Genesis and decay of baroclinic eddies in the seasonally ice-covered Arctic Ocean

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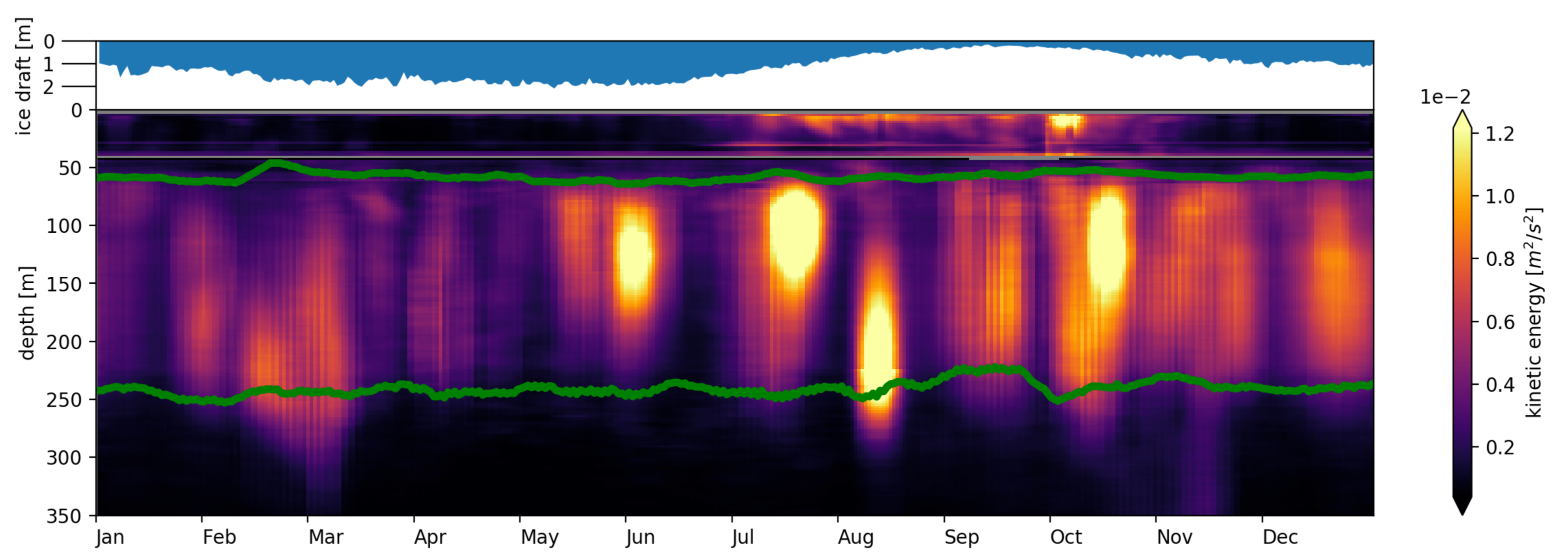
Key points

- Because of the presence of multiple stratification maximums, independent baroclinic eddies are generated at different depths in the Arctic.
- Surface intensified eddies interact with the ice and are subject to a strong seasonal cycle: after growing during summer, they are dissipated by friction against the ice cover during winter.
- Subsurface intensified eddies, obtaining their energy from the halocline, persist all year long independently of the presence of sea ice.

Observations

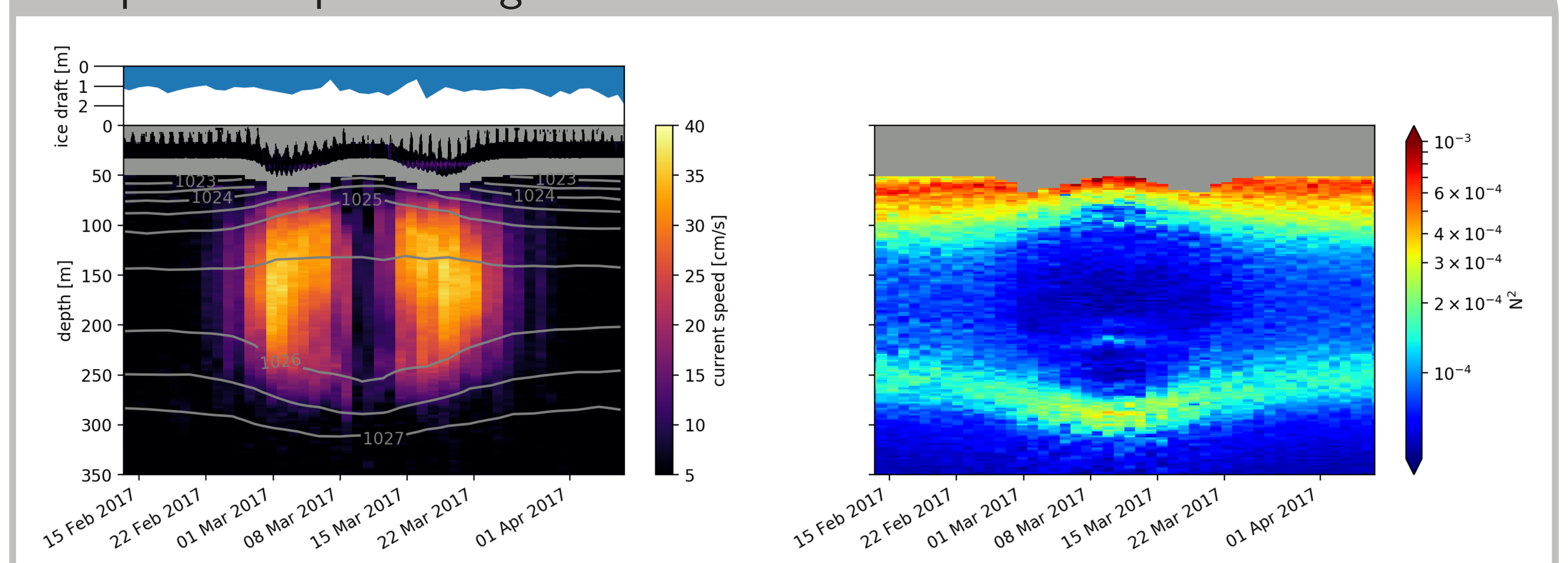


Observations of ocean currents in the Arctic show a curious, and hitherto unexplained, vertical distribution of eddy kinetic energy, shown in the panel above (data from the Beaufort Gyre Exploration Project's mooring A in the central Canadian Basin). A marked seasonal cycle is found close to the surface: strong eddy activity during summer, observed from both satellites and moorings, is followed by very quiet winters. In contrast, subsurface eddies, confined within two peaks in stratification $N^2 = -\frac{g}{\rho_0}\frac{\partial \rho}{\partial z}$ (gray lines), persist all year long within the deeper halocline, and below.

Informed by baroclinic instability analysis, and a high resolution pan-Arctic ocean model, we explore the origin and evolution of baroclinic eddies in the seasonally ice-covered Arctic Ocean. Mesoscale activity is fundamentally different between ice-covered and ice-free regions, with important consequences on the evolution of the Arctic eddy field under a changing ice cover and stratification.

Mooring data were collected and made available by the Beaufort Gyre Exploration Program based at the Woods Hole Oceanographic Institution (http://www.whoi.edu/beaufortgyre) in collaboration with researchers from Fisheries and Oceans Canada at the Institute of Ocean Sciences.

Dissipation of preexisting eddies



LEFT: MOORING OBSERVATION OF CURRENT SPEED AND ISOPYCNALS (GRAY LINES) FOR AN ANTICYCLONIC EDDY PASSING BY THE MOORING IN WINTER 2017. ICE DRAFT IS SHOWN IN CYAN. RIGHT: STRATIFICATION N^2.

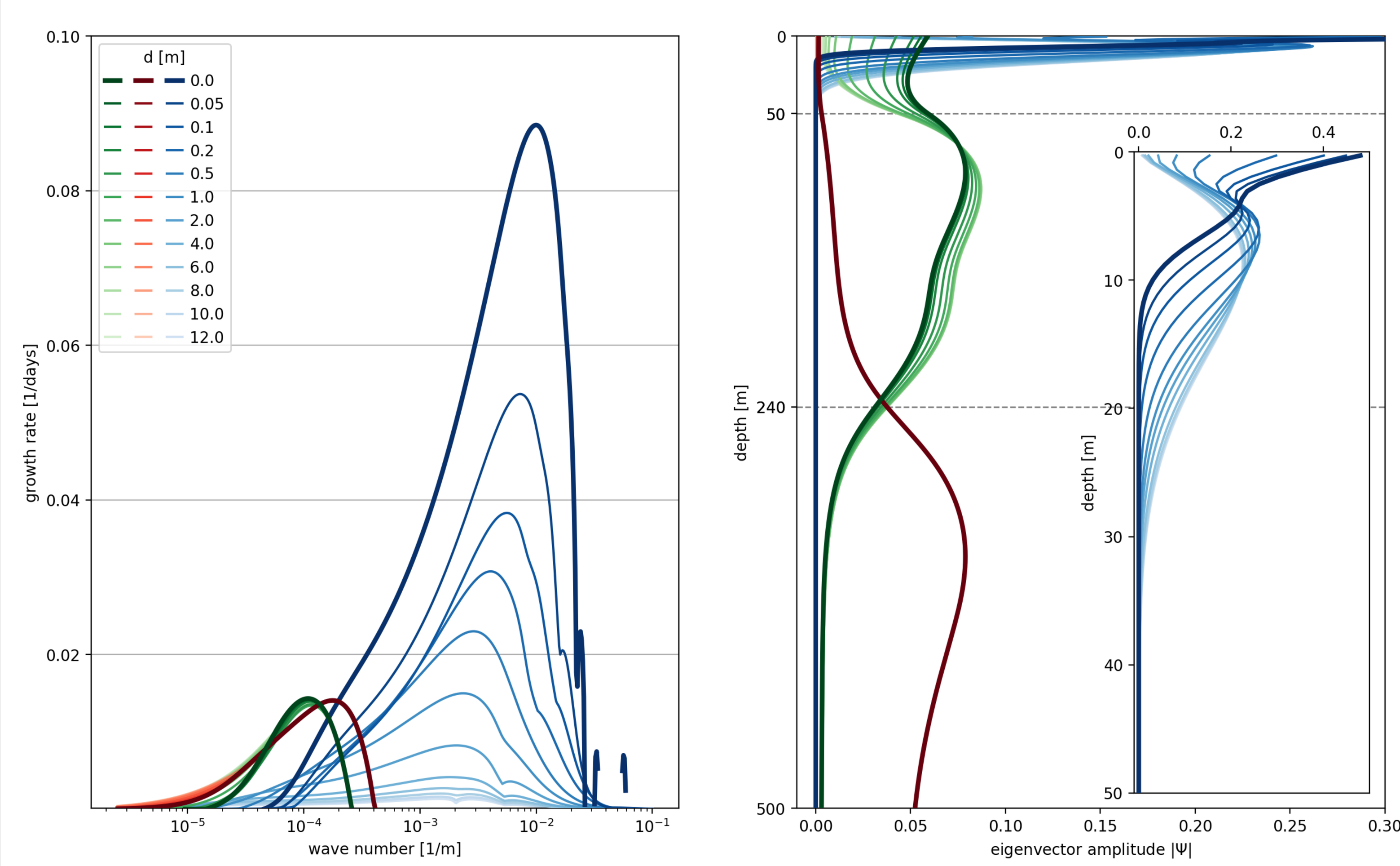
Surface eddies generated, e.g., in ice free regions, dissipates fast once they encounter ice. Their spindown time scale can be estimated by energetic consideration as

$$T_\nu = \frac{K}{\dot{W}} = \frac{H}{d} f^{-1}$$

where K is the kinetic energy of the eddy, $\dot{W}$ is the power dissipated by friction, H is the depth of the eddy and d is the Ekman layer length scale. If we consider a characteristic vertical scale $H \approx 30\,\mathrm{m}$, characterizing the surface eddies, and an Ekman layer depth of order 1 m, the resulting time scale is about 2 days, with larger Ekman layer depths dissipating eddies even faster. Surface layer eddies cannot travel long distances while in contact with the ice cover.

In contrast subsurface eddies are insulated from the ice cover by a strong stratification (red region in the right panel). The eddy shown in the figure above (left panel) appears unaffected by the presence of thick sea ice (cyan curve) during its one month-long transit. A thermal-wind driven velocity shear, due to the displacement of the isopycnal (gray lines) in combination with the strong stratification (right panel), reduces the current speed to zero already at a depth of 50 m. No Ekman layer, and no frictional dissipation, is induced by the passing eddy, which can then move undisturbed under ice.

Baroclinic eddies



GROWTH RATES (LEFT) AND VERTICAL STRUCTURE OF THE PERTURBATIONS (RIGHT) FOR THE SURFACE LAYER (RED) AND HALOCLINE (BLUE) EDDIES. SELECTED VALUES OF EKMAN LAYER DEPTH d.

— Theory —

We analyzed the growth of baroclinic perturbation by solving the eigenvalue problem associated with the linearized, quasi-geostrophic potential vorticity (PV) equation

$$\frac{Dq}{Dt} = -\nabla Q \cdot \boldsymbol{u}$$

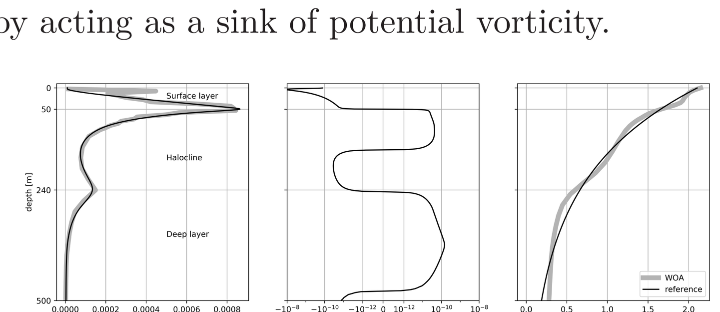
$$q = \nabla^2\psi + \frac{\partial}{\partial z}\left(\frac{f_0^2}{N^2}\frac{\partial \psi}{\partial z}\right) + \beta_0 y$$

where Q is the background potential vorticity. Boundary conditions are provided by the quasi-geostrophic density equation

$$\frac{D}{Dt}\left(\frac{\partial \psi}{\partial z}\right) = \frac{\partial \boldsymbol{U}}{\partial z}\cdot\nabla\psi + \frac{N^2}{f_0}\underbrace{\frac{d}{2}\nabla^2\psi}_{w_E}$$

where $w_E = \frac{d}{2}\nabla^2\psi$ is the Ekman pumping driven by frictional effects against the ocean floor and sea-ice (d is the Ekman layer depth).

Note how perturbations can be driven by the interior potential vorticity gradient ∇Q and/or the velocity shear at the boundaries $\frac{\partial U}{\partial z}$. In contrast, friction against sea ice, modeled by the Ekman pumping w_E, hinders the growth of perturbations by acting as a sink of potential vorticity.



REFERENCE STRATIFICATION N^2, POTENTIAL VORTICITY GRADIENT ∇Q AND VELOCITY U PROFILES USED IN THE STABILITY COMPUTATIONS (BLACK). GRAY LINES MARK PROFILES OBTAINED FROM THE WORLD OCEAN ATLAS

— Results —

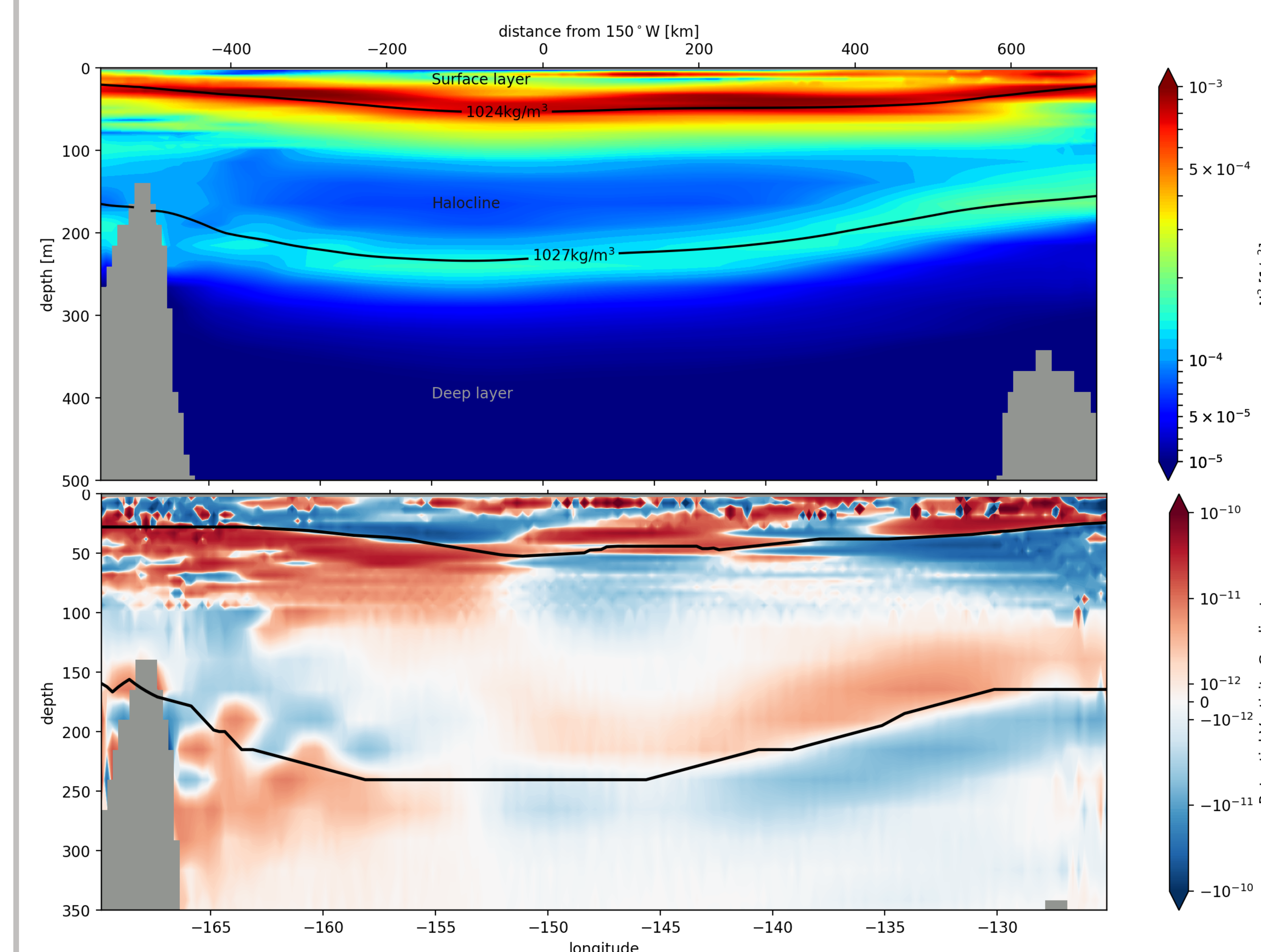
What happens as the ice cover changes over the seasonal cycle? We show above the growth rate (left) and vertical structure (right) of instabilities as a function of the Ekman layer depth d, a proxy for the ice ability to sustain internal stress.

In the absence of friction (thick lines, $d = 0$), the surface mode (blue) grows on a time scale of order ten days, and is characterized by an horizontal length scale of order 100 m. Its signature is concentrated between the surface and the shallower peak in stratification located at 50 m depth (see inset). The halocline mode (green) is characterized by a slower time scale of order two months, and a larger horizontal length scale of order 10 km; its signature reaches its maximum in the halocline between the two peaks in stratification at 50 m and 240 m, but in the absence of friction its imprint is still visible at the surface. The deep mode (red), characterized by similar time and length scales, has its maximum below the deepest peak in stratification, and decays to zero across the halocline before reaching the surface.

Friction has a strong effect on the surface mode (blue). An Ekman layer depth of 2 m is enough to reduce the growth rate by more than an order of magnitude, while the length scale of the fastest growing mode increases. The vertical structure of the surface mode shows the effect of dissipation. Increasing friction drives the mode to zero at the surface, and a subsurface peak in the streamfunction amplitude is developed, but is still contained in the surface layer, as can be seen in the inset.

This has to be contrasted with its effect on the halocline (green) and the deep (red) modes. Their growth rate and length scale are barely affected by increased friction. The only visible effect is a reduction in the surface amplitude of the halocline modes, but the bulk of the perturbations, lying within the halocline below the stratification peak at 50 m, is unchanged.

Conclusions

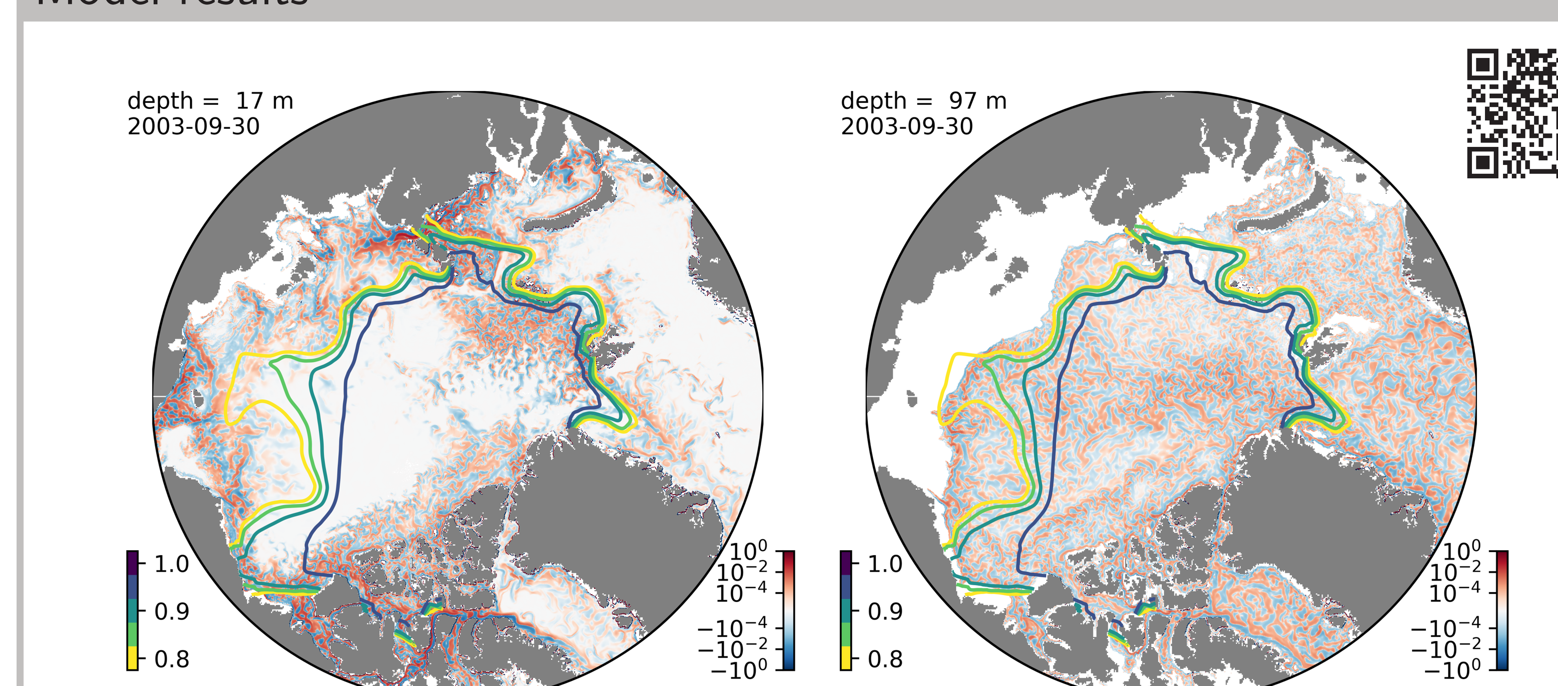


TOP: STRATIFICATION N^2 ACROSS THE CANADA BASIN AT 75°N. BOTTOM: POTENTIAL VORTICITY GRADIENT $\nabla f \frac{\partial \rho}{\partial z}$ IN THE CANADA BASIN. BLACK LINES MARK THE PEAKS IN STRATIFICATION. DATA FROM THE 2005-2017 WOA CLIMATOLOGY.

Observations, baroclinic instability analysis and an eddy resolving numerical model show that

- Different modes grow indipendently at different depths, bounded vertically by local maximums in stratification N^2.
- Surface layer eddies are strongly affected by the presence of the ice cover. Even moderate friction reduces growth rates to levels too low for eddies to actually develop. When developed over the summer or in ice free regions, these eddies will be dissipated on a time scale of a few days; they cannot travel under ice.
- Halocline, and deeper, eddies are unaffected by the presence of the ice. Their growth rate is of the order of a few months, and their length scale or order 10 km. Insulated from the ice above by a strong stratification peak, they can potentially travel long distances without being dissipated.

Model results



NORMALIZED RELATIVE VORTICITY AT 17 m AND 97 m DEPTH FOR SEPTEMBER 2003. NOTE LOGARITHMIC COLOR SCALE. CONTOURS REPRESENT ICE CONCENTRATION RANGING FROM 95% (PURPLE) TO 80% (YELLOW).

An eddy-resolving confirms our baroclinic instability results. Surface intensified eddies in the central Arctic rapidly disappear with the formation of the ice in autumn (left), with relative vorticity dropping by more than four orders of magnitude across the marginal ice zone. In contrast, halocline eddies are largely unaffected by the presence of ice.

We use results from a simulation based on the CREG12 configuration (Dupont et al. 2015), encompassing the Arctic and parts of the North Atlantic. It is based on the NEMO (Madec et al. 2014) and LIM3 (Rousset et al. 2015) numerical models for the ocean and sea ice components, respectively. LIM3 uses an EVP (Elasto-Viscous-Plastic) rheology (Hunke and Dukowicz 1997). The configuration has a high vertical (75 levels) and horizontal (3-4km) resolution in the Arctic Ocean.