



Distinct and synergistic influences of sea ice rheology and seabed stress on simulated Arctic fast ice

Augustin Lambotte¹, Thierry Fichet¹, François Massonnet¹, Pierre Rampal², Laurent Brodeau², Jean-François Lemieux³, Frédéric Dupont⁴, Martin Vancoppenolle⁵, Antoine Barthélemy¹, and Pierre-Yves Barriat¹

¹Université catholique de Louvain (UCLouvain), Earth and Life Institute (ELI), Earth and Climate Research Center (ELIC)

²IGE/CNRS, Grenoble, France

³Recherche en Prévision Numérique Environnementale/Environnement et Changement Climatique Canada, 2121 route Transcanadienne, Dorval QC, Canada

⁴Service Météorologique Canadien, Environnement et Changement Climatique Canada, 2121 route Transcanadienne, Dorval QC, Canada

⁵LOCEAN, IPSL, Paris

Correspondence: Augustin Lambotte (augustin.lambotte@uclouvain.be)

Abstract. Fast Ice (FI) refers to sea ice that remains mechanically immobilised along the coast for extended periods. In the Arctic, its presence influences the stability of the halocline by displacing polynyas offshore and by altering the mixing of river plumes. In numerical ocean-sea-ice models, the realism of simulated FI depends on how grounding processes and sea ice rheology are accounted for. Indeed, grounding parameterisations define if, where, when and how anchor points establish in shallow seas, whereas sea ice rheology dictates the way sea ice dynamics respond to the mechanical forcings. Here, we assess the direct impact of the rheological formulation on the simulation of Arctic FI, as well as its indirect effect when an additional grounding scheme is included. Using the Nucleus for European Modelling of the Ocean - Sea Ice modelling Integrated Initiative (NEMO-SI³) on a 0.25° global grid, we conduct experiments using two different rheological frameworks—the adaptive Elastic-Viscous-Plastic with tensile strength (aEVPts) and the brittle Bingham-Maxwell (BBM), each tested with and without a state-of-the-art grounding scheme. The results show the primary importance of the grounding scheme to simulate Arctic FI. We also find that grounding and rheology interact: BBM produces more rigid sea ice than aEVPts, which favors the formation of FI, but also ridges less easily, which inhibits grounding, and ultimately reduces FI coverage. Comparison with in situ thickness measurements further highlights the importance of accurate timing and duration of immobilisation: when ice is locked early and primarily grows through thermodynamic processes, modelled thickness errors are significantly reduced.

1 Introduction

Fast ice (FI hereafter) refers to sea ice that remains mechanically immobilised along the coast for several weeks. Depending on coastal features and bathymetry, the Arctic FI can extend hundreds of kilometers offshore, such as in the Laptev and East



Siberian Seas. In winter, Arctic FI contributes to approximately 12% of the total sea ice area (Yu et al., 2014) and is mostly found in shallow waters regions near the coasts of the Beaufort, Chukchi (Mahoney et al., 2007b, a), East Siberian, Laptev, and Kara Seas (Polyakov et al., 2003). Its presence affects the underlying Arctic Ocean by influencing the stability of the Arctic halocline, the location of coastal polynyas, and the mixing of river plumes (Itkin et al., 2015; Sterlin et al., 2024; Liu et al., 2025). FI is immobile and mainly grows through thermodynamic processes, which limits its thickness by inhibiting dynamic thickening. Arctic communities, which rely on FI for travel and fishing (George et al., 2004), are exposed to a reduction in the duration of the FI season (Cooley and Ryan, 2024; Yu et al., 2014). Recent efforts have focused on developing seasonal forecasts of the Arctic FI to support on-ice activities (Bieniek et al., 2022; Petrich et al., 2012). More accurate forecasts of FI formation and breakup could help the Arctic communities adapt to ongoing changes. In addition, FI serves as a habitat for animals that depend on immobile sea ice for breeding and hunting (Meier et al., 2014; Cooley et al., 2020; Zhai et al., 2021; Cooley and Ryan, 2024; Fraser et al., 2023).

To form a stable FI cover, sea ice needs anchor points to attach to. Those can be provided by either the coasts or the seabed. FI encroached on the coast can be found in the Kara Sea, where the FI cover is kept by a chain of islands (Divine et al., 2005; Ólason, 2016), or in narrow passages such as the Nares Strait (Kwok et al., 2010). By contrast, seabed anchoring is achieved through the grounding mechanism, when sea ice pressure ridges reach the seabed in shallow water. This mechanism is crucial for the presence of FI in Arctic shallow seas, such as the Beaufort Sea (Mahoney et al., 2007b) or the Laptev Sea (Haas et al., 2005).

To capture the effect of the sea ice grounding mechanism in numerical models, Lieser (2012) forced all sea ice with thickness above a given threshold and found in water shallower than 30 m to remain immobile. A more physical approach was then proposed in Lemieux et al. (2015) by adding a basal stress term (or seabed) to the equation of sea ice momentum. This term is effective when the local ice thickness exceeds a critical value relative to the local bathymetry. The maximum seabed stress is then proportional to the weight of the sea ice above that of the displaced seawater. Dupont et al. (2022) further improved this parameterisation by taking into account the dynamical history of sea ice within each grid cell through the subgrid-scale ice thickness distribution (or Ice Thickness Distribution (ITD)). These parameterisations have proven to be highly effective in shallow waters, enabling accurate FI simulations in such regions. Note that, to account for subgrid-scale coastal geometry, Liu et al. (2022) proposed replacing the usual no-slip condition at the coast with a lateral drag term in the sea ice momentum equation, though this parameterisation is not employed in the present study.



The mechanical resistance of sea ice, whether it is anchored to the coast or to the seabed, is the second physical process that must be accurately represented to simulate a realistic FI cover. In numerical models of sea ice, this behaviour is governed by the rheology, i.e., the constitutive law that describes how the material deforms under mechanical stress. Moreover, the intensity of the deformation affects the sea ice thickness field by increasing (through ridging) or limiting (through opening of leads) dynamical thickening. Different rheological modelling frameworks exist for being applied to sea ice, but the point to which they may differently affect the realism of the simulated FI cover has not been assessed yet.

The most widely used rheologies are derived from the Viscous-Plastic (VP) framework introduced by Hibler (1979), with an elliptical yield curve and a normal-flow rule. In this formulation, sea ice behaves as a highly viscous material under low stresses and deforms plastically when the stresses reach critical values defined by the elliptical yield curve. Although not changing the physical assumptions of the VP rheology, the Elastic-Viscous-Plastic (EVP) approach (Hunke and Dukowicz, 1997) approximates its solution, but improves the computational efficiency of the implicit methods. The computational cost is even further reduced with the use of the adaptive Elastic-Viscous-Plastic (aEVP) scheme (Kimmritz et al., 2016). Dumont et al. (2009) showed that the arching of sea ice (i.e., the typical arc-like structure that sea ice takes in narrow straits, under the action of the winds) can be reproduced in the EVP framework if the aspect ratio of the elliptic yield curve is reduced compared to its usual value originally proposed by Hibler (1979) and equal to 2. This modification effectively increases the uniaxial tensile strength and cohesion. Adding tensile strength has also been found to improve FI simulations in deeper waters (Beatty and Holland, 2010; Lemieux et al., 2016).

Another family of rheologies, based on brittle mechanics, has been developed for sea ice modelling. Following findings by Rampal et al. (2008) and an early Elasto-Brittle (EB) rheology (Girard et al., 2011), Dansereau et al. (2016) introduced the Maxwell Elasto-Brittle (MEB) rheology that was later implemented into a large scale sea ice model, NeXtSIM, by Rampal et al. (2016). With this rheology, sea ice behaves as an elasto-viscous material that can experience damage under critical stresses. In practice, both its elastic modulus and viscosity depend on the degree of damage, which evolves over time according to the local level of internal ice stresses and thermally driven processes. Ólason et al. (2022) extended the MEB framework to develop the BBM, reducing computational cost, and introducing a new threshold to constrain permanent deformation under compression or, in other words, the dynamical thickening of the sea ice cover. Brodeau et al. (2024) implemented this Brittle Bingham Maxwell (BBM) rheology in the Eulerian sea ice model Sea Ice modelling Integrated Initiative (SI3) (Vancoppenolle et al., 2023), the sea ice component of the Nucleus for European Modelling of the Ocean (NEMO) ocean model (Madec et al., 2023). In particular, brittle-based rheologies have been shown to be effective in simulating discontinuities in the sea ice pack, such



as fractures and leads, and in reproducing the observed scaling properties of sea ice deformation (Rampal et al., 2016, 2019; Ólason et al., 2021, 2022).

How does the formulation of rheology affect the simulated Arctic FI? What are the synergistic effects of the rheology and grounding assumptions on the FI representation? Are the effects season- or region-dependent? What affects the simulated thickness of the FI? Which features of the different rheological formulations most strongly influence FI simulation? To answer these questions, we performed four global NEMO-SI3 simulations at 0.25° horizontal resolution, using adaptive Elastic-Viscous-Plastic with tensile strength (aEVPts) and BBM, each run with and without the probabilistic grounding scheme of Dupont et al. (2022). Our analysis focuses on Arctic FI, as Antarctic FI formation depends on different mechanisms such as grounded icebergs (Fraser et al., 2023; Pirlet et al., 2025; Van Achter et al., 2022) which are not represented in our model. We compare the experiments with satellite-derived FI coverage and drill-hole thickness in situ measurements from several coastal stations. This paper is structured as follows: Section 2 describes the numerical model, aEVPts and BBM, the grounding scheme, and the observational data used for validation. Section 3 presents the results, comparing the simulated extent and thickness of the FI with observations. Section 4 discusses the results and their implications.

2 Methods

2.1 Numerical model

In this study, we use the modelling framework NEMO (Madec et al., 2023), version 4.2.2 which includes the Sea Ice modelling Integrated Initiative (SI3) model as a sea ice component (Vancoppenolle et al., 2023) featuring the implementation of BBM by Brodeau et al. (2024). NEMO is a finite-difference, hydrostatic, primitive equation model using the Boussinesq approximation. SI3 is a dynamic-thermodynamic continuum sea ice model. This model has a discrete subgrid-scale ITD (Thorndike et al., 1975). The dynamical processes of sea ice simulated by SI3 include the conservation momentum, with a non-linear internal stress term (rheology), ridging and rafting, and horizontal advection. Regarding thermodynamic processes, the model includes heat conservation, salt dynamics, solar radiation penetration and transmission, a surface albedo function of ice temperature, thickness, snow depth, cloud fraction and melt ponds, a lateral melting parameterisation based on ice floe size, diagnosed from ice concentration (Bitz et al., 2001), snow-to-ice conversion and level-ice melt ponds including frozen lids (Hunke et al., 2013).



105 2.2 Configuration of the numerical model

Both the sea ice and the ocean model share the same horizontal global curvilinear tripolar grid eORCA025 at a quarter of a degree of horizontal resolution. In the ocean, the 75 vertical levels used are unevenly spaced, starting from one meter thickness at the surface to hundreds of meters at the bottom. The advective time step is 720 s for both the ocean and sea ice models. The coupling between the two is done at each advective time step. The sea ice model ITD uses five thickness categories with the boundaries set at 0, 0.45, 1.13, 2.14, 3.67 and 99 m (Massonnet et al., 2019).

All simulations are initialized with the same ocean and sea ice restart files, corresponding to Jan 1st, 1990. These restart files have been obtained after a 15-year spin-up, starting from rest with the ocean temperature and salinity fields from the World Ocean Atlas 2018 (WOA) (Zweng et al., 2019; Boyer et al., 2018; Locarnini et al., 2019). The sea surface salinity is restored toward the WOA climatology. The restoring timescale is 227 days, assuming a mixed layer depth of 50 m. This restoring is commonly used in ocean models to prevent unrealistic model drift. The restoring strength decreases linearly with increasing sea ice concentration to avoid interfering with ocean-ice interactions, so that this restoring is inactive where the ocean is fully covered by ice. The atmospheric forcing is the European Center for Medium Range Weather Forecasts (ECMWF) ReAnalysis v5 (ERA5, Hersbach et al. (2023)). An extensive description of the model configuration can be found in Richaud et al. (2026). The only differences with Richaud et al. (2026) are the time step, reduced here to 720 s, and the ones describe hereafter.

125 2.3 Sea ice rheologies and grounding scheme

In this section, we briefly present the two rheologies (aEVpts and BBM) and the grounding scheme (Dupont et al. (2022), hereafter referred to as D22) used. For a more complete presentation of these rheologies, we refer to Hibler (1979). Hunke and Dukowicz (1997). Kimmritz et al. (2016). Lemieux et al. (2016) for aEVpts and to Ólason et al. (2022). Brodeau et al. (2024) for BBM.

The vertically integrated sea ice momentum equation is given by

$$m \frac{D\mathbf{u}}{Dt} = \nabla \cdot \boldsymbol{\sigma} + \boldsymbol{\tau}_a + \boldsymbol{\tau}_w + \boldsymbol{\tau}_b - m f \mathbf{k} \times \mathbf{u} - m \hat{g} \nabla H, \quad (1)$$

130 where m is the mass of sea ice and snow per unit area, $\mathbf{u} = u\mathbf{i} + v\mathbf{j}$ the sea ice velocity vector with u and v the zonal and meridional velocities, respectively, and $\mathbf{i}$ and $\mathbf{j}$ the orthogonal horizontal units vectors. $\boldsymbol{\sigma}$ is the internal stress tensor, $\boldsymbol{\tau}_a$ and $\boldsymbol{\tau}_w$



are the atmospheric and oceanic stresses, respectively, and τ_b the seabed stress calculated following the D22 parameterisation, f is the Coriolis parameter, $\mathbf{k}$ the vertical unit vector, $\hat{g}$ is the gravitational acceleration and H is the sea surface height.

- 135 The rheology constitutive equation defines the relationship between the internal stress in the ice σ and its deformation rate $\dot{\epsilon}$. σ and $\dot{\epsilon}$ are both second order tensors with three independent components in two dimensions:

$$\sigma = \begin{pmatrix} \sigma_{11} & \sigma_{12} \\ \sigma_{12} & \sigma_{22} \end{pmatrix} \quad (2)$$

$$\dot{\epsilon} = \begin{pmatrix} \dot{\epsilon}_{11} & \dot{\epsilon}_{12} \\ \dot{\epsilon}_{12} & \dot{\epsilon}_{22} \end{pmatrix} = \begin{pmatrix} \partial_x u & \frac{1}{2}(\partial_y u + \partial_x v) \\ \frac{1}{2}(\partial_y u + \partial_x v) & \partial_y v \end{pmatrix}$$

Two invariants (σ_I and σ_{II}) can be constructed from the three elements (σ_{11} , σ_{22} and σ_{12}) of the stress tensor σ :

$$\begin{cases} \sigma_I = \frac{1}{2}(\sigma_{11} + \sigma_{22}) \\ \sigma_{II} = \sqrt{\left(\frac{\sigma_{11} - \sigma_{22}}{2}\right)^2 + \sigma_{12}^2} \end{cases} \quad (3)$$

where σ_I is the isotropic normal stress and σ_{II} the maximum shear stress.

- 140 Note that the stresses are vertically integrated and are therefore expressed in N m^{-1} , as is usually done in the viscous-plastic framework. In the original description of BBM (see Ólason et al. (2022) and Brodeau et al. (2024)), the stresses are expressed in N m^{-2} . Consequently, the stress term in the sea ice momentum equation is written as $\nabla(h \cdot \sigma)$ rather than $\nabla \sigma$, as done here. Our presentation of BBM therefore differs slightly from the one presented in Ólason et al. (2022) and Brodeau et al. (2024) because we consider vertically integrated stresses.

145 2.3.1 Adaptive elastic-viscous-plastic rheology with tensile strength

In the VP rheological framework (Hibler, 1979), sea ice behaves as a highly viscous material under low stress and as a plastic material when the stresses reach critical values defined by a yield curve. The constitutive equation of EVP for an elliptical yield curve with tensile strength is

$$\dot{\epsilon}_{ij} = \frac{1}{E} \frac{\partial \sigma_{ij}}{\partial t} + \frac{1}{2\eta} \sigma_{ij} + \frac{\eta - \zeta}{4\eta\zeta} \sigma_{kk} \delta_{ij} + \frac{P - T}{4\zeta} \delta_{ij} \quad (4)$$

- 150 where η and ζ are the shear and bulk viscosities, respectively, and δ_{ij} is the Kronecker delta. The Young's modulus E is not a physical property of sea ice but is chosen such that elastic waves dissipate before the forcings are updated. The compressive



strength P and tensile strength T depend on the grid-cell-mean thickness h and concentration A , and are defined as:

$$\begin{cases} P = P^* h e^{-C(1-A)} \\ T = k_t P, \end{cases} \quad (5)$$

where P^* and C are constant parameters, and $k_t \in [0, 1]$ controls the amount of tensile strength. A value of $k_t = 1$ gives sea ice equal strength in tension and compression, while $k_t = 0$ recovers the original EVP formulation without tensile strength. The

155 viscosities are expressed as follows:

$$\begin{cases} \zeta = \frac{P+T}{2\Delta} \\ \eta = \frac{\zeta}{e^2} \end{cases} \quad (6)$$

where $\Delta = \sqrt{(\dot{\epsilon}_{11}^2 + \dot{\epsilon}_{22}^2)(1 + e^{-2}) + 4e^{-2}\dot{\epsilon}_{12}^2 + 2\dot{\epsilon}_{11}\dot{\epsilon}_{22}(1 - e^{-2})}$ is a measure of the deformation rate and e is the ellipse ratio of the elliptical yield curve. A lower value of e increases the uniaxial tensile strength and cohesion of sea ice. A lower limit, Δ_{min} , is imposed to Δ to prevent the viscosity from tending to infinity. The corresponding maximum viscosity, as studied in Ólason (2016), is given by $\zeta_{max} = \frac{1+k_t}{2\Delta_{min}} P$.

160 2.3.2 Brittle Bingham-Maxwell rheology

In BBM (Ólason et al., 2022), sea ice behaves as an elastic material under low stress and as a viscous-elastic material under high stress. The constitutive equation is

$$\frac{\partial \sigma_{ij}}{\partial t} = \frac{E}{1-\nu^2} \left[(1-\nu)\dot{\epsilon}_{ij} + \nu\delta_{ij}\dot{\epsilon}_{kk} \right] + \frac{1}{\lambda} \sigma_{ij} \left(1 + \tilde{P} + \frac{\lambda}{1-d} \frac{\partial d}{\partial t} \right), \quad (7)$$

with ν the Poisson ratio, $d \in [0, 1]$ a damage scalar representing the subgrid-scale fracturation of sea ice and $\tilde{P}$ a parameter, defined as a purely compressive stress threshold, which controls the transition between purely elastic and viscous-elastic regimes. The elastic modulus E and viscosity λ decrease with increasing damage d , following

$$\begin{cases} E = E_0(1-d)h e^{-C(1-A)}, \\ \lambda = \lambda_0[(1-d)e^{-C(1-A)}]^\alpha, \end{cases} \quad (8)$$

where E_0 and λ_0 are the elasticity and viscosity of undamaged sea ice ($d = 0$), respectively, and $\alpha > 1$ is a constant. The parameter $\tilde{P}$ is defined as:

$$\tilde{P} = \begin{cases} 0 & \text{if } \sigma_I > 0 & \text{(viscous-elastic divergence),} \\ -1 & \text{if } -P_{max} < \sigma_I < 0 & \text{(elastic convergence),} \\ \frac{P_{max}}{\sigma_I} & \text{if } \sigma_I < -P_{max} & \text{(viscous-elastic convergence.)} \end{cases} \quad (9)$$



170 P_{max} , the parameter controlling sea ice thickening through sub-grid scale ridging, depends on local thickness h and concentration A as

$$P_{max} = P_0 \left(\frac{h}{h_0} \right)^{\frac{3}{2}} h e^{-C(1-A)} \quad (10)$$

where h_0 and P_0 are constant parameters. The damage scalar d increases if the stress state exceeds the Mohr-Coulomb criterion $\sigma_{II} > -\mu\sigma_I + c$, with μ the internal friction coefficient and c the cohesion of the ice. The cohesion scales with the model resolution following

$$c = c_{ref} h \sqrt{\frac{l_{ref}}{\Delta x}} \quad (11)$$

where c_{ref} is the reference cohesion at resolution $l_{ref} = 0.1$ m and Δx is the local grid spacing of the model.

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To account for the refreezing of the subgrid-scale fractures (represented by damage), a healing mechanism is implemented. The strength of healing is proportional to the difference between the basal and surface temperatures of the ice ΔT_h and a healing constant k_{th} following

$$\frac{\partial d}{\partial t} = -\frac{\Delta T_h}{k_{th}}. \quad (12)$$

2.3.3 Probabilistic seabed ice keel interaction

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With D22, the seabed stress term τ_b in the sea ice momentum equation (Equation 1) is calculated as:

$$\tau_b = -\tau_b^{max} \frac{\mathbf{u}}{|\mathbf{u}| + u_0} e^{-\alpha_b(1-A)}, \quad (13)$$

with α_b a constant and u_0 a small velocity scale that ensures a smooth transition between static and dynamic regimes. The calculation of τ_b^{max} is based on the probability of contact between sea ice and the seabed, assuming that both are random variables with a known Probability Density Function (PDF) on the grid cell scale. Taking into account $g(x)$, the PDF that gives the likelihood of finding a sea ice thickness x within a model grid cell and $b(y)$ the probability of finding a depth y , τ_b^{max} is given by:

$$\tau_b^{max} = \mu_s \hat{g} \int_0^\infty dx \int_0^{D(x)} dy (\rho_i x - \rho_w y) g(x) b(y), \quad (14)$$

where μ_s is a static friction coefficient, $D(x) = \rho_i x / \rho_w$ is the sea ice draft, ρ_i and ρ_w are the densities of sea ice and seawater, respectively.



190 To represent the tail of the ITD, which is crucial for grounding, the PDF $g(x)$ used to calculate interactions with the seabed is a log-normal distribution fitted to the actual simulated ITD, so that they share the same mean and variance. The bathymetry PDF is assumed to be Gaussian with a mean equal to the local bathymetry and a constant standard deviation of $\sigma_b = 2.5$ m. The ice thickness PDF is truncated at the 99.7th percentile and the bathymetry at $\pm 3\sigma_b$. These parameter values were obtained in Dupont et al. (2022) after a tuning based on experiments using aEVPts.

195 2.4 Experimental setup

We perform four simulations with the model configuration detailed above between 1990 and 2016. Two experiments use aEVPts and the other two BBM. For each pair of simulations, one simulation has the grounding scheme activated and the other one does not.

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In order to have realistic, but also comparable, mean drift speed with the two rheologies, we use distinct atmospheric sea ice drag coefficients (C_d^i). This approach is similar to the one used in Brodeau et al. (2024) and ensures a fair comparison of the two rheologies. The values of this coefficient are $C_d^i = 1.3 \times 10^{-3}$ for aEVPts and $C_d^i = 1.6 \times 10^{-3}$ for BBM. To obtain these values, we compared the simulated mean sea ice drift in winter (October-April) over the Arctic pack ice region between 205 1991 and 1994 with satellite-based estimates from the Ocean and Sea Ice Satellite Application Facility (OSI-SAF) data set (OSI SAF, 2022) (see Appendix A). Except for this parameter, the rheological formulation, and the use of a grounding scheme or not, all the simulations are identical. Table 2 recaps the values of each parameter relevant to this study. The large-scale performances of the four experiments are discussed in the Appendix B.

210 BBM currently has slightly higher computational cost than aEVPts for the same number of iterations (sub-timesteps in aEVPts and sub-cycles in BBM). In our 27-year experiments without grounding, the coupled model with BBM last 1.27 times longer than with aEVPts. The extra cost is due to the grid demultiplication used in BBM (the E-augmented C-grid approach described in Brodeau et al. (2024)) and the additional computations required to advect the damage tracer and stress tensor components.



Table 1. The four performed experiments in this study.

Experiment name	Rheology	Grounding scheme
aEVpts_no_grounding	aEVpts	None
BBM_no_grounding	BBM	
aEVpts_grounding	aEVpts	D22
BBM_grounding	BBM	

Table 2. Values of principal parameters used in this study.

Parameter	Definition	Value
General parameters		
C_d^o	ice-ocean drag coefficient	5×10^{-3}
μ_s	static friction coefficient for grounding scheme	0.7
Δt	advective time step	720 s
C	compaction	20
aEVpts related parameters		
P*	ice strength parameter	2×10^4 Pa
e	ellipse eccentricity	1.4
k_t	ratio between tensile and compressive strengths	1/20
Δ_{min}	lower limit for deformation rate	$2 \times 10^{-9} \text{ s}^{-1}$
N_{aEVpts}	number of aEVpts subcycles	100
C_d^i	ice-atmosphere drag coefficient	1.3×10^{-3}
BBM related parameters		
k_{th}	healing constant for damage	24 K s
c_{ref}	cohesion at reference scale	2×10^6 Pa
Δt_{BBM}	dynamical time stepping for BBM	7.2 s
E_0	elasticity of undamaged sea ice	5.96×10^8 Pa
λ_0	viscosity of undamaged sea ice	10^7 s
μ	internal friction coefficient	0.7
P_0	scaling parameter for elastic regime limit	10^4 Pa
ν	Poisson's ratio	1/3
α	damage parameter	5
C_d^i	ice-atmosphere drag coefficient	1.6×10^{-3}



2.5 Comparing simulated FI to observational data products

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The FI simulated with both rheologies, with and without grounding, is validated against two types of observational products: charts of FI presence and FI thickness from drill-hole measurements in the Canadian Arctic Archipelago (CAA) and the Russian Arctic.

220 Bi-weekly FI charts released by the National Snow and Ice Data Center (NSIDC) (U.S. National Ice Center, 2006), available on a 25 km grid from 1972 to 2007, are used to evaluate the simulated FI extent, i.e., the FI covered surface. For the period between 2007 and 2016, we use an extension of this dataset produced by Dupont et al. (2022). These charts are manually produced by ice chart specialists without a consistent criterion. Consequently, comparing them with model outputs is not straightforward. Lemieux et al. (2015) identified a model grid cell as FI covered if its two-weekly averaged speed is below a

225 threshold of $5 \times 10^{-4} \text{ m s}^{-1}$. The two-week averaging ensures that immobile sea ice caused by temporary low wind events is not misclassified as FI. However, slow drifting sea ice that is too far from the coast can occasionally meet this criterion and be considered as FI, although it would not be accounted for by the NSIDC specialist. For instance, sea ice between Novaya Zemlya and the mainland, in the western Kara Sea, occasionally meets this criterion in our simulations but is excluded from the NSIDC dataset due to its too large distance from the coast. Since no predefined FI region mask is available from NSIDC

230 and the classification is operator-dependent, we constructed a mask (black line in Fig. 1) using all points with at least one FI observation in the NSIDC dataset between 1990 and 2016. We therefore identify a model grid cell as FI covered if the two weeks averaged sea ice drift in that grid cell is lower than the velocity threshold and is within the mask described above. Note that the use of this mask has barely no impact on the seasonal cycles of FI extent presented in this paper.

235 The FI simulated with the two rheologies, with and without grounding, is compared to the NSIDC data set. Following Yu et al. (2014) and Lemieux et al. (2016), we calculated the frequency of occurrence of FI between January and May, the period of the year when the FI cover is stable (Fig. 2). The frequency of occurrence, defined as the fraction of days during which a grid cell is classified as FI over the total number of days in the January-May period, was averaged over 1991-2016, and mapped (Fig. 2). The 10% contour of this frequency of occurrence obtained from the NSIDC dataset is also shown for comparison.

240 The seasonal cycle of FI extent is also evaluated for nine Arctic regions over 1991-2016, for the four simulations and the observational NSIDC reference dataset (Fig. 1). Since simulations do not account for tidal processes, we define the CAA_NT (Canadian Arctic Archipelago No Tides) region by excluding a portion of the CAA where Lemieux et al. (2018) showed that tides have large effects on the simulated FI.



245 For FI thickness, we use drill-hole thickness measurements from stations in the CAA and the Russian Arctic. The CAA
measurements are available at <https://www.canada.ca/en/environment-climate-change/services/ice-forecasts-observations/latest-conditions/archive-overview/thickness-data.html> (last accessed: 2024-11-14). We identified six stations with measurements
performed on sea ice and excluded those taken on lakes. These measurements are performed weekly and span the entire period,
with a gap in 2001. In the Russian Arctic, we used data from 15 stations with only the annual maximum thickness without the
250 measurement date (Polyakov et al., 2017). The coordinates of the stations are listed in Table 3, and their locations are shown
in Fig. 1.

Comparing on-site measured sea ice thicknesses with modelled ones is done as follows. The drill-hole measurements are made
on level ice, which corresponds to the mode (i.e., the most frequent thickness) of the local sea ice thickness distribution.
255 Therefore, comparing this measurement with the mean sea ice thickness of the grid cell may not be ideal, as the mean also
accounts for the thickness of pressure ridges and thin ice. For comparison, we interpolated the model ITD, with five categories,
to an ITD of ten thousand categories, assuming a log-normal distribution with the same mean and variance. The mode of this
interpolated ITD is what we use to compare with the drill-hole measurement.

3 Results

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In this section, we compare the simulated Arctic FI in our four experiments with each other and with observational references.
The FI extent and spatial distribution are compared to satellite-based products in various Arctic regions. We also compare the
simulated FI thickness with drill-hole FI measurements from 21 coastal stations in the Arctic.

3.1 FI occurrence frequency and extent

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In the two simulations with grounding, the simulated seasonal cycle of FI is reasonably well captured, regardless of the
rheological formulation (Fig. 3). Without grounding, the only region where the simulated FI is close to the observations is
the CAA_NT. In all other regions, the extent of FI is largely underestimated in both **aEVPts_no_grounding** and **BBM_no**
_grounding. This suggests, as already shown in previous works, that grounding is essential to the formation of Arctic FI
270 (Lemieux et al., 2015, 2016; Dupont et al., 2022). In the East Greenland Sea, all simulations largely underestimate the ob-
served FI extent. A potential explanation lies in the absence of icebergs in our numerical setup. While icebergs are known to
act as anchoring points for FI in general (Fraser et al., 2023), and northeast Greenland is characterised by significant iceberg

Table 3. List of drill-hole stations and their coordinates used for FI thickness validation.

Station Name	Latitude	Longitude	Location	Thickness observations
Eureka	79.98°N	85.95°W	CAA	~ Weekly measurements, with date.
Resolute	74.71°N	94.98°W		
Cambridge Bay	69.11°N	105.13°W		
Hall Beach	68.78°N	81.23°W		
Iqaluit	63.75°N	68.55°W		
Coral Harbour	64.13°N	83.26°W		
Maliye Karmakuly	72.38°N	52.73°E	Russian Arctic	Annual maximum, without date.
Amderma	69.77°N	61.68°E		
Belyi	73.33°N	70.05°E		
Dikson	73.50°N	80.23°E		
Sterlegova	75.42°N	88.90°E		
Uedinenia	77.50°N	82.23°E		
Golomiannii	79.55°N	90.62°E		
Cheluskin	77.72°N	104.30°E		
Sannikova	74.67°N	138.90°E		
Kotelnii	76.00°N	137.90°E		
Shalaurova	73.18°N	143.93°E		
b. Tiksi	71.58°N	128.92°E		
Chetirekhstolbovii	70.63°N	162.40°E		
Valkarkai	70.08°N	170.93°E		
Vrangelya	70.98°N	181.47°E		

and FI presence (Reeh et al., 2001; Deakin et al., 2026), the precise role of icebergs in the FI formation of this specific region remains to be established.

275

In **BBM_no_grounding**, the simulated FI extent increases generally faster compared to **aEVPts_no_grounding**, and reaches generally higher levels throughout the FI season. The exceptions are the East Siberian Sea, where both rheologies produce similar coverages, and the Laptev, Chukchi and Barents Seas, for which the curves of FI extents converge by the end of the season (Fig. 3). The maps of the FI frequency of occurrence (Fig. 2) show that the maximum extents are comparable, although a higher occurrence frequency within these areas is found for **BBM_no_grounding** compared to **aEVPts_no_grounding**. Differences in FI extent across rheologies are smaller when grounding is activated. In other words, aEVPts seems less natively capable of simulating immobilised sea ice than the BBM in the absence of grounding, but its stronger response to the grounding scheme compensates for this initial difference. However, we note that differences remain even when the grounding is taken

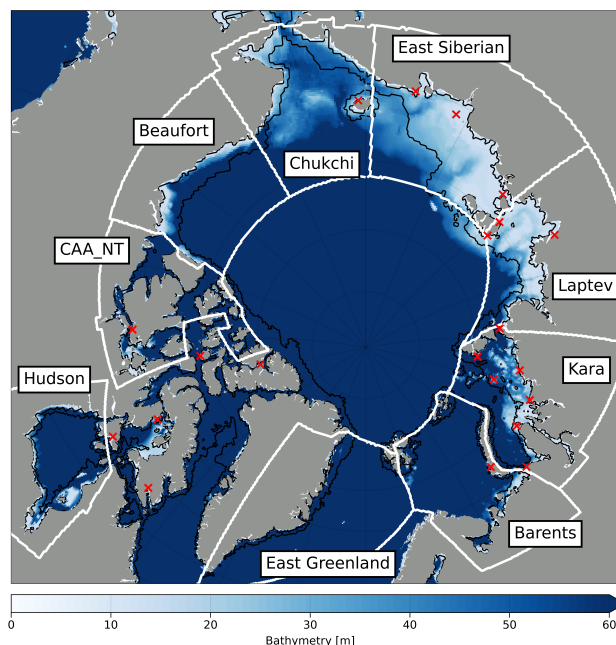


Figure 1. Shallow water bathymetry of the Arctic Ocean. The nine regions defined in this study are delineated in white. The red crosses are the locations of the drill-hole stations listed in Table 3. The black line is the mask generated from all the points with at least one FI observation in the NSIDC dataset between 1990 and 2007 and the dataset extension of Dupont et al. (2022) between 2007 and 2016.

into account. For instance, the maximum FI extent generally peaks later in **aEVpts_grounding** compared to **BBM_grounding**
 285 in the regions where the grounding has a significant impact - Kara, Laptev, East Siberian, Beaufort, Chukchi and Beaufort
 Seas. Also, the largest differences in the FI extent and frequency of occurrence between the **aEVpts_grounding** and **BBM**
grounding simulations are obtained in the East Siberian Sea.

With BBM, the simulated FI cover has a higher resistance to compression than with aEVpts. Indeed, BBM is characterised
 290 by a pure elastic (i.e. not allowing for any permanent deformation) behaviour under low compression stress and an elasto-
 viscous regime under high compression stress. For highly damaged sea ice, the elasto-viscous regime can lead to important
 deformations, but those are limited when the damage decreases. By contrast, the aEVpts compression regime is viscous under
 low compression stresses, with a high viscosity which prevents any detectable deformation, and fails plastically under high
 compression stresses. aEVpts is therefore not able to sustain any compression higher than its limits (given by Equation 5),
 295 where it deforms instantaneously due to plastic failure. On the other hand, being outside the BBM compression limit (Equation
 10) only results in the addition of a viscous term, allowing permanent but not instantaneous deformations.



This higher resistance of the FI to compression with BBM has two impacts on the simulated FI cover, one direct and one indirect. The direct impact is highlighted by the two simulations without grounding. The higher extent of FI simulated in **BBM**
 300 **_no_grounding** compared to **aEVpts_no_grounding** is the consequence of the more rigid behaviour of BBM. The indirect impact arises from the dependence of the grounding scheme to the sea ice thickness. In the BBM framework, the FI higher resistance in compression reduces the dynamical thickening, e.g., when the wind pushes sea ice against the coast. This thinner FI cover leads to a weaker grounding and, therefore, to a FI cover that extends less and breaks up earlier in the melting season. This explains why **aEVpts_grounding** has a FI extent that peaks later in the season compared to **BBM_grounding**. This
 305 is also why, while **BBM_no_grounding** simulates greater extents of FI compared to **aEVpts_no_grounding**, a systematic difference is no longer discernible between **BBM_grounding** and **aEVpts_grounding**.

The influence of the BBM limited dynamical thickening on the grounding scheme is particularly pronounced in the East Siberian Sea. Figure 4 shows the frequency of occurrence of days with non zero seabed stress term in Equation 1 for the two
 310 experiments with grounding. The smaller number of grounding events in **BBM_grounding** compared to **aEVpts_grounding** explains the lower frequency of occurrence of FI observed in this region (Fig. 2). Weaker grounding in **BBM_grounding** is associated with thinner sea ice in this region (panel **a** in Fig. 5). As expected, the difference in sea ice thickness is most pronounced along the East Siberian Sea islands, where the wind (arrows on the same map) pushes the sea ice against the coasts.

315

This thickness difference is explained by the higher resistance of BBM in compression. Figure 5 (panel **b**) displays the density distribution of the stress state in the invariant stress plane (σ_I , σ_{II} , see Equation 3) for **BBM_grounding**, during the months of January to April in the region represented by the blue box in panel **a**. In that same figure, specific characteristics of the two rheologies are shown for different thicknesses ranging from half a meter to 2 m (the range of FI thickness in the region). Those
 320 are the Mohr-Coulomb failure envelope (diagonal lines) and the compressive stress threshold (vertical lines) in BBM in one hand, and the elliptical *failure* envelope (curved lines) assumed in aEVpts in one other hand. The distribution of BBM stress state shows that sea ice with BBM withstands higher compression than with aEVpts, being for a significant time outside of what would be allowed by aEVpts yield curve. When the compression corresponds to plastic deformation with aEVpts, BBM is most of the time in elasto-viscous regimes. In other words, in situation where the compressive stress would be high enough to
 325 trigger plastic deformations with aEVpts, it deforms more gently following BBM viscous-elastic regime, attenuating possible ridging.



The plastic deformations occurring with aEVPTs are triggered at the beginning of the FI season in the East Siberian Sea, allowing for greater initial thickening of the FI cover. In Fig. 5, we show the mean divergence (Panel **c**) and the mean sea ice thickness (Panel **d**), between September 1995 and September 1997, calculated over the area defined in Panel **a** of the same figure. The beginning of the season is marked by stronger convergence rates with aEVPTs, leading to more thickening than with BBM. A larger prevalence of sea ice in the thickest categories, associated with thicker ice, provides more anchoring points and a more stable FI cover, which can therefore grow thermodynamically. Conversely, the FI simulated with BBM, with less ice in the thickest categories, has fewer anchoring points and is therefore less stable, with occasional breakup marked by divergent events and thinning in the area, further limiting the mean FI thickness. The same figure shows how the thicker FI cover obtained with aEVPTs leads to later breakup and even a possible persistence of sea ice through summer.

Changing the rheological framework affects not only the amplitude of the FI extent seasonal cycle, but also the timing of the FI season. As shown in Fig. 3, the FI decline starts earlier in **BBM_grounding** than in **aEVPTs_grounding**. This timing difference stems from the thicker sea ice found in compression zones with aEVPTs, a direct consequence of its smaller resistance to compression compared to BBM. At the onset of the melting season, this thicker ice sustains mass loss for longer before reaching the critical thickness at which it becomes ungrounded, delaying FI breakup. This results in a difference between the timing of FI breakup between the two rheologies, with the one producing thinner coastal sea ice facing earlier FI breakup.

3.2 FI thickness at observation stations

345

We compare in this section simulated FI thicknesses to in situ drill-hole measurements at Arctic stations (see Table 3 for station's location). The bi-weekly measurements in the CAA allow direct comparison to the simulated seasonal cycle of FI thickness at those stations (Fig. 6). The CAA seasonal cycle of FI is relatively close across the four simulations. Some small differences can be found when changing rheology at Coral Harbour and Hall Beach. The largest bias is found at Eureka, where the four simulations present a persistent positive bias of about 1 m, likely due to the unrealistic year-round persistence of FI in the simulations. The error histograms (Fig. 7) confirm that neither rheology nor grounding significantly affect the thickness of the FI in the CAA where it is simulated even in the default model configuration. The only difference is due to the fact that with BBM, higher occurrence of FI is found at the stations resulting in a greater number of data points in the distributions. This higher occurrence of FI is consistent with the difference in frequency of occurrence in CAA between the two rheologies (Fig. 2).

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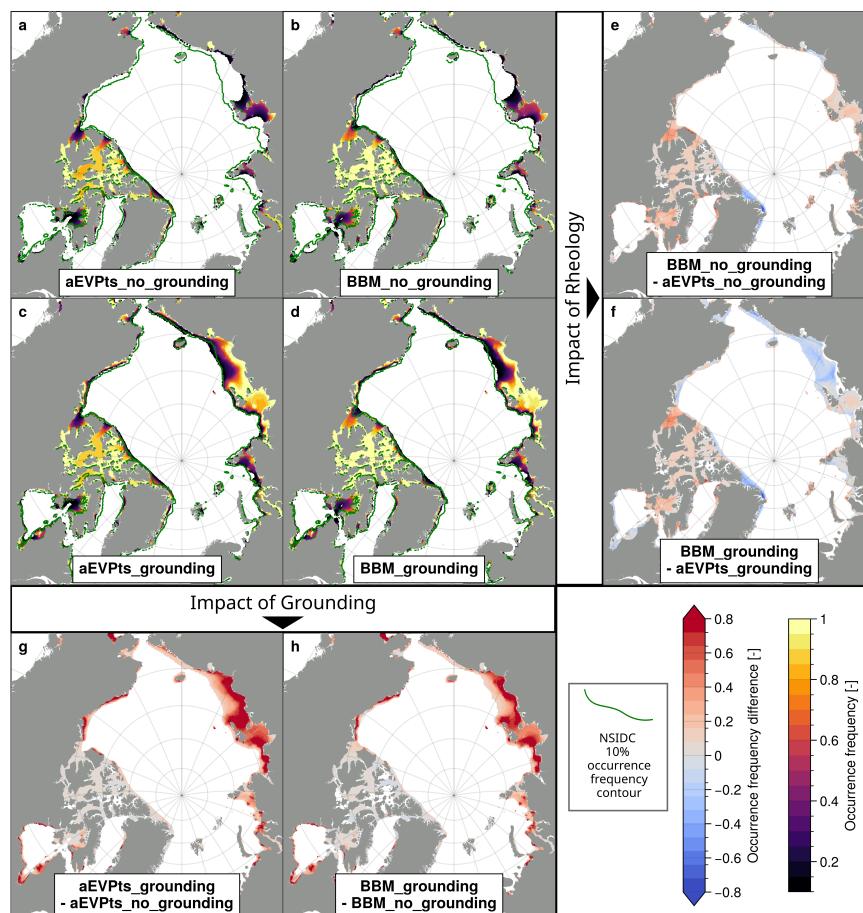


Figure 2. The occurrence frequency of simulated FI (a: aEVPTs_no_grounding, b: BBM_no_grounding, c: aEVPTs_grounding and d: BBM_grounding), between January and May, averaged from 1991 to 2016. The occurrence frequency is calculated as the fraction of days during which a grid cell is classified as FI over the total number of days in the January-May period. The last four panels show the impact of grounding with both rheologies (differences between g: aEVPTs_grounding and aEVPTs_no_grounding, h: BBM_grounding and BBM_no_grounding) and the impact of rheology, with and without grounding (difference between e: BBM_no_grounding and aEVPTs_no_grounding, f: BBM_grounding and aEVPTs_grounding). The green contour shows the 10% occurrence frequency from satellite observations.

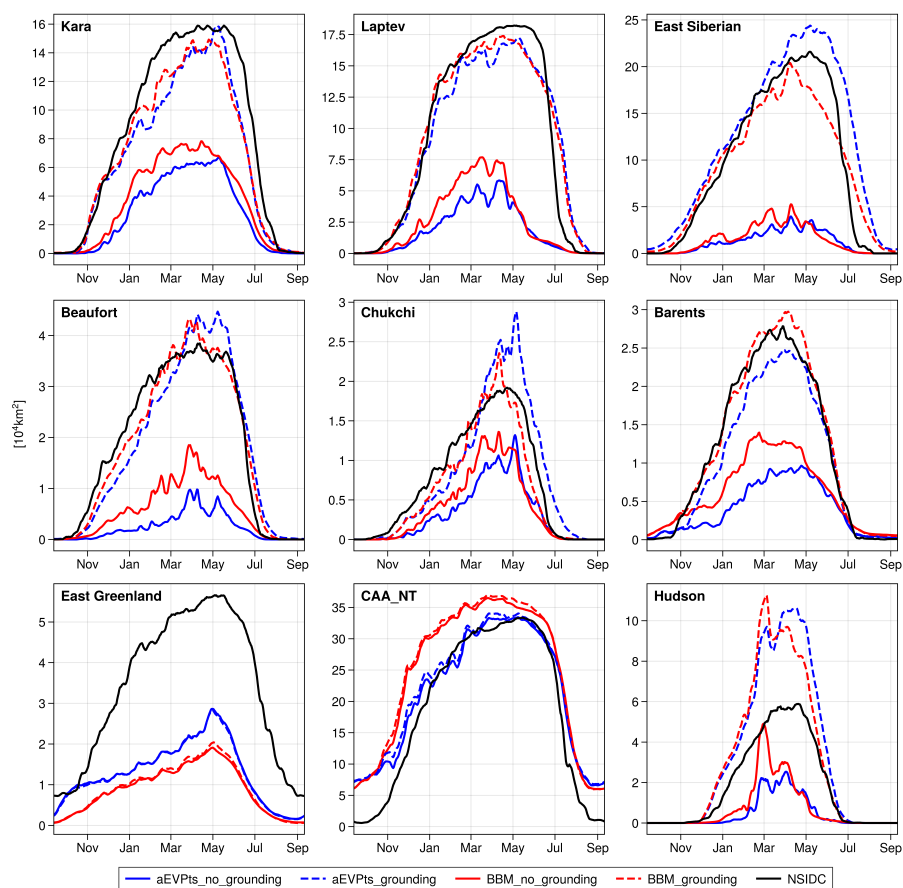


Figure 3. Seasonal cycle, averaged between 1991 and 2016, of FI extent in the nine Arctic regions defined in Fig. 1. The black line corresponds to the ice charts, the blue solid line to **aEVpts_no_grounding**, the red solid line to **BBM_no_grounding**, the blue dashed line to **aEVpts_grounding** and the red dashed line to **BBM_grounding**. A five days running mean filter has been applied to all curves for readability.

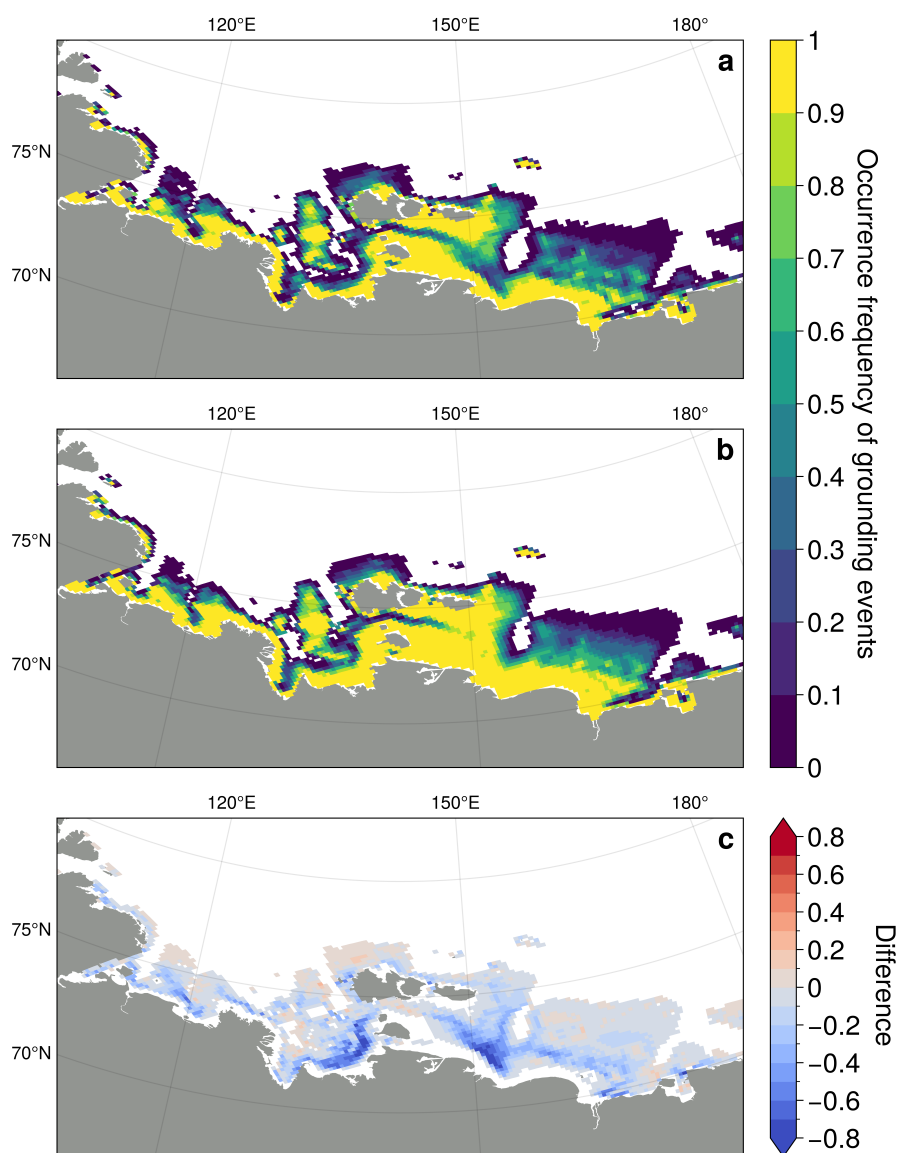


Figure 4. Effect of rheological formulation on grounding frequency occurrence in the Laptev and East Siberian Seas. Frequency of occurrence of days with non zero seabed stress (see Equation 1,) from January to April between 1991 and 2016, for **BBM_grounding** (panel a), **aEVpts_grounding** (panel b) and the difference between **BBM_grounding** and **aEVpts_grounding** (panel c).

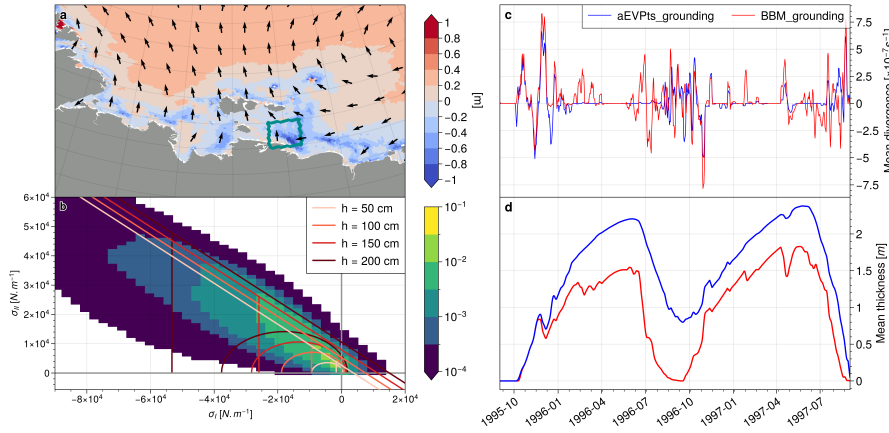


Figure 5. **Panel a:** zoom in on the Laptev and East Siberian Seas. Colors indicate the mean sea ice thickness differences between **BBM_grounding** and **aEVpts_grounding** from January to April, between 1991 and 2016. The arrows represent the mean wind direction from ERA5, the atmospheric forcing, over the same period. **Panel b:** density plot of the stress state in **BBM_grounding**, during the same period and within the cyan box displayed in **panel a**. The diagonal lines define the Mohr-Coulomb criterion, the brittle limit for BBM. The vertical lines represent the threshold separating the BBM elastic and viscous-elastic regimes (Equation 10). The ellipses represent the plastic failure limit of aEVpts. Each is evaluated at different thickness values and at the maximum authorized concentration in the model (99.7%). **Panel c:** The mean daily divergence calculated over the cyan box in **panel b** between September 1995 and September 1997. **Panel d:** The mean sea ice thickness simulated over the cyan box in **panel b** between September 1995 and September 1997. A five days running mean filter has been applied to all curves for readability.

In the Russian Arctic, where only the annual maximum of FI thickness is available, we cannot evaluate the simulated FI thickness against observations over a full annual cycle. Figure 7 (bottom panels) presents histograms of the maximum FI thickness differences between model and observations. Compared to the CAA, the model-observation FI thickness differences are larger in the Russian Arctic and more sensitive to rheology and grounding. Without grounding, the rheological formulation has a large influence on the shape of the distributions of thickness error, their Root Mean Square Error (RMSE)s, the number of simulated FI points and the bias (Table 4). Indeed, although both **aEVpts_no_grounding** and **BBM_no_grounding** are negatively biased, the bias is more marked in the latter (-0.18 m in **aEVpts_no_grounding** and -0.48 m in **BBM_no_grounding**). However, the modes of both distributions are similar and the smaller bias obtained in **aEVpts_no_grounding** is caused by the greater positive tail of the distribution compared to **BBM_no_grounding** (Fig. 7 bottom left panel). The highest h_{max} errors are around 1.80 m in **BBM_no_grounding**, and reach 4 m in **aEVpts_no_grounding**. This is a direct consequence of the more rigid behaviour of BBM detailed above, those extreme values errors being evidence of large dynamical thickening at the coasts. The highest number of data points in the distributions in **BBM_no_grounding** (see Table 4) supports this interpretation.



Table 4. Statistics on the FI thickness distributions presented in Fig. 7: the bias, the Root Mean Square Error (RMSE) and the number of data points (where both model and observations indicate FI presence). The statistics are shown for the four experiments (**aEVpts_no_grounding**, **BBM_no_grounding**, **aEVpts_grounding**, **BBM_grounding**) and separated between the Canadian Arctic Archipelago (CAA) region and the Russian Arctic (RU).

	Bias [m]	RMSE [m]	Number of data points
CAA			
aEVpts_no_grounding	0.19	0.51	2558
BBM_no_grounding	0.18	0.49	2736
aEVpts_grounding	0.19	0.51	2621
BBM_grounding	0.20	0.48	2756
RU			
aEVpts_no_grounding	-0.18	1.04	128
BBM_no_grounding	-0.48	0.75	168
aEVpts_grounding	-0.37	0.58	237
BBM_grounding	-0.39	0.62	240

Without grounding, both rheologies show a negative thickness bias in the Russian Arctic, because the ice is not immobilized
 370 long enough to reach a realistic maximum thickness through thermodynamic growth.

No significant difference in the FI thickness error distributions at observation stations is noticed between the two experiments with the grounding scheme activated (Fig. 7, bottom right panel). The large positive thickness bias observed in **aEVpts_no_grounding** is not present in **aEVpts_grounding**. Grounding enables the FI edge to move further offshore, thereby preventing
 375 important deformations of the FI located at coastal stations. Nevertheless, both distributions still exhibit a negative bias of ~ 0.38 m.

At coastal stations, far from the FI edge, two factors can lead to errors in the simulated maximum FI thickness (h_{max}): an incorrect FI season duration and biased heat fluxes through the ice. Indeed, shielded from dynamical thickening, the FI grows
 380 almost exclusively through thermodynamic processes. Therefore, the maximum FI thickness is driven by the duration of the immobilisation and the heat exchange between the ocean and the atmosphere, through the ice and snow. Considering only



thermodynamic growth, h_{max} is given by the Stefan's law:

$$h_{max} = \underbrace{\sqrt{\frac{2k}{L\rho_i}}}_C \sqrt{\underbrace{\langle\Delta T\rangle(d_{max} - d_{formation})}_{FDD}} \quad (15)$$

where L is the latent heat of fusion, ρ_i the density of the sea ice, k the effective thermal conductivity, and ΔT the temperature difference between the bottom of the sea ice and the surface. d_{max} is the day at which the FI has reached its maximum thickness and $d_{formation}$ is the first day of FI formation. $\langle\Delta T\rangle$ is the averaged of ΔT over the period bounded by $d_{formation}$ and d_{max} . The factor C depends on the structural properties of the sea ice and the snow covering it and is about 3.5 cm per $[\text{°C-days}]^{1/2}$. The Freezing Degree Days (FDD) term depends on the duration of the FI season and on the temperature difference between the ocean and the atmosphere. A misrepresentation of either of these, or of the snow cover or ocean heat fluxes, leads to an incorrect seasonal maximum of FI thickness, which in turn affects the timing of FI breakup.

390

In the bottom panel of Fig. 8 we display the relation between the simulated h_{max} and the FDD, for a given value of C , from our four experiments. This figure shows that the relation in Equation 15 holds for sea ice that is thin when becoming FI. The top panel indicates that the error in h_{max} decreases with FDD, but large errors can persist at high FDD if the sea ice has thickened dynamically before becoming FI. Therefore, accurately capturing the maximum FI thickness requires both a correct FDD and
 395 model physics that allow sea ice to lock early enough to be shielded from excessive dynamical thickening.

The positive FI thickness bias in the CAA is the result of the excessive immobilisation of the FI, which we documented in the previous section. A cold bias in the reanalysis, in the mountainous part of the CAA, caused by the spectral representation smoothing out mountainous areas over the sea, could further enhance the positive bias. This results in surface air temperatures
 400 being more representative of high terrain - i.e. too cold (Dupont, 2012). In the Russian Arctic, a negative maximum thickness bias is still present in the two simulations with grounding, whereas the simulated durations of the FI seasonal cycle are similar to the observations (Fig. 3). This can be partly explained by the warm Arctic winter bias of ERA5 (Zampieri et al., 2023; Michalezyk et al., 2025), the atmospheric forcing used in this study.

4 Discussion

405

The key difference between BBM and aEVPts, for the representation of FI, is the behaviour in compression. With BBM, sea ice can withstand higher compression than with aEVPts. This increased resistance helps to more readily immobilise FI between anchor points resulting in greater extents of FI in **BBM_no_grounding** compared to **aEVPts_no_grounding** (Fig.

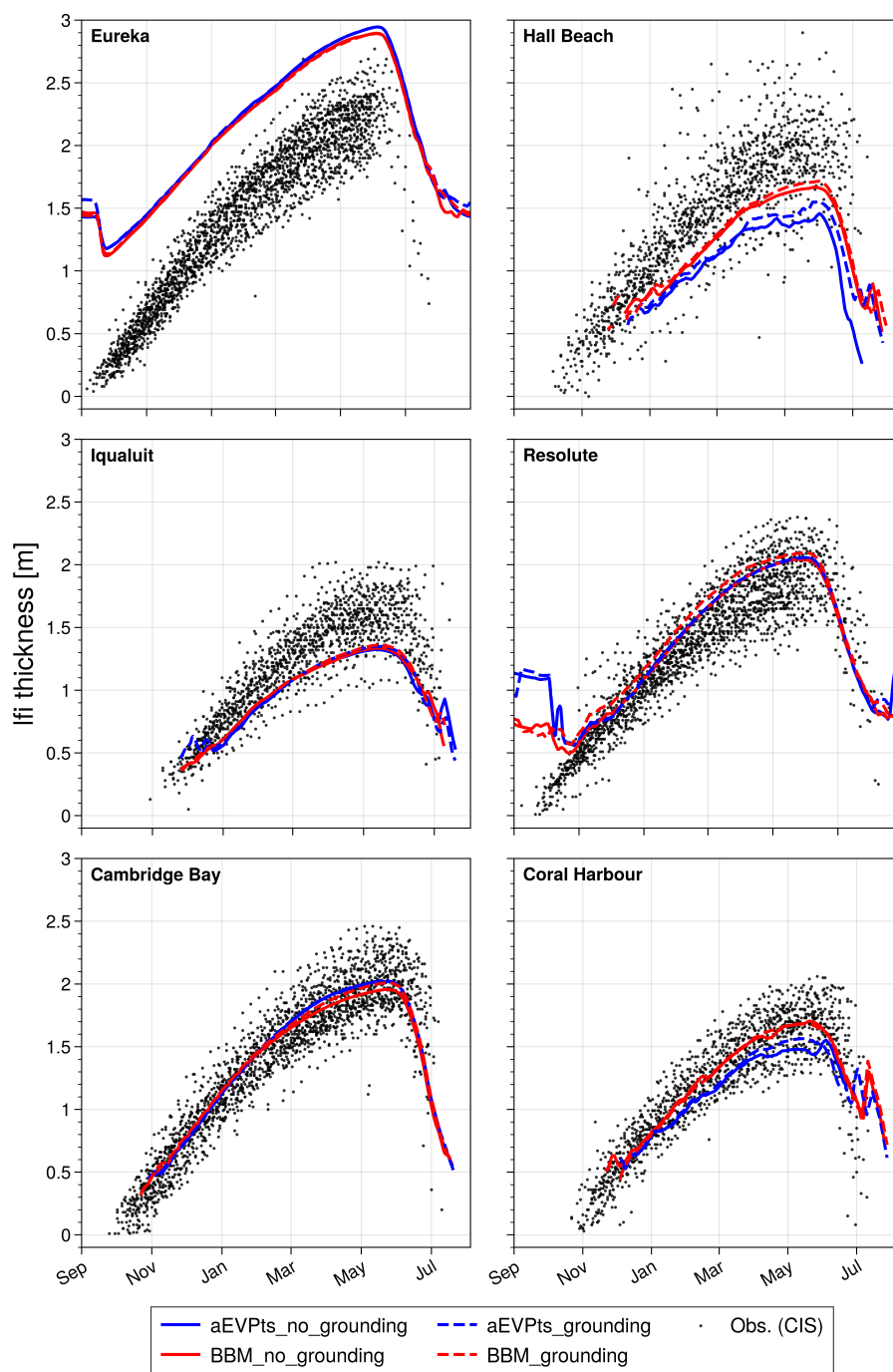


Figure 6. Seasonal cycle, averaged between 1991 and 2016, of the FI Ice Thickness Distribution (ITD) mode of the at six stations in the CAA (see Fig. 1 for station locations and Table 3 for their coordinates). The blue solid line corresponds to **aEVpts_no_grounding**, the red solid line to **BBM_no_grounding**, the blue dashed line to **aEVpts_grounding** and the red dashed line to **BBM_grounding**. A five days running mean filter has been applied to all curves for readability. The black dots are the bi-weekly drill-hole measurements from CIS.

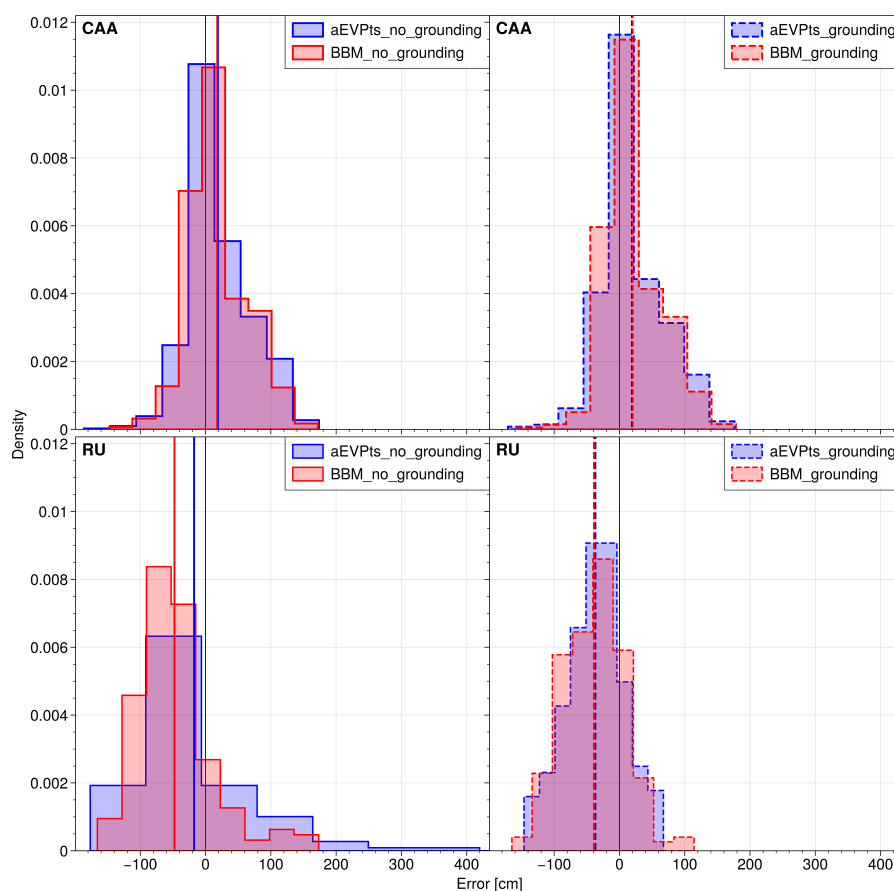


Figure 7. Histogram of FI Ice Thickness Distribution (ITD) mode error at stations in the CAA (**top**) and the Russian Arctic (**bottom**). The left panels correspond to experiments without grounding and the right panels to experiments with the grounding activated. The blue solid line corresponds to **aEVPTs_no_grounding**, the red solid line to **BBM_no_grounding**, the blue dashed line to **aEVPTs_grounding** and the red dashed line to **BBM_grounding**. The bias, the RMSEs and the number of data points (where both model and observations indicate FI presence) for those distributions are given in Table 4.

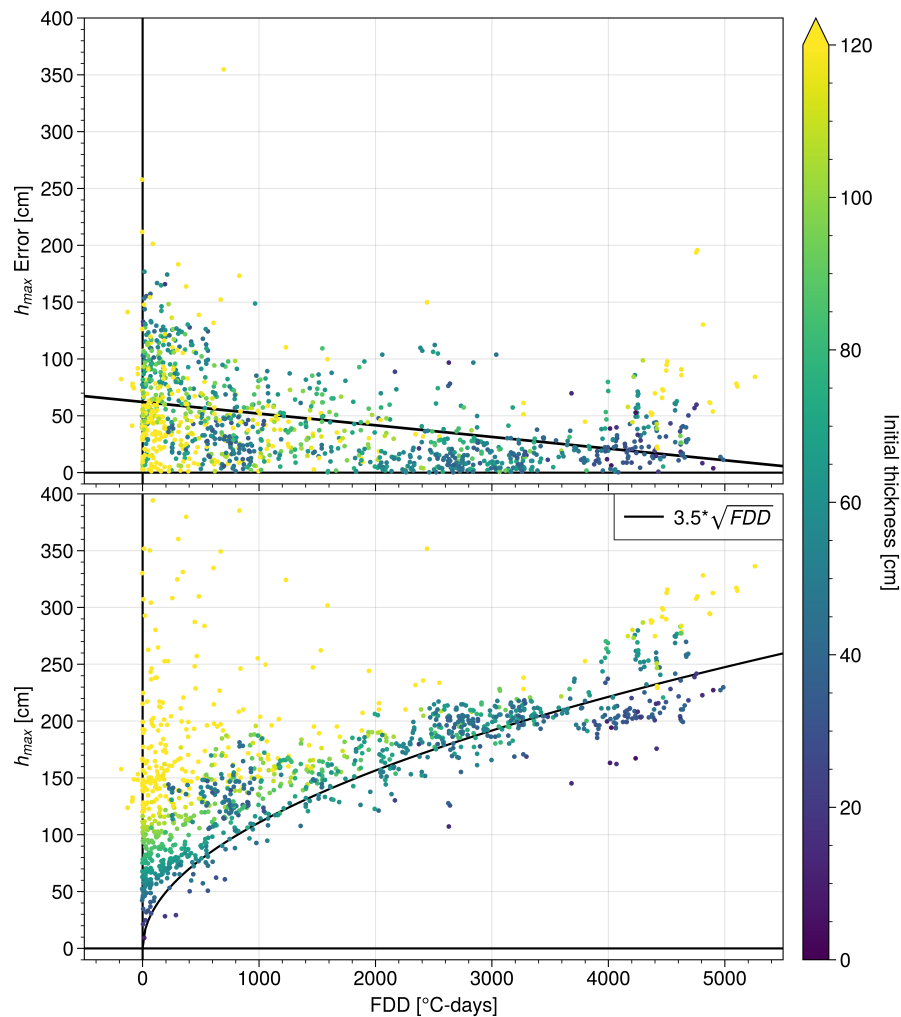


Figure 8. Top: Scatter plot of absolute error in maximum FI Ice Thickness Distribution (ITD) mode versus FDD. The black line is the statistically significant trend (p-value $< 10^{-31}$). **Bottom:** Scatter plot of maximum FI thickness versus FDD. The black curve represents the theoretical relation from Stefan's law (Equation 15) with $C = 3.5 \text{ cm } ^\circ\text{C-days}^{-1/2}$. For both panels, the color corresponds to the initial thickness (at the moment when the corresponding grid cell become FI covered).



3). When comparing simulated FI thicknesses to drill-hole measurements along the Siberian shelf, we found that **aEVpts_no**
410 **_grounding** produces excessively thick FI, indicating excessive dynamical thickening. In contrast, **BBM_no_grounding** does
not exhibit such disproportionately thick sea ice at coastal station locations. However, since the greater BBM rigidity limits
the dynamical thickening of sea ice, it reduces the efficiency of the grounding scheme in generating new anchor points. This
explains why more FI is found in **BBM_no_grounding** compared to **aEVpts_no_grounding** while no systematic difference
is found between **aEVpts_grounding** and **BBM_grounding**. **aEVpts_grounding** generates more FI in shallow water regions,
415 where the importance of grounding is particularly pronounced, while **BBM_grounding** generates more FI in the deeper water
regions.

The development of a ridging and rafting scheme dedicated to BBM would be beneficial for FI simulation with this rheology.
Indeed, in its absence, we used the scheme developed for aEVpts. However, as we have seen, aEVpts experiences stronger
420 convergence events than BBM. A dedicated ridging and rafting scheme would allow to produce more last thickness category
sea ice during convergence events, enabling more effective grounding without a loss of rigidity.

Accurate simulation of FI thickness depends not only on the correct duration of FI immobilisation but also on the heat flux
through the ice which is itself affected by atmospheric forcing, the snow cover representation and ocean heat fluxes. Therefore,
425 the negative bias observed in the simulated maximum FI thickness in the Russian Arctic in the experiments with grounding is
likely partly due to the warm Arctic winter bias present in the ERA5 reanalysis used to force our model (Zampieri et al., 2023;
Michalezyk et al., 2025).

The FI thickness has a large influence on the timing of breakup and therefore on the FI season length. Indeed, the thicker FI
430 obtained with aEVpts postpones the mean breakup date by up to two months in the East Siberian and Chukchi Seas compared
to BBM (Fig. 3). It can also lead to sea ice persistence through summer in one rheology but not the other (Fig. 5, panel d). This
highlights the importance of accurately representing FI thickness, and therefore of capturing dynamical thickening correctly
and using realistic atmospheric forcing.

5 Summary and Conclusion

435

In this study, the influence of grounding and rheological formulations on the Arctic FI characteristics simulated by a coupled
sea ice-ocean model was assessed using sensitivity experiments and available observational reference products. A grounding



scheme (Dupont et al., 2022) and two rheological formulations were tested: aEVPTs, which had previously been used in FI simulations, and the more recently developed BBM based on brittle mechanics.

440

Our results confirm that the grounding scheme is of primary importance for simulating realistic Arctic FI coverage, regardless of the rheology employed, everywhere except in the CAA. The rheology is also influential on FI coverage: without a grounding mechanism, a more rigid rheology, such as BBM, increases FI coverage. However, our results also show that the effects of grounding on FI characteristics depend on the rheological formulation used: more rigid ice increases FI formation but ridges
 445 less easily, which reduces the number of anchoring points, inhibits grounding, and may ultimately reduce FI extent in shallow water regions. This interaction also affects season length, as a more solidly grounded FI cover breaks up later and can even persist through summer. It motivates the development of dedicated ridging and rafting scheme when developing more rigid rheologies. Finally, the realism of FI thickness depends on the duration of the immobilisation, but also on the heat fluxes through the ice, and therefore on the realism of the atmospheric forcing and snow representation in the model.

450 *Code and data availability.* The original code of NEMO-SI3 (v 4.2.2) is available from https://forge.nemo-ocean.eu/nemo/nemo/-/tree/4.2.2?ref_type=tags with the associated manual <https://doi.org/10.5281/zenodo.7534900>. The BBM implementation is available from <https://github.com/brodeau/bbm4si3>. All the relevant model outputs used in this study, as well as the simulations namelist and the code used to run the simulations, are accessible in <https://doi.org/10.5281/zenodo.22939885>.

Appendix A: Tuning of the atmospheric drag coefficient

455

Using the same atmospheric sea ice drag coefficient (C_d^i) with different rheologies results in different mean sea ice drift speeds. To allow a fair comparison between the two rheologies, we tuned the coefficient C_d^i so that the mean drift speeds were similar for the four experiments. For tuning, we compared the simulated sea ice drift to the one obtained from OSI-SAF (OSI SAF, 2022) during the months between October and April (the period for which observational data are available from OSI-SAF),
 460 from 1990 to 1994, and limited to the pack ice region (excluding the coasts and the marginal ice zone, as delineated by the black line on the inset map in Fig. A1). We applied a spatial coarsening of the model outputs to match the spatial resolution of the OSI-SAF product and performed a nearest-neighbour interpolation from these coarsened model outputs to the OSI-SAF grid. Finally, we only retained grid cells for which there were data in both the four simulations and the observational data set, ensuring that all distributions had the same number of data points. We found the following values: $C_d^i = 1.3 \times 10^{-3}$ for aEVPTs
 465 and $C_d^i = 1.6 \times 10^{-3}$ for BBM. The speed histograms obtained using the method described above are shown in Fig. A1. The distributions obtained with the two rheologies exhibit different shapes, those with BBM presenting a peak of nearly immobile sea ice, counterbalanced by a larger tail of the distributions. In contrast, the two distributions of sea ice speed obtained from

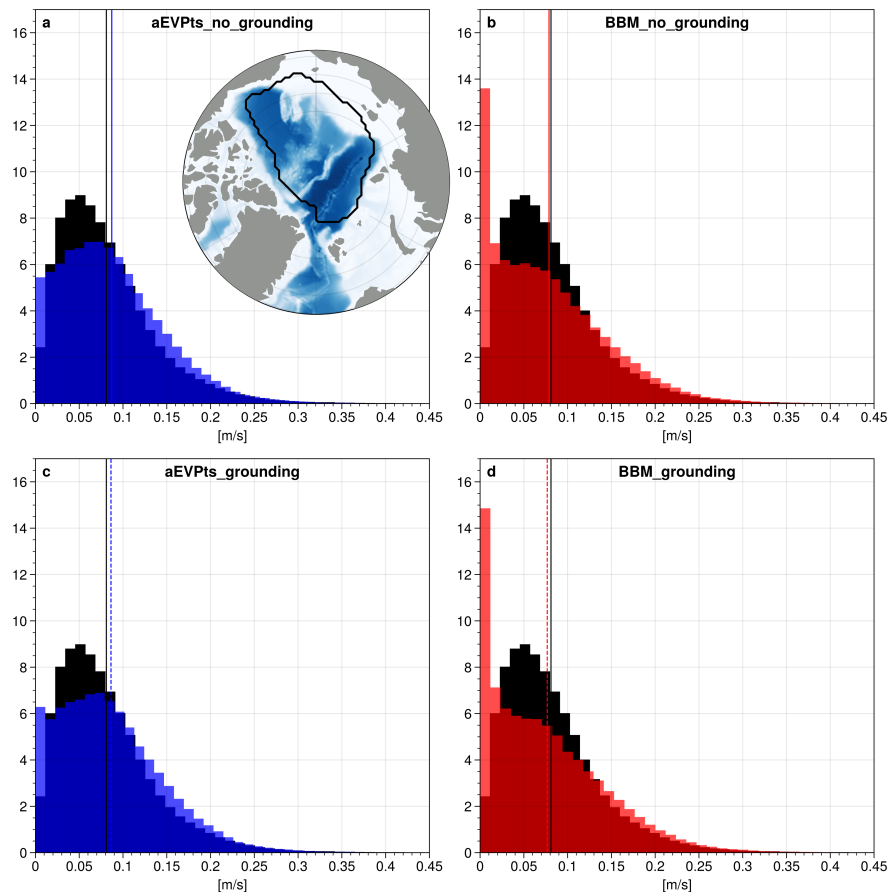


Figure A1. Histograms of sea ice speed between October and April from 1990 to 1994, limited to the region outlined by the black line on the inset map. On each panel, the black histogram corresponds to OSI-SAF data. The blue and red histograms are respectively those of: **aEVpts_no_grounding** (Panel a), **BBM_no_grounding** (Panel b), **aEVpts_grounding** (Panel c), **BBM_grounding** (Panel d). The vertical lines show the means of the distributions.

the simulations with aEVpts peak at 0.7 m s^{-1} and have half as much nearly immobile sea ice as the experiments with BBM. However, and despite their different shapes that are inherent to the existing differences of physical assumptions in the two classes of rheology, the four distributions have a similar mean drift speed that is close to the observations.

Appendix B: Large-scale performances of the four experiments

Figure B1 shows the mean Arctic sea ice thickness in March and September, the months of maximum and minimum sea ice volume, respectively. The mean thickness fields obtained using aEVpts are smoother than those simulated using BBM.



475 Experiments with the two rheologies produce comparable mean sea ice thickness fields in terms of magnitude. However, the
 fields obtained with BBM present a lower maximum thickness north of Greenland in August, as well as slightly thicker sea ice
 in March. Figure B2 shows the seasonal cycle of Sea Ice Extent (SIE) and Sea Ice Volume (SIV) from the four experiments,
 along with the Nimbus-7 SMMR and DMSP SSM/I-SSMIS Passive Microwave Data, Version 2 dataset (DiGirolamo et al.,
 2022) from NSIDC and the Pan-Arctic Ice Ocean Modeling and Assimilation System (PIOMAS) dataset (Zhang and Rothrock,
 480 2003; Schweiger et al., 2011). PIOMAS is not an observational product but a reanalysis that assimilates sea ice concentration
 and sea surface temperature. The two rheologies agree well for SIE, with a little overestimation of the maximum and a little
 underestimation of the minimum compared to NSIDC. The two rheologies also produce similar SIV seasonal cycles, with the
 BBM SIV being greater by up to 7.4% compared to the aEVPts one during winter. Both rheologies underestimate the SIV
 compared to PIOMAS during summer. No significant differences are observed in the SIV and SIE seasonal cycles between
 485 experiments with and without grounding for a given rheology.

Author contributions. AL, FM and TF outlined the study. AL performed the simulations, the analyses and the figures, and all the co-authors
 contributed to the interpretation of the results. AL wrote the manuscript with inputs from all co-authors.

Competing interests. At least one of the (co-)authors is a member of the editorial board of The Cryosphere.

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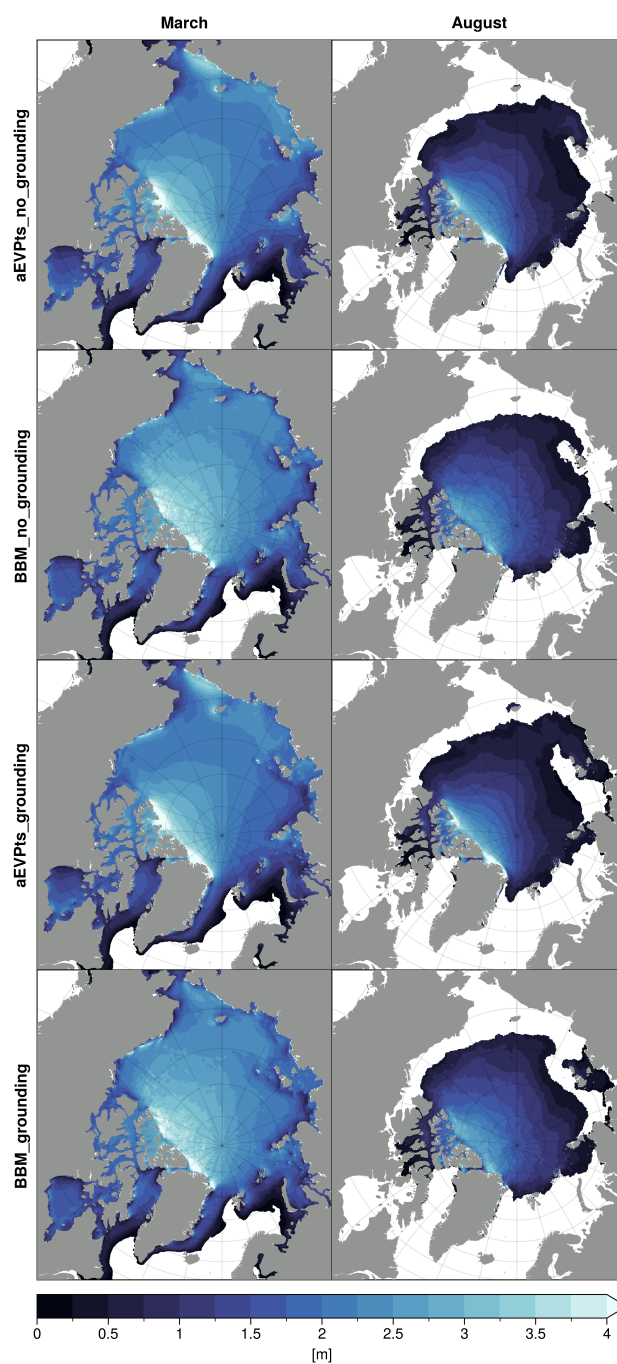


Figure B1. Sea ice thickness maps at the month of maximum (March, left panels) and minimum (August, right panels) sea ice volume for the four experiments. Thickness in grid cell with sea ice concentration below 15% is not shown.

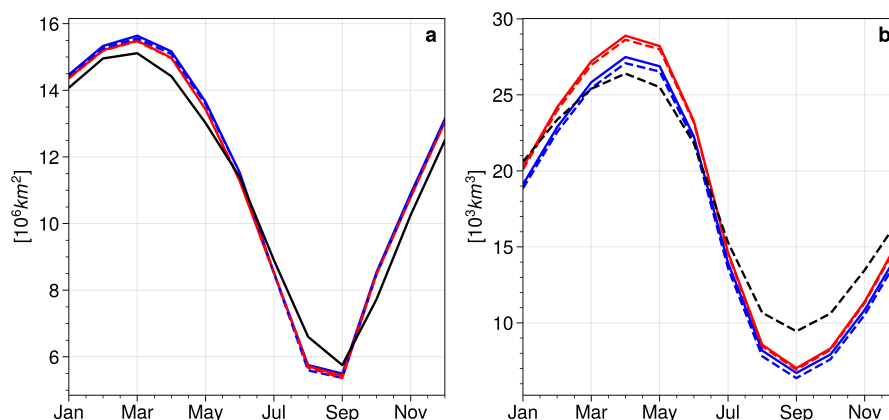


Figure B2. Arctic sea ice extent (left panel) and volume (right panel) seasonal cycles, averaged over 1991-2016. The solid black line corresponds to the NSIDC dataset and the dashed black line to the PIOMAS dataset. The blue solid line corresponds to **aEVPts_no_grounding**, the red solid line to **BBM_no_grounding**, the blue dashed line to **aEVPts_grounding** and the red dashed line to **BBM_grounding**.

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