Modeling the Dynamics of the North Water Polynya Ice Bridge

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ABSTRACT

The North Water polynya, the largest polynya in the world, forms annually and recurrently in Smith Sound in northern Baffin Bay. Its formation is governed in part by the formation of an ice bridge in the narrow channel of Nares Strait below Kane Basin. Here, the widely used elastic–viscous–plastic elliptical rheology dynamic sea ice model is applied to the region. The idealized case is tested over a range of values for $e = [1.2, 2.0]$ and initial ice thicknesses from 0.75 to 3.5 m, using constant northerly winds over a period of 30 days, to evaluate long-term stability of different rheological parameterizations. Idealized high-resolution simulations show that the formation of a stable ice bridge is possible for $e \leq 1.8$. The dependence of the solution in terms of grid discretization is studied with a domain rotated 45°. A realistic domain with realistic forcing is also tested to compare time-variant solutions to actual observations. Cohesion has a remarkable impact on if and when the ice bridge will form and fail, assessing its importance for regional and global climate modeling, but the lack of observational thickness data during polynya events prevents the authors from identifying an optimal value for $e$.

1. Introduction

Polynyas are regions of ice-covered oceans where low sea ice concentration anomalies are observed. They are the location of enhanced biological productivity and ocean–atmosphere energy exchange in polar oceans. One of the largest polynyas and most productive ecosystems in the world is located in the North Water (NOW), northern Baffin Bay (Deming et al. 2002). Numerous observational (Melling et al. 2001; Barber et al. 2001) and modeling (Mysak and Huang 1992; Darby et al. 1994; Heinrichs 1996; Biggs and Willmott 2001; Yao and Tang 2003) studies have been conducted to describe and understand the opening mechanisms of the NOW polynya. There is strong agreement that the main driving mechanism is the wind-forced advection of sea ice downwind of an ice bridge that forms seasonally and recurrently, between Greenland and Ellesmere Island (Fig. 1). The polynya existence essentially depends on the formation of this ice bridge (also often called the ice arch). The possibility of an ice bridge not forming because of a variable or changing environment may impact the whole ecosystem and at the very least the local climate (Marsden et al. 2004). Interannual and interdecadal variability have been characterized by Barber et al. (2001); recent remote observations suggest that its formation may greatly be affected by multiyear ice depletion (Belchansky et al. 2004). Typically, the polynya exists on the order of several weeks in late spring (March to June) but can form during late fall or anytime during winter depending on sea ice and meteorological conditions. The ice bridge location is highly correlated with the coastline features and typically lies at the constriction point between the two landmasses. The ice edge shape is variable, but it is always concave and archlike. Adequately reproducing the sea ice behavior in such a constrained area constitutes one step toward the understanding of the effect of climate on the polynya and its marine ecosystems through ocean–sea ice coupled modeling.

Pioneer studies of sea ice arching were greatly inspired from soil mechanics studies. The problem of sea ice flow in constrained channels or rivers is very similar to the gravity flow of granular material through vertical
channels, hoppers, and silos. This last problem has been extensively studied using the granular theory for bulk solids (Richmond and Gardner 1962; Walters 1973; Savage and Sayed 1981). In contrast to fluids, bulk solids can transmit shear stresses while at rest. Cohesive materials (e.g., damp sand), which are able to support higher static shear stresses, are capable of forming a self-obstruction to flow (Walker 1966). Accordingly, ice arching has been observed and modeled assuming sea ice to be a plastic (discontinuous flow under continuous forcing) and cohesive (with the ability to maintain its integrity while submitted to tensile forces) material. Sodhi (1977) compared ice arches forming in Bering Strait and in the Amundsen Gulf with a cohesive Mohr–Coulomb granular rheology and found a good correspondence between the modeled ice arch form and the observed profile. He followed the analytical analysis of Morrison and Richmond (1976) that relates the arching process with the cohesive strength of the granular material. Ip (1993) studied the problem of ice arching in converging channels with different rheologies and showed that plastic yield curves lying on or crossing the principal stress axes allow ice arch formation, given the adequate loading and thickness. His main conclusion is that only cohesive materials are able to form arches.

The elliptical yield curve for sea ice, first introduced by Hibler (1979), has become the most widely used dynamic sea ice model in climate and process studies. It represents a cohesive material because a part of it covers the second and fourth quadrants of the principal stress space (Fig. 2). It does not allow pure tension, which occurs when both principal stress components are positive. Shear strength relative to compressive strength is scaled by the major to minor axis ratio parameter $e$. Cohesion increases along with shear strength as $e$ decreases. A value of $e = 2$ was originally chosen by Hibler (1979) to approximately match the ratio of energy dissipation in shear to energy dissipation in sea ice ridging calculated by Rothrock (1975). Hibler (1979) showed that decreasing $e$ stiffens the ice flow throughout the Arctic Basin but does not significantly alter spatial patterns of modeled ice thickness. On the other hand, such a change has a drastic impact on both the flow and spatial distribution of ice thickness in smaller enclosed areas. Kubat et al. (2006) used the ellipse and showed that modifications of the shear strength (cohesion) may
lead to ice arch formation in an idealized converging channel. A value of $e = 1.2$, representing a higher shear strength, successfully simulated an ice arch while the original value of $e = 2$ did not. They used an ice strength value $P^o = 10^4$ Pa, which led to the formation of very thick-ridged ice (up to 15 m) at the converging coastlines. The elliptical yield curve with $e = 2$ is also used in dynamic river ice transport and ice jam–formation modeling (Shen et al. 2000). Sea ice granularity in constrained areas like the channels of the Canadian Arctic Archipelago or rivers is greatly affected by the spatial scale that limits the maximum floe size and increases the importance of floe–coast interactions.

The goal of this paper is to examine the particular case of the Nares Strait ice arch formation and to characterize sea ice behavior as a function of shear strength. The channel exit width and the converging slope are held fixed even though they influence the arch formation (see Ip 1993). Because cohesion is fundamental for ice arching, its effect is explored and compared to synoptic observations obtained mainly from satellite imagery. The elliptical yield curve is used because of its success in modeling sea ice at large scales, because it allows sea ice to be cohesive, and because cohesion and shear strength are easily varied with the free parameter $e$. First, a sensitivity study is conducted in an idealized domain as a function of initial ice pack thickness and shear strength. Then, simulations are performed in a realistic domain with realistic wind forcing, and the impact of wind stress is assessed.

Details of the rheology, a description of the model, initial conditions, and external forcing are presented in section 2. Section 3 presents results obtained in idealized conditions. The sensitivity of the results to thickness, cohesion, grid orientation, and boundary roughness is explored. Section 4 presents the results from realistic simulations of the North Water ice bridge, and a conclusion is provided in section 5.

2. Model

a. Momentum equations and sea ice rheology

Our goal is to investigate the dynamical behavior of sea ice in environmental conditions similar to those of the North Water when an ice bridge is formed. Sea ice thermodynamics are not considered while ice dynamics are modeled following the elastic–viscous–plastic (EVP) approach of Hunke and Dukowicz (1997). We use the Geophysical Fluid Dynamics Laboratory (GFDL) code version, which is coupled to the Modular Ocean Model (MOM), release 4.0. Although ice arch formation has been studied with constant applied forcing (Ip 1993) or in equilibrium situations (Sodhi 1977), simulated sea ice should quickly respond to rapidly varying forcing. Hunke and Zhang (1999) showed that the explicit time stepping scheme based on elastic waves mechanisms responds very well to daily stress forcing variations.

The two-dimensional momentum conservation can be written as

$$
\frac{m}{\partial t} \frac{\partial u_i}{\partial x_j} = \frac{\partial \sigma_{ij}}{\partial x_j} + \tau_{ai} + \tau_{wi} + \epsilon_{ij} \frac{mHu_j}{\partial x_i} - \frac{mg}{\partial x_i},
$$

with $m$ being the ice mass per unit surface, $\sigma_{ij}$ the ice internal stress tensor, and $\tau_{ai}$ and $\tau_{wi}$ the components of the stress imposed on the ice by the wind and the ocean. These terms are respectively expressed as

$$
\tau_{ai} = \rho_w C_{da} |u| (u \cos \theta + k \times u \sin \theta),
$$

$$
\tau_{wi} = \rho_w C_{dw} |u_w - u| (u_w - u) \cos \theta + k \times (u_w - u) \sin \theta),
$$

with $\rho_a$ and $\rho_w$ being the air and water densities, $C_{da}$ and $C_{dw}$ the ocean–ice and air–ice drag coefficients, $u_w$ the geostrophic wind, and $\theta$ the turning angle. The two last terms of the right-hand side of (1) represent the Coriolis pseudoforce and the gravity on a tilted ocean surface $H_0$, respectively. In the idealized experiments discussed in section 3, these two forces are set to zero for the sake of symmetry and simplicity. An ocean drag of $C_{dw} = 3.24 \times 10^{-5}$ is considered with an initially static ocean. The model boundaries are closed and ocean currents are not prescribed.

Solving for $u_i$ requires a constitutive relation that relates the stress and strain rates. This relation contains a yield curve in the stress space and a flow rule describing the strain-rate orientation with respect to the
stress. The viscous–plastic (VP) constitutive relation based on the normal flow rule takes the following form:

$$\sigma_{ij} = 2\eta \dot{e}_{ij} + (\zeta - \eta)\dot{e}_{kk}\delta_{ij} - \frac{P}{2}\delta_{ij},$$

(4)

where

$$\dot{e}_{ij} = \frac{1}{2} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right)$$

(5)
is the strain-rate tensor. Here, $\zeta$ and $\eta$ are the bulk and shear nonlinear viscosities defined in terms of the strain-rate tensor components as

$$\zeta = \frac{P}{2\Delta},$$

(6)

$$\eta = \frac{P}{2\Delta e^2},$$

(7)

$$\Delta = [(\dot{e}_{11}^2 + \dot{e}_{22}^2)(1 + e^{-2}) + 4e^{-2}\dot{e}_{12}^2 + 2\dot{e}_{11}\dot{e}_{22}(1 - e^{-2})]^{1/2},$$

(8)

where $e$ is the ratio of major to minor axes for the ellipse.

The ice internal resistance is denoted by $P$, which depends on ice equivalent thickness $h$ and concentration $c$ as

$$P = P^* h \exp[-C(1-c)],$$

(9)

where $P^*$ and $C$ are constant parameters.

To integrate the solution for $u_t$ using the EVP scheme, (4) is inverted and an elastic term is added:

$$\dot{e}_{ij} = \frac{1}{E} \frac{\partial \sigma_{ij}}{\partial t} + \frac{1}{2\eta} \sigma_{ij} + \frac{\eta - \zeta}{4\eta^2} \sigma_{kk}\delta_{ij} + \frac{P}{4\zeta}\delta_{ij},$$

(10)

The first term on the right-hand side of the EVP constitutive relation (10) describes an elastic response of the strain to a given stress, where $E$ is Young’s Modulus. As detailed by Hunke and Dukowicz (1997), this term is introduced to significantly improve the integration efficiency of the explicitly discretized form of (1) and (10) toward the VP behavior and does not correspond to a physical property of sea ice. Here, $E$ depends on ice mass and grid resolution and is independent of rheological parameters. Hunke and Zhang (1999) and Arbetter et al. (1999) demonstrated that in large-scale realistic simulations of the Arctic, the EVP formulation produces dynamical ice behavior that is equivalent to the VP formulation, particularly for long time scales. At each model time step $\Delta t$, Eqs. (1) and (10) are sub-stepped $N$ times to damp the elastic waves toward a stationary state, in which case (10) reduces to (4). Viscosities are updated at each elastic loop, following Hunke (2001), while the elastic parameter $E$ is defined according to Hunke and Dukowicz (1997). The minimum number of sub time steps for reaching the VP solution is roughly determined by the ratio of the viscous to elastic time scales $T_v$ and $T_e$, such that $N \approx T_v/T_e$:

$$T_v = \frac{m}{\zeta} \Delta x^2 \quad T_e = \frac{m}{E} \Delta x,$$

(11)

where $\Delta x$ is the grid spacing. Here $N$ is chosen so that it is approximately 4 times higher than the minimum acceptable value to damp the elastic waves (Table 1). Maximum and minimum bulk viscosities $\zeta_{\text{max}}$ and $\zeta_{\text{min}}$ values are introduced to regularize the case of vanishing strain rates and to avoid numerical instabilities, respectively. Corresponding limiting values are set for $\eta$ using (6) and (7). Table 1 provides a summary of the parameter values used in the simulations.

Figure 2 represents the elliptical yield curve in the principal stress space ($\sigma_1, \sigma_2$), where $\sigma_1$ is the maximum normal stress and $\sigma_2$ is the minimum normal stress defined as

$$\sigma_1 = \sigma_1 + \sigma_{12} = \frac{1}{2}(\sigma_{11} + \sigma_{22}) + \frac{1}{2}\sqrt{(\sigma_{11} - \sigma_{22})^2 + 4\sigma_{12}^2},$$

(12)

$$\sigma_2 = \sigma_1 - \sigma_{12} = \frac{1}{2}(\sigma_{11} + \sigma_{22}) - \frac{1}{2}\sqrt{(\sigma_{11} - \sigma_{22})^2 + 4\sigma_{12}^2},$$

(13)

Here, $\sigma_1$ and $\sigma_{11}$ are stress invariants representing the average normal stress and the maximum shear stress, respectively. In the context of soil mechanics, cohesion is the ability for a material to sustain shear stress under zero confining pressure. Here, we choose the uniaxial compressive strength $\sigma_{uc}$ of the material as an indication of cohesion. It is defined as the maximal value of $\sigma_1$ that the material can sustain when $\sigma_2 = 0$ (Fig. 3). The dependence of $\sigma_{uc}$, normalized by the ice strength parameter $P$, versus $e$ is given by

$$\sigma_{uc} = \sqrt{2}(1 + e^2)^{-1}$$

(14)
and is depicted in Fig. 4. High shear strength (low \( e \)) means high uniaxial compressive strength \( \sigma_{uc} \) and thus high cohesion. This definition is also valid for other rheologies such as cohesive Mohr–Coulomb or the teardrop yield curves. Note that sea ice is submitted to tension whenever the maximum normal stress \( \sigma_1 \) is positive.

b. Sea ice conditions in Nares Strait

The spatial distribution of ice thickness in Kane Basin and Smith Sound is not well known, mainly because of the lack of data and the high temporal and spatial variability. Wind and ocean currents transport thick multiyear ice floes formed in the ridging zone of the Arctic Basin into Nares Strait. In Kane Basin, these multiyear ice floes mix with first-year ice of medium thickness (0.7–1.2 m) and very thin ice (0.1–0.7 m), which continuously forms due to winter conditions (Tang et al. 2004). A significant portion of the Kane Basin area is occupied by landfast ice with a typical thickness of 1 to 2 m, according to observational (Mundy and Barber 2001) and modeling (April 2006) studies. Taking into account the absence of measurements in the Robeson Channel in the recent period and numerous estimates from previous studies, Kwok (2005) uses a value of 4 m (±0.5) for multiyear sea to calculate the volume flux through Nares Strait. Based on these estimations, we consider that the ice thickness in Kane Basin during late winter polynya events is somewhere between 0.75 (thin first-year ice) and 3.5 m (thick multiyear ice mixed with first-year ice). These extreme values define the interval over which the sensitivity study is conducted.

c. Atmospheric forcing

All winter long, Nares Strait is the scene of strong northerly winds, favoring opening of the polynya when-ever the ice pack is weak enough to rupture at the ice bridge location and drift southward. Ito (1982) showed that the wind speed and direction in Smith Sound are strongly constrained by the steep topography of Greenland and Ellesmere Island. Observational ship data support this fact [see Fig. 2 of Ingram et al. (2002) where the wind direction is parallel to the channel orientation]. Idealized forcing characterized by strong, uniform, and constant winds is thus a reasonable approximation of reality. However, low-resolution global reanalyses usually display large errors in the wind speed and/or direction in the narrow channels of the region. For the realistic simulations presented in the second part of the paper, we use the Canadian operational weather forecast Global Envi-ronmental Multiscale (GEM) model (Côté et al. 1998) daily reanalyses with a resolution of 1.0° × 0.25° (approx-imately 25 km in Smith Sound). GEM wind data have been validated using ship data by April (2006), who found a good correspondence for both direction and speed during the spring and summer seasons of 1998 (i.e., during legs 1 to 4 of the International NOW Polynya Study). The wind direction frequency rose for winter–spring 1998 is shown in Fig. 5 and demonstrates the prevalence of northerly winds during the winter–spring period.

3. Idealized simulations

a. Sensitivity study

The sensitivity study is performed using an idealized domain that mimics the main characteristics of NOW (Fig. 6): a wide rectangular basin representing Kane...
Basin converges linearly into a 46-km-wide channel and opens up again into a wider area. The resolution is such that there are 14 grid cells across the narrowest part of the channel. The initial ice pack spans from the northern end of the domain down to the constriction point and has uniform concentration $c = 1$ (i.e., 100% ice cover, no open water) and a specified thickness $h_0$. Initially, there is no ice in the lower portion of the domain. A northerly wind stress is applied uniformly over the entire domain. We use a constant value of $0.20 \text{ N m}^{-2}$, representing strong wind conditions (approximately 10 m s$^{-1}$) typical of the late spring period in the North Water polynya (Ingram et al. 2002). Initial ice thickness in Kane Basin and shear strength (determined by varying $e$) are free model parameters. Initial thickness $h_0$ takes values from 0.75 to 3.5 m while $e$ is varied from 1.2 to 2.0.

The stability of the ice bridge is assessed after 30 days and is considered stable if the following three criteria are satisfied: 1) the ice edge remains clearly defined as a steplike change in the concentration (from below 0.2 to above 0.8 over one grid cell); 2) the shape of the ice bridge is concave and attached to the coast at the constriction point; and 3) the ice edge position is static or at least moves slowly compared to sea ice free-drift velocity.

Figure 7 shows results of the sensitivity study. Over one hundred simulations were performed to cover the defined range of $e$ and $h_0$ and to optimize model parameters such as the number of ice categories and the viscous, elastic, and advective time steps (see Table 1). Three main behaviors are identified based on previously defined stability criteria. Filled dots represent stable ice arch simulations. Squares represent simulations of unstable ice bridges where ice, at some point during the simulation, is flushed out through the narrow channel. They define, for a given value of $e$ and wind stress, a critical thickness under which sea ice cannot resist the internal stresses imposed by the wind. Finally, triangles represent situations in which the initial ice pack stays rigid and undeformed. Triangles displayed on the graph define an upper boundary for ice thickness, above which the ice is too resistant to form an arch. The maximum thickness increases with decreasing cohesion. Similarly, the squares on the graph define a lower boundary, below which ice is too thin to form a stable arch. The minimum thickness also decreases with cohesion. These boundaries intersect near ($e = 1.8–1.9, h_0 = 2.5–3.0 \text{ m}$), meaning that less cohesive sea ice can never form a stable ice bridge regardless of thickness; it will be either

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**FIG. 5.** Wind direction frequency for January to May given by the GEM 1998 reanalysis. The circle indicates a 25% frequency. The area over which the wind vector speed has been averaged is shown in Fig. 6. Mean wind speed is $8.57 \text{ m s}^{-1}$.

**FIG. 6.** (left) NOW geographical area and (right) the idealized domain used in the simulations. The dashed area represents the region where the wind field has been sampled and averaged (see Fig. 5).
too resistant to deformation or not cohesive enough to be stable.

Stable cases are identified by diagnosing the southward mass transport across a zonal section located upstream of the ice bridge (Fig. 8, top). For $e = 1.2$ to $1.5$ ($h_0 = 1$ m), the upstream transport is constant and stays below $10^5$ kg s$^{-1}$. This value is two to three orders of magnitude smaller than would be the transport induced by free-drifting ice having a velocity of 0.1 m s$^{-1}$. For higher values of $e$ ($>1.5$), the upstream transport increases by two orders of magnitude corresponding to the breakup of the ice flow restriction. The case $e = 1.6$ shows the formation of a stable ice bridge for the first 20 days, which breaks up around day 25. Values of $e$ higher than 1.6 never form a stable ice bridge when the initial thickness is 1.0 m. For each value of $e$, a stable ice arch forms for a range of initial thicknesses, defining an “arching thickness range.” The average thickness at which arching is observed increases with $e$ (decreases with cohesion) until $e$ reaches 1.9. For $e > 1.8$, ice arches form, live for a few days, and break up.

In every simulation, sea ice flow never completely vanishes, even in stable cases. Although the cause has not been investigated here, we suspect numerical diffusion associated with the upstream advective scheme to be mainly responsible [see Ip (1993) for a discussion about the impact of advective schemes in sea ice models]. An ice edge corresponds to a high spatial discontinuity in ice dynamical fields. A slight decrease in sea ice concentration at the ice edge leads to a significant decrease of the sea ice strength $P$ [Eq. (9)], which can switch the flow state from rigid (inside the ellipse) to plastic (on the ellipse). Note that the ocean drag adds to the wind stress near the ice edge. It is strongest at the ice edge, reaching 10% of the total force acting on sea ice, and decreases exponentially with an e-fold distance of 20 km. This mechanical perturbation adds up to the numerical diffusion to explain the nonvanishing transport across the ice edge. Nonetheless, the maximum ice edge displacement velocity is less than 1.5 km day$^{-1}$ (0.017 m s$^{-1}$), and the typical time scale of a strong wind event (a few days) is significantly shorter than the period of stability defined here (30 days).

Diagnosing the mass transport across a zonal transect located downstream of the ice bridge provides additional information on the behavior of the ice edge. In addition to the upstream mass transport, it includes ice

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**Fig. 7.** Sensitivity of ice arching in terms of major to minor axis ratio $e$ and initial thickness $h_0$ (or $P_0 = h_0 P^*$, right vertical axis) using an idealized domain of the North Water polynya. The initial ice pack is forced by a constant and uniform southward wind stress of 0.20 N m$^{-2}$. Three main behaviors are observed: stable ice arches (filled circles), unstable ice arches or plastic flow (squares), and undeformed ice pack (triangles).

**Fig. 8.** Southward ice mass transport (top) upstream of the ice edge and (bottom) downstream for different values of $e$ and $h_0 = 1$ m. Solid lines represent stable ice arch simulations while dashed lines represent unstable cases. The stability is assessed mainly by the upstream transport. Export, characterized by upstream transport, is negligible for $e < 1.6$. 

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detaching from the ice bridge, which affects the form and position of the ice edge. Downstream transport for $h_0 = 1$ m and various values of $e$ are shown in Fig. 8 (bottom). As cohesion increases, the average level of downstream transport decreases, indicating a slower ice edge displacement. For stable cases ($e < 1.6$), the downstream transport has a decreasing trend, suggesting that the ice edge position would stabilize eventually. Some oscillations are observed as cohesion increases, corresponding to a pulse-like displacement of the ice edge. The ice edge draws a concave shape of an arch joining the two landmasses at the constriction point and is resolved by a drop of 0.6–0.9 in ice concentration across one grid cell. Figure 9 shows the ice edge position for different values of $e$. The last panel ($e = 1.6$) represents the ice bridge during breakup at day 28.

\[ \tan 2\psi = \frac{2\sigma_{12}}{\sigma_{11} + \sigma_{22}}. \] (15)

**b. Stress state**

The internal stress profile in the ice pack is similar to what is observed in granular materials stocked in vertical hoppers. Figure 10 (left) shows the principal ($\sigma_1, \sigma_2$) and invariant ($\sigma_I, \sigma_{II}$) stress profiles along the central vertical axis of the channel for $h_0 = 1$ m and two values of $e$ (1.2 and 1.5) in which a stable ice bridge is formed. The orientation $\psi$ of the first principal stress component $\sigma_1$ with respect to the zonal direction is obtained with the following expression:

Figure 10 (right) shows the orientation of the major axis of the ellipse after 30 days for $e = 1.2$ and $e = 1.5$ with $h_0 = 1$ m. In the center of the channel, the major principal stress is directed horizontally, perpendicular to the wind direction. From north to south, the average normal stress $\sigma_I$ reaches a negative plateau and then decreases toward the stress-free surface. The maximum shear stress $\sigma_{II}$, positive by definition, reaches its maximum value right before the stress-free location point due to convergence of the coast. The ice edge forms at the point where the maximum allowable shear stress ($\sigma_{II}$) is reached. As shear strength decreases with increasing $e$, the ice edge forms farther north, where the shear stress is smaller. Below the maximum, stresses rapidly fall to zero and sea ice flows freely downwind of the ice edge. At the ice edge, the major principal stress $\sigma_1$ is positive (tensile) while $\sigma_2$ is negative (compressive), corresponding to a cohesive state. This situation is also expressed as $|\sigma_{II}| \geq |\sigma_I|$, meaning that the maximal shear stress that sea ice can sustain is larger or equal to the average normal stress applied. Internal friction, as defined for granular materials characterized by cohesionless Mohr–Coulomb rheology (Tremblay and Mysak 1997), can no longer explain alone the shear resistance of sea ice. The maximum internal friction angle value that is physically acceptable is $90^\circ$, which represents a triangular yield curve lying exclusively in the third quadrant of the principal stress space (Figs. 2 and 3). Cohesion, which corresponds to the area of the ellipse lying outside the third quadrant, is thus necessary to explain the behavior of sea ice in this particular regime.

From Fig. 10 (right), we also observe a discontinuity in the major principal stress orientation, delimiting two types of regions: a central zone, where the lines are describing an arching profile, and side zones, where the lines are nearly parallel to the converging coastline. This feature is similar to what is observed in funnel flow.
conditions when the granular material inside a hopper only flows in the center part while material adjacent to the walls is jammed (see, e.g., Nguyen et al. 1980). In this context, the side zones are often called “dead zones” because there is no flow in that part. The angle of the line separating the two zones differs for different values of $e$, and thus depends on cohesion. A more cohesive material produces a steeper angle. This phenomenon creates an inner converging channel made of sea ice that explains why the flow state and the arching process are not highly affected by the angle of the converging channel. This was noted by Ip (1993). Similar results have also been observed by Gutfraind and Savage (1998) using a noncohesive Mohr–Coulomb rheology implemented with a smoothed-particle hydrodynamics numerical scheme. However, in their case, the noncohesive rheology never leads to flow obstruction.

c. Coastline definition and grid orientation

In this section, we test the robustness of the solution presented in the previous section, namely that stable ice arching is observed with a clearly defined ice edge in a domain equivalent to the North Water and that it depends on material cohesion. The model domain used so far is characterized by a stair-like converging slope, and it is likely to be the case for an arbitrary realistic finite difference model domain. Here, we build a second idealized domain that has similar dimensions.

![Figure 10](image-url)

Fig. 10. (left) Principal ($\sigma_1$, $\sigma_2$) and invariant ($\sigma_{II}$, $\sigma_{III}$) stress profiles in the center of the channel, and (right) major principal stress ($\sigma_1$) orientation for two values of $e$: (top) $e = 1.2$ and (bottom) $e = 1.5$. 

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567.0x756.0
but rotated by 45°. The converging part of the channel has boundaries perpendicular to each other as in the previous case but are aligned with the grid rather than stairlike. Small differences pertaining to the nonuniform grid aspect ratio of the spherical coordinates, such as the channel width and the wind stress orientation, may affect the solution details, but the main characteristics should not be affected. To test the arching sensitivity to cohesion, the initial ice pack is 1-m thick and different values of \( e \) are used to identify the boundary between stable and unstable arching.

Figure 11 shows the ice concentration after 30 days. The simulation with \( e = 1.5 \) leads to a stable arch while the case with \( e = 1.6 \) goes unstable near the end of the simulation, similar to the nonrotated domain. This shows the invariance of the solution to the grid orientation and the way the coastline is discretized. In section 4, we provide another example of an arbitrary domain.

### 4. The North Water

Although constant and uniform wind conditions are useful to characterize the dynamical response, the model behavior should be tested with a more realistic wind forcing characterized by higher-frequency fluctuations. The model response must be fast, and the ice bridge long-term (one month) stability criterion can be relaxed to the typical time scale of wind events (a few days). The wind field used here is the GEM 1998 daily averaged reanalysis dataset in which the winter winds are characterized by gales producing stresses up to 0.6 N m\(^{-2}\). The initial ice pack is the same as in the idealized cases except that it extends farther south, down to 77°N.

Figure 12 presents the daily wind forcing and the corresponding meridional extent of the ice edge for the first 60 days of 1998 for three values of \( e \) (1.4, 1.7, and 2.0) and three values of \( h_0 \) (1.0, 1.5, and 2.0 m). Points where the ice edge is not resolved by a concentration discontinuity (from below 0.2 to above 0.8) across one grid point are not plotted. The vertical axis of the three last panels spans from 78.4° to 79.1°N, which corresponds to the approximate range of locations where the ice bridge forms as assessed by satellite imagery. The ice edge moves northward in successive steps following strong wind events, corresponding to detachments from the ice pack and restabilizations farther north. Sometimes, the ice edge stops being clearly defined, corresponding to events where sea ice cannot sustain the wind-induced load. During these breakup events, sea ice in the central part of the domain drifts southward and escapes from Kane Basin into Smith Sound (see Fig. 1 for geographical locations). Along the converging coasts, in regions identified as dead zones (see section 3b), sea ice fails by compression and ridges, which increase ice thickness. Compression zones are clearly seen in Fig. 13, especially at low thickness and low cohesion. As a consequence of the breakup, a lead opens in northern Kane Basin and another ice arch forms at the local constriction point. For \( h_0 = 1 \) m, breakup events are observed at day 28 (\( \tau \sim 0.6 \) N m\(^{-2}\)) for the three cases and at day 53 (\( \tau \sim 0.3 \) N m\(^{-2}\)) only for \( e = 1.7 \) and \( e = 2.0 \). The number of days during which the ice bridge is present increases with increasing cohesion (decreasing \( e \)) and increasing thickness. On the other hand, the position of the ice edge is less clearly dependent on cohesion. Figure 13 reveals that even if the ice edge is clearly defined, the ice thickness distribution varies significantly. The strength of the ice bridge is assessed by the amount of sea ice that has flowed out during the entire simulation, which is proportional to the amount of open water upstream of the ice edge. Ice bridges simulated using \( e = 2 \) are relatively weak, even for 2-m-thick ice, compared

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**Fig. 11.** Ice concentration field after 30 days for an initial 1-m-thick ice pack submitted to a 0.20 N m\(^{-2}\) wind stress parallel to the channel. Arching stability is lost when \( e = 1.6 \), as in the case for the nonrotated grid.
to ice bridges formed at $e = 1.4$ and $e = 1.7$. When the ice bridge fails, it is mainly sea ice located in the central part that flows, reminiscent of a funnel flow type. Indications that funnel flow may happen in the North Water can be seen in Fig. 1 where the central part appears to be more granular compared to the more uniform—thus less mobile—portions near the coasts. As opposed to most granular materials, sea ice is compressible and increasingly resistant as sea ice thickness increases due to ridging. Low cohesion leads to more ridging, thus stiffening the ice pack and increasing its ability to obstruct the flow again.

5. Summary and conclusions

Ice arching is observed in geographically constrained areas of the peripheral Arctic and has a major impact on the overall sea ice and freshwater export. The ice bridge that forms in Nares Strait controls the winter sea ice export from the Arctic Basin into Baffin Bay and eventually the northwest Atlantic. It also leads to the opening of the North Water polynya, the largest recurrent polynya in the world. Ice arching and flow obstruction are directly related to the hydromechanical properties of sea ice. Viscous–plastic dynamical models are thus well suited to simulate such behavior. In this study, we have shown for the first time that a clearly resolved ice arch that obstructs sea ice flow can be adequately simulated using an EVP dynamic sea ice model. The EVP model provides a computationally efficient and well-behaved alternative to the classic VP model. Its Eulerian formulation makes it better adapted to climate studies compared to Lagrangian models that require tremendous computational resources when applied to large domains.
Granular materials in general have the ability to support static shear stresses. However, cohesive granular materials are capable of further supporting static stresses with one stress-free surface, leading to a flow obstruction (Walker 1966). The elliptical yield curve, widely used in sea and river ice modeling, is found to successfully simulate the formation of ice arches. We confirm in this paper that cohesion is a necessary and sufficient condition to form arches under continuous wind forcing in a converging channel. We conducted a sensitivity study of the arching process versus shear strength—which determines the level of cohesion within the ellipse—and sea ice thickness, in an idealized domain representing the North Water. It showed that a stable ice arch forms when the major to minor axis ratio $e$ of the elliptical yield curve is lower than 1.8 for ice thicknesses between 0.75 and 3.5 m under a constant 0.2 N m$^{-2}$ wind stress. This corresponds to a uniaxial compressive strength $\sigma_{uc} \approx 0.33 P$ [from Eq. (14) with $e = 1.8$]. Values of $e$ greater than 1.8 lead to unstable ice arches for thicknesses up to 3.5 m under constant and uniform wind stresses greater than or equal to 0.2 N m$^{-2}$ (approximately 10 m s$^{-1}$). There exist minimum and maximum thicknesses between which an ice bridge may form; these define an arching thickness interval for a given cohesive yield curve and wind stress. Those extreme values decrease as cohesion increases. The ice bridge becomes unstable below the minimum thickness, and the ice pack remains undeformed beyond the maximum thickness.

When the domain is rotated by 45°, the converging part of the channel boundaries are resolved by straight lines rather than staircases. The loss of stability for
$h_0 = 1.0 \text{ m}$ as a function of cohesion occurs when $\varepsilon = 1.6$ in both the rotated (Fig. 11) and the nonrotated grid (Fig. 7), showing that the solution is invariant with respect to grid orientation and coastline definition.

When applied to a more realistic representation of the North Water area, the simulated ice edge location and shape compare very well with satellite observations. The ice edge line joins Greenland and Ellesmere Island at the closest point between the two landmasses. We show that the type of ice flow depends on the cohesion, affecting the final ice thickness spatial distribution. For example, when thickness and cohesion are low, compression zones are observed along the converging boundaries where ice ridges and becomes more resistant. As a consequence, only the central part of the ice pack flows out through the narrow channel, which is similar to the funnel flow observed in grain silos under certain conditions. Sea ice thickness measured during ice bridge periods together with simultaneous wind stress data would make a good proxy to evaluate cohesion by providing information about the type of flow and thickness spatial distribution. The ice edge position, on the other hand, is less affected by material properties and stabilizes at a certain distance north of the constriction point ($79^\circ$N). Ice arches formed at large cohesion (low $\varepsilon$) are more resistant to the same wind forcing because they let less ice exit Kane Basin.

Based on idealized simulations of the North Water, the traditional value of $\varepsilon = 2$ does not seem appropriate for high-resolution regional sea ice modeling and ice bridge formation in particular. The conclusion is less clear when one looks at realistic simulation results where this value seems to give reasonable results as well. It is not yet possible to identify a single value that would uniquely improve sea ice modeling performance from local to regional to global scales. Additional exploration and validation work would be necessary to determine the correct level of cohesion to include in a yield curve. In the climate modeling community, cohesion has always been regarded as a constant sea ice property. However, it is reasonable to hypothesize that cohesion can depend on thermodynamic conditions as well. For example, for a given ice thickness, cold winter conditions can glue individual floes together and increase the overall ice pack cohesion and resistance to shear stresses, while warmer conditions could render the ice pack fragile and decrease its capacity to sustain high static shear stresses. There is still a need for sound thickness data before and during polynya events to further explore rheological parameters. Other model parameters remain to be explored as well with respect to the problem of ice arching, mentioning the effect of ocean currents on the ice bridge stability, the sensitivity of the solution with respect to model resolution and the number of ice categories, and the effect of numerical integration schemes of this highly nonlinear problem.

Now that the ice bridge can be prognostically modeled, our future work will focus on the ocean response to the presence of the ice edge, in contrast with previously published studies where the ice edge was prescribed. Turning on the thermodynamics will allow the study of the ice bridge formation, duration, and break up, questions that are becoming increasingly important for local and regional stakeholders in a context of Arctic warming, increasing marine traffic in ice-covered seas, and the uncertain fate of sea ice–dependant ecosystems.

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