state variable) marine ecosystem model coupled with a one-dimensional physical model (Fig. 1; Fujii et al., 2002, 2005, 2007; Yamanaka et al., 2004). The ecosystem model is based on NEMURO (North Pacific Ecosystem Model for Understanding Regional Oceanography; e.g., Kishi et al., 2007). In this model, phytoplankton are categorized into two groups: diatoms (PL) and non-diatom small phytoplankton (PS) including coccolithophorids. Phytoplankton components utilize nitrate (NO\(_3\)) and ammonium (NH\(_4\)) in the process of photosynthesis and produce soft tissue in the form of particulate organic nitrogen (PON). Along with photosynthesis, diatoms utilize silicate (Si(OH)\(_4\)) to produce frustules of biogenic silica. Zooplankton are categorized into three groups: microzooplankton (ZS) including foraminifera, mesozooplankton (ZL) and predatory zooplankton (ZP). Seasonal vertical migration of mesozooplankton is taken into account: the mesozooplankton vertically migrate out of the model domain below 330 m-depth at the end of August, and 20% of the mesozooplankton return to the euphotic zone at the beginning of April every year, which is similar to the procedure used by previous studies (Kishi et al., 2001). Coccolithophorids and foraminifera produce hard laths of calcium carbonate (CaCO\(_3\)). Total alkalinity (TAlk) is calculated by balances of CaCO\(_3\), NO\(_3\) and NH\(_4\). Dissolved inorganic carbon (DIC) is calculated by balances of TAlk, NO\(_3\) and NH\(_4\) (with a carbon:nitrogen ratio of 6.625; Redfield et al., 1963). Effects of the other materials and ions on TAlk and DIC are relatively small and negligible. Partial pressure of CO\(_2\) at the sea surface \((p_{\text{CO}_2})_{\text{sea}}\) is calculated and the air-sea CO\(_2\) flux can be estimated by this model. Values of biogeochemical parameters in the model are the same as those of Fujii et al. (2007).

Air-sea CO\(_2\) flux is calculated in the model using the transfer velocity-wind speed relationships of Wanninkhof (1992) as follows:

\[
\text{Air-sea CO}_2 \text{ flux} = 0.31U^2 \sqrt{660/ScL} \left( (p_{\text{CO}_2})_{\text{sea}} - (p_{\text{CO}_2})_{\text{air}} \right),
\]

where \(U\) is the wind speed at 10 m height (m s\(^{-1}\)), \(Sc\) is the Schmidt number for CO\(_2\), expressed as:

\[
Sc = 2073.1 - 125.62\text{SST} + 3.6276\text{SST}^2 - 0.043219\text{SST}^3.
\]

\(L\) is the solubility of CO\(_2\) calculated from temperature and salinity (Weiss, 1974). The \((p_{\text{CO}_2})_{\text{air}}\) is the partial pressure of CO\(_2\) in the atmosphere and the monthly in situ observational data collected in the Shemya Island, Alaska, USA (SHM; 53\(^\circ\) N, 174\(^\circ\) E) are used (Keeling et al., 1982; Conway et al., 1994).

The model is driven by the wind and solar radiation at the sea surface, and the temperature and salinity at the surface.
and at the bottom of the model domain (330 m). We use the National Centers for Environmental Prediction (NCEP) objectively analyzed data (Kalney et al., 1996) for the daily (every six hours) wind and solar radiation at the sea surface, the Reynolds weekly data for the SST (Reynolds and Smith, 1995), and the KNOT time-series observations for the temperature at the bottom and the salinity at the surface and bottom (Fujii et al., 2002; Tsurushima et al., 2002). The model performance of reproducing the observed results were double-checked by referring to observational data obtained by the joint Japan-Canada monitoring program using ships-of-opportunity (e.g. Nojiri et al., 1999; Wong et al., 2002a, b; Zeng et al., 2002; Chierici et al., 2006).

We define storm events as those in which the wind speed is more than 2σ value from a 30-day running mean. To examine effects of storms on biogeochemistry, we carried out two experiments, namely Exp-1 (with storms) and Exp-2 (without storms). In Exp-1, the model is driven by the winds including more than 2σ values. In Exp-2, the model is driven by the winds in which more than 2σ values are filtered out. The simulation is calculated from 1982 to 2000, and the 19-year monthly-mean model results are presented below.

3 Results and discussions

Most of storm events last for no more than one day, and there are no storm events that last for more than five days (Table 2). All the storms have the wind speed of more than 10 (m s⁻¹) with the highest wind speed of 26.3 (m s⁻¹) for the 19 years. However, the storm events which have the wind speed of more than 25 (m s⁻¹) are very limited (Table 3), which is different from situations in the subtropical regions in which typhoons and hurricanes have higher wind speeds more than 30 (m s⁻¹) (e.g. Bates et al., 1998a, b; D’Asaro and McNeil, 2007; Wanninkhof et al., 2007).

The model reproduces well the observed seasonal changes of both physical environments and biogeochemistry at Station KNOT (Fig. 2). The strong wind in winter causes the MLD, defined as the depth at which the vertical diffusive coefficient is 1.0×10⁻⁴ m² s⁻¹, to deepen dramatically to more than 100 m in late winter (Fig. 2a, b). The strong wind and pCO₂sea elevated by storm-induced DIC (normalized to a constant salinity of 35 psu) injections into the surface waters cause tremendous sea-to-air CO₂ efflux in late winter (Fig. 2a, d, e, f). The net community production is low in winter because of strong light limitation on the phytoplankton growth due to vertical dilution of the phytoplankton, along with low irradiance in this season (Fig. 2c). The net community production is relatively high in spring through early autumn when the light limitation on the phytoplankton growth is alleviated by high irradiance and stratification.

Table 2. Number of storms which last for 1, 2, 3 and 4 days, respectively, from 1982 to 2000 at Station KNOT.

<table>
<thead>
<tr>
<th></th>
<th>1 day</th>
<th>2 days</th>
<th>3 days</th>
<th>4 days</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt;2σ</td>
<td>156</td>
<td>34</td>
<td>7</td>
<td>1</td>
</tr>
<tr>
<td>&gt;3σ</td>
<td>23</td>
<td>4</td>
<td>1</td>
<td>0</td>
</tr>
</tbody>
</table>
of the surface waters in these periods. The oceanic region functions as a sink of the atmospheric CO$_2$ in spring through autumn, supported by large biological uptake of CO$_2$, as reported by previous observations (e.g. Tsurushima et al., 2002).

By comparing the model results between Exp-1 (with storms) and Exp-2 (without storms), we find that the storms contribute to air-sea exchange of CO$_2$ (Fig. 2f). The sea-to-air CO$_2$ efflux is enhanced by the storms in late spring, early summer and autumn (Fig. 2f-2). For example, the $p$CO$_{2\text{sea}}$ abruptly increases by 51 $\mu$atm during a storm passage in mid-June 1994 in which the wind speed is more than 3$\sigma$ value from the 30-day running mean (Fig. 3b). The $p$CO$_{2\text{sea}}$ reaches one of its annual peaks during the storm, although the annual maximum generally appears in late winter in the subarctic western North Pacific (Fig. 2e; Tsurushima et al., 2002). The sea-to-air CO$_2$ flux abruptly increases during the storm passage (Fig. 3c), accounting for 3% of the annual sea-to-air efflux in 1994 for two days. The sudden increase in $p$CO$_{2\text{sea}}$ results from the increase in DIC by 22 mmolC m$^{-3}$ (Fig. 3a) because of the storm-induced vertical mixing, which dominates the counteracting effect of the

<table>
<thead>
<tr>
<th>Wind speed [m s$^{-1}$]</th>
<th>10$^{\sim}$15 m$^{-1}$</th>
<th>15$^{\sim}$20 m$^{-1}$</th>
<th>20$^{\sim}$25 m$^{-1}$</th>
<th>$&gt;$25 m$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt;2$\sigma$</td>
<td>6488 h (40.0%)</td>
<td>8850 h (54.5%)</td>
<td>888 h (5.5%)</td>
<td>6 h (0.0%)</td>
</tr>
<tr>
<td>&gt;3$\sigma$</td>
<td>795 h (23.4%)</td>
<td>1752 h (51.5%)</td>
<td>847 h (24.9%)</td>
<td>6 h (0.2%)</td>
</tr>
</tbody>
</table>

Table 3. Hours for which storms had the wind speed of 10$^{\sim}$15 m$^{-1}$, 15$^{\sim}$20 m$^{-1}$, 20$^{\sim}$25 m$^{-1}$ and (4) $>$25 m$^{-1}$, respectively, from 1982 to 2000 at Station KNOT. Percentage shows the ratio to total duration of the storms.
storm-induced sea surface cooling of 0.8°C. This is consistent with a result based on mooring buoy data deployed in the East China Sea during the passage of three typhoons in 1995 that for changes in $pCO_{2sea}$ the effect of DIC increase dominated that of SST decrease (Nemoto et al., 1999).

On the contrary, the sea-to-air efflux is reduced slightly by the storms in late summer, because the net community production is enhanced by the storm-induced nutrient injections into the surface waters (Fig. 2c, f-2). The model result of the decrease in the $pCO_{2sea}$ in late summer (Fig. 2e) is consistent with a ship-based result of Bates et al. (1998b), who observed a sudden decrease in $pCO_{2sea}$ in the Sargasso Sea in the subtropical gyre during hurricane Felix passage in August, 1995. Wanninkhof et al. (2007) also showed a decrease in $pCO_{2sea}$ by 12 $\mu$atm over a period of hurricane Frances passage in the Caribbean Sea in September, 2004. However, the cause of the storm-induced decrease in the sea- to-air CO$_2$ efflux is different from these studies in the subtropical regions: They observed only a slight increase in DIC, and the decrease in $pCO_{2sea}$ was followed by a decrease in SST. The difference in responses of $pCO_{2sea}$ to storms between this study and Bates et al. (1998b) and Wanninkhof et al. (2007) results from differences in the upper ocean structure and hydrographic conditions between the subarctic and subtropical gyres. In the subarctic gyre, the pycnocline is relatively shallow, and storms occasionally mix the surface waters with DIC-rich deep waters below the pycnocline. In the subtropical gyre, on the other hand, the pycnocline is usually deep and storm-induced vertical mixing does not reach to the deep waters. The magnitude of the sea surface cooling also depends on the upper ocean structure and hydrographic conditions, as noted in previous studies (e.g. Cornillon et al., 1987; Sakaida et al., 1998). Storm-induced sea surface cooling is more efficient in the subtropical ocean than in the subarctic ocean because of the permanently warm surface waters in the subtropical ocean.

The model result shows that the annual air-to-sea CO$_2$ influx is lower in Exp-1 (211 [mmolC m$^{-2}$ yr$^{-1}$]) than in Exp-2 (218 [mmolC m$^{-2}$ yr$^{-1}$]) (Table 4), and therefore, that the storms reduce the annual oceanic uptake of the atmospheric CO$_2$ efflux by 3%. It is possible that previous studies, in which the flux was calculated using the climatological wind, sea level pressure (which is necessary in converting $xCO_{2air}$ to $pCO_{2air}$) and $pCO_{2sea}$ data (e.g. Takahashi et al., 2002; Tsurushima et al., 2002), underestimated the effect of storm events on air-sea CO$_2$ exchange and overestimated the role of the entire subarctic western North Pacific in taking up atmospheric CO$_2$.

Our model results also suggest that storm events, tropical or extra-tropical, could potentially have a large effect on
net community production in the other seasons is reduced through early autumn because of the storm-induced nutrient injections into the surface waters. On the contrary, the storms enhance the net community production in summer because of large biological uptake of CO$_2$ during storm passage in various oceanic regions to elucidate which effects dominate the storm-induced air-sea CO$_2$ exchange. Therefore, to reduce uncertainties in the global oceanic CO$_2$ uptake, changes in these parameters caused by episodic atmospheric disturbances should be measured continuously.

While climate changes associated with global warming may influence the frequency and intensity of storms (e.g. Emanuel, 1987, 2005; Beersma et al., 1997; Saunders and Harris, 1997; Sugi et al., 2002; Geng and Sugi, 2005; Yoshimura and Sugi, 2005; Webster et al., 2005; Yoshimura et al., 2006), storm-induced biogeochemical activity may also contribute to climate. Therefore, to predict future climate change, it is essential that we elucidate the biogeochemical responses to storms. Although it is difficult to conduct direct observations during the rough weather, we would very much like to have high-frequency, in-situ biogeochemical observations with which we could conduct more accurate simulations.
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