

Revisiting mesoscale eddy genesis mechanism of nonlinear advection in a marginal ice zone

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Abstract

A three-dimensional (3-D) ocean model is coupled with a two-dimensional (2-D) sea ice model, to revisit a nonlinear advection mechanism, one of the most important mesoscale eddy genesis mechanisms in the marginal ice zone. Two-dimensional ocean model simulations suggest nonlinear advection mechanism is more important when the water gets shallower. Instead of considering the ocean as barotropic fluid in the 2-D ocean model, the 3-D ocean model allows the sea ice to affect the current directly in the surface layer via ocean-ice interaction. It is found that both mesoscale eddy and sea surface elevation are sensitive to changes in a water depth in the 3-D simulations. The vertical profile of a current velocity in 3-D experiments suggests that when the water depth gets shallower, the current move faster in each layer, which makes the sea surface elevation be nearly inverse proportional to the water depth with the same wind forcing during the same time. It is also found that because of the vertical motion, the magnitude of variations in the sea surface elevation in the 3-D simulations is very small, being only 1% of the change in the 2-D simulations. And it seems the vertical motion to be the essential reason for the differences between the 3-D and 2-D experiments.

Key words: nonlinear advection, mesoscale eddy, marginal ice zone, ocean-ice interaction

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1 Introduction

The marginal ice zone (MIZ) is a transition zone between the open ocean and packed sea ice (Johannessen et al., 1987). A typical MIZ in the circumpolar Antarctica during winter extends over 200 km from the edge of the packed ice to the open ocean (Wadhams and Holt, 1991). The MIZ in other regions, however, is usually much smaller. Frazil/pancake ice, which is the major component in the MIZ, is typically 1–2 m in width and thus allows ignorance of Young's modulus of elasticity. Mesoscale eddies in the ocean are typically 20–40 km horizontally and extends hundred meters in the vertical. Most of the eddies rotate cyclonic, with a maximum orbital velocity of 0.5–0.7 m/s and can be easily distinguished in the synthetic aperture radar images (Dumont et al., 2011; Johannessen O M et al., 1987). The mesoscale eddies always play an important role in a meridional heat transport at mid-high latitude (Yang and Dai, 2015). However, the generation mechanism of the mesoscale eddies still remains unclear. With observational results, Johannessen et al. (1987) suggested four possible mechanisms: (1) nonlinear advection, (2) topographic control, (3) barotropic instability, and (4) baroclinic instability.

Previous studies (Hibler, 1979; Williams et al., 2013a, b) have employed numerical modeling to investigate the generation mechanism of the mesoscale eddies. Two-dimensional (2-D) ocean models (Røed and O'Brien, 1983; Häkkinen, 1986; Liu et al., 1993) consider the ocean as a barotropic fluid and thus the velocity does not vary in the vertical direction. On the other hand,

three-dimensional (3-D) ocean models consider the ocean as a multiple-layer fluid, and allow velocity changes with a depth.

When we establish a numerical ocean-ice system, many physical processes should also be considered. As a wind stress drives sea ice more efficiently than the current, sea ice always moves faster than the current (Røed and O'Brien, 1983). This discrepancy in the velocity causes an interfacial stress (McPhee, 1975). When gravity waves propagate into the sea ice, the wave energy is dissipated, and offers the sea ice wave radiation stress (Liu and Mollo-Christensen, 1988; Liu et al., 1991; Wadhams et al., 1988, 2002; Squire et al., 1995; Squire, 2007). Generally, an ice internal stress resists an ice convergence and rigidity, which resists deformation and convergence (Hibler, 1979; Hunke and Dukowicz, 1997; Hunke and Zhang, 1999).

Given that the sea ice moves faster than the ocean current, the interfacial stress acts to transfer the momentum from the sea ice to the current. This results in the faster movement of current beneath the sea ice than that in the open ocean, and causes Ekman pumping at the ice boundary. Using a wavy ice edge, Häkkinen (1986) reproduced mesoscale eddies with a horizontal scale comparable to observations (20–40 km). She suggested that the nonlinear advection mechanism was crucial for the generation of mesoscale eddies when the water depth was 25 m. For the rest mechanisms: a strong jet, which is produced by the wind or wave, may lead to strong barotropic instability and generating the mesoscale eddies. If there is a strong horizontal shear in the initial stratification, the unstable state can produce a great num-

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ber of available potential energy (APE) and eddy available potential energy (EAPE). With a small perturbation, the mesoscale eddies can grow quickly, because an eddy kinetic energy (EKE) can be generated from the EAPE; this is the so-called baroclinic instability mechanism. If the bottom topography has a great horizontal gradient, the water column can be shrunk or stretched, because the Coriolis parameter does not change in the mesoscale motion, and the relative vorticity will change according to the absolute vorticity conservation law, which may induce the mesoscale eddies. This is the topographic controlled mechanism (Gula et al., 2015).

In this study, we revisited the role of the nonlinear advection mechanism (Häkkinen, 1986) in the generation of mesoscale eddies. In the pioneering work by Häkkinen (1986), she reproduced the mesoscale eddies with a wavy ice boundary and a northeastward wind stress. However, the sea surface elevation was as much as 10 m high in her model, which is unrealistic. Usually, only a storm can induce such a huge sea surface elevation. We suspect such a high elevation was needed, because some physical process is missing in her 2-D ocean model. Will a 3-D ocean model overcome this problem? Suppose that the sea surface height (SSH) can be smaller in the 3-D model for the mesoscale eddies to appear, what magnitude should the SSH be? Since Häkkinen (1986) assumed the nonlinear advection might be the dominant mechanism in shallow water, we are interested in how the nonlinear advection changes with the water depth. The purpose of this work is to revisit the nonlinear advection mechanism with the 3-D ocean model. We ask: can we reproduce the mesoscale eddies using the nonlinear advection mechanism in the 3-D ocean model? Also we will try to find difference between 2-D and 3-D model results and the reasons behind the differences. We also intend to find out how the nonlinear advection mechanism works differently when the water gets shallower.

This paper is arranged as follows: in Section 2, we introduce a model and experiments; in Section 3, we analyze the results from both 2-D and 3-D ocean models, both of which are coupled to a 2-D sea ice model, and in Section 4, we provide a summary and identify the future research topics.

2 Model and experiments

The model used in this study is the regional oceanic modeling system (ROMS) of University of California, Los Angeles (UCLA). It is a coupled ocean-ice-wave system. The ocean component model (Shchepetkin and McWilliams, 2005) solves 3-D hydrostatic primitive equations in vertical hybrid z -sigma (Lemarié et al., 2012) and horizontal curvilinear coordinates. The system uses the Wentzel-Kramers-Brillouin (WKB) wave model (Uchiyama et al., 2011), and a one-layer frazil/pancake ice model. This system also includes current-surface gravity wave interaction (McWilliams et al., 2004; Lane et al., 2007), sea ice-surface gravity wave interaction, and K-profile parameterization (KPP) for vertical mixing (Large et al., 1994).

In this study, the wave part is turned off, since the nonlinear advection mechanism has nothing to do with the waves. Thus, we give a theoretical frame as follows.

(1) Sea ice model:

$$\rho_i M_i \left(\frac{\partial u_i}{\partial t} + u_i \frac{\partial u_i}{\partial x} + v_i \frac{\partial u_i}{\partial y} \right) = \rho_i M_i f v_i + A_i (\tau_{ai}^x + \tau_{wi}^x) + F_i^x, \quad (1)$$

$$\rho_i M_i \left(\frac{\partial v_i}{\partial t} + u_i \frac{\partial v_i}{\partial x} + v_i \frac{\partial v_i}{\partial y} \right) = -\rho_i M_i f u_i + A_i (\tau_{ai}^y + \tau_{wi}^y) + F_i^y, \quad (2)$$

$$\frac{\partial A_i}{\partial t} + \frac{\partial(A_i u_i)}{\partial x} + \frac{\partial(A_i v_i)}{\partial y} = 0 \quad (0 \leq A_i \leq 1), \quad (3)$$

$$\frac{\partial M_i}{\partial t} + \frac{\partial(M_i u_i)}{\partial x} + \frac{\partial(M_i v_i)}{\partial y} = 0. \quad (4)$$

(2) Three-dimensional ocean model:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} = -g \frac{\partial \eta}{\partial x} + f v + (1 - A_i) \tau_{aw}^x - A_i \tau_{wi}^x, \quad (5)$$

$$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} = -g \frac{\partial \eta}{\partial y} - f u + (1 - A_i) \tau_{aw}^y - A_i \tau_{wi}^y, \quad (6)$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \quad (7)$$

$$\frac{\partial \eta}{\partial t} + \vec{v}_{\text{baro}} \cdot \nabla \eta = 0. \quad (8)$$

(3) Two-dimensional ocean model:

$$\frac{\partial U}{\partial t} + \frac{\partial(U^2/H)}{\partial x} + \frac{\partial(UV/H)}{\partial y} = -\frac{g^*}{2} \frac{\partial H^2}{\partial x} + f V + ((1 - A_i) \tau_{aw}^x - A_i \tau_{wi}^x) / \rho_w + A_H \nabla^2 U, \quad (9)$$

$$\frac{\partial V}{\partial t} + \frac{\partial(UV/H)}{\partial x} + \frac{\partial(V^2/H)}{\partial y} = -\frac{g^*}{2} \frac{\partial H^2}{\partial y} - f U + ((1 - A_i) \tau_{aw}^y - A_i \tau_{wi}^y) / \rho_w + A_H \nabla^2 V, \quad (10)$$

$$\frac{\partial H}{\partial t} + \nabla \cdot (\vec{v}_{\text{baro}} H) = 0. \quad (11)$$

In the above equations u , v and w is current velocities in the x , y and z directions, respectively; $\vec{v}_{\text{baro}}$ is a barotropic velocity vector, which equals to $\vec{v}$ in the 2-D ocean model; u_i and v_i are ice velocities in the x and y directions, respectively; η is the sea surface elevation; f is the Coriolis parameter, here we use $f=0.00014$; $M_i = A_i h_i$, is the ice mass; where A_i ($0 \leq A_i \leq 1$) is the ice concentration and h_i is the ice thickness; F_i is the ice internal stress; τ_{ai}^x and τ_{ai}^y are wind stresses exerted on sea ice in the x and y directions, respectively; similarly, τ_{aw}^x and τ_{aw}^y are wind stresses exerted on currents in the x and y directions, respectively; τ_{wi}^x and τ_{wi}^y are interfacial stresses between ocean and ice in the x and y directions, respectively, which can be written as $\vec{\tau}_{wi} = c_{wi} |\vec{v} - \vec{v}_i| (\vec{v} - \vec{v}_i)$, where c_{wi} is a drag coefficient between ocean water and ice; $\vec{v}$ and $\vec{v}_i$ are current and sea ice velocity vectors, respectively. In the ice free region ($A_i=0$), the external forcing to the current is the wind stress only.

The system is configured on a domain of 100 km×70 km, with horizontal resolution 1 km × 1 km (following Häkkinen, 1986). In the vertical direction, the water depth is set at 400 m (deep water) and 80 m (shallow water). To avoid influence by stratification or available potential energy, the temperature, salinity and density are uniformed and constant everywhere. To compare with previous work, we designed four sensitivity experiments, 2-D ocean model with 400 m depth (Exp 2-D400), 2-D ocean model with 80 m depth (Exp 2-D80), 3-D ocean model with 400 m depth (Exp 3-D400), and 3-D ocean model with 80 m depth (Exp 3-D80). The SSH, the current velocity and the ice velocity are set to 0 at the initial state. The ice thickness is 2 m, and the ice distribution is set as two rectangles (Fig. 1, shading). For the forcing condition, the northeastward wind with a speed of 1.8 m/s (Fig. 1, vector) is

used in the full domain. To simplify the boundary condition, both east-west periodic boundary and north-south periodic boundary are applied. All the simulations were run for three model days, and we take the results at the end of the third day for analysis.

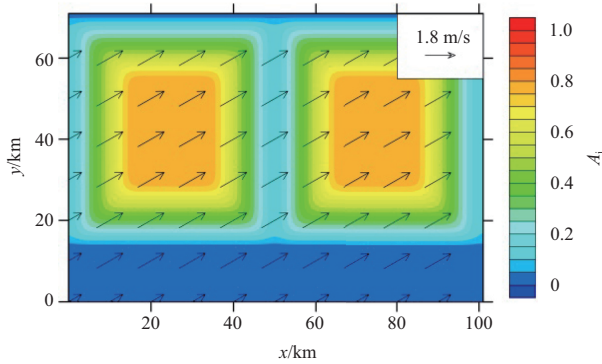


Fig. 1. Initial sea ice distribution (shading) and external wind forcing (vector) over the domain.

3 Results

3.1 Two-dimensional experiments

As the wind forcing is exerted on the whole domain, both sea ice and currents gain a momentum over time. After 3 d wind forcing, the sea ice distribution still remains two rectangles, though smoother at the ice edge (Figs 2a and b, shading). Sea ice moved eastward uniformly with a peak velocity about 3.6 cm/s (Figs 2a and b, vector), which implies that the interfacial stress between the current and sea ice has little influence on sea ice with other external forcing. An eastward velocity induces southward Coriolis force (Eq. (2)), which counteracts the northward component of the wind stress.

The efficiency of momentum transfer between the atmosphere and sea ice is twice larger than that between the atmosphere and the ocean. As a result, the sea ice moves faster than

the current. By the current-sea ice interaction, the current beneath the sea ice moves faster than that in the open ocean. Thus, when the sea ice moves towards the open ocean, there is a water mass convergence at the ice edge, the sea surface elevation increases (Figs 3a and b, warm color), and vice versa (Figs 3a and b, cold color). Compared with Häkkinen (1986), where the sea surface elevation can reach as high as 10 m, the sea surface elevation can go up by 0.15 m. This is because our wind forcing is only 1/44 of that in her experiments. On the other hand, we run the model for twice of her time but use three point two times water depth. We reproduced her results with the same condition. When eddies are generated, the current is cyclonic (anticyclonic) in the northern (southern) half of the domain. If we reduce the ocean depth from 400 to 80 m, which means there is less water mass, the current velocity will increase by the same momentum. Thus, the vorticity, as well as the diffusivity, grows, which leads to the sea surface elevation increasing.

3.2 Three-dimensional Experiments

Similar to 2-D experiments, the sea ice still appears as two patches at the end of the third day in our 3-D simulations, although it spreads to the surrounding area comparing to initial field (Figs 2c and d, shading). The ice field (Figs 2c and d, vector) moved eastward uniformly with a peak velocity of 3.5 cm/s. It is surprised to find, even though the interfacial stress is calculated with different velocities in 2-D and 3-D simulations, the momentum transferred from the ice to the ocean is the same. Check the barotropic velocity in the 2-D simulation and the sea surface velocity in the 3-D simulation, both of which are much smaller than the ice velocity. Thus, we consider the sea ice always offer the same momentum to the current. The current is accelerated by the momentum from the sea ice and the wind forcing. If we check the ocean barotropic velocity (Figs 3c and d, vector), there are eddies produced by the nonlinear advection mechanism, and the sea surface elevation (Figs 3c and d, shading) rises (lowers) south (north) of the sea ice.

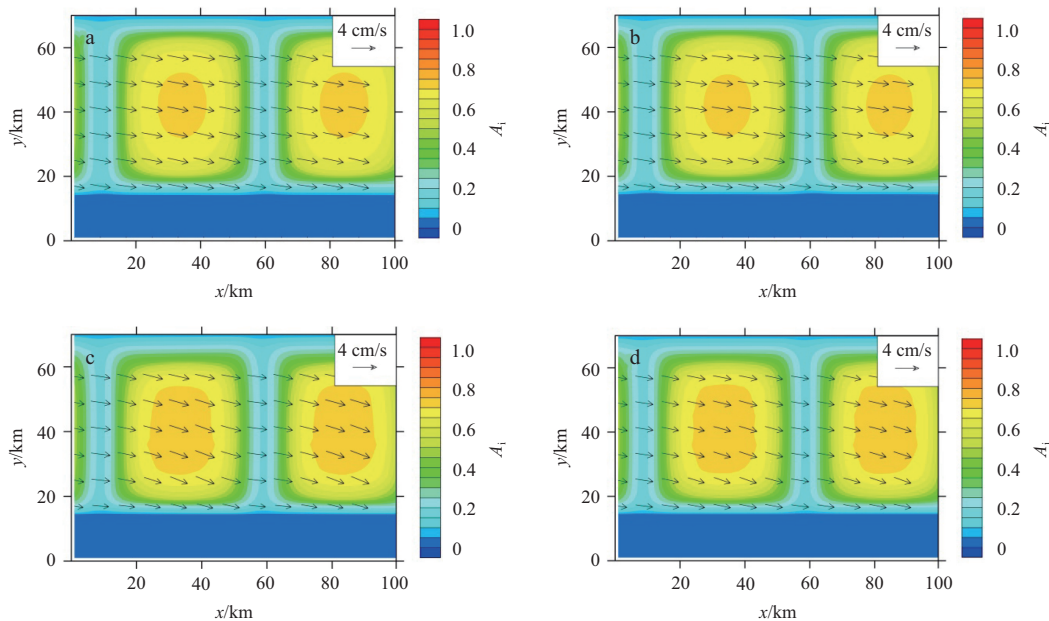


Fig. 2. Distributions of sea ice fraction (shading) and sea ice velocity ($\vec{v}_i$) at the end of the 3rd day in 2-D80 (a), 2-D400 (b), 3-D80 (c), and 3-D400 (d).

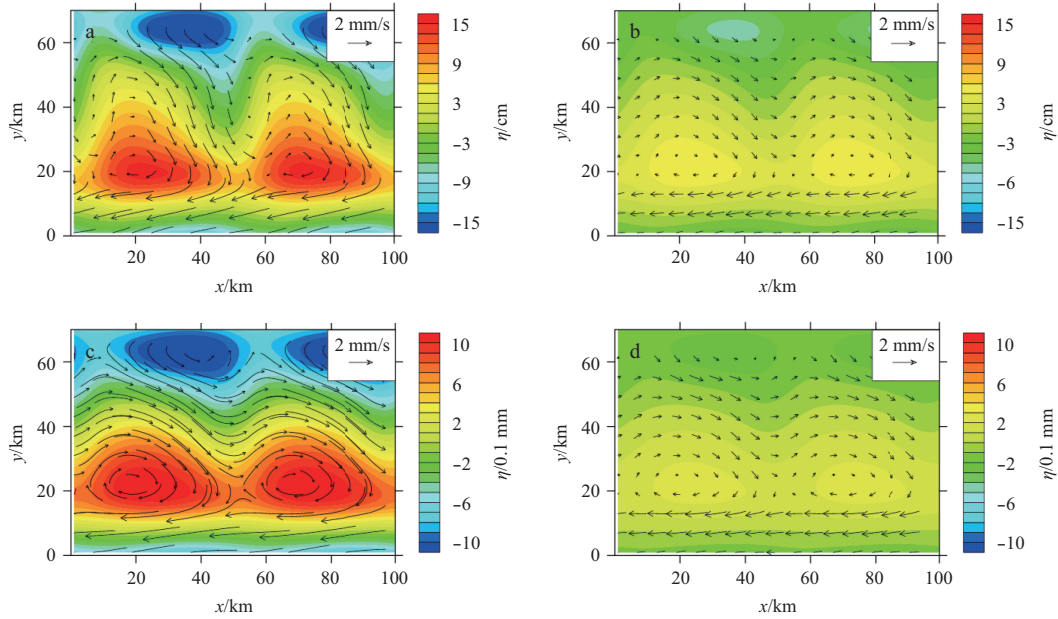


Fig. 3. Distributions of sea surface elevation (shading in the top panels, and on the bottom) and barotropic velocity ($\vec{v}_{\text{baro}}$) at the end of the 3rd day in 2-D80 (a), 2-D400 (b), 3-D80 (c), and 3-D400 (d).

3.3 Comparison of 3-D and 2-D experiment results

In the 3-D experiments, the sea surface elevation is only 0.1 mm, which is much less than that in the 2-D experiments (about 0.15 m). An important reason for this result is the vertical velocity (w) in the 3-D ocean model. When the sea surface elevation rises, the vertical velocity forces the sea surface back toward the sea level. However, in the 2-D ocean model, the sea surface elevation is only decided by the horizontal advection; thus, the elevation is much easier to rise.

Another issue is when the water depth is reduced in the 2-D experiments (Figs 3a and b, shading), the current is faster, the mesoscale eddies are stronger and the sea surface elevation becomes larger. This process can be explained as follows: with uniform wind forcing, the sea ice velocity is also uniform, which is much faster than the current. As a result, the sea ice always gives the same interfacial stress; in other words, the ocean gains the same momentum. When the water depth or mass increases, the current moves more slowly, which induces a smaller velocity gradient. As the sea surface elevation is proportional to the horizontal velocity gradient, the sea surface elevation decreases. The opposite happens: the water depth drop induces the sea surface elevation increase.

Similar to the 2-D results, the sea surface elevation, as well as the mesoscale eddy strength, seems to be influenced by the wa-

ter depth change in the 3-D experiments (Figs 3c and d, shaded). This result confirmed the 2-D experiments' result, that is, the nonlinear advection mechanism is more important in the shallow water. We mentioned this in the beginning of this section that the vertical velocity in the 3-D ocean model can decrease the sea surface elevation. When we check it carefully, the strongest vertical motion appears close to the center of the sea surface elevation (Fig. 4 shading). Furthermore, although the sea surface elevation is larger in 3-D80, the surface vertical motion (Fig. 4 contour) is even stronger in 3-D400, especially the downwelling close to the center of the positive sea surface elevation. Thus, we conclude that it is the vertical motion determines the magnitude of the sea surface elevation, despite of the horizontal advection in the 3-D simulation.

Besides current speed, there are still differences between 3-D400 and 3-D80 simulations. Although the surface current did not present cyclonic pattern in both simulations (Figs 5a and b), we can see different cyclonic currents when goes into a lower layer (Figs 5c, d, e and f). In 3-D400, the current becomes cyclonic at a depth deeper than 80 m (Fig. 5f); however, the cyclonic current appears only at a depth of 40 m in 3-D80 (Fig. 5c). It is clear that the speed of the surface current is much larger than the current speed in lower layer, which implies that most of the momentum are trapped in upper layer in 3-D400. Another issue we notice is

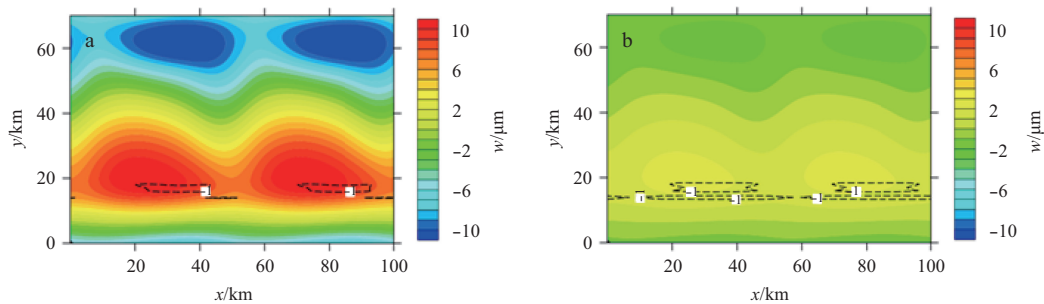


Fig. 4. Distributions of ice fraction (shading) and surface vertical velocity (contour) at the end of the 3rd day in 3-D80 (a) and 3-D400 (b).

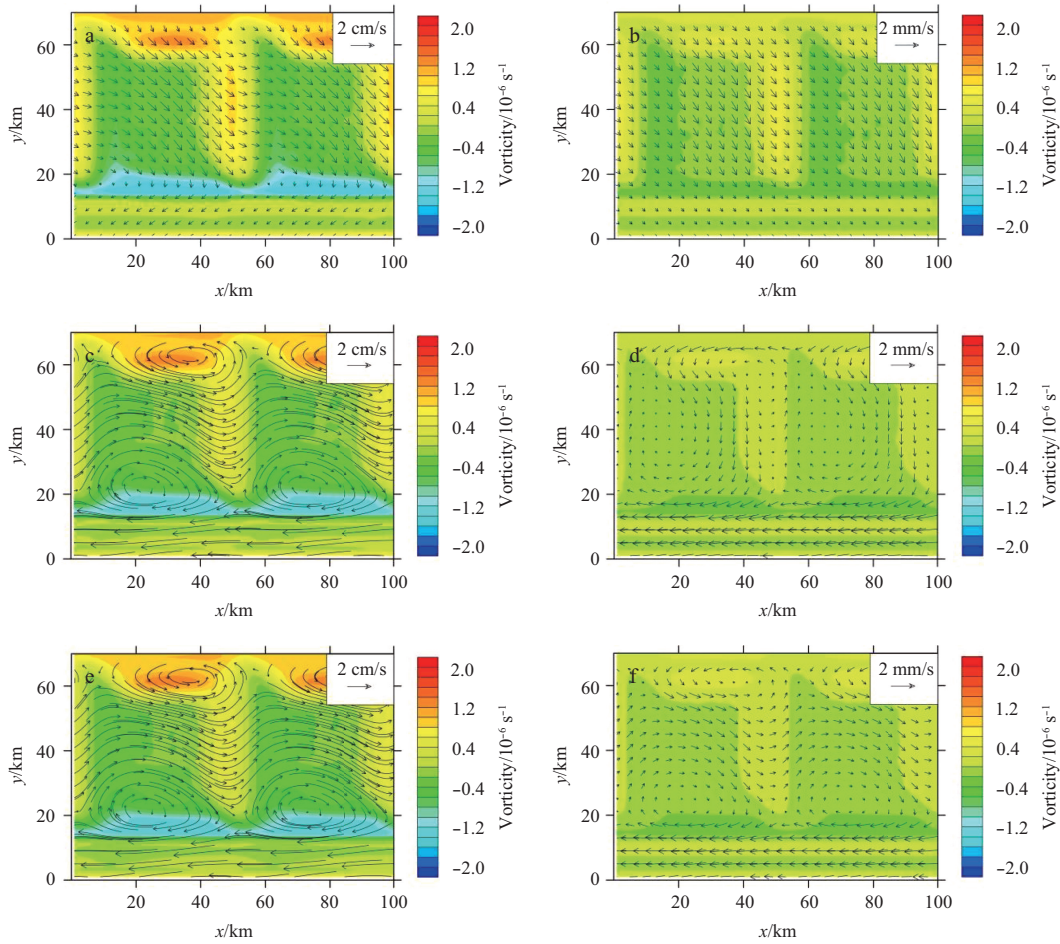


Fig. 5. Distributions of current velocity ($\vec{v}$) and vorticity (shading) in different layers: surface (a, b), 40 m (c, d), and 80 m (e, f). The left panels represent 3-D80, and the right panels represent 3-D400.

that the current speed in 3-D80 is always times of that in 3-D400 in lower layers.

Here we offered another way to understand the nonlinear eddy genesis mechanism. Since the bottom is flat, and the sea surface elevation can be neglected compared with the water depth in the 3-D simulation, the water column is always constant. Consider the Coriolis parameter is uniform; the relative vorticity can be induced by the gradient of the wind stress, the interfacial stress and the bottom drag with the vorticity equation. In the 3-D simulation, the wind stress and the interfacial stress are proportional to the ice distribution, thus, their gradient is caused by the ice distribution gradient. The bottom drag is proportional to the bottom velocity; we also found that the obvious bottom velocity gradient induced the relative vorticity (Fig. 6), which will also contribute to the relative vorticity.

4 Discussion and summary

In this study, we revisited one of the most important eddy genesis mechanisms—the nonlinear advection. To confirm that this mechanism does exist in the real world, we employ a 3-D ocean model. Following 2-D ocean-ice modeling study (Häkkinen, 1986), the pressure term did not contribute to the sea ice momentum directly, but it made no difference when we added it (not shown).

The nonlinear advection mechanism states that under forcing of the same uniform wind, the discrepancy of momentum

exchange efficiency between air-to-ocean and air-to-ice, as well as the ocean-ice interfacial stress, induces faster current beneath the ice than that in the open ocean, this leads to convergence or divergence at the ice edge and thus allows the genesis of mesoscale eddies.

In this study, we discussed the nonlinear advection mechanism further than any previous studies in several ways.

(1) The nonlinear advection mechanism is able to induce mesoscale eddies in both 2-D and 3-D ocean models. But when we checked the results carefully, there are a few differences between the 2-D and 3-D experiments. Although, in both systems, sea ice spread to surround, and moved eastward uniformly, although the interfacial stress calculation is different in the 2-D and 3-D simulations.

(2) The importance of the nonlinear advection mechanism in shallow water confirmed the results in Häkkinen (1986). Although the sea surface elevation is still inverse proportional to the water depth, it is much smaller than Häkkinen's results, since vertical motion counteracts with the sea surface elevation.

(3) Compared with the barotropic instability mechanism, the nonlinear advection mechanism has some specific requirements. For example, we always have the sea ice mass gradient in both x and y directions. If there is only the meridional or zonal sea ice mass gradient, there are no mesoscale eddies any more. On the other hand, we can reproduce the mesoscale eddies with a simple ice edge in the barotropic instability mechanism. Thus,

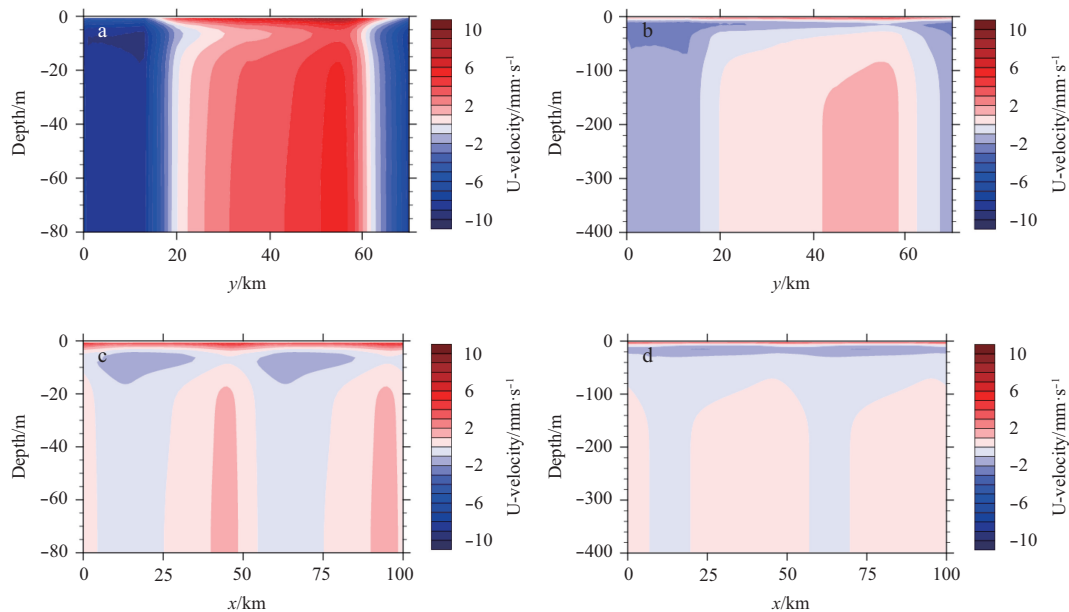


Fig. 6. *U*-velocity vertical profile on *y*-*z* panel (a, b) and *x*-*z* panel (c, d), in 3-D80 (a, c) and 3-D400 (b, d).

the barotropic instability might be a more common mechanism.

We obtained some new results, but there are many issues to be studied. There are mainly four mesoscale eddy genesis mechanisms in the MIZ. We only discussed the nonlinear advection in this study, and we also have some conclusions about the barotropic instability in our latest studies. Gula et al. (2015) discussed the topographic control in the Atlantic, but we think it should be the same scenario in the MIZ. Thus, we still need to discuss the baroclinic instability. When we concern the baroclinic instability, an important issue we should carefully think about is the stratification. How to make an unstable stratification, how to relate it to the real world? Later on, more complex scenario should be discussed. For example, two or three mechanisms may appear at the same time. As in the real world, the four mechanisms may occur at the same time. How to quantify which one is more important, and even to predict which mechanism is more likely to happen? These are all challenging topics.

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