Air-sea carbon flux simulated in an upper ocean biogeochemical model

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Introduction

The partial pressure of CO_2 (pCO_2) in the surface ocean shows spatial and temporal variations larger than the atmospheric counterparts, caused by the mixed-layer development and biological productivity. In the subpolar regions of the oceans, observations show that the oceanic pCO_2 and the concentration of total inorganic carbon are higher during winter than summer. The ocean is a CO_2 source for the atmosphere in winter and a CO_2 sink in summer. This seasonal evolution may be mainly accounted for by the winter mixing of deep waters, which are rich in total inorganic carbon, and the biological production in summer. On the other hand, in the subtropical region, the ocean is a CO_2 sink in the winter and a CO_2 source in the summer. This variation is governed primarily by temperature.

Method

In this study, we focus the mechanism of carbon cycle in the surface ocean. We employ a three-dimensional ocean model including the bulk mixed-layer model [*Kraus and Turner,* 1967; *Niiler and Kraus,* 1977]. Air-sea carbon flux is calculated using a this model, in which the distributions of total inorganic carbon, phosphate, and carbonate alkalinity are simulated. The model has the upper 300 m of the ocean. The model is initialized with historical data in the North Pacific and forced by European Centre for Medium-Range Weather Forecasts (ECMWF) data and National Centers for Environmental Prediction (NCEP)/ National Center for Atmospheric Research (NCAR) data. Velocity structures are obtained from wind driven Ekman transport and geostrophic ocean currents diagnostically induced by the observed density structures. This model retains the following processes: air-sea carbon flux at the sea surface, the biological pump, the development of mixed layer, horizontal and vertical advections and horizontal and vertical diffusions. The mixed layer development occurs by cooling and wind stress through the sea surface. The entrainment velocity, W_e, is determined from the energy balance. A gain of potential energy at the base of the mixed layer balances with a sum of the turbulent kinetic energy and the potential energy input at the surface. When the atmospheric cooling changes to the heating (e.g. summer), the entrainment velocity is set to zero.

The equations for conservation of total inorganic carbon C_m, carbonate alkalinity A_m, and phosphate P_m in the mixed layer are

$$
\begin{aligned}
\frac{\partial C_m}{\partial t} &= W_e\frac{\triangle C}{h} - \frac{F_{CO_2}}{h} - \frac{106}{1}S_{P_m} \\
&\quad - \left(u + u_e\frac{h_e}{h}\right)\frac{\partial C_m}{\partial x} - \left(v + v_e\frac{h_e}{h}\right)\frac{\partial C_m}{\partial y} \\
&\quad + K_H\left(\frac{\partial^2 C_m}{\partial x^2} + \frac{\partial^2 C_m}{\partial y^2}\right), \qquad (1)
\end{aligned}
$$

$$
\begin{aligned}
\frac{\partial A_m}{\partial t} &= W_e\frac{\triangle A}{h} - \frac{106}{1}\frac{1}{10}S_{P_m} \\
&\quad - \left(u + u_e\frac{h_e}{h}\right)\frac{\partial A_m}{\partial x} - \left(v + v_e\frac{h_e}{h}\right)\frac{\partial A_m}{\partial y} \\
&\quad + K_H\left(\frac{\partial^2 A_m}{\partial x^2} + \frac{\partial^2 A_m}{\partial y^2}\right), \qquad (2)
\end{aligned}
$$

$$
\begin{aligned}
\frac{\partial P_m}{\partial t} &= W_e\frac{\triangle P}{h} - S_{P_m} \\
&\quad - \left(u + u_e\frac{h_e}{h}\right)\frac{\partial P_m}{\partial x} - \left(v + v_e\frac{h_e}{h}\right)\frac{\partial P_m}{\partial y} \\
&\quad + K_H\left(\frac{\partial^2 P_m}{\partial x^2} + \frac{\partial^2 P_m}{\partial y^2}\right), \qquad (3)
\end{aligned}
$$

where $\triangle C$, $\triangle A$ and $\triangle P$ are the discontinuities in total inorganic carbon, carbonate alkalinity, and phosphate at the base of the mixed layer. u, v and w are the horizontal and vertical velocity components, u_e and v_e are the Ekman velocities, w_e is the vertical velocity (Ekman pumping or suction) associated with convergence or divergence of the Ekman transport, h is the mixed-layer depth, h_e is the Ekman depth, and K_H is the horizontal diffusion ($= 10^2$ m^2 s^{-1}). The formula for the air-sea carbon flux, F_{CO_2} is

$$F_{CO_2} = k_s\gamma(pCO_{2oc} - pCO_{2atm}), \qquad (4)$$

where k_s (cm h^{-1}) is the piston velocity [*Wanninkhof*,1992], γ (mol kg^{-1} atm^{-1}) is the solubility of CO_2 in seawater, and pCO_{2oc} (μatm) and pCO_{2atm}(μatm) are the partial pressures of CO_2 in the surface ocean and in the atmosphere, respectively. The oceanic pCO_2 is related to total inorganic carbon and carbonate alkalinity under the assumptions of chemical equilibrium.

The biological pump S_{P_m} is considered as a function of phosphate concentration in the mixed layer :

$$S_{P_m} = \frac{L}{T_e}P(t)\frac{30}{h}, \qquad (5)$$

where L is the coefficient at which the biological production occurs ($L = 1$) or does not occur ($L = 0$), and the T_e is the e-folding time scale of phosphate reduction (or biological production). The coefficient L is 1 in summer and 0 in winter, with the transition times given as a function of the latitudes. The values of L and T_e control the magnitude and the period of the biological production, respectively. The values of T_e are set as 60 days west of 165°E and 180 days east of 165°E.

The equations for total CO_2 C_l, carbonate alkalinity A_l, and phosphate P_l in the lower layer are

$$\begin{aligned}\frac{\partial C_l}{\partial t} &= \frac{106}{1} S_{P_l} \\ &\quad -u\frac{\partial C_l}{\partial x} - v\frac{\partial C_l}{\partial y} - (w + w_e)\frac{\partial C_l}{\partial z} \\ &\quad +K_H\left(\frac{\partial^2 C_l}{\partial x^2} + \frac{\partial^2 C_l}{\partial y^2}\right) + K_Z\left(\frac{\partial^2 C_l}{\partial z^2}\right),\end{aligned} \tag{6}$$

$$\begin{aligned}\frac{\partial A_l}{\partial t} &= \frac{106}{1}\frac{1}{10} S_{P_l} \\ &\quad -u\frac{\partial A_l}{\partial x} - v\frac{\partial A_l}{\partial y} - (w + w_e)\frac{\partial A_l}{\partial z} \\ &\quad +K_H\left(\frac{\partial^2 A_l}{\partial x^2} + \frac{\partial^2 A_l}{\partial y^2}\right) + K_Z\left(\frac{\partial^2 A_l}{\partial z^2}\right),\end{aligned} \tag{7}$$

$$\begin{aligned}\frac{\partial P_l}{\partial t} &= S_{P_l} \\ &\quad -u\frac{\partial P_l}{\partial x} - v\frac{\partial P_l}{\partial y} - (w + w_e)\frac{\partial P_l}{\partial z} \\ &\quad +K_H\left(\frac{\partial^2 P_l}{\partial x^2} + \frac{\partial^2 P_l}{\partial y^2}\right) + K_Z\left(\frac{\partial^2 P_l}{\partial z^2}\right),\end{aligned} \tag{8}$$

where K_Z is the vertical diffusion coefficient ($= 10^{-5}$ m^2 s^{-1}). The vertical profile of the mineralization S_{P_l} is given by

$$S_{P_l} = \frac{S_{P_m} h}{1000} exp(-(z - h)/1000), \tag{9}$$

where z is the depth of water. Thus, the organic carbon flux reduces exponentially to 1/e at 1000 m.

Results

We describe the result of difference of the effect of the biological production between the eastern and western regions. The horizontal map of annual mean air-sea carbon flux is shown in Figure 1 (a). Positive values denote the region where the ocean is a CO_2 sink in one year. Especially, in the vicinity of the Japan coast, the ocean is a strong CO_2 sink because the strong biological production occurs in summer. In the limited position of the subpolar region, the ocean is a CO_2 source where vertical mixing of deep waters, which are rich in total inorganic carbon, occurs rapidly in winter.

In the mean balance, the biological pump is nearly canceled by the advection effects in the subpolar region, and the minor residual takes the atmospheric CO_2 to the ocean (Figure 1(b)). The advection effects in the subpolar region are mainly due to the Ekman upwelling, increasing CO_2 by transporting the high CO_2 from the lower layer. To the subtropical region, the large CO_2 in the subpolar region is transported by the southward Ekman flow. All three components are smaller in the subtropical region. The seasonal variability of air-sea carbon flux is controlled by the biological pump in spring and summer and vertical mixing to the mixed layer in winter along with the secondary effect of the Ekman upwelling which is stronger in winter (Figure 2).

References

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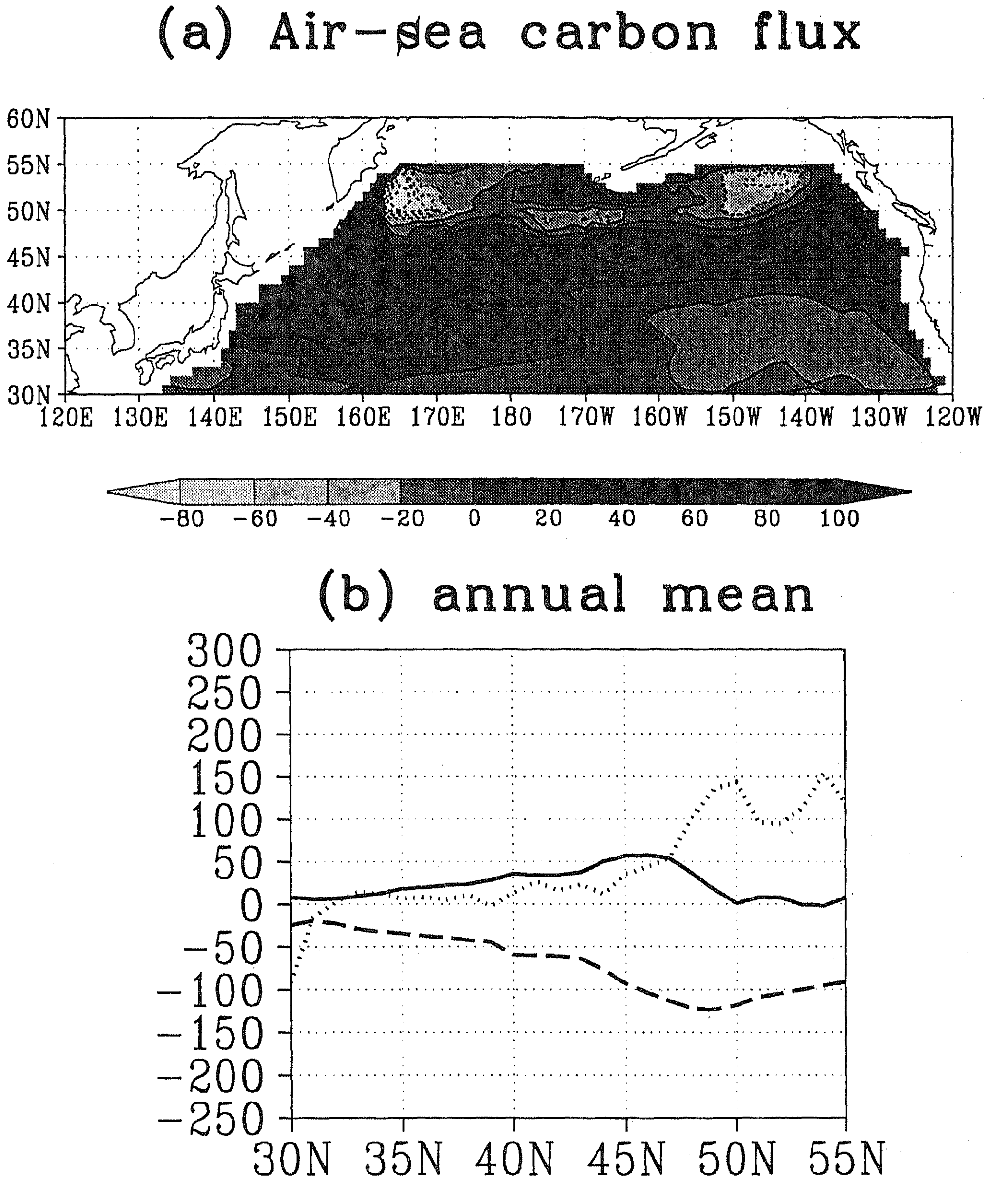


Figure 1. Annual mean distribution of (a) air-sea carbon flux (mgC m^{-2} d^{-1}) in the North Pacific. The contour interval is 20 mgC m^{-2} d^{-1}. Positive values are where the ocean is a CO_2 sink, and negative values are where the ocean is a CO_2 source. (b) Annual mean carbon flux (mgC m^{-2} d^{-1}), which vertical integrated from sea surface to 200 m depth, along the latitude. The solid line denotes air-sea carbon flux. The short dashed line denotes biological pump which exports organic carbon for the lower layer. The dots line denotes the advections which transports carbon in the surface layer.

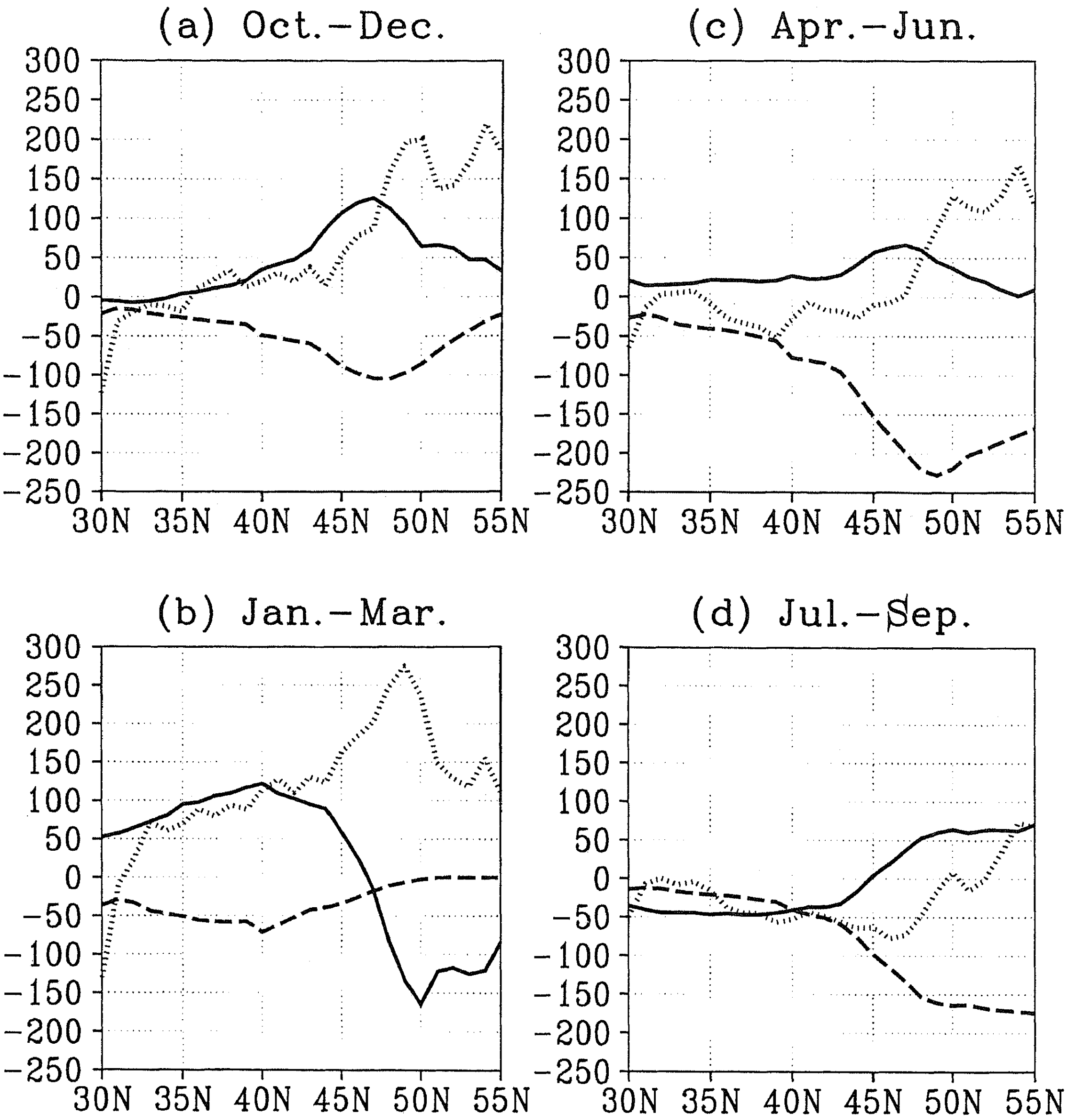
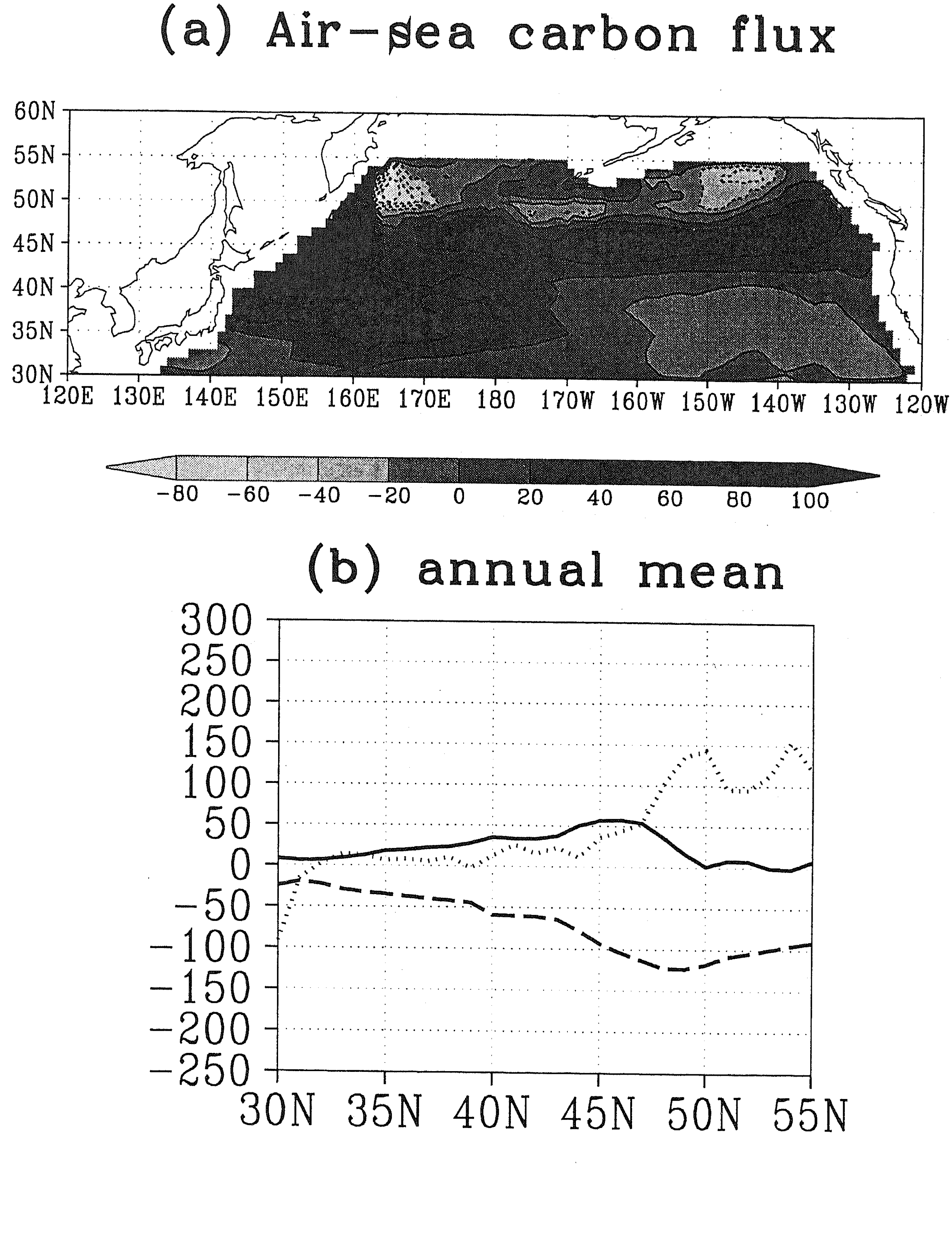
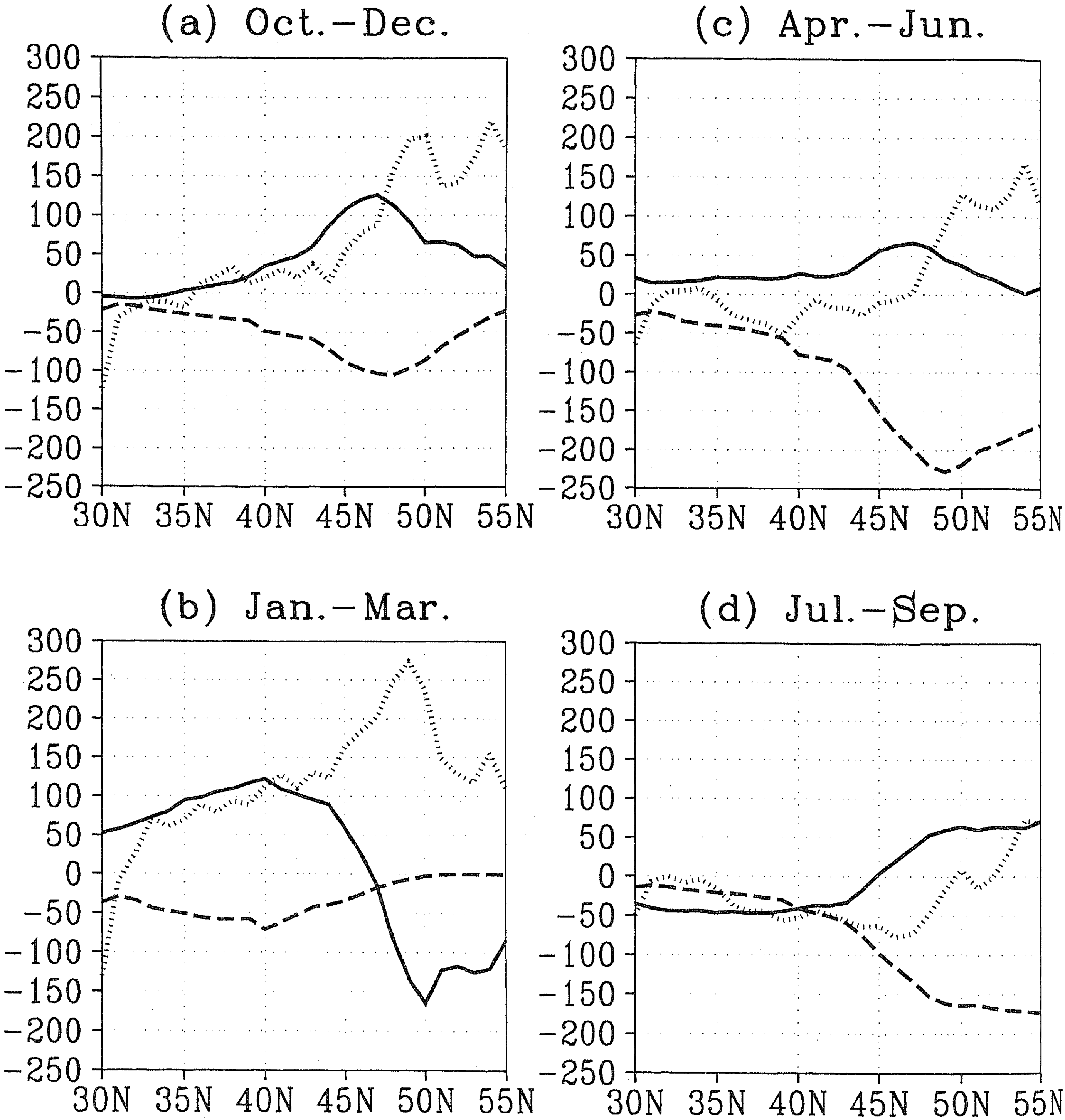


Figure 2. Same as Figure 1 (b) but for the seasonal mean carbon flux (mgC m^{-2} d^{-1}).


(a) Air-sea carbon flux
60N
55N
50N
45N
40N
35N
30N
120E 130E 140E 150E 160E 170E 180 170W 160W 150W 140W 130W 120W
-80 -60 -40 -20 0 20 40 60 80 100
(b) annual mean
300
250
200
150
100
50
0
-50
-100
-150
-200
-250
30N 35N 40N 45N 50N 55N


(a) Oct.-Dec.
(c) Apr.-Jun.
(b) Jan.-Mar.
(d) Jul.-Sep.
300
250
200
150
100
50
0
-50
-100
-150
-200
-250
30N
35N
40N
45N
50N
55N