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Characterizing the seasonal cycle of upper-ocean flows under multi-year sea ice

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ABSTRACT

Observations in the Arctic Ocean suggest that upper-ocean dynamics under sea ice might be significantly weaker than in the temperate oceans. In particular, observational evidence suggests that currents developing under sea ice present weak or absent submesoscale ($\mathcal{O}(1)$ Rossby number) dynamics, in contrast with midlatitude oceans typically characterized by more energetic dynamics at these scales. Idealized numerical model results of the upper ocean under multi-year sea ice, subject to realistic forcing, are employed to describe the evolution of the submesoscale flow field. During both summer and winter under multi-year sea ice, the simulated submesoscale flow field is typically much less energetic than in the midlatitude ice-free oceans. Rossby numbers under sea ice are generally consistent with geostrophic dynamics ($Ro \sim \mathcal{O}(10^{-3})$). During summer, ice melt generates a shallow mixed layer ($\mathcal{O}(1)$ m) which isolates the surface from deeper, warmer and saltier waters. The Ekman balance generally dominates the mixed layer, although inertial waves are present in the simulations during weakening and reversals of the ice-ocean stress. During winter, mixed-layer deepening (to about 40 m depth), is associated with convection driven by sea-ice growth, as well as ice-ocean shear-driven entrainment at the base of the mixed layer. Submesoscale activity is observed to develop only rarely, when winter convective mixing is laterally inhomogeneous (i.e., in the presence of sea-ice leads or spatially inhomogeneous sea-ice thickness) and when this coincides with weak ice-ocean shear-driven mixing. These submesoscale features are diagnosed with particular focus on their implications for ocean-to-ice heat fluxes.

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1. Introduction

Submesoscale (SM) flows, characterized by Rossby number $\mathcal{O}(1)$ and horizontal scales between 100 m and 10 km in the midlatitudes, have been shown to play an important role in the upper-ocean dynamics of most major ocean basins. Submesoscales features are known to develop in weak stratification regimes, such as the ocean mixed layer, and in the presence of a source of available potential energy such as: at major ocean fronts (e.g., Gula et al., 2014; Veneziani et al., 2014), along density gradients generated by river outflows (e.g., Luo et al., 2016), around mesoscale eddies (e.g., Mensa et al., 2013; Sasaki et al., 2014), and in regions of coastal upwelling (e.g., Capet et al., 2008d).

Baroclinic instabilities developing in the presence of weak stratification and a reservoir of available potential energy can generate features with large Rossby numbers and strong vertical velocities (Stone, 1966; Boccaletti et al., 2007). Due to their ageostrophic nature, SM features are thought to play an important role in the ocean energy budget, providing a pathway for forward energy dissipation (Muller et al., 2005; McWilliams, 2008; Molemaker et al., 2010). Given the large vertical velocities that SM flows can generate, SM features are also thought to be responsible for a significant fraction of the observed primary production in the oceans (Lévy et al., 2001; Mahadevan and Tandon, 2006; Mahadevan et al., 2008; McGillicuddy et al., 2007; Klein and Lapeyre, 2009), driving for vertical transport of nutrients into the euphotic zone. SM features are also of significance to lateral material transport in the ocean (e.g., Lumpkin and Elipot, 2010; Poje et al., 2014). Lateral transport by SM flows has important practical consequences to the fate of pollutants (e.g., oil spills) pointing to the need for numerical simulations and observations to resolve these small scale flows (Mensa et al., 2015).

While there has been much progress in developing an understanding of SM dynamics in the midlatitude ice-free oceans, little is known about this flow regime in the Arctic Ocean. SM features may impact on sea ice, primary production, lateral transport of nutrients and pollutants, and the upper-ocean heat budget. Sea-ice cover is influenced by heat fluxes at the base of the ocean mixed layer where a reservoir of ocean heat exists underlying a rela-

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tively cool fresh mixed layer (e.g., Maykut, 1982; Wettlaufer, 1991; McPhee, 1992; Perovich and Elder, 2002). Small-scale flows associated with vertical velocities that might enhance ocean-to-ice heat fluxes could have serious consequences to sea-ice cover. Similarly, lateral transport of ocean heat by SM flows could enhance sea-ice melt, a process largely underestimated by general circulation models (Serreze, 2007; Stroeve et al., 2007; Rampal et al., 2011) and of crucial importance in the understanding of present and future global climate (e.g., Walsh, 1983; Budikova, 2009). Further, in an Arctic undergoing rapid change, and expansion of activities such as oil exploration, understanding pollutant dispersal in the upper ocean is of utmost importance (National Academies Report (NAS), 2014).

Observations in the Arctic Ocean suggest that the energetics of upper-ocean flows under sea ice, and in ice-free regions of the Chukchi Sea, might be different from those of the temperate oceans (Timmermans and Winsor, 2013; Timmermans et al., 2012). In contrast with midlatitude oceans characterized by energetic upper ocean dynamics at submesoscales (Capet et al., 2008b; Mensa et al., 2013; Callies et al., 2015), observations suggests that submesoscale dynamics are weak under sea ice, however the generality of this statement remains unknown. For example, Timmermans et al. (2012) show that although some frontal activity is present under sea ice, horizontal wavenumber (k) spectra of potential density variance in the mixed layer exhibit steep slopes – scaling as k^{-3} for wavelengths between around 5 and 50 km, compared to the k^{-2} scaling more typical of the midlatitudes.

The presence of sea ice affects both the dynamics and thermodynamics of the upper ocean and possibly impacts the development of SM instabilities. Ice cover effectively limits the propagation of surface gravity waves and mediates wind-forced internal waves (e.g., Levine et al., 1985; Dosser et al., 2014). On the other hand, sea-ice drift can generate significant upper-ocean stresses, and enhance turbulent mixing near the surface (e.g., Denbo and Skyllingstad, 1996; Backhaus and Kampf, 1999; Skyllingstad, 2001; Cole et al., 2013). Buoyancy fluxes are also mediated by sea ice, with sea-ice melt and freshwater input generating a stabilizing buoyancy flux during summer, and sea-ice growth and brine rejection generating a destabilizing buoyancy flux during winter.

In this paper, we present results from an idealized high-resolution numerical simulation of the upper Arctic Ocean with the intention of exploring the seasonal cycle of the upper ocean under multi-year sea ice. Results are limited to regions of the Arctic Ocean that are permanently ice covered (i.e., areal concentration of sea ice is never below about 75%). A transition from fully ice covered to ice-free, and the intermediate marginal ice zone dynamics, are outside the scope of this study and likely present significantly different results. Here we retain the simplest possible framework avoiding the uncertainties associated with additional sea-ice parameterizations (e.g., Smedsrud and Martin, 2015). Our simulation is forced with realistic atmospheric forcing and the model produces seasonal cycles of sea ice and upper-ocean properties that are consistent with observations (Section 3). In Section 4, we show how the development of small-scale flows under sea ice does not present the typical submesoscale *soup* (Gula et al., 2014) characteristic of the midlatitudes. Instead, small-scale flows in the mixed layer under sea ice are dominated by Ekman dynamics and convective processes with little interaction between the surface mixed layer and underlying interior ocean. Two examples that demonstrate occasional development of SM features are also presented in this section; the corresponding background settings for these provides context and motivation for future study. Also in Section 4, we test the applicability of a submesoscale parameterization in the under-ice setting modeled here. In Section 5, we summarize and discuss our results, comparing and contrasting the well-known midlatitude SM regime with ocean dynamics under sea ice.

2. Numerical model configuration

Our study employs the MIT general circulation model (MITgcm, Marshall et al., 1997; Adcroft et al., 2014) in an idealized configuration. The square domain spans 256 km by 256 km in the horizontal and has a fixed depth of 700 m. Boundary conditions are doubly-periodic at the sides, with free-surface and free-slip at the surface and bottom boundaries, respectively. The model has a horizontal resolution of 500 m and vertical resolution varying from 0.2 m near the surface (over the mixed layer, mean vertical resolution ranges between 0.5 m in summer and 1.6 m in winter) to 50 m near the bottom, for a total of 52 layers. Computations were performed on the Center for Computational Sciences clusters at the University of Miami.

The model consists of an ocean component and a sea-ice component, with communication between the modules at the ice-ocean interface. The ocean component is configured to solve the hydrostatic, Boussinesq equations. We use an f-plane approximation as the β-effect is negligible near the poles. Horizontal tracer diffusivities are implicit while horizontal viscosity in the momentum equations is biharmonic and uses a Leith eddy viscosity coefficient (Leith, 1996; 1968). Vertical eddy viscosity and diffusivities are set by the non-local K-profile parameterization (KPP, Large et al., 1994). KPP enhances vertical viscosity and diffusivities when shear instabilities and convection generate mixing in the boundary layer (i.e., the mixed layer). The mixed-layer or boundary-layer depth (BLD) is set as the depth at which the bulk Richardson number equals a critical value of 0.3. A nonlinear seawater equation of state is used to compute density (see Jackett and Mcdougall, 1995), which in the cold polar oceans is primarily a function of salinity (e.g., see the discussion in Timmermans and Jayne, 2016).

The ice model consists of both a dynamic and thermodynamic component, generating sea-ice stress acting on the ocean surface and buoyancy fluxes associated with its growth and melt. The sea-ice dynamics implements the model of Zhang and Hibler (1997), where internal stresses are described via a viscous plastic model (Zhang and Hibler, 1997). Ice dynamics mediates the transfer of atmospheric stresses to the ocean. Ice thermodynamics follows the model by Hibler III (1980). The model uses a zero-layer formulation, in which heat conductivity across the ice is parameterized assuming a linear ice temperature profile together with a constant ice conductivity (Semtner, 1976). Although this formulation has the tendency to underestimate the amplitude of the seasonal cycle in sea-ice thickness and extent (Semtner, 1984), we find a sea-ice seasonal cycle in good agreement with the observations described by Timmermans (2015) for the same region and time. The thermodynamics model computes sea-ice thickness (H_I) and fractional area of sea ice (A_I, defined as the area of each grid cell covered by sea-ice), and the fluxes at the ice-ocean interface.

2.1. Initial conditions and model spin-up

We initialize the ocean module with temperature and salinity typical of a weak surface front under sea ice in August in the Arctic Ocean's Canada Basin (see Timmermans et al., 2012). The initial surface front (configured as a filament, Fig. 1a) is characterized by a surface (i.e., mixed layer) horizontal density gradient of 5×10^{-7} kg m^{-4}. The filament consists of fresher (and less dense) surface waters, while surface waters either side are saltier (and more dense). The entire surface is at the freezing temperature, which means that the filament waters are slightly warmer than the waters either side.

The front is allowed to relax (unforced and without sea ice) for 100 days in a 1 km horizontal resolution simulation; the resulting state exhibits a fully developed eddy field and is integrated

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