The Impact of Surface Mixing on the Arctic River Water Distribution and Stratification in a Global Ice–Ocean Model

YOSHIKI KOMURO

Japan Agency for Marine-Earth Science and Technology, Kanagawa, Japan

(Manuscript received 6 February 2013, in final form 20 January 2014)

ABSTRACT

The impact of oceanic vertical mixing on the near-surface vertical structure of the Arctic Ocean is investigated in a global ice–ocean model with a passive tracer. Lowering surface background vertical diffusivity and ignoring the effects of surface wave breaking under sea ice improves the model simulation of the horizontal Arctic river water distribution. This improvement is largely responsible for the freshening of the Arctic surface salinity in the model. Although these modifications in the model vertical mixing scheme are applied over the whole global ocean, the change in the surface salinity over the Arctic is larger than that in the rest of the global ocean by one to two orders of magnitude. In contrast, when a reduced background vertical diffusivity is used at all depths, the Arctic vertical salinity stratification is improved below the surface as well as in the surface layer, but the vertical structure and deep circulation in the rest of the global ocean are also strongly affected. Weaker surface vertical mixing in the Arctic Ocean also causes sea ice to thicken even without changes in the parameters for the sea ice component.

1. Introduction

The Arctic and global climate systems interact with each other. Global climatic changes, such as global warming, can have a major impact on the Arctic; examples include the recent decline in Arctic sea ice (Stroeve et al. 2007; Kwok and Rothrock 2009) and Arctic amplification (Serreze and Barry 2011). At the same time, changes in the Arctic affect the global climate via the atmosphere and ocean (e.g., Komuro and Hasumi 2005, 2007; Jahn et al. 2010; Overland and Wang 2010).

Numerical modeling is key to deepening our understanding of the interaction between the Arctic and global climates. Large uncertainties remain, however, in present climate models, particularly in the Arctic. There is a wide spread in the sea ice extent simulated by the Coupled Model Intercomparison Project phase 3 (CMIP3; Meehl et al. 2007) and phase 5 (CMIP5; Taylor et al. 2012) models (e.g., Stroeve et al. 2012; Massonnet et al. 2012), while the stratification and circulation of the upper Arctic Ocean is poorly reproduced in ice–ocean models (Holloway et al. 2007; Karcher et al. 2007; Proshutinsky et al. 2011). Hence, a better representation of the Arctic in global climate models is necessary.

Some studies have focused on the representation of the low-salinity surface mixed layer and underlying cold halocline. Rudels et al. (1996) proposed that low-salinity shelf water, including much of the river runoff, covers the top of the water column, limiting the winter convection and eventually producing the observed vertical structure. Zhang and Steele (2007) employed extremely low background vertical diffusivity, \( K_V = 10^{-6} \text{ m}^2 \text{ s}^{-1} \), in an ice–ocean coupled model to achieve a better simulation of stratification in the upper layer and Atlantic Water layer. They pointed out that realistic stratification over these layers leads to the model reproducing the observed cyclonic circulation in the Atlantic Water layer. The representation and parameterization of subgrid-scale brine rejection has also been found to improve model simulations (Matsumura and Hasumi 2008; Nguyen et al. 2009).

This study focuses on the representation of the Arctic river water outflow and its impact on surface stratification. It is found that specifying weak surface mixing in a global ice–ocean coupled model has a considerable
impact on the Arctic river water distribution and consequently on the Arctic stratification. The remainder of this paper is organized as follows. The model description is provided in section 2 and the simulation results are shown in section 3. A discussion and conclusions are presented in section 4.

2. Model description

The ice–ocean coupled model used in this study is Center for Climate System Research (CCSR) Ocean Component Model, version 4.5 (COCO4.5), which has also been employed as the ice–ocean component for the Model for Interdisciplinary Research on Climate, version 5 (MIROC5; Watanabe et al. 2010). A brief description is given here. For detailed information on the oceanic and sea ice components, see Watanabe et al. (2010) and Komuro et al. (2012), respectively.

The model domain is global, with a horizontal grid size of approximately 60 km in the Arctic Ocean, and 17 vertical levels in the uppermost 250 m. The model geometry in the northern high latitudes is shown in Fig. 1a. COCO4.5 employs a second-order moment advection scheme (Prather 1986) that significantly reduces numerical diffusion (Tatebe and Hasumi 2010). The sea ice component includes a subgrid-scale sea ice thickness distribution, which greatly improves Arctic atmosphere–ocean heat exchange (Komuro and Suzuki 2012).

A turbulent closure scheme of Noh and Kim (1999), which is a derivative of the level-2.5 turbulent closure scheme of Mellor and Yamada (1982), is used as a surface mixed layer parameterization. Based on vertical shear and stratification, the level-2.5 turbulent closure scheme predicts turbulent kinetic energy (TKE) and diagnoses vertical diffusivity ($K_V^b$) and vertical viscosity. A difference of this scheme from that of Mellor and Yamada (1982) is to give TKE input at the sea surface, which represents the effect of near-surface mixing by surface wave breaking. This surface TKE flux is proportional to the cube of the friction velocity $u_*$ defined by

$$u_* = \sqrt{\frac{\tau}{\rho_0}},$$

where $\tau$ is wind stress and $\rho_0$ is the density of seawater. Vertical diffusivity $K_V$ actually used for the integration is determined as the larger value of $K_V^b$ and a prescribed background vertical diffusivity, $K_{bV}$:

$$K_V = \max(K_V^b, K_{bV}).$$

The vertical profile of $K_{bV}$ (Fig. 2) follows the case III profile of Tsujino et al. (2000), by using the global overturning circulation that is well reproduced.

The control experiment (CTRL) with the standard settings described above, generally reproduces the large-scale temperature and salinity structures, flow patterns, and horizontal sea ice distribution. However, the CTRL results have some biases in the Arctic Ocean. In particular the mixed layer is deeper than in reality, and its improvement is one of the motivations of this study. The horizontal extent of the Beaufort gyre is also too small. In addition, the relatively coarse resolution puts limitations on the reproducibility of features such as coastal currents.

Three sensitivity experiments are performed: low $K_V^b$ at all depths (LALL), low $K_V^b$ at the surface (LSFC), and low $K_V^b$ at surface with no wave-breaking mixing.
LALL employs a background value of $K^b_V = 10^{-6} \text{m}^2\text{s}^{-1}$ at all depths, which is approximately an order of magnitude smaller than that for CTRL in the surface layer. This $K^b_V$ profile is the same as that of the case KPP0.01 of Zhang and Steele (2007), which was an optimal setting for reproducing the Arctic Ocean stratification in their study. The value of $K^b_V$ for LSFC is $10^{-6} \text{m}^2\text{s}^{-1}$ in the uppermost 22.5 m, gradually increasing to $10^{-5} \text{m}^2\text{s}^{-1}$ between 22.5- and 50-m depth, and is the same as in CTRL below 50 m. These vertical $K^b_V$ profiles are depicted in Fig. 2. LSNW uses the same background $K^b_V$ profile as LSFC, but additionally assumes no surface wave breaking and resultant near-surface mixing where the surface is covered by sea ice. In LSNW, only wind stress on open water contributes to $\tau$ of Eq. (1) in calculating $u_s$; in the other experiments, $\tau$ also includes the stress between ice and seawater. Note that the treatment of horizontal momentum flux into the ocean through the ice is the same in all cases.

In all the experiments, the model is integrated over 250 years from an initial state at rest with climatological temperature and salinity. The model is driven by a daily atmospheric climatology based on the Common Ocean Reference Experiment (CORE) normal year forcing (Large and Yeager 2004). The river runoff given to the ocean in the northern high latitudes is depicted in Fig. 1b. Following the method of Large and Yeager (2004), the runoff is spread near the modeled coastline. Sea surface salinity (SSS) is restored to the monthly climatology for all the regions except the Arctic Ocean and Norwegian Sea with a time scale of 10 days over the 2.5-m-thick uppermost layer. This SSS restoring is not applied to the region to the south of 55°N after the 200th model year (Fig. 1c) in order to evaluate an impact of the $K^b_V$ settings on surface salinity of the global ocean. A passive tracer is added in the last 50 years of the integrations: its source is the same as the river runoff flux into the Arctic Ocean (the region enclosed by the thick white line in Fig. 1a), and it resets to zero outside the Arctic Ocean. Hence, the tracer can be considered as the Arctic river water tracer.

3. Results

a. Salinity and river water distribution in the Arctic Ocean

Salinity cross sections of the uppermost 250 m across the Arctic basin are shown in Fig. 3 for the four runs and for the Polar Science Center Hydrographic Climatology (PHC; Steele et al. 2001). The climatological data show that surface water with low salinity (around 31 psu) extends out from the continental shelves. In CTRL, the volume of low-salinity water is very small and it is trapped at the shelf break (Fig. 3b). In the other three experiments (LSFC, LSNW, and LALL), low-salinity water spreads out from the shelves, although to a smaller extent than in the climatology (Figs. 3c–e). The vertical salinity gradient at around 100 m in LALL is sharper and closer to the PHC data than in the other cases because of the low background vertical diffusivity and is consistent with previous modeling studies with low vertical diffusion (Zhang and Steele 2007; Nguyen et al. 2009).
Sea surface salinity in and around the Arctic Ocean (Fig. 4) shows similar behavior. Low-salinity water spreads out from the shelves in LSFC, LSNW, and LALL, especially in the latter two cases (Figs. 4c–e). In these experiments, fresher surface water is also conspicuous in the coastal areas, where river water is added to the sea surface. The low-salinity water in the central Arctic Ocean seems to be connected to the fresh coastal water. These spatial patterns are also found in the PHC data (Fig. 4a).

Figure 5 depicts the annual-mean fraction of the Arctic river water tracer along the section indicated in Fig. 3. The fraction is the lowest in CTRL, where the

![Figure 3](image-url)
FIG. 4. (top) Sea surface salinity in the northern high latitudes, and (bottom) differences from the PHC. (a) PHC, (b) CTRL, (c) LSFC, (d) LSNW, and (e) LALL.
maximum appears below the sea surface (Fig. 5a). In the other cases, the fraction is higher and the maximum is at the surface (Figs. 5b–d). Estimates of river water fraction based on ship observations (Guay et al. 2009; Roeske et al. 2012) indicate that the river water is concentrated within the top 50 m, particularly in the Canadian Basin, and that the fraction exceeds 10% in some regions. Thus, the river water distribution in LSFC, LSNW, and LALL is more realistic than that in CTRL, with LALL giving the best results.

The vertical flux of the river water tracer at 50-m depth and its difference from that in CTRL are shown in Fig. 6. In this figure, positive (negative) flux means upward (downward) river water transport. Table 1 summarizes the downward river water transport integrated over the areas of the Arctic Ocean categorized by bottom depth. Note that the total river water input into the Arctic Ocean is 0.108 Sv (1 Sv = 10^6 m^3 s^-1). Intense localized downward river water transport is found over the Siberian shelf and the coastal shelf area of the Beaufort Sea in CTRL (Fig. 6a). It means strong vertical mixing of the river water on these shelf areas. This vertical mixing on the shelves is suppressed in the other cases (Figs. 6b–d). The downward river water transport integrated over the region with depths less than 500 m, which corresponds approximately to the shelf area (cf. Fig. 1a), is smaller in LSFC, LSNW, and LALL compared with that in CTRL (Table 1). In contrast, the downward transport in the rest of the Arctic Ocean is larger in the former three cases than in CTRL (Table 1). These changes in vertical transport suggest enhancement of horizontal spreading of river water in the surface layer in LSFC, LSNW, and LALL. This enhancement is consistent with the surface enrichment of the river water in the central basin (Figs. 5b–d).

Annual-mean salinity and river water fraction averaged over the Arctic Ocean are calculated for each case, and the differences between CTRL and each of the other three cases are shown in Figs. 7a and 7b. The largest differences are found from the surface to 30 m in all cases. Significant changes below 100 m are found only in LALL; however, the surface magnitude in LALL is similar to that in LSNW. The differences in river water fraction (Fig. 7b) are also largest in the surface layer. The impact of the changes in the river water

---

**Fig. 5.** Annual-mean fraction of the Arctic river water tracer along the section shown in Fig. 3 for (a) CTRL, (b) LSFC, (c) LSNW, and (d) LALL.
fraction on the salinity is also indicated in Fig. 7b by assuming that the salinity of the seawater is 31 psu, which roughly corresponds to the Arctic sea surface salinity. Comparing Figs. 7a and 7b shows that the increase in the river water fraction in the surface layer is largely responsible for the surface freshening in these three cases.

Changes are also found in sea ice thickness among the four cases, although the parameter settings for the sea ice component of the model are the same in all the experiments. Figure 8a depicts Arctic summer (July–September) sea ice thickness in CTRL. The thickness averaged over the Arctic Ocean is 242 cm. Figures 8b–d show differences in the thickness between CTRL and the other three cases. The sea ice thickness over most of the Arctic ice extent in all the cases. The fresher the surface salinity, the thicker the summer sea ice: the summer thickness averaged over the Arctic Ocean compared with that for CTRL is 17, 28, and 39 cm greater for LSFC, LSNW, and LALL, respectively.

In summary, the representation of the surface low-salinity water is improved in the Arctic Ocean in all the cases with lower background $K_b^V$. LALL is the most realistic, but LSNW also performs well, particularly in the surface layer. The distribution of the river water tracer suggests that the concentration of river water in the surface layer leads to these improvements. In addition, the Arctic sea ice thickness is also sensitive to the change in background $K_b^V$. 

Fig. 6. (a) Annual-mean vertical flux of the Arctic river water tracer at 50-m depth for CTRL. Positive values mean upward flux. (b) Difference between LSFC and CTRL vertical river water tracer flux. (c) As in (b), but for LSNW. (d) As in (b), but for LALL.
b. Comparison of the impacts in the Arctic Ocean and the global ocean

Figure 7c shows differences in annual-mean salinity averaged over the rest of the global ocean, excluding the Arctic Ocean. The impacts on global ocean salinity are smaller than those for the Arctic Ocean (Fig. 7a) by at least an order of magnitude in all three cases. The impacts in LSFC and LSNW are almost identical. At the surface, the global impact on salinity is approximately 50 times smaller than the Arctic impact in LALL and is more than two orders of magnitude smaller in LSFC and LSNW.

The horizontal distribution of \( K^V \) is responsible for the contrast between the global and Arctic salinity changes. Figure 9 shows diagnosed winter and summer \( K^V \) at 22.5-m depth for CTRL, LSFC, and LSNW. In this figure, hatching indicates areas where \( K^V \) parameterized according to Noh and Kim (1999) is smaller than the background value \( K^b \). At this level the background \( K^b \) is set to \( 13.4 \times 10^{-6} \) m² s⁻¹ for CTRL and \( 1.0 \times 10^{-6} \) m² s⁻¹ for the other cases (Fig. 2b). In most areas, \( K^V \) is larger than the background coefficient \( K^b \), meaning that the \( K^V \) overwrites the background value in these regions. Consequently, the change in background \( K^b \) has an impact only in regions where it is larger than \( K^V \) and adopted as the model \( K^V \) (i.e., hatched areas in Fig. 9).

The Arctic Ocean is the most obvious of such regions, probably due to the strong surface density stratification, originating from the large amount of river runoff and additional sea ice meltwater in summer. In CTRL and LSFC, the areas where \( K^b \) is adopted as \( K^V \) are similar in each case (Figs. 9a–d), but the smaller \( K^b \) in LSFC causes the salinity impact described in the previous section. In LSNW, lower TKE input across the sea ice cover leads to smaller \( K^V \) there and a consequent wider area of \( K^V \) (Figs. 9e,f), which has a larger impact on river water distribution and salinity in the Arctic Ocean.

On the other hand, the low background \( K^b \) in LALL affects not only the Arctic surface salinity but also the global ocean below the surface layer. The change in globally averaged salinity is much larger in LALL than in the other cases (Fig. 7c). The deep ocean circulation is also changed considerably in LALL; for example, the northward Antarctic Bottom Water transport across the

---

**Table 1. Downward river water transport across 50-m depth integrated over the areas of the Arctic Ocean categorized by bottom depth. Unit is Sverdrups (1 Sv = 10⁶ m³ s⁻¹).**

<table>
<thead>
<tr>
<th>Case</th>
<th>Bottom depth &lt; 500 m</th>
<th>Bottom depth &gt; 500 m</th>
</tr>
</thead>
<tbody>
<tr>
<td>CTRL</td>
<td>0.050</td>
<td>0.020</td>
</tr>
<tr>
<td>LSFC</td>
<td>0.040</td>
<td>0.032</td>
</tr>
<tr>
<td>LSNW</td>
<td>0.038</td>
<td>0.036</td>
</tr>
<tr>
<td>LALL</td>
<td>0.032</td>
<td>0.031</td>
</tr>
</tbody>
</table>

---

**Fig. 7.** (a) Salinity difference with respect to CTRL as a function of depth, averaged over the Arctic Ocean for LSFC, LSNW, and LALL. (b) As in (a), but for river water fraction. The corresponding change in salinity is also shown (see text for details). (c) As in (a), but the salinity is averaged over the global ocean excluding the Arctic Ocean. Note that the line of LSFC is almost identical to that of LSNW in (c).
Pacific equator is 5.5 Sv in LALL, which is smaller than the 8.2–8.9 Sv in the other cases (not shown).

4. Discussion and conclusions

The above results demonstrate that the surface salinity of the Arctic Ocean is sensitive to the following model settings for vertical mixing. First, employing a low surface background $K_b^v$ suppresses surface mixing and improves the horizontal Arctic river water distribution. This improvement is largely responsible for the freshening of the Arctic surface salinity in the model used in this study. This result suggests that ocean models with a surface background $K_b^v$ of $O(10^{-5})$ m$^2$ s$^{-1}$ underestimate the horizontal spreading of Arctic river waters. Second, comparison of the results of LSFC and LSNW shows that ignoring the TKE flux input through the ice-covered ocean surface weakens Arctic surface mixing and leads to an enhancement of river water spreading, which also causes surface salinity improvement in this study’s model. Ocean surface mixed layer parameterizations such as that proposed by Noh and Kim (1999) are intended to add to the surface TKE input contributions from processes such as surface wave breaking, which are suppressed for an ice-covered sea surface. A very small value of vertical diffusivity ($2 \times 10^{-7}$ m$^2$ s$^{-1}$) during ice-covered months was estimated from observations made at a mooring at a seasonally ice-covered location in the Arctic Ocean (Rainville and Woodgate 2009). Such studies justify the TKE flux treatment of the present study. Third, comparison of LSFC and LALL shows that weak vertical mixing below the surface layer gives
sharper vertical stratification below the mixed layer. The surface and subsurface vertical salinity gradients averaged over the Arctic Ocean of LALL are the most realistic of all four cases (not shown). This is consistent with previous Arctic modeling studies with low background vertical diffusivity (Zhang and Steele 2007; Nguyen et al. 2009).

The weak surface vertical mixing in the Arctic Ocean also causes sea ice to thicken as a consequence of ice–ocean interaction. The relation between the surface

---

**FIG. 9.** Parameterized vertical diffusion coefficient $K^V$ at 22.5-m depth for (a) CTRL in winter (January–March), (b) CTRL in summer (July–September), (c) LSFC in winter, (d) LSFC in summer, (e) LSNW in winter, and (f) LSNW in summer. Areas where the parameterized $K^V$ is smaller than the background value $K^V_b$ are hatched. Note that $K^V_b$ at this level is $13.4 \times 10^{-6} \text{m}^2 \text{s}^{-1}$ for CTRL, and $1.0 \times 10^{-6} \text{m}^2 \text{s}^{-1}$ for the other cases.
salinity freshening and the increase in sea ice thickness suggests that the increase occurs by suppression of the heat flux below the mixed layer by a sharper pycnocline, which is realistic over much of the Arctic Ocean (e.g., Aagaard et al. 1981; Shaw et al. 2009).

Although LALL reproduces sharp vertical salinity gradients in the Arctic surface and subsurface layers, it also has the largest change in global stratification. Moreover, $K_V^{b}$ in LALL, $10^{-6}$ m$^2$s$^{-1}$ at all depths, is too small compared with a thermoline diffusivity of $O(10^{-5})$ m$^2$s$^{-1}$ based on observational estimates (e.g., Gregg 1989; Ledwell et al. 1993) and a globally averaged abyssal value of $O(10^{-4})$ m$^2$s$^{-1}$ inferred by Munk (1966) and Munk and Wunsch (1998). Some previous modeling studies (e.g., Hasumi and Suginohara 1999; Tsujino et al. 2000) reported that extremely low $K_V^{b}$ in the deep ocean caused unrealistically weak deep circulation. Thus, the parameter settings used in LALL are not suitable for global models, particularly for long integration times.

On the other hand, the effects of the parameter settings used in LSFC and LSNW are confined mainly to the Arctic, although they are uniformly applied to the global ocean. Therefore, the parameter choices in these cases, particularly the combination of the low surface background $K_V^{b}$ and suppression of TKE input in ice- covered regions, can improve the simulation of the Arctic Ocean without large changes over the rest of the global ocean and without any change in parameter values for sea ice models.

The present results show that surface background $K_V^{b}$ of $O(10^{-6})$ m$^2$s$^{-1}$ is actually applied only to the Arctic Ocean. This suggests that the surface background $K_V^{b}$ should represent that in the Arctic, not in the global ocean. Previous studies that observed internal wave fields in the Arctic Ocean (Levine et al. 1985; D’Asaro and Morison 1992; Halle and Pinkel 2003; Rainville and Woodgate 2009) showed weak mixing compared with that in lower latitudes. However, the number of such observations, especially in the surface layer under sea ice cover, is limited. Thus, more observations of Arctic Ocean mixing are required to identify an appropriate value of surface background $K_V^{b}$ for global models with horizontally uniform background mixing, as is the case with present ocean models. Such observations are also meaningful to understand more suitable treatment for near-surface mixing under sea ice cover, which is neglected in LSNW as an extreme assumption.

The sensitivity of Arctic ice thickness in this study might be underestimated, since the atmospheric boundary condition used for ice–ocean coupled models constrains the distribution of sea ice. Hence, a logical next step would be to study the impact of weak surface mixing in the Arctic on an atmosphere–ice–ocean coupled model. The horizontal distribution and spreading processes of water masses other than river water, such as the Pacific water, which also has an important role in Arctic stratification and an effect on sea ice thickness (e.g., Steele et al. 2004; Shimada et al. 2006), should also be investigated. Such studies would contribute to more reliable simulations and projections from climate models.

Acknowledgments. I acknowledge two anonymous reviewers for their constructive comments on this paper. Prof. Hiroyasu Hasumi and Dr. Hiroaki Tatebe made constructive comments and suggestions. This study was supported by the GRENE Arctic Climate Change Research Project conducted by the Ministry of Education, Culture, Sports, Science and Technology of the Japanese Government. Figures in this paper were produced using the GFD-DENNOU graphics library.

REFERENCES


—, and —, 2007: Effects of variability of sea ice transport through the Fram Strait on the intensity of the Atlantic


